The sedimentology and lithostratigraphy of the Upper Jurassic Lourinha formation, Lusitanian Basin, Portugal

Thesis

THE SEDIMENTOLOGY AND LITHOSTRATIGRAPHY OF THE UPPER JURASSIC LOURINHÃ FORMATION, LUSITANIAN BASIN, PORTUGAL

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The Open University

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The Open University

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Enclosures

THE SEDIMENTOLOGY AND LITHOSTRATIGRAPHY OF THE UPPER JURASSIC LOURINHÃ FORMATION, LUSITANIAN BASIN, PORTUGAL

A thesis submitted for the degree of Doctor of Philosophy

in Earth Sciences

by Graham Hill BSc. Hons. (Aston)

Friday 30th. September 1988

To my Parents Roland E. & Beryl A. Hill

For they are ultimately responsible !

Brightly coloured mudrocks characterise distal alluvial fan facies of the Praia da Amoreira member, Lourinhã formation. Porto Novo to Porto Dinheiro section, Estremadura, Portugal.

EX12

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ABSTRACT

The Lusitanian Basin of West-Central Portugal is one of a family of formerly adjacent North Atlantic margin basins. Like others of its type (e.g. the Jeanne d'Arc and Western Approaches basins) early phases of rifting and subsidence occurred as far back as the Triassic.

During the Late Jurassic rifting phase in the development of the Lusitanian Basin marine shelf and slope systems were replaced by alluvial fan and fluvial environments which prograded from the north during the late Kimmeridgian. Continental environments, periodically punctuated by marine incursions, persisted over northern and central areas of the basin at least until the end of the Jurassic. The deposits of these continental and shallow marine systems are here defined as the Lourinhã formation, which is sub-divided into five members.

Sedimentary structures, textures and sand body geometries suggest that the Praia da Amoreira member was deposited in a distal alluvial fan setting where sheet-flooding preceded the establishment of channelised flow. The presence of calcretes suggests a semi-arid/sub-tropical climate with markedly seasonal rainfall. The fan systems, confined to the western margin of the basin, passed laterally into meandering fluvial systems, the deposits of which form the Porto Novo member.

A marine transgression from the south resulted in the establishment of shoal-water delta systems, the deposits of which form the Praia Azul (lower delta plain) and Assenta (upper delta/fluvial plain) members. The largely continental Assenta member is punctuated by thin, laterally extensive marine horizons which suggest rapid relative rises in sea level. In northern and eastern parts of the basin the deposits of sandy fluvial systems are replaced by those of the gravelly Santa Rita member, which appear to have been sourced from the east.

In a number of the above units evidence for the operation of active extensional processes is to be found. The deposition of the Lourinhã formation preceded the final Aptian rifting event which resulted in the formation of the North Atlantic Ocean. However, it may have heralded, and been contemporaneous with, an earlier phase of sea floor spreading, later abandoned, thought to have occurred in the area of the present day Tagus Abyssal Plain.

ACKNOWLEDGEMENTS

When one considers that a PhD. is supposed to be very much an individual effort it comes as somthing of a suprise to find that a veritable army of individuals are responsible for your ability to complete the damn thing.

This is my army:

I thank my supervisors Chris Wilson, Tim Astin and Rick Hiscott for their friendship, advice and support both in the field and in the production of this document; no blame can be laid at their feet for any inadequacies herein.

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For logistical support the Department of Earth Sciences as an institution has to be acknowledged and in particular; John Holbrook, Ian Chaplin and their crew for technical back-up; John Taylor and Andrew Lloyd for their wealth of advice over drafting and for not being too offended when I turn up to do diagrams, mug in hand, at 'tea time'; and the secretarial staff for their backchat, in particular Beryl West for her Mac-advice and typing the paper, and Janet Dryden and Carol Whale for the **gallons** of coffee I have consumed over the last three years. John and Debbie in the Open Univerity photographic department deserve thanks for thier assistance.

All my contemporaries are thanked for their friendship, particularly Matthew's Willis and Watkinson for life support whilst doing field work *esta bem* technical chats and witticisms; Dave and Leon for having to live with me and particularly hardly ever seeing me over the last couple of months (possibly not such a bad thing); Liz Harper for risking life and limb, or at least her fingers, whilst chopping up photographs; and Christopher MacLeod, my compatriot in the completion marathon, who never once winced at the possibility of being lapped.

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Thank you all. If I have forgotten anyone then please forgive me-its the PRESSURE!

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CHAPTER 1

INTRODUCTION AND STRATIGRAPHY

The Lusitanian basin of West Central Portugal is an Atlantic Marginal Basin in which early phases of rifting and subsidence occurred as far back as the Triassic (Ribeiro et al. 1979; Wilson et al. in press). It is one of a family of such basins that today are situated off Iberia, Canada and NW Europe (Figure 1.1). It extends some 300km north-south and up to 180km east-west with a considerable extension offshore to the west and north. Originally it may have been contiguous with the Western Approaches Basin to the south west of the United Kingdom (Masson and Miles 1986).

The basin is bordered to the east by Hercynian basement rocks of the Iberian Meseta, which crop out extensively inland to the north and east (Figure 1.1). A few kilometres offshore, fault-bounded blocks forming the Berlengas and Farilhões Islands are part of a system of horsts which mark the basin's ill defined western boundary (Figure 1.2, Boillot et al. 1975, 1978). In other offshore areas, basement rocks are often only shallowly buried (Willis 1988).

The exceptionally good Mesozoic outcrops available for study all appear in a NE-SW trending high flanked by two Tertiary sub-basins (Figure 1.2). This area, formerly buried by over 1.5km of Tertiary cover, was inverted during the Miocene by compressional movements linked to the Betic Orogeny (Willis 1988; Wilson et al. in press).

1.1 AIMS OF THE PROJECT

The principle aim of the study was the detailed description and interpretation of Upper Jurassic, predominantly continental clastic rocks of the southern part of the Lusitanian Basin. Fieldwork accounts for the majority of the data input and has involved detailed examination of both vertical and lateral facies changes. In the absence of a formal lithostratigraphy a scheme was erected (see section 1.6) which formed the basis of this study and is followed in the breakdown into chapters of this thesis. The lithostratigraphy has been determined from reconnaissance and detailed study of coastal outcrop, plus discussion with supervisors, particularly R. Hiscott. It has been integrated with the scheme proposed by Leinfelder (1986) for approximately the same time interval in the Arruda region, where the sequence is rather different.

In the concluding chapter, the results are integrated with those of other workers to contribute to the picture of the basin's development during the Mesozoic. On a broader scale, comparisons of the Lusitanian Basin's evolution, with that of other Iberian marginal basins and those on the formerly adjacent eastern seaboard of North America will be made.

1.2 STRUCTURE OF THE LUSITANIAN BASIN

The main structural features of the basin are faults, folds and diapirs, the latter cored by Triassic evaporites (Figure 1.3). Most of the diapirs follow a NNE-SSW trend which mirrors that of major basement lineaments (Ribeiro et al. 1979). They have had a complex and long-lived history which include significant strike slip movement during the Tertiary (Willis 1988).

The inverted block of Mesozoic strata is bounded by structures with a Hercynian trend and is flanked by Tertiary sub-basins (Figure 1.2). Thermal maturity data indicate that Upper Jurassic sediments within the inverted area were once buried to at least 1.5km depth (Willis 1988). The inversion is thought to be due to the Betic orogeny (Willis op cit.). This event was fortunate in that it exposed at the surface today pre- and syn-rift passive margin sequences (sensu Sibuet et al. 1984), which in other Atlantic margin basins are usually buried on continental shelves.

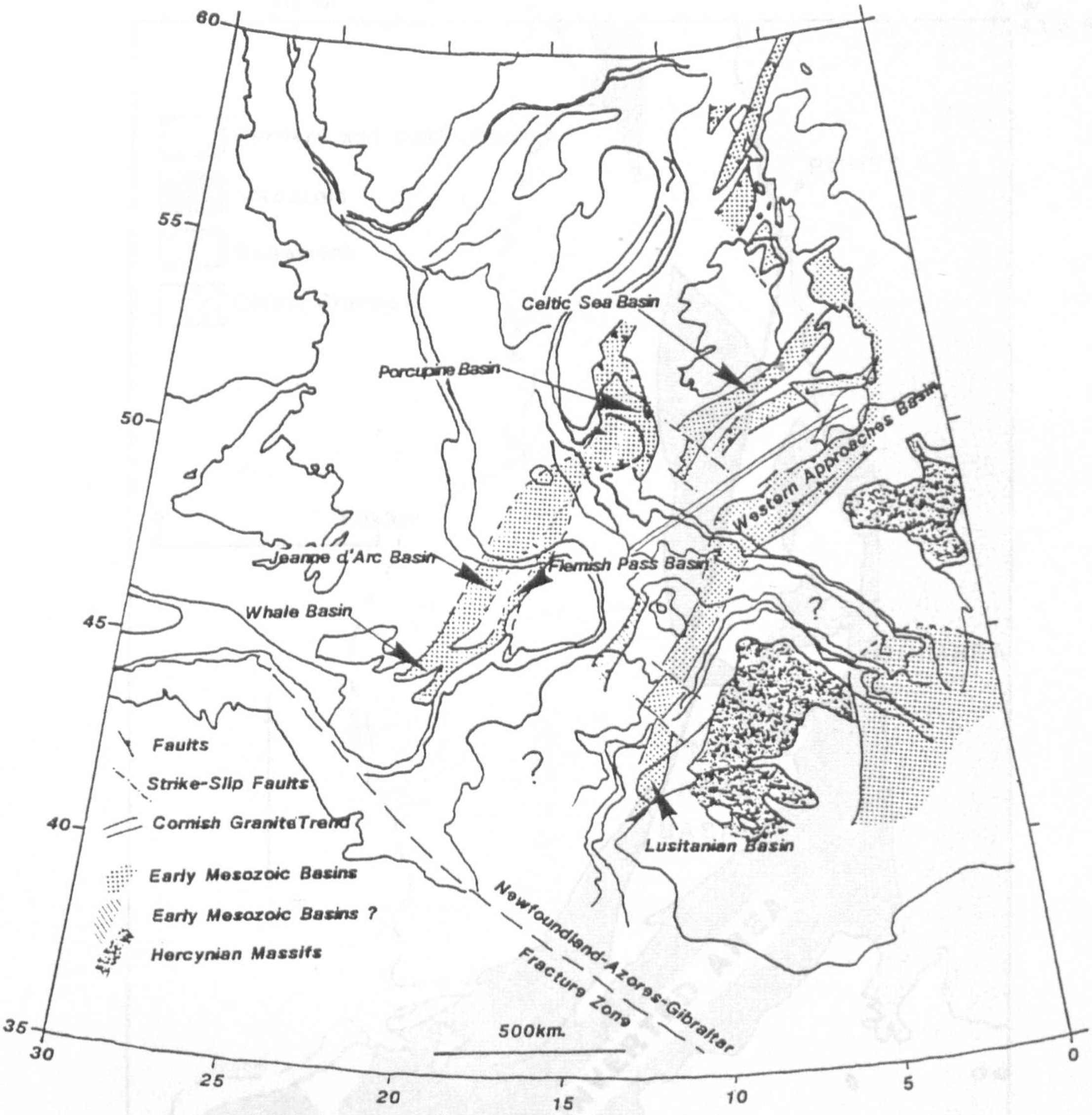


Figure 1.1 Early Mesozoic reconstruction of the North Atlantic rift system, after Masson and Miles (1986). The Lusitanian Basin may originally have been connected with the Western Approaches Basin.

Crude Mesozoic and Tertiary isopachs (Figure 1.4), indicate the main axis of subsidence to have a gross NNE-SSW trend. This parallels a major set of basement faults and suggests that these structures may have controlled early Mesozoic subsidence and sedimentation patterns. The true picture is rather more complicated as revealed by modern seismic data (Wilson et al. in press). The northern part of the basin can be divided into three sub-basins (Figure 1.3) :

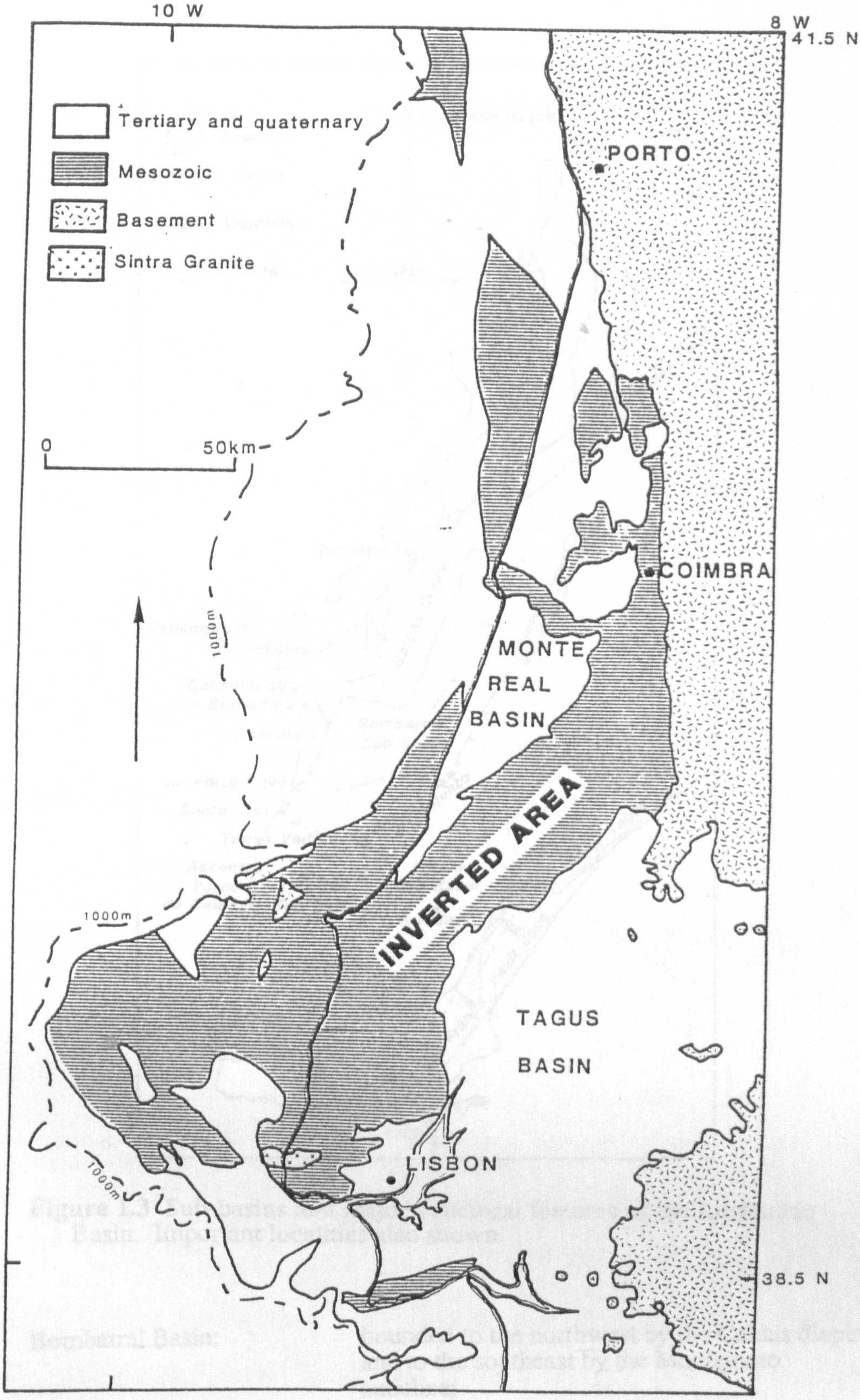


Figure 1.2 Geological map of the on- and offshore parts of the Lusitanian Basin (based on Boillot et al. 1978). Coastline is marked by the double line. The NE-SW trending strip of Mesozoic sediments separating two Tertiary sub-basins was inverted during the Miocene (Willis 1988; Wilson et al. in press). The two larger areas of basement rocks mapped offshore mark the sites of the Berlengas and Farilhões Islands. The smaller more southerly area is submerged. Over much of the offshore area basement rocks are only shallowly buried (Willis 1988).

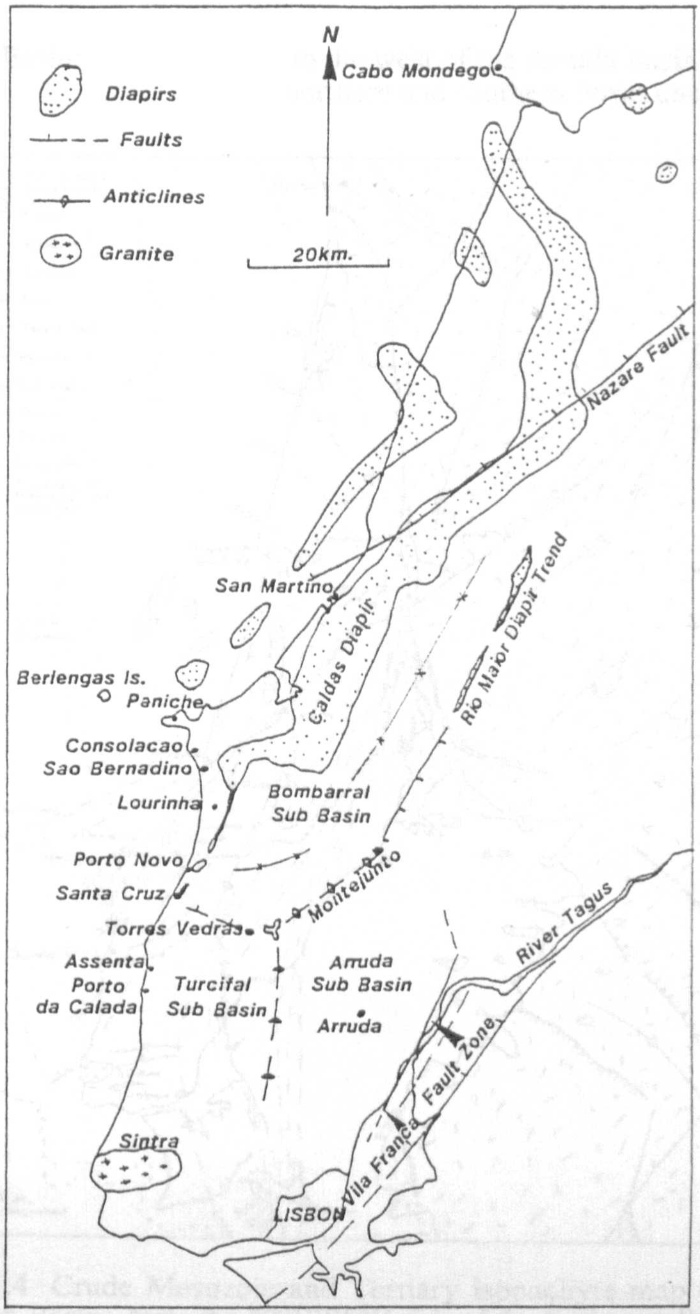


Figure 1.3 Sub-basins and major structural features of the Lusitanian Basin. Important localities also shown.

Bombarral Basin: bounded to the northwest by the Caldas diapir and to the southeast by the Montejunto anticline;

Arruda Basin: half graben, bounded to the north by the Montejunto anticline, nature of southern margin unknown;

Turcifal Basin: to the west of the Arruda Basin, half graben, northern and southern limits uncertain;

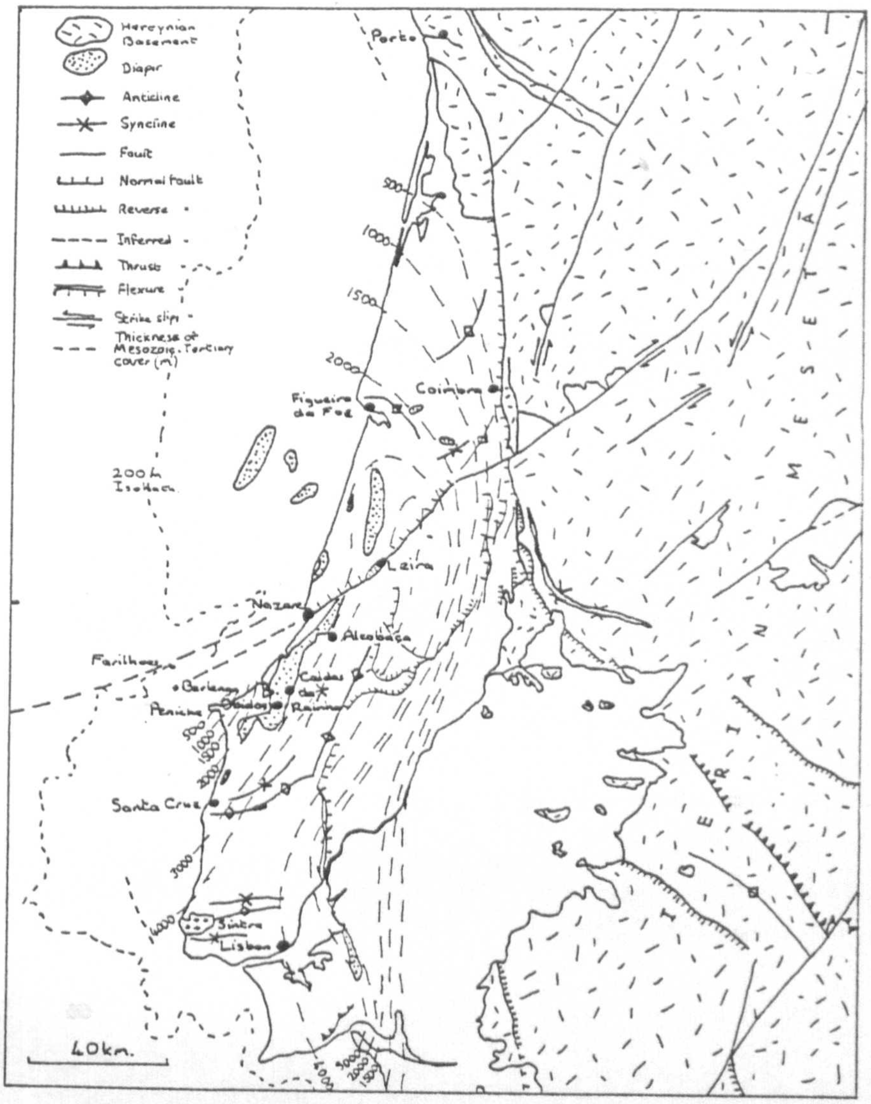


Figure 1.4 Crude Mesozoic and Tertiary isopachyte map of the Lusitanian Basin. Note the NNE-SSW axis of the basin parallels the structural trends in the underlying basement. This trend is also followed by salt diapirs. Based on the Tectonic map of Portugal Ribeiro et al. (1972).

Diapiric structures are confined to the northern part of the study area and this probably reflects the original distribution of the Hettangian evaporitic deposits (Wilson et al. in press). Where these are thick, diapirs have developed over the reactivated Hercynian structures. The Bombarral sub-basin is situated between the Caldas and Rio Maior diapirs (Figure 1.3) and takes the form of a large open syncline (Willis 1988). The axis of this structure sweeps westwards at its southern end. The formation of this basin, which at its centre contains over 1000m of Upper

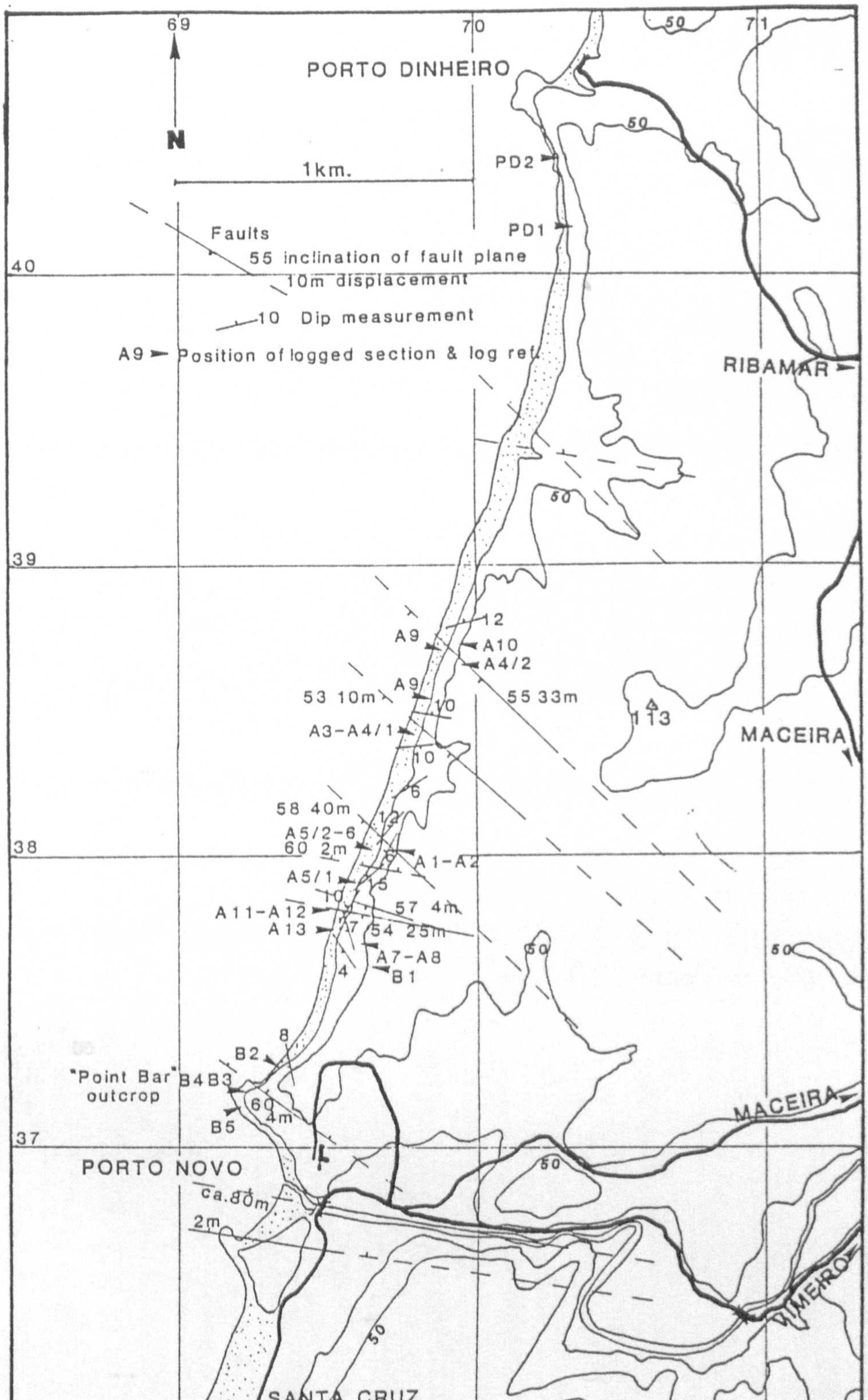


Figure 1.5 Sketch map of the section between Porto Dinheiro and Porto Novo, Showing faults, dip measurements and the position of logged sections. The 'point bar' outcrop is discussed in Chapter 3. Fault trends such as those illustrated consistently appear in many of the coastal sections examined. The structure of this particular section is an open anticline with an axis trending approximately east-west. For much of the section Praia da Amoreira member sediments (see 1.6.6) crop out. These are overlain by Porto Novo member sediments at P. Novo and P. Dinheiro. Heights in metres.

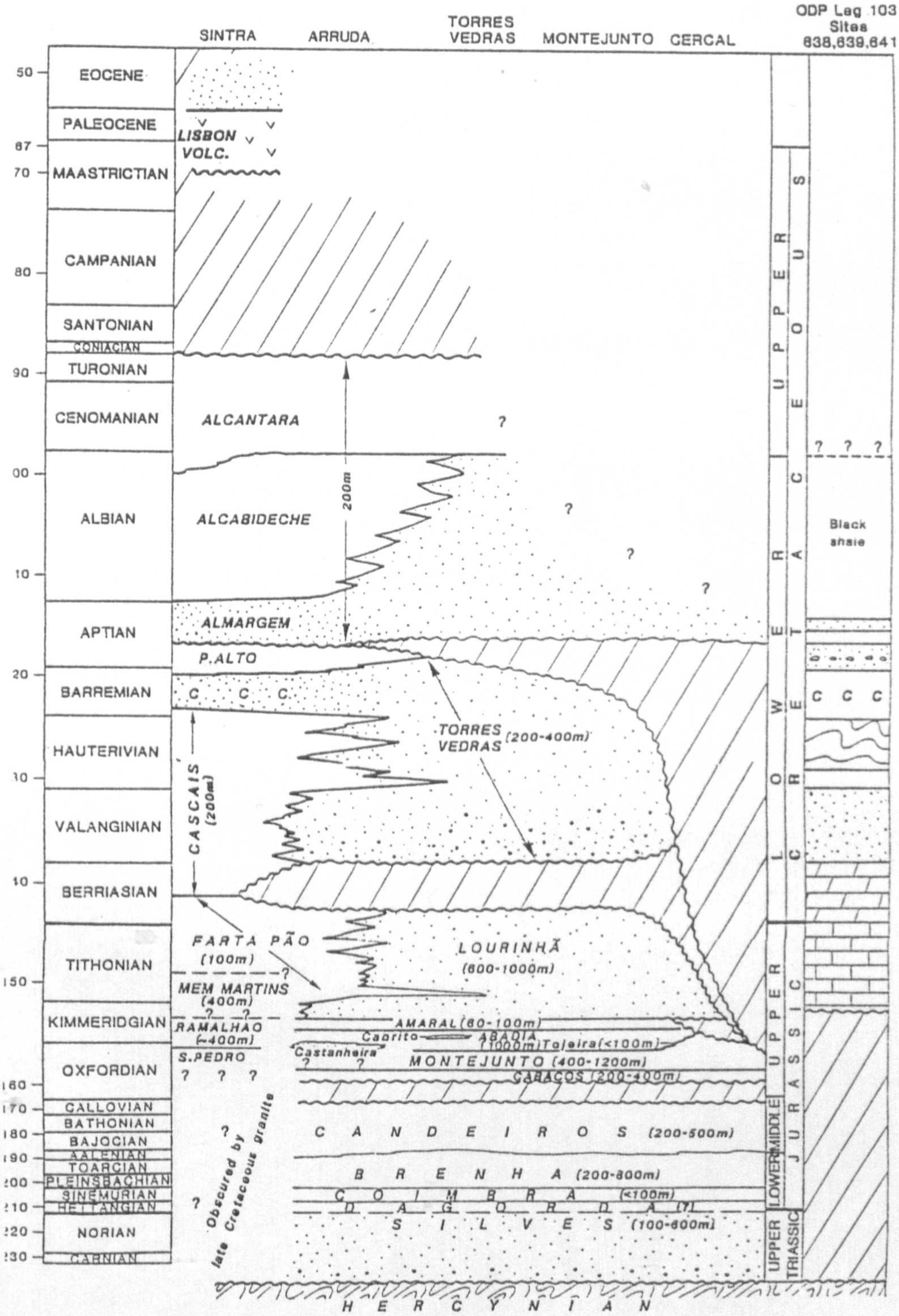


Figure 1.6 Outline of Mesozoic and early Tertiary lithostratigraphy of the Lusitanian Basin (siliciclastics stippled) and for comparison, the succession drilled during ODP Leg 103 (Boillot et al. 1987). Note changes in time scale (after Kent and Gradstein 1985). The nomenclature is informal and currently under discussion. Formations in upper case, members in lower case. The Lourinhã formation is the unit studied by the author. Locations shown on Figure 1.7. Diagram from Wilson et al. in press.

Jurassic continental clastic sediments, is partly due to salt withdrawl. Salts migrated laterally into the diapirs which flanked the structure (Zbyszewski 1959; Ribeiro et al. 1979; Willis 1988; Wilson et al. in press).

An anticlinal structure, the Montejunto high, which separates the Bombarral from the two southern sub-basins (Figure 1.3) also has a salt pillow beneath it. On seismic lines late Jurassic sediments are observed to onlap the structure (Wilson et al. in press).

In the absence of significant thicknesses of evaporites the structural styles south of Montejunto (Figure 1.3) are rather different. In this area, fault movements along Hercynian structures propagated into the Mesozoic and Tertiary cover. The Arruda and Turcifal sub-basins consequently have a half-graben style (Wilson et al. in press). The bounding faults of both these structures lie on their eastern margins and downthrow to the west.

On the coast, the Upper Jurassic sections examined are often cut by normal faults with a NW-SE or E-W trend. Displacements are of the order of a few metres to several tens of metres. Particularly good examples crop out in the sections north of Porto Novo (Figure 1.5) and south of San Martino do Porto. Faulting styles vary from simple single planar faults to complex zones of syn- and antithetic fractures. A number of these structures are injected by small dykes of intermediate composition which have been dated by Ar^{40} techniques at 130-140Ma (Willis 1988). Willis also observed that Cretaceous sediments, thought to be of Valanginian age (Rey 1972), are not affected by this deformation. In addition, observations made by the author on the coastal sections south of Peniche suggest that some of these faults may have acted as growth faults during the late Jurassic (see Chapter 3).

1.3 BASIN DEVELOPMENT IN THE MESOZOIC

A simplified basin stratigraphy is shown in Figure 1.6. The scheme results from the efforts of a number of co-workers based at the Open University, and in Portugal and West Germany. As the lithostratigraphic names are not yet formalised, lower case titles ie. formation/member are used throughout. The late Triassic-middle Jurassic stratigraphy of the basin is similar to that of other Iberian and Canadian marginal basins. During the Triassic to lowermost Jurassic, a largely continental clastic sequence was deposited. The Silvés Sandstone formation consist of fluvial sandstones, mudstones and conglomerates (Ribeiro et al. 1979). These interfinger to the southwest with the Dagorda marls and evaporites, probably the deposits of restricted continental seas.

The basin was probably characterised by significant relief and differential subsidence, resulting in the accumulation of variable thicknesses of sediment (Willis 1988). Over 2km. of evaporites were deposited in places and these have formed chains of diapirs along the NNE-SSW basement structural trend (Figure 1.3). These were particularly active during the Oxfordian / Early Kimmeridgian and affected facies distributions during this period (Ellwood 1987).

A widespread marine transgression is indicated by the overlying, largely dolomitised, carbonate ramp sequence of the Coimbra formation (M. Watkinson pers comm.). In contrast to the Silvés and Dagorda formations the Coimbra formation exhibits relatively constant thicknesses (Willis 1988).

Open marine conditions ensued due to a transgression from the northwest during the lower Jurassic. This led to the development of a carbonate ramp, with ammonitic mudstones and limestones of the Brenha formation to the west, and progressively shallower carbonate facies of the Candieiros formation eastwards (Wright and Wilson 1984). The maximum extent of the transgression occurred in the Bajocian and Toarcian stages.

During the Toarcian-Aalenian a shallow-water cabonate shelf existed on the western margin of the basin. This shed debris eastwards into the deepwater areas around Peniche, forming a carbonate submarine fan (Wright and Wilson 1984). There is also faunal evidence (Mouterde et al. 1979), indicating an open sea to the west which connected to both the sub-Boreal and Tethyan realms. This supports the suggestion of Masson and Miles (1986) of a connection with the Western Approaches Basin, in addition to an open southern end to the basin.

The Upper Jurassic commenced with a major regressive event which resulted in a widespread hiatus and local unconformity across which strata of latest Callovian/early Oxfordian age are missing (Ribeiro et al. 1979). A feature of many French and Iberian North Atlantic marginal basins, and possibly the Grand Banks (R.C.L. Wilson pers comm.), this suggests a widespread eustatic fall in sea level and lowstand during the lower to mid. Oxfordian. Several authors, notably Vail et al. (1977), have proffered evidence for such an event from elsewhere in the world.

Renewed submergence in the late Oxfordian led to bituminous freshwater, playa and peritidal sediments being deposited locally over karstified middle Jurassic carbonates. The base of the Kimmeridgian generally coincides with a change from carbonate to siliciclastic deposition. An early Kimmeridgian rifting event caused the break up of Oxfordian Montejunto carbonate buildups and the resedimentation of some of these carbonates as debris and turbidity flows into basinal areas (Guery et al. 1986; Wilson et al. in press). Platform carbonates still existed locally over diapiric highs, the orientation of which parallel NNE-SSW trending Hercynian basement faults and which are cored by Triassic evaporites (Figure 1.3).

Increasing subsidence rates in the Kimmeridgian resulted in the replacement of most of the Oxfordian carbonate facies with the deepwater siliciclastic facies of the Abadia formation. The thickness of this formation is 800-900m in most of the basinal areas (Ellwood 1987). This second phase of rifting continued into the Cretaceous. It resulted in the development of the sub-basins described earlier, and heralded the opening of the North Atlantic in the Albian/Aptian (Boillot et al.1985; Boillot 1986; Masson and Miles 1986). On the eastern margin of the Arruda sub-basin, horst blocks capped by shallow marine carbonates were bypassed by a siliciclastic fan system which prograded westwards. Supplied by basement material from the east, over 2.5km of sediment accumulated in the sub-basin during the Kimmeridgian (Leinfelder and Wilson, in press).

In contrast, to the south around Sintra a carbonate ramp developed, fed by a carbonate shelf system on the western margin of the basin. The deepwater carbonate debris flows of the Ramalhão formation are contemporaneous with the Abadia Marl formation. These are overlain by the Mem Martins formation which include deepwater reefal limestone (Ellis 1984). The Mem Martins formation ranges from middle Kimmeridgian to early Tithonian in age.

Inland around Torres Vedras and Montejunto, the Abadia formation is capped by up to 60m. of shallow marine sandstones and oolites (the Amaral formation (Ellwood 1987; Leinfelder 1987)). To the northwest, the Abadia formation passes into a platform marginal sequence at Consolação and continental, deltaic and occasional shallow marine facies at San Martino and Cabo Mondego (Ellwood 1987) (Figure 1.3 for locations). In coastal outcrops oolitic facies of the Amaral formation are not developed. At Santa Cruz, a shallowing-up siliciclastic sequence is terminated by an 8m-thick, intensely bioturbated, muddy sandstone. This unit is erosively overlain by continental sediments the subject of this thesis. This influx of continental clastic sediment during the late Kimmeridgian/Tithonian was widespread in the northern parts of the basin, reflecting a significant regressive event.

The uppermost Jurassic (-Cretaceous?) continental clastic facies are the subject of interest in this project. Formerly informally called the Grés Superiores (Upper Sandstones) their stratigraphic nomenclature is confused and age poorly constrained due to a lack of biota. In this thesis, the unit is termed the Lourinhã formation and divided into members. In the coastal sections examined, the Lourinhã formation is

600 to 1100m thick. To the east, around Arruda, clastics interfinger with shallow marine carbonates which in this region are far more prevalent (Leinfelder 1986). This is also the case in the Serra da Arrabida to the south of Lisbon (Figure 1.2) where easterly derived clastics interfinger with marine carbonates (Fürsich and Schmidt Kittler 1980; Felber et al. 1982).

Cretaceous strata are rather thin, only 200-300m in all, and reflect a much simpler facies pattern. From the Valanginian to the Cenomanian, shallow marine carbonates to the southwest interfinger northwards with coarse siliciclastics of high and low-sinuosity fluvial systems (R. Hiscott pers. comm.). These lie unconformably upon Lourinhã formation sediments, the uppermost units of which may be Berriasian in age (Rey 1972). Small amounts of igneous activity resulted in the extrusion of alkali olivine basalts in the Lisbon area and the intrusion of the Sintra granite (85 My.) during Santonian to Maastrichtian times.

1.4 AREA AND NATURE OF OUTCROP

Jurassic (Kimmeridgian/Tithonian) alluvial fluvial and deltaic, siliciclastic sediments crop out over a broadly triangular area of the Lusitanian Basin (Figure 1.7). Overall they have a shallow, generally southerly dip. The northern-most outcrop occurs in the Alcobaça area. To the south-east in the Arruda sub-basin, clastics interfinger with marine carbonates (Leinfelder 1986). In the south-west, as far as Porto da Calada, the sequence is rather more clastic dominated .

The Arruda sub-basin (Figure 1.3) has been mapped and studied by Leinfelder (1986), who also erected a lithostratigrapic scheme for the sequences found there. I have not examined this area in detail, although the implications of Leinfelder's work will be considered in the regional considerations at the end of this thesis.

Fieldwork for this study was confined to the broad strip of outcrop from Alcobaça south-eastward to Porto da Calada on the coast, an area some 70km by 13km

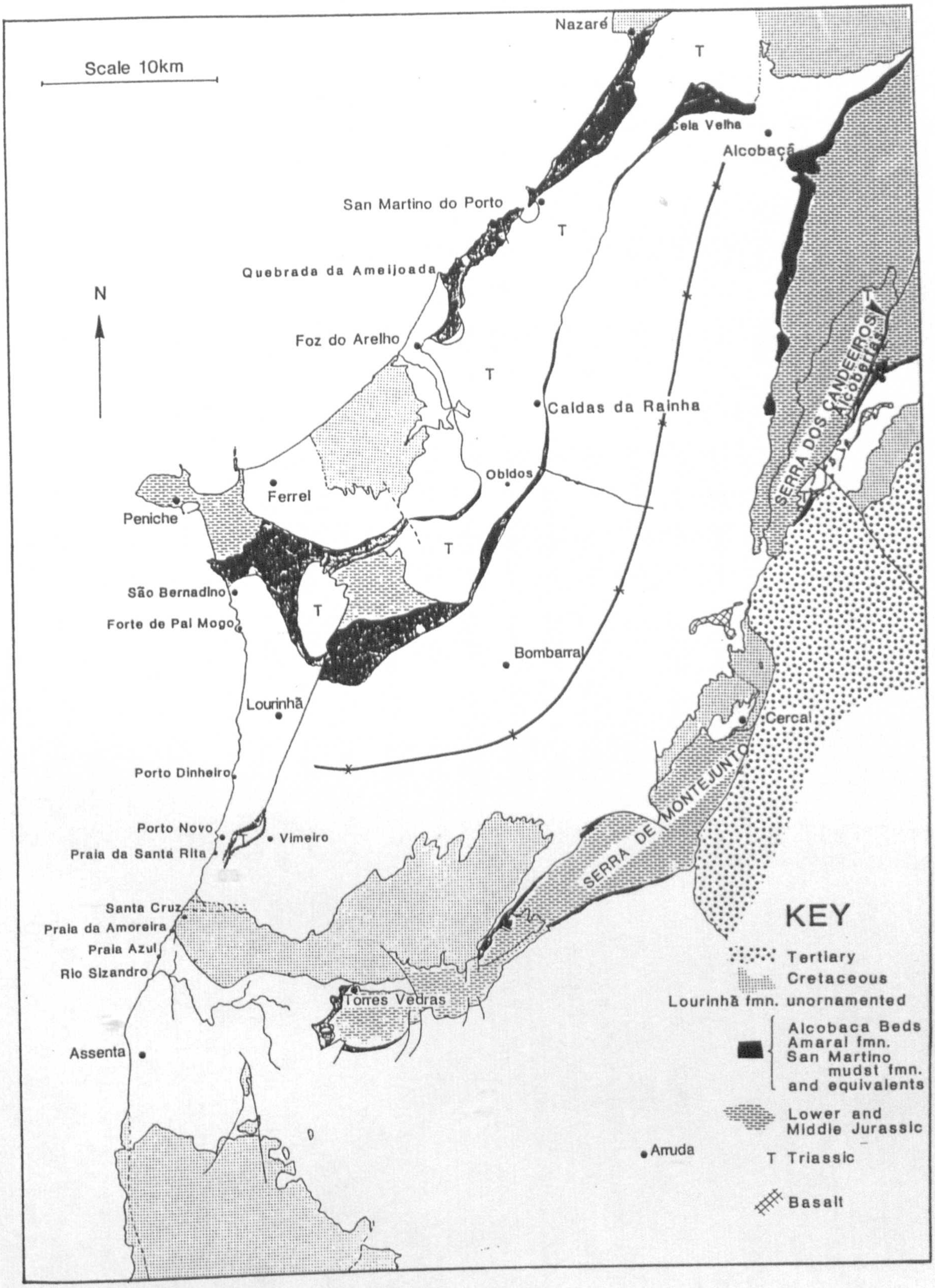


Figure 1.7 Simplified and modified geological map illustrating the outcrop area, in white, of Upper Jurassic alluvial, fluvial and deltaic siliciclastic sediments (Lourinhã fmn.) plus the units bounding these facies. Based on the geological maps of the Serviços Geologicos de Portugal sheets 26B,C,D and 30A,B,C. Important localities and major structural features also illustrated.

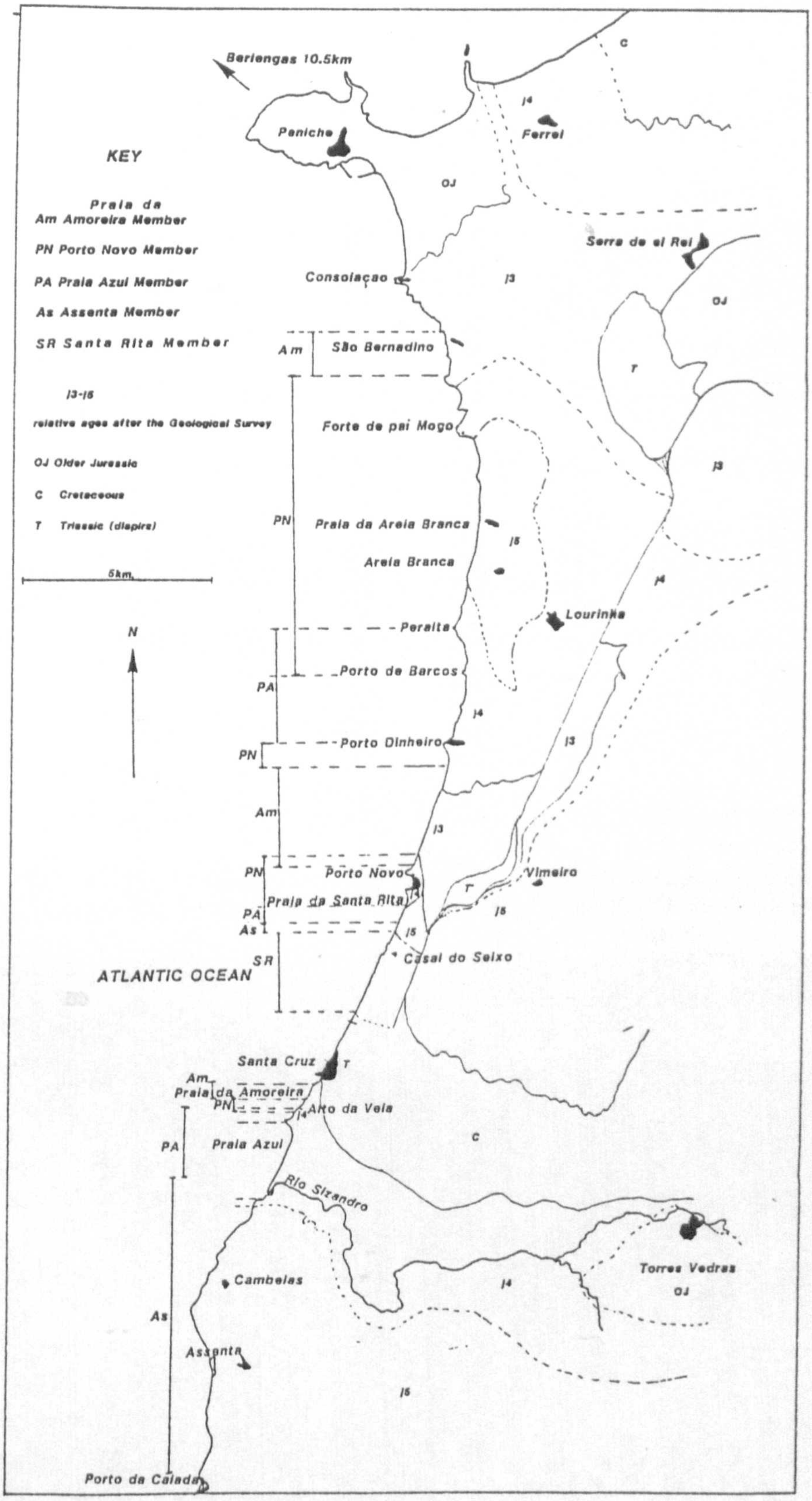


Figure 1.8 Simplified and slightly modified map showing the distribution of Mesozoic rocks as mapped by the Serviços Geologicos de Portugal. Particularly important are the purported relative ages of Kimmeridgian/Tithonian strata (J3-J5) and the deformation associated with the diapirs (T). Important locations and the distribution of lithostratigraphic units along the coastal sections are shown.

(Figure 1.7). By far the best outcrop occurs on the 30km of coastline south of Peniche, between São Bernadino and Porto da Calada (Figure 1.8). The almost continuous cliff outcrops, up to 70m high, allows the two and occasionally, three dimensional nature of the sequence to be documented. The majority of fieldwork concentrated on the detailed examination of parts of this section. There are a number of major breaks in the section which are associated either with the intersection of a diapiric trend with the coast or with a large fault. Inland, the only obvious break is provided by the Montejunto high. To the west of the southern limb of this structure a fault occurs and intersects the coast north of Santa Cruz (Figure 1.7). Elsewhere, outcrop in the clastic sequences inland is very poor, so that data was difficult to gather. However a study of of drainage patterns by Willis (1988) suggests that they are pervasively fractured. Three weeks were spent mapping facies distributions inland. The quality of data only allowed generalisations to be made regarding the nature of facies changes and the distribution of the members defined below.

Other, smaller coastal outcrops occur to the north of Peniche, north of Ferrel and between San Martino do Porto and Foz do Arelho (Figure 1.7). The Ferrel section has been logged and the data gathered is largely used to compare and contrast with facies developed in the more southerly sections. The base of the formation is not exposed at Ferrel but the section is otherwise almost continuous with only minor breaks, except for the upper 200m where outcrop is poor. The unconformity below the Cretaceous is well exposed.

The San Martino / Foz do Arelho section forms a continuous outcrop, faulted in places, which dips at 20° or more westwards into the sea. It is difficult to negotiate at most states of the tide and was only reconnoitred, no attempt was made to log it in its entirity partly due to the difficult nature of the outcrop and also the time available. Most attention was paid to the basal transition from carbonate dominated facies to siliciclastics (see section 1.6.3) and to general observations re the scale of channels and differences when compared with outcrops further south.

1.5 TECHNIQUES

Field techniques employed included the graphic logging of sequences to establish overall thicknesses and vertical facies changes. Additional more detailed short logs allowed the lateral changes of facies and differences in vertical succession to be determined. The amount of detail gathered depended partly on the quality of outcrop and partly on the time available. The most detailed work was concentrated along the section from Porto Dinheiro to about 2km south of the Rio Sizandro (Figure 1.8). In total 7 months fieldwork was carried out over 4 visits between November 1985 and July 1987.

The geometries of sand bodies are particularly useful in the interpretaion of environment. These were studied using field sketches and graphic logs backed up by photographs plus detailed examination of grain-size, structures and their variation. Much of the excellent coastal exposure was photographed from a boat, because scale distortion occurred in photographs taken close to the cliffs on the narrow beaches. This allowed the spatial relationships of different facies to be evaluated. A detailed examination of the architecture of the sequence (sensu Allen and Williams 1982) was not possible because of the paucity of marker horizons in all but the most southern sections.

Inland, efforts were concentrated on establishing the nature of facies transitions, particularly at the base of the Lourinhã formation and also any gross differences between the exposed rocks and those found in coastal outcrops.

Maps used included the 1:50 000 geological maps published by the Servicos Geologicos du Portugal (SGP) and the corresponding (and more recently updated) topographic maps of the Instituto Geográfico & Cadastral (sheets 26 B,C,D and 30 A,B,C). For the coastal outcrops more detailed 1:25 000 maps of the Carta Militar de Portugal were utilised. All grid references given refer to these latter maps. The former do not have a superimposed grid system.

Representative samples of the various lithologies, and certain of the fauna were collected from the coastal sections. Thin sections of a selection of these were cut and a few polished blocks examined. A comprehensive petrographic study has not been carried out on these specimens largely due to time constraints.

1.6 STRATIGRAPHIC NOMENCLATURE

1.6.1 Introduction

The Upper Jurassic stratigraphy of the Lusitanian Basin was reviewed by Wilson (1979) and Ellwood (1987). New terms were proposed by Ellwood (1987) for a number of the units below the Lourinhã formation.

The present stratigraphic terminology for the region is a complicated mixture of bio- and lithostratigraphy which can be confusing. This is partly due to the wide variety of facies present in the basin. Biostratigraphic maps published by the SGP apparently attempt to use lithostratigraphic sub-divisions where ages are poorly constrained.

Much of the terminology used by the SGP dates back to Choffat (1901). His *Lusitanian* (lower calcareous) and *Neojurassic* (upper clastic) stages were originally defined in the Torres Vedras area where the Upper Jurassic has a large marine component. For the continental-clastic parts of the Lourinhã formation which are lacking in biota, this scheme appears to have been applied to lithological units. Consequently there is an admixture of terms such as Lusitaniano superior, Pteroceriano, Portlandiano and Kimeridgiano, with different terms apparently used for the same unit on different maps. In addition local, probably lithostratigraphic, (eg.'Camadas de Alcobaca') and general ('Grés Superiores') terms are used on map keys. The Grés Superiores appears variously at the top of the Lusitaniano and above the Lusitaniano. The Lusitaniano appears in the mid. and lower Kimeridgiano or below the Kimeridgiano. It is not clear if this implies diachroneity. A table summarising the confused terminology was published by Wilson (1979). In short, the age of the Grés Superiores is poorly constrained as upper Kimmeridgian/Tithonian and the nomenclature confused. Ages attributed to lithological units are not consistent and must be regarded as unreliable. Recent work by co-workers at the Open University and in Portugal and West Germany has begun to erect what is hoped to be a formal lithostratigraphy for the basin (Figure 1.6). The following section deals with the terminology proposed for the Upper Jurassic continental clastic sequences studied by the author.

1.6.2 The Lourinhã formation

The lithostratigraphic scheme outlined in Figure 1.9 was originally based on the coastal sections south of Peniche. Exposure is more-or-less continuous for 30 km between São Bernadino and Porto da Calada (Figure 1.8). There are a few small breaks in section and one stretch north of Santa Cruz, where Cretaceous and Triassic sediments crop out. Subsequent mapping of inland and further coastal outcrops has confirmed the general applicability of the scheme. Figure 1.10 is a schematic N-S cross section related to the Ferrel section plus those to the south of Peniche. The section is based upon the correlation of graphic logs and other field data. Naturally there are a large number of assumptions (listed in Appendix 1) made in such a cross section.

The sequence is approximately 600 to 1100m thick. The base is defined by the onset of continental fluviatile conditions and is exposed at São Bernadino, Santa Cruz and also south of San Martino do Porto. The upper boundary, at the base of the Cretaceous continental clastic Torres Vedras formation, is exposed at Porto da Calada and north of Ferrel. Neither boundary is well exposed inland, but facies transitions over a limited distance have been observed and often have a topographic expression.

The main characteristics of Grés Superiores sediments are channelised coarse sediments always found in sequences characterised by significant volumes of, frequently red, mudrock. Soil profiles of caliche type, developed to various degrees, are common.

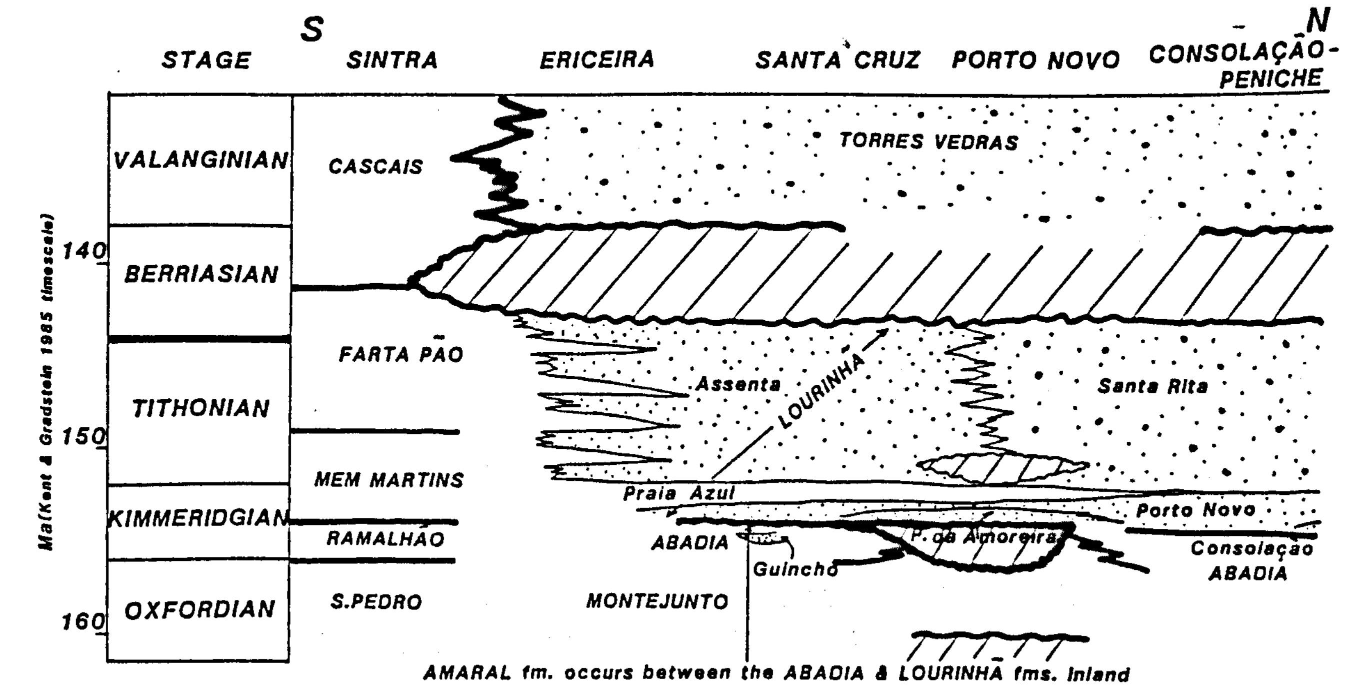


Figure 1.9 Simplified and as yet informal lithostratigraphy for the Upper Jurassic and Lower Cretaceous of the Lusitanian Basin. Cross-section is essentially north-south along the coast. Continental facies stippled, marine strata unornamented. Proposed formational boundaries are heavy lines. The Lourinhã formation is comprised of five members.

It is proposed to convert the Gres Superiores to formation status, with member status attributed to variable lithological units which characterise different sedimentary environments. The name of the town Lourinhã (G.R.782 438, Figure 1.8) is selected as the formation name. Lourinhã is situated on the continental sediments and is centrally placed in relation to the coastal type sections outlined in the following discussion.

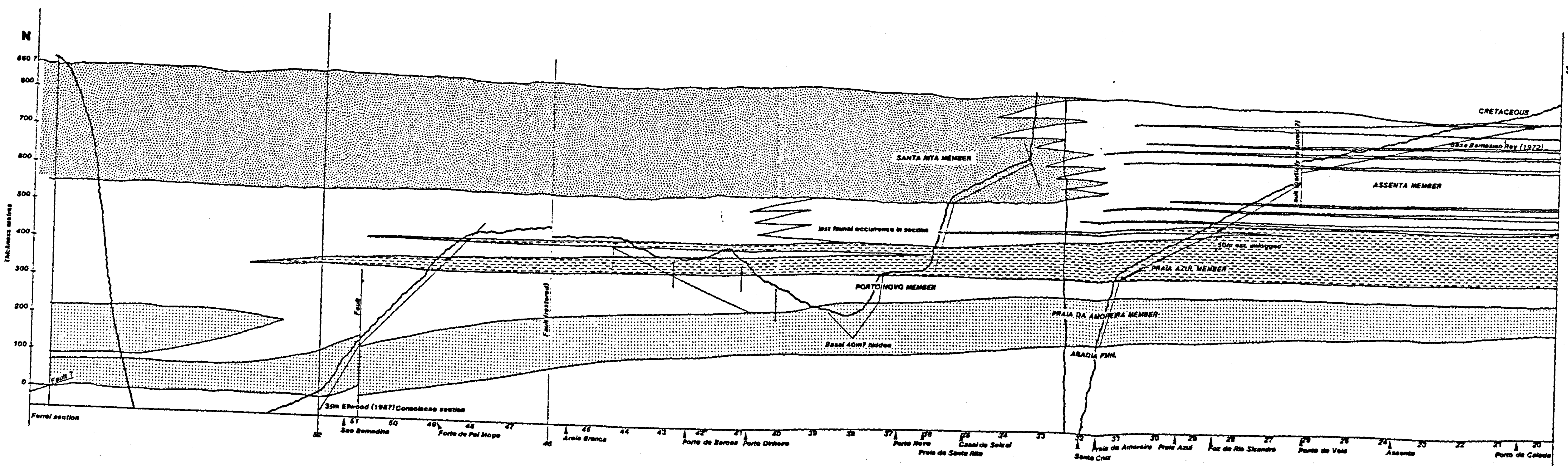


Figure 1.10 Schematic N-S cross section of the Lourinhã fmn. based on the correlation of graphic logs, plus other field observations. Assumptions made are listed in Appendix 1. The base of the formation may be diachronous as it thickens significantly northwards. The section is 'hung' on the base Praia Azul and Santa Rita member markers. Vertical and diagonal lines (unless otherwise stated) represent logged section. That at '52' denotes the start of the São Bernadino to Porto da Calada section. Marine horizons denoted by heavy black lines. Numbers represent grid northings from military maps. Irregular heavy line marks the present level of erosion in the sections. Note that the break between the Porto Novo and Assenta members is probably rather gradual, but that between the Santa Rita and Assena members is probably rather abrupt (see Chapters 5&6).

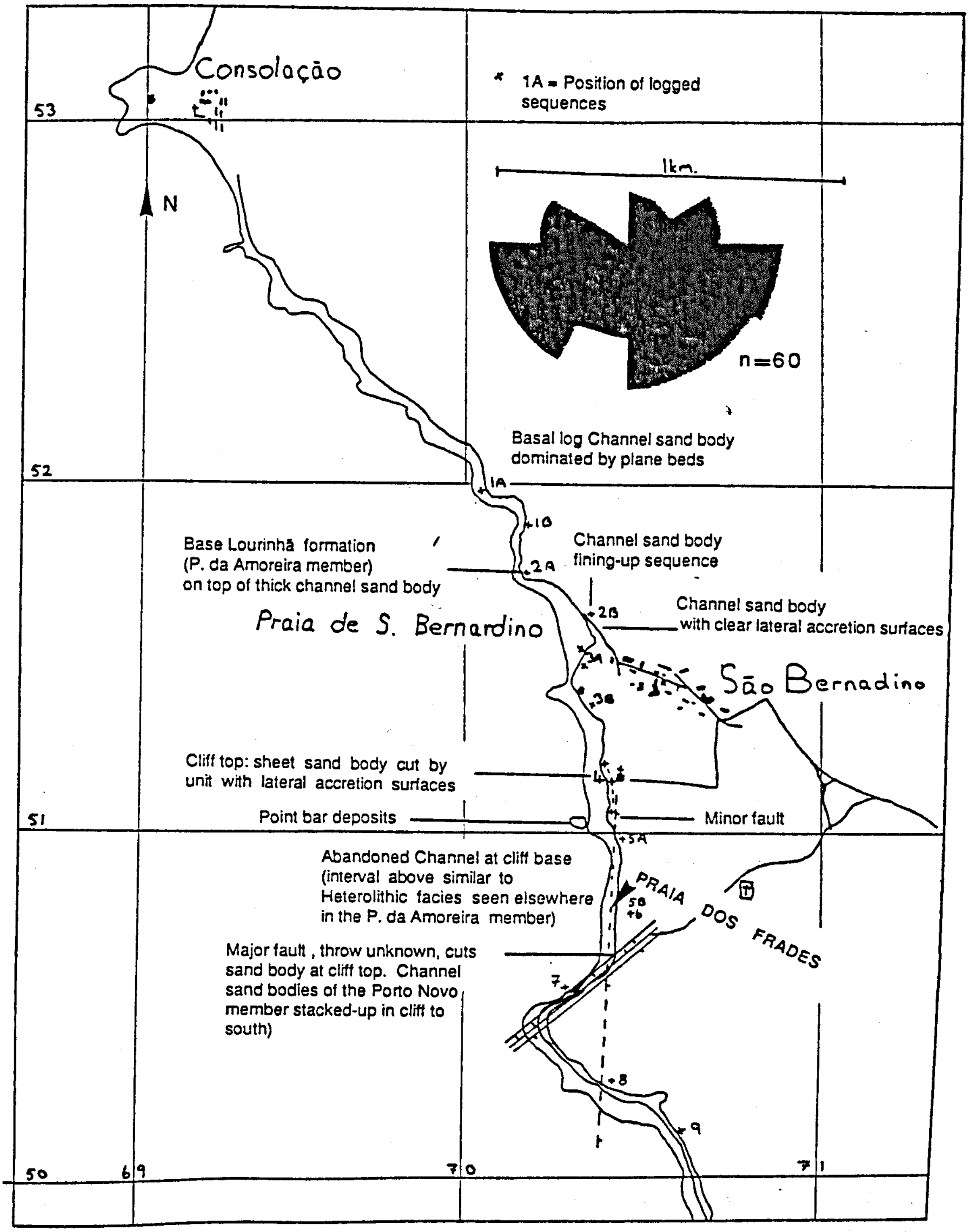


Figure 1.11 Sketch map of the São Bernadino section. The locations of logged sections and important features are indicated. Palaeocurrent data presented is that collected from the Praia da Amoreira member outcrop between the major fault and the base of the Lourinhã formation. The fault marks the base of the Porto Novo member outcrop which runs to the south. It was probably active during the deposition of those sediments (see Chapter 3).

1.6.3 The basal boundaries of the Lourinhã formation

São Bernadino

Logs SB 1&2 cross the formational boundary. Just north of Praia da São Bernadino (Figure 1.11) is a 9 metre-thick fluvial channel sandstone, interpreted as a delta distributary by Ellwood (1987) (Log SB 1). Planar bedding and trough cross bedding dominate various parts of the channel which is composed of coarse yellow sandstone. In its upper part are a number of large planar–tabular cross beds up to 1m thick. The foresets of these units occasionally show small slump and water-escape features, indicative of rapid rates of sedimentation. Large blocky mudclasts up to 100mm square are common in the lower half of the channel.

About mid-way along its outcrop the lower third of the channel is cut by a deep, steep-sided scour. Rather more variable sediments fill the scour, including caliche and mudclast-pebble conglomerates rich in organic debris, plus a number of logs up to 1m in diameter, interbedded with thin cross-bedded sands. The top of the channel is characterised by decimetre bedded fine, cross-bedded sands, interbedded with thin, light grey marls. In places these sediments are structureless and rather mottled, probably a pedogenic modification.

Immediately above this channel are a series of much coarser arkosic, rather kaolinitic sands interbedded with brown and green mottled silts. At the cliff top at GR. 7012 5175 a caliche soil profile is developed (44m Log SB 2), consisting of irregular carbonate nodules up to 10mm in diameter in mottled silts. The petrographic change, mudrock reddening and appearance of caliche mark the onset of continental sedimentation. Therefore, the base of the Lourinhã formation lies directly above the major channel sand body described earlier i.e. the 36m level in log SB 2A. This boundary modifies slightly that defined as the top of the Abadia formation by Ellwood (1987) at this locality. He placed it at the *base* of the major channel. The top is felt to be more appropriate because:

- there is a distinct petrographic change from yellow coarse sands to the coarse, granular, arkosic and kaolinitic sands of the basal member of the Lourinhã formation.
- yellow sands are similar to those found in channels below in the Abadia formation
- soils occur above, but not below the channel.
- mudrocks are grey beneath the channel and red, brown and green above.
- abundant organic debris and the large logs are not a major feature of the basal member of the Lourinhã formation in coastal outcrops south of Peniche.

Santa Cruz

The section to the west and south of Santa Cruz (Figure 1.12) lies on the margin of a salt diapir. As a result of the diapirism the beds dip steeply to the west. Severely disrupted red Dagorda marls are faulted against the Abadia marl formation. Described and interpreted by Ellwood (1987), the Abadia marl formation consists of deep basinal marls with thin turbidites and occasional, spectacular, coarse submarine channel fills.

Storm beds and synsedimentary slump scars denote a progressive shallowing-upward, which terminates in a thick shallow-water submarine channel unit, to the south of Praia Formosa (Figure 1.12). The top 8 metres of this unit, (Santa Cruz Section Basal Log 0-8 m) consists of grey, coarse, granule-rich, muddy sandstone largely homogenised by bioturbation. A few parallel and ripple laminated, sideritic, dark grey silts and brown fine sandstones, are preserved in the uppermost part of this bed. Lignitic debris is common and bed surfaces are colonised by small clusters of nerineid gastropods. This is erosively overlain by the first of a number of coarse to very-coarse grained, fluvial channel sandstones (Figure 1.13). The basal sand body is rather structureless and iron stained but those above are clearly trough and planar cross bedded. In addition, lignitic debris is largely absent. Interbedded with the channel sands are dark red mudstones with caliche and decimetre bedded sands and silts. The transition between marine and terrestrial facies is quite distinct at this location. The base of the Louinhã formation being coincident with the base of the first fluvial channel sand body.

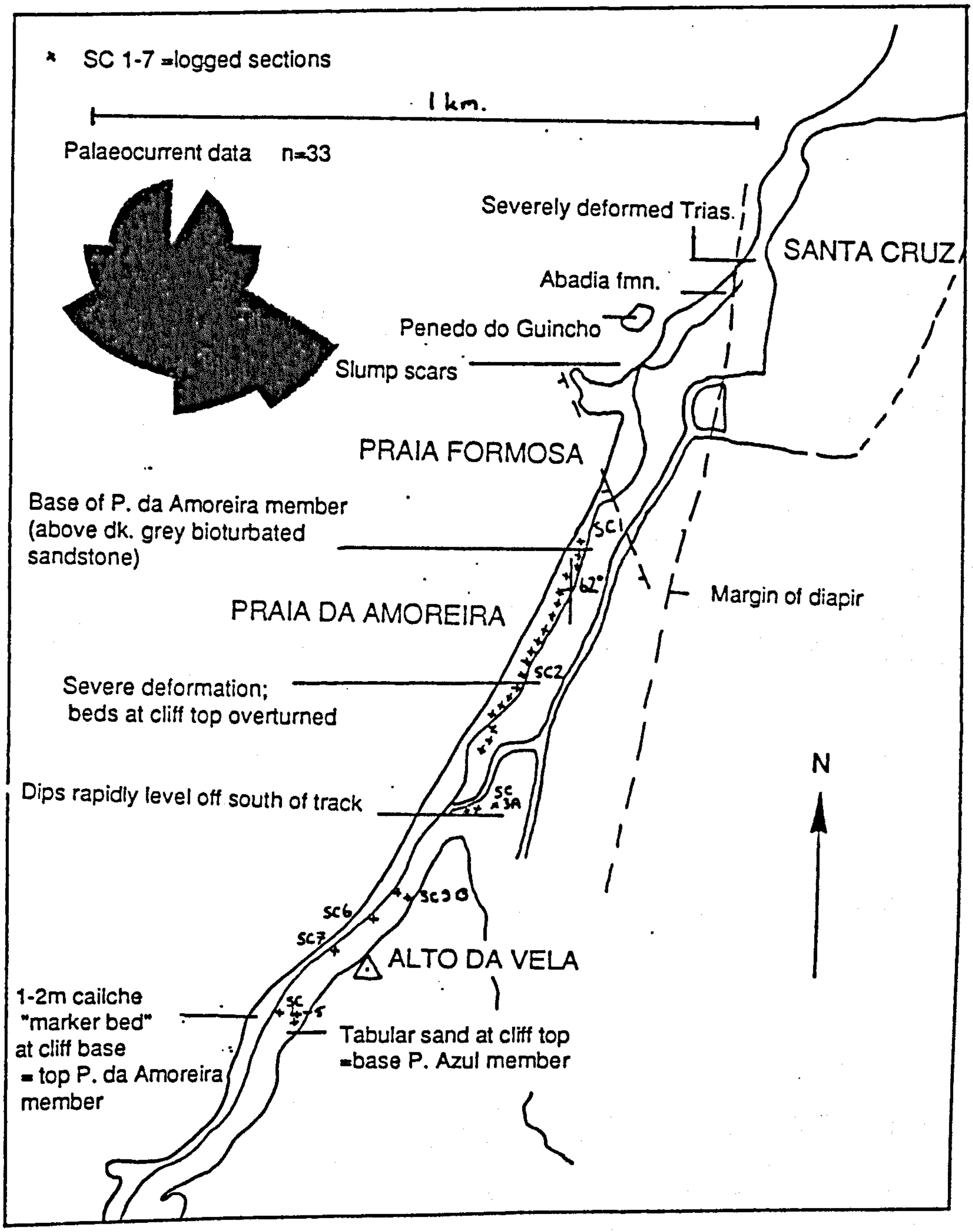


Figure 1.12 Sketch map of the Praia da Amoreira section. Showing the location of logged sections and other important features. Palaeocurrent data indicates a moderately sinuous system with a SSE mode. Position of diapir and fault after Ellwood (1987). Heights in metres.

Quebrada da Ameijoada

The base of the formation is exposed at one other coastal locality, Quebrada da Ameijoada,which lies 3.5km north of Foz do Arelho and 7.5km south of the inlet at San Martino do Porto (Figure 1.7). Log FA 1 records the nature of the succession immediately beneath the Lourinhã formation.

Figure 1.13 GR. 6675 3150 Praia da Amoreira, base of the Lourinhã formation, facing due S. There is a clear contrast between the first erosive-based, pale, kaolinitic fluvial sand and the dark grey marine sands beneath. Trough cross bedding can clearly be seen in the second fluvial channel. Intense bioturbation has resulted in completely homogenised beds on the left.

The sediments south of San Martino are essentially a repeat of the succession to the north of the inlet described by Ellwood (1987) and dip at ca.20° to the west. A continental fluviatile unit,the San Martino mudstone formation, is sandwiched between two shallow marine carbonate/clastic units. The lower of these, termed the

Pholadomya protei beds (Ellwood op cit.) were not examined by the author. Log FA 1 starts at the top of the San Martino mudstone formation. Continental sediments are initially interbedded with and then abruptly overlain by massively bedded micritic carbonates containing a sparse but varied fauna of corals, gastropods, bivalves, algae, stromatoporoids and echinoderms. Bed bases exhibit large thalassinoides burrows (Figure 1.14).

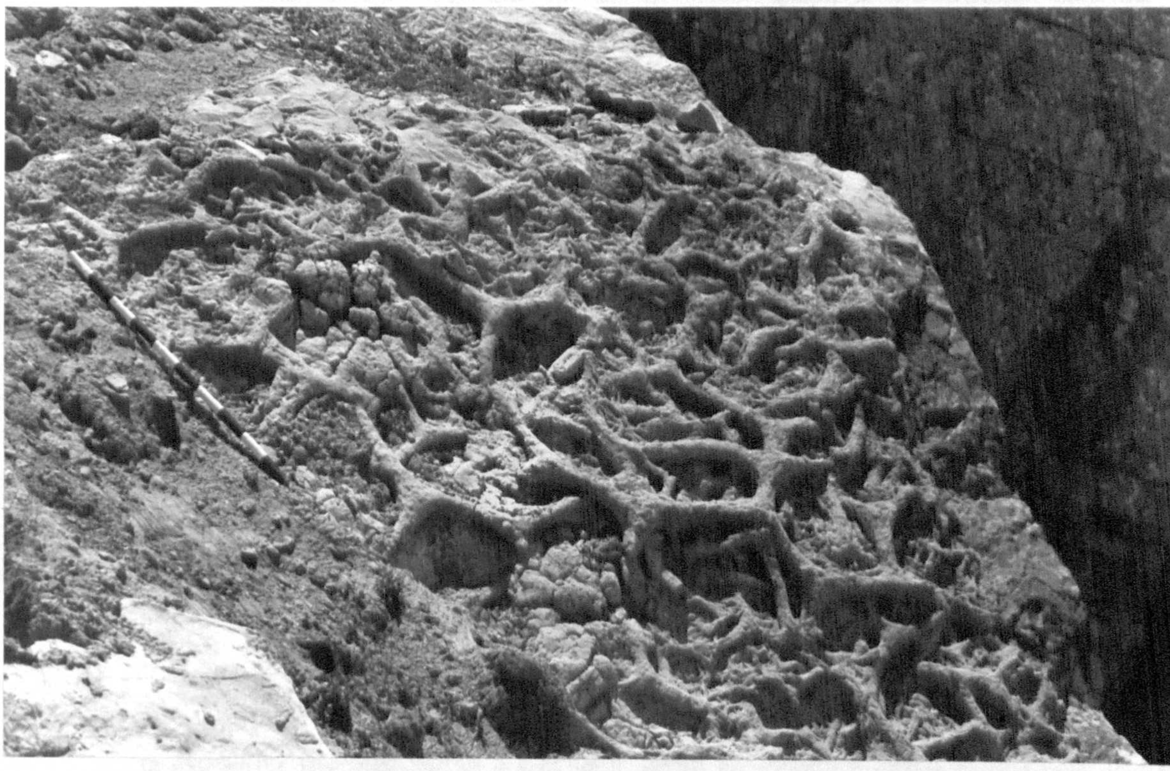


Figure 1.14 Large scale Thalassinoides burrows in micritic limestones on the base of an upturned block, Alcobaça beds, Quebrada da Ameijoada. Scale bar = 1m.

Designated the Alcobaça beds by Camarate França and Zbyszewski (1963) in the map memoir, this terminology was followed by Ellwood (1987). It is possible that they are the facies equivalent of the Amaral formation to the south. The massive micrite beds have thin marl partings which locally exhibit faint laminae.

Near the base of the Alcobaça beds (at 11 and 13.5m on the log) are two slightly thicker units of very fine siliciclastic sand containing lignitic debris. Both units have at their bases intraclasts of lithified micrite. The lower unit also exhibits inclined bedding (Figure 1.15) and may represent a small bar or possibly a sinuous channel cutting across a carbonate platform.

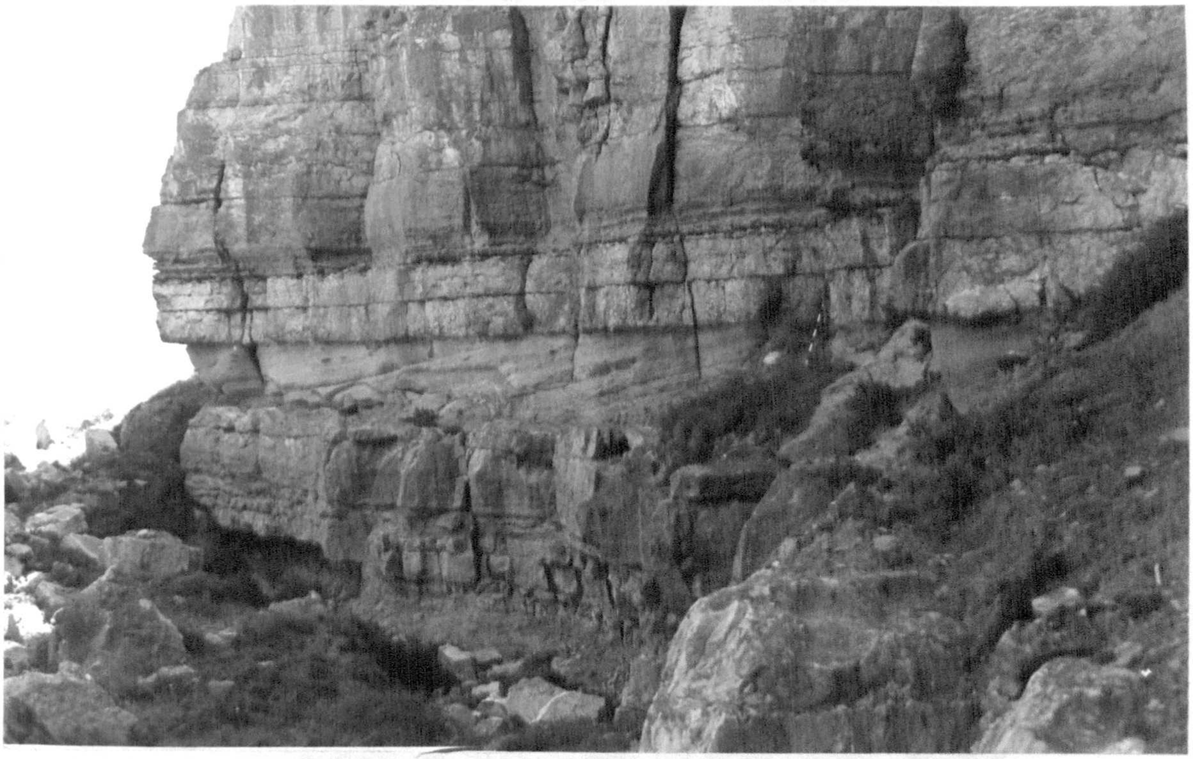


Figure 1.15 Cross bedded siliciclastic sand containing micrite intraclasts within the dominantly carbonate Alcobaça beds. This unit is probably the deposit of a bar or a small sinuous channel, which migrated across a carbonate platform. Quebrada da Ameijoada. Scale bar = 1m.

The upper 15m of the Alcobaça beds, which are in total some 38m thick, consists dominantly of grey laminated marls exhibiting both wave and current ripples. The transition to dominantly terrestrial fluviatile sediments is largely obscured by debris. In the cliff section (Figure 1.16) the transition takes the form of massive red-brown silts with occasional, erosive based, lenticular sandbodies 1-3m thick and tens of metres wide. Lateral accretion bedding has been identified in a number of these units. Marginal to these sandbodies are a number of laterally persistent, structureless, decimetre - scale fine sand beds, interbedded with mudrock.

Figure 1.16 Transition between the Alcobaça beds and the base of the Lourinhã fmn.(Porto Novo mbr.) at Quebrada da Ameijoada. Camera facing due south. The Alcobaça beds form an irregular carbonate pavement in the foreground and the Lourinhã fmn. crops out in the vertical cliff face. Prominent bed just below the two large blocks is the last micrite, at 37m, on Log FA1. Above this unit is a rapid transition from grey to red-brown massive mudrocks with small, lenticular channel sandbodies (eg. at centre right arrowed) and laterally persistent decimetre-metre scale, structureless sheet sands. The spacing of these latter units is greatly reduced in the upper part of the cliff face. These are incised into by channel sands exhibiting probable lateral accretion surfaces (also arrowed). Cliff height 100m.

Log FA 2 passes through two of the lenticular sandbodies, interpreted as channel deposits, which crop out some 10m above the marine/terrestrial transition. From their characteristics the channels clearly form part of the Porto Novo member (see section 1.6.6).

The base of the Lourinhã formation inland

As stated earlier, the actual base of the formation has not been directly observed inland. It is however possible to map the boundary, using topographic features and occasional closely spaced outcrops, plus abrupt changes in the colour of soils in the fields.

Eastern and western margins of the Bombarral sub-basin : To the east and west of Alcobaça the published map (Camarate França and Zbyszewski 1963), indicates the presence of the Alcobaça beds (Camadas de Alcobaça)(Figure 1.7). In the east this unit, which is rather variable, is generally very poorly exposed. It often exhibits a low topographic relief, forming a flat ploughed area with grey soils between the scrubby irregular limestone terrain of the Montejunto formation and the rolling relief and red soils of the Lourinhã formation. Sedimentary facies consist of grey marls and thin buff limestones with a mixed fauna of bivalves (including oysters, *I. lusitanicum* and *E. securiformis* (see Chapter 4)), stromatoporoids, gastropods, small solitary corals, colonial corals and echinoderms. Carbonate lithologies include biomicrites and slightly sparry, peloidal grainstones. Similar features occur as far south as the Bolhos diapir SE of Peniche (Figure 1.8).

On the eastern flank of the Caldas diapir, outcrop is generally very poor indeed. Immediately east of Celha Velha in a steep walled road cutting (Figure 1.7) carbonate beds, presumed to equate with the Alcobaça beds elsewhere, dip at ca. 35° to the east. Lithologies consist of massively bedded oolitic and pisolitic grainstones. A number of these are reddened and rather brecciated. Up dip thinner bedded micrites crop out. The brecciation of some of the carbonates suggests faulting at the margin of the diapir. Elsewhere there is clear evidence of this, eg. at GR 8620 5545 in a road cutting 1.5km south of Obidos (Figure 1.7), where a variety of carbonate lithologies are severely brecciated. These sediments abruptly pass into red marls of probable Triassic age at their base.

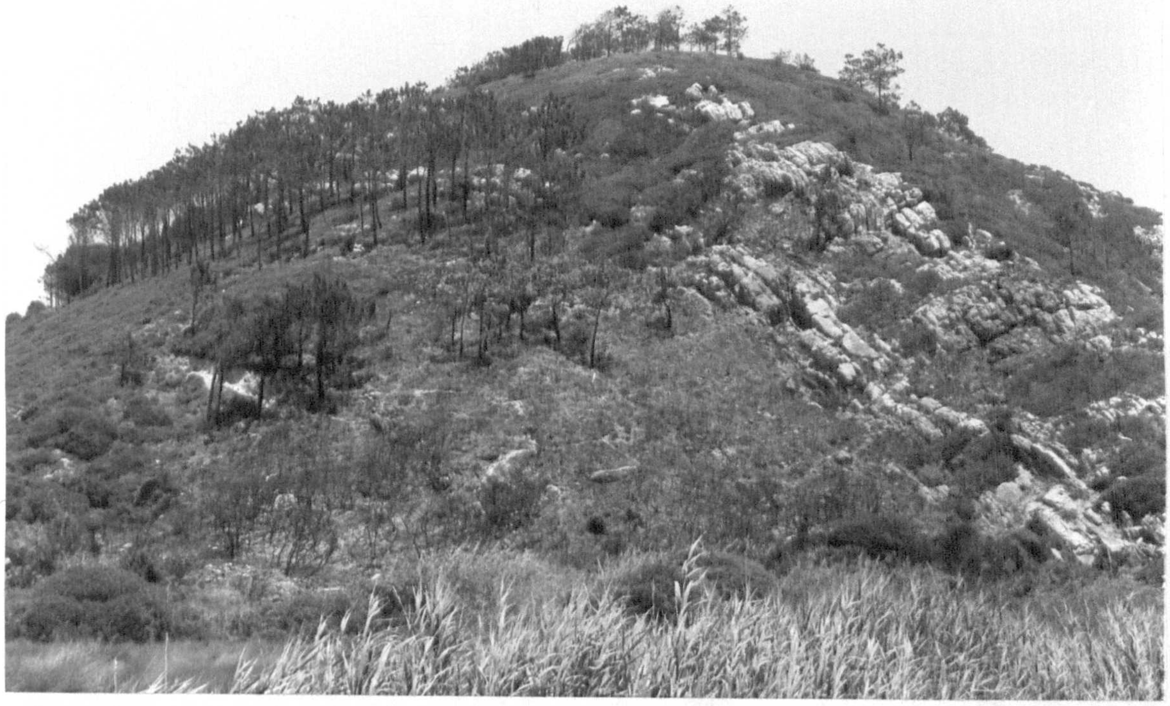


Figure 1.17 Southwest flank of the Vimeiro diapir. Camera facing NE. Hill height ca. 50m. Slightly overturned Kimmeridgian carbonate beds pass into P.da Amoreira member sediments across the white scar in the small stand of trees at lower left. It is possible that locally the contact is not faulted but has an erosive nature (see text).

All the above examples pass vertically into continental sequences dominated by red silts interbedded with buff micaceous channelised sandstones. The scale and dominant grain sizes of the channel sand bodies is rather variable. In the west they are generally small, only 1-3m thick, fine grained and of limited lateral extent. The scale of sand bodies increases significantly up-section. To the south and east the scale of sand bodies and grain sizes tend to be greater at lower levels. Sandbodies also tend to be sheet like. Based on their lithological characteristics the basal units of the Lourinhã formation in these inland localities seem to belong to the Porto Novo member (see section 1.6.6).

The Vimeiro diapir : This structure lies close to the coast to the east of Porto Novo (Figure 1.8). Its flanks consist of steeply dipping and overturned thickly bedded carbonates (Figure 1.17). Lithologies include shelly micrites, both wackestones and mudstones, plus peloidal packstones and mudstones. These units are of probable

Upper Kimmeridgian age (M. Ramalho pers comm.) based on the presence of the dasycladacean algae *Campibelliela striata* (Ramalho 1971). As such they are probably coeval with the Alcobaça beds to the north and the upper part of the Abadia formation and the Amaral formation to the south.

Figure 1.18 Maceira road section on the western flank of the Vimeiro diapir, facing due north. Vertically dipping, highly coloured P. da Amoreira mbr. sediments. Scale bar = 1m. No difference in dip can be demonstrated between these units and the carbonates to the east.

Willis (1988) mapped the diapir and found it complex in terms of both structure and facies variation. On its western margin, where outcrop is at its best, though still rather poor, the carbonates are observed to pass into coarse kaolinitic sands and red mudrocks which also have a steep, near vertical dip (Figure 1.18). The contact, though not well exposed, does not appear to be disturbed by faulting everywhere, particularly in the cutting on the road to Maceira (Figure 1.5).

An outcrop at GR 7003 3680 to the north of the Vimeiro road (Figure 1.17), reveals no difference in dip across the boundary and some rounded limestone pebbles at the contact. This may indicate re-working rather than brecciation in a fault zone. The lithological characteristics of the continental sediments above the carbonates indicate them to belong to the Praia da Amoreira member (see section 1.6.6).

The Amaral formation : This unit crops out below the Lourinhã formation on both flanks of the Montejunto structure (Figure 1.2b, Leinfelder 1986) and to the south and west of Torres Vedras (Ellwood 1987, Figure 1.7). It is a unit of variable thickness (11-60m Ellwood op cit.). Lithologies present include marls, sandstones and quartz cored oolites, with local patch reefs (Ellwood op cit.). Ellwood interpreted the sequence as a mixed carbonate/siliciclastic shelf system into which fluvially dominated deltas prograded. The Amaral formation lies above the marine slope systems of the Abadia marl formation and represents the transition betwen marine and terrestrial facies.

The Lourinhã formation is particularly poorly exposed in areas of Amaral formation outcrop. From the limited data available, fine to medium buff sands and red mudrocks, the Amaral formation appears to be overlain by sediments pertaining to the Porto Novo member.

1.6.4 The upper boundaries of the Lourinhã formation

Porto da Calada

The long, shallow dipping sections of the Lourinhã formation between the Rio Sizandro (GR. 6555 2845) and Porto da Calada (GR. 6400 2075)(Figure 1.8), consist of dominantly continental sediments with occasional marine intercalations (see section 1.6.6 and Chapter 5) including a number of thin nodular limestones and a coral bed. Large volumes of red and grey mudrock characterise the Lourinhã formation.

Rey (1972) believed there was a conformable transition into a 22m thick, Berriasian peritidal sequence, of ferroan dolomites, fine parallel laminated sands and thin marls, cut by channelised, lignite-rich, cross-bedded sands, up to 6 m thick. This is overlain by coarse fluivial sands with strongly erosive bases. Composed of coarse, white, kaolinitic sandstone rich in quartz pebbles, the fluvial unit crops out extensively inland and is quite distinctive. It is thought to be of Valanginian age. Rey (1972) attributed a Berriasian age to the peritidal sequence because of its position beneath the Valanginian sands. In addition he cites the presence of a fauna including the large foram *Anchispirocyclina lusitanica* and an abundant species of *Myophorella* different to that found in the Lourinhã formation to the north.

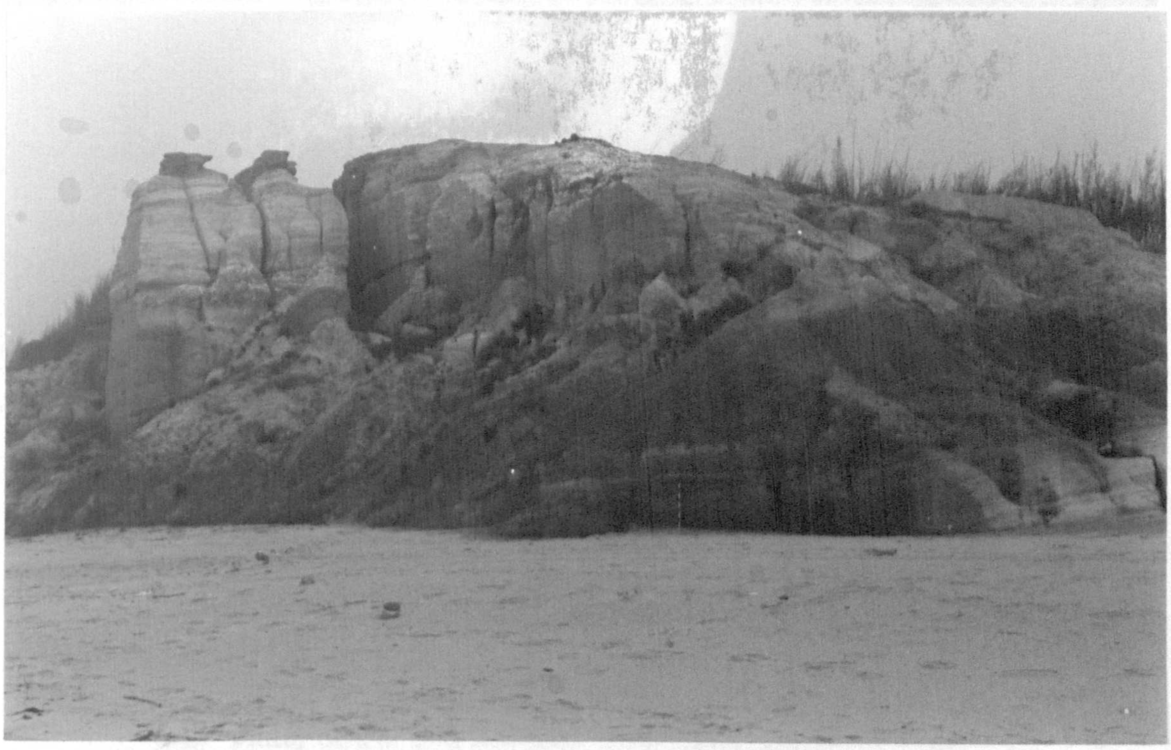


Figure 1.19 Top of the Lourinhã fmn. north of Ferrel. Scale (1m) sits by the last channel sand body of the Santa Rita member overlain by scarlet mudrocks. The massive sands above mark the base of the Cretaceous Torres Vedras fmn. (see Figure 1.20). It is not possible to prove an angular unconformity at this locality.

The sequence at Porto da Calada has recently been studied by R. Hiscott and the following is the result of personal communication. He concluded that there were insufficient differences to separate Rey's Berriasian from the upper Jurassic lithostratigraphically and in addition palynological data indicates possible ages ranging from early Tithonian to late Berriasian. In the light of such uncertanty, it is proposed that the upper boundary of the Lourinhã formation at Porto da Calada is marked by the strongly erosive, probably unconformable base of the (fluvial) Torres Vedras formation.

The Ferrel section

Figure 1.20 Typical appearance of the Torres Vedras fmn. Light grey-white, quartz and kaolinite rich sandstones with large scale cross beds. Mudrocks are largely absent. The sediments are probably the deposits of low sinuosity (braided ?) fluvial systems (R. Hiscott pers comm.). Sand bodies are clearly multi-storey (note change in scale of structures across the intraclast-lined erosion surface). Scale bar = 1m.

The contact with the base of the Cretaceous is also exposed to the north of Ferrel, at GR 7440 6000 (Figure 1.19). Coarse gravelly channel sediments and maroon mudrocks belonging to the Santa Rita member (see section 1.6.6 and Chapter 6), are erosively overlain by white pebbly sandstones identified as possibly Neocomian in age by Camarate França et al. (1960). These Cretaceous facies are quite distinct (Figure 1.20), being quartz rich, rather kaolinitic and white. Mudrocks are absent. It is not possible to prove a difference in dip between the Santa Rita member and the overlying units.

Inland outcrop

Cretaceous continental facies overly the Lourinhã formation in a number of areas, notably north of Alcobaça, around Cercal, south and east of Vimeiro, southeast of Alcobertas and south and east of Assenta (Figure 1.7). The transition, though rarely well exposed often has an expression in both relief and land use. Low rolling relief and agriculture on red soils associated with the Lourinhã formation, often gives way to steep relief and forestry (notably eucalyptus and coniferous stands) on pale acid soils. Where the outcrop pattern is indented e.g. to the east of Vimeiro (Figure 1.7) flat lying valleys used for agriculture are flanked by low, white, forested hills.

1.6.5 Summary of the formational boundaries

The upper and lower boundaries of the Lourinhã formation are each exposed at a number of localities in coastal sections. The base of the formation, denoted by the first influx of terrestrial fluviatile sediments, crops out north of São Bernadino, south of San Martino and also south of Santa Cruz. At the former it reflects a transition from shelf-edge deltas, at Quebrada do Ameijoada a transition from shallow marine carbonate shelf to muddy fluviatile systems and the latter a coarse terrestrial influx over an emergent shallow submarine channel.

The transitions observed in most cases, both on the coast and inland seem to reflect shoreline progradation into a shallow marine shelf area. In contrast, the coastal outcrops at São Bernadino and Santa Cruz are the result of a regression, apparently quite rapid, which may have been accentuated by uplift at the basin margin which lies to the west, off the present day coastline (see Chapter 2). It is not known if the base of the formation is significantly diachronous due to the problems of dating outlined earlier. However, this is likely particularly in the north where the formation is significantly thicker. In the coastal outcrops to the south of Peniche and inland around the Vimeiro diapir, the Praia da Amoreira member marks the base of the formation. Elsewhere the Porto Novo member appears at the base.

The top of the formation, marked by a sequence indicating frequent fluctuation in sea level at Porto da Calada, is rather different further north, consisting of gravelly continental fluvial deposits. The Cretaceous Torres Vedras formation lies unconformably on the Lourinhã formation at many localities both inland and on the coast. The fact that this boundary is unconformable is confirmed by examination of the published maps, notably Zbyszewski (1966), where cross cutting relationships are clearly visible. At Cercal (Figure 1.7), the Torres Vedras formation rests upon Oxfordian carbonates the Lourinhã formation being locally absent (Rey 1972). Cercal lies on the northeastern limb of the Montejunto anticline which separates the Bombarral and Arruda sub-basins. It is probable therefore, that this was a structural high late in the upper Jurassic, possibly due to halokinesis. This has a significance when the distribution of the younger members of the formation are discussed (see section 1.6.6 and Chapters 5&6).

1.6.6 The five members of the Lourinhã formation

There are five members of the Lourinhã formation each distinguished on a number of features:

- the petrography and grain size of the coarse fractions.
- the provenance of pebble-grade clasts.
- the geometry and typical structures of channel sandstones.
- the relative abundance of plant debris.
- the presence or absence of a macrofauna.
- types of trace-fossil found.
- colour of mudrocks.

- the dominant type of soil profile developed.

A summary of the salient features of the five members is shown in Table 1.1.

Figure 1.8 shows the location of the type sections and the distribution of the members in other parts of the main coastal sections. Two of the members (Praia da Amoreira and Porto Novo) occur at the same level ie. the base of the Lourinhã formation. The remaining three members, Praia Azul, Santa Rita and Assenta, reflect changes in environment in vertical succession and also laterally, as the sequence is progressively more marine influenced towards the south.

The Praia da Amoreira member

Location

Both the base and top of this member are exposed in the type section. The outcrop stretches from the northern end of Praia da Amoreira, to the southern end at the base of the cliff below Alto da Vela (GR's. 6675 3150 and 6635 3095 respectively). There are two means of approach to the section (Figure 1.12). From the car park east of Penedo do Guincho, down the steps and round the base of the cliff on the south side of Praia Formosa to the base of the section. Alternatively drive/walk south along the track at the cliff top and take the steep track down to the right. This brings you to mid-section.

Description of section

The lower part of the section dips steeply at over 60° to the west (Figure 1.21) and in places is overturned towards the cliff top. The deformation is due to the presence of the Santa Cruz diapir. Dips rapidly decrease to the south of the intersection of track with beach (Figure 1.21). The base is defined by the first fluvial sandstone above a grey, intensely bioturbated, muddy marine sandstone .

Laterally very extensive channel sands up to 11m thick, with locally rather irregular bases, characterise the lower half of the section. These are interbedded with dark

Table 1.1 Characteristics of the five members of the Lourinhã formation

Member	Lithology of cs. fraction	Sand body geometries	Plant debris	Mudrocks	Fossils	Other
P. da Amoreira	coarse arkosic & kaolinitic sandstones-granulestones; granite and phyllite pebbles	lenticular 'winged' 3-5m thick; sheet sands developed locally; margins rather diffuse	not abundant except locally in reduced (green) mudrocks	highly coloured bright reds & greens	*Scoyena* trace fosil assemblage	often distinctive heterolithic facies & well developed stage III calcretes
Porto Novo	coarse-fine sandstones few granite pebbles buff/grey in colour	sheet sand bodies 3-8m thick some development of fining-up	locally abundant particularly within sand bodies	reds, greens, greys	dinosaur bones and teeth simple *Scoyenia ?* burrows	Caliche pebble and mudclast conglomerates common at sandbody bases
Praia Azul	medium-fine buff/grey sandstone	lenticular and lenticular 'winged' geometries <3m thick	fine lignitic debris abundant in mudrocks	mostly grey locally reddened	variety of bivalves plus echinoderms gastropods and ostracods; minor reptile remains	*Diplocrateiron* *Thallasinoides* poorly developed palaeosols
Assenta	medium-fine buff/grey sandstone	sheet sand bodies <8m thick and small lenticular 'winged' sand bodies <3m thick	locally abundant in sand bodies and mudrocks	reds, greens, greys	as per P. Azul plus corals and a rather different bivalve fauna in uppermost part of section	V. distinctive thin marine bands of nodular biomicrite plus dolmites and marine sands in upper part
Santa Rita	coarse matrix-supported pebble and cobble conglomerates; clasts of basement-derived quartzite & greywacke	sheet sand bodies up to 11m thick often multi-storey	low abundance	highly coloured scarlets & purples; locally grey with plant debris	dinosaur bones	calcretes locally well developed e.g. at P. da Santa Rita

red mudstones, which, marginal to channels, are interbedded with thin coarse sand sheets. Structures within channel sand bodies are dominated by large trough cross bed sets and planar cross beds up to 1m thick.

Figure 1.21 Praia da Amoreira section facing due S. Steeply dipping fluvial sands and interbedded muds of the Praia da Amoreira member on the flanks of the Santa Cruz diapir. Dips rapidly decrease away from the diapir. High point on cliff in background is Alto da Vela. Channel in foreground (hammer for scale) is that at 17 to 20m (Santa Cruz basal log), with rhizocretions developed on its upper surface. At beach level mudrocks beneath are cut into by the steep margin of an earlier channel.

Channel sands are coarse to very coarse and granule-rich. They are white to grey, due to the presence of much intergranular kaolinite and generally very poorly lithified. Particularly characteristic are basement-derived pebbles of granite and phyllite with lesser amounts of milky vein quartz and psammite.

Soil profiles, consisting of small, diffuse, irregular carbonate nodules, occur in the mudrocks, occasionally being preferentially developed in the sandier fractions. The upper parts of channels are occasionally pedogenically modified, being structureless, colour mottled, and having what are apparently carbonate rhizocretions developed upon them.

Towards the upper part of the section channels become thinner (1-2m) and of much more restricted lateral extent. They have a symmetrical, lenticular, winged cross-section with width-thickness ratios of less than 15/1 and therefore are 'ribbon' sands (Friend et al. 1979). Grey mudrocks dominate the upper part of the section. Within these are a number of better developed caliche soil profiles with larger nodules at higher densities than lower in the section. Purple discolouration and deeply penetrating sand-filled desiccation cracks are also associated with soil horizons.

A prominent, well developed caliche soil horizon in grey silts, marks a significant break in sedimentation and the top of the Praia da Amoreira member (140m log No. 3, 8.5 to 10.5m log No. 4). The caliche is overlain by a sheet-like channel sandstone of dominantly planar bedded, yellow, medium sands, rich in lignitic debris. The top of the Praia da Amoreira member is marked by a distinct change in channel morphology, lithological colour and grainsize, sedimentary structures and abundance of plant debris. The caliche profile can be designated as a marker bed (Holland et al. 1978) and crops out at the base of the cliff below Alto da Vela (G.R. 6635 3095). As its type locality the Praia da Amoreira member is 130m thick (Santa Cruz logs 1 to 3).

Reference section

The type section is not perfect. It is deformed by the diapir, heavily jointed and, particularly in its lower parts, discoloured by iron staining. Much of the section is inaccessable or poorly exposed. The reason for its selection is the fact that it is the only complete sequence.

Because of the above, a reference section ('Hypostratotype' Hedberg 1976) is desirable. The best coastal outcrop of the Praia da Amoreira member occurs 6 to 9km northwest of Santa Cruz, between Porto Novo and Porto Dinhero (Figure 1.5) (between GRs. 6945 3740 and 7035 4020). Only the top 80m of the member appear here but the outcrop is excellent, with low dips and a few normal faults (see iii).

The marker bed (caliche) described above, occurs at beach level at the more southerly location referenced above. In the northern part of the section it is incised by a sand body at the base of the Porto Novo member. Thick sheet-sand bodies are less common and generally never as thick as those found at Praia da Amoreira. Few are thicker than 1 to 3m and none thicker than 5m. There are a number of differences between this and the type section. In paticular, mudrocks are highly coloured, either red or green, and glaebular caliche soil profiles are particularly well developed ('type B' profiles, Allen 1974). Marginal to channel sand bodies are extensive bodies of decimeter bedded coarse sediment and mudrock. Despite the differences the two sections below the marker bed can be correlated, using such features as sandbody density, soil profiles and thick mudrock units (see Chapter 2).

Other outcrops of the Praia da Amoreira member

There are two further outcrops of the Praia da Amoreira member in addition to minor inland outcrops on the flanks of the Vimeiro diapir. A fault-bounded unit some 220m thick, sits at the base of the Ferrel section (see Figure 6.3). It is punctuated by a tongue of the Porto Novo member some 10m thick 70m above the base of the section.

The other outcrop, is that at São Bernadino. The base of the formation is exposed here (see section 1.6.3) and 140m of the Praia da Amoreira member crop out above. Lithologically the sediments clearly resemble the Praia da Amoreira member, but facies are rather variable (see Chapter 2). The contact with the Porto Novo member is faulted (see Chapter 3).

Conclusions

From the observations recorded in section 1.6.5 it is clear that the Praia da Amoreira member is confined to the western margin of the basin. It almost certainly feathers-out to the east (NB part of this expectation is based upon the environmental interpretation of the member see Chapter 2) and interfingers with the Porto Novo member. The Porto Novo member lies above the Praia da Amoreira member on the coast. The Praia da Amoreira member clearly represents a very diverse suite of sediments but there are certain key features which enable its identification. The most characteristic features are, the petrography of the coarse sediment, the abundance of kaolinite, the provenence of the pebble-grade clasts, and the generally low abundance of plant debris. Caliches are at their most advanced stages of development and mudrocks most highly coloured.

The Porto Novo member

Location

This unit crops out directly above the Praia da Amoreira member at all coastal localities south of the Ferrel section. It is extensively exposed to the south of São Bernadino at the Forte de Pai Mogo, Areia Branca, Porto de Barcos and Porto Dinheiro (Figure 1.8). Extensive outcrops appear to the north of Ferrel and north of Foz do Arelho. At the latter locality the member clearly lies at the base of the Lourinhã formation. Inland, the member outcrops extensively, particularly north and west of Bombarral. Access at Porto Novo is achieved by walking round the headland northwest of the Golf Mar Hotel (Figure 1.5). This is difficult at high tide.

Description of the type section

Above the (caliche) marker bed there is an abrupt change in the petrography of channel sandstones and also in the geometry and density of sand bodies. The erosive-based channel sandstone bodies 3 to 7m thick, have a sheet-like geometry (sensu Friend et al. 1979). The sands are yellow-grey in colour and generally no more than coarse grained. Planar beds and trough cross-beds dominate. Lignitic debris, large logs and re-worked clasts of caliche are common, particularly as lag deposits.

In some sand bodies good fining-upward cycles are developed with channel-levee-floodplain sequences readily indentified. Large inclined lateral accretion surfaces are evident in a number of the channel sandstones. A particularly good example crops out on the headland to the northwest of the Golf Mar Hotel. Mudrocks are generally grey-green and reddened at certain horizons where caliches may be developed. Soil profile development is characterised by carbonate rhizocretion development, or carbonate nodule growth. The very well developed glaebular caliches seen in the Praia da Amoreira member north of Porto Novo are largely absent.

The upper boundary of the Porto Novo member with others of the Lourinhã formation is marked either by evidence of marine conditions, often a 0.5m thick shell bed or by a significant coarse clastic input depending on whether the Praia Azul or Santa Rita member is encountered vertically. A shell bed caps the type section at the cliff top G.R. 6935 3720 and also occurs halfway-up the cliff at Porto de Barcos (Figure 1.8 see Figure 3.1b). At the northern end of Praia Azul G.R. 6610 3030 at the cliff base, it is marked by a thick (approx. 2 m) tabular sandstone with thin shell bands.

The Porto Novo Member is 70m thick at its type locality and thins southwards, being only 47m thick below Alto da Vela. This is the most southerly outcrop of the member, partly because of the wedging out and partly because of the regional southerly dip.

Relationship to other members

North of Lourinhã on the coast, the Porto Novo member re-appears above the Praia Azul member and is up to 350m thick in total. It is replaced southwards by the

Assenta member (Figure 1.10). This transition has not been observed at outcrop as the present coastal exposure is below the level where this would have appeared.

In outcrops to the north, at Ferrel and south of San Martino do Porto, the Porto Novo member is also significantly thicker than at its type locality, at 330m and 800m respectively. The estimate made for the section to the south of San Martino has room for significant error due to the difficult nature of the outcrop, particularly the possibility of significant undetected fault displacements on the flanks of the Caldas diapir. The Porto Novo member is overlain by the gravelly Santa Rita member in these more northerly sections. Below the Santa Rita member, the Porto Novo member frequently exhibits significantly coarser grain sizes than are found in the type section (Figure 1.22). This, particularly in the Foz do Arelho/San Martino section, makes the identification of a distinct member boundary difficult.

The most characteristic features of the member are the colour and grainsize of channel sands, the development of fining-upward cycles, point bars, abundance of plant debris, re-working of caliche and lack of marine fossils.

The Praia Azul member

Introduction

The proposed type section for this member at Praia Azul (G.R. 660 295, Figure 1.8) has been studied in detail by the author and also by Fürsich (1981b). Fürsich (op cit.) concentrated on the biofacies and interpreted the sequence as that of a delta with protected bays and lagoons.

Description of the type section

The sequence, dominated by grey mudrocks, also consists of laterally extensive banks of the large semi-infaunal bivalve *Isognomon lusitanicum* plus *Myophorella lusitanicum* and sand beds rich in the shallow infaunal bivalve *Eomiodon securiformis*. Elsewhere, there are patch reefs of oysters and cidaroid (echinoderm)

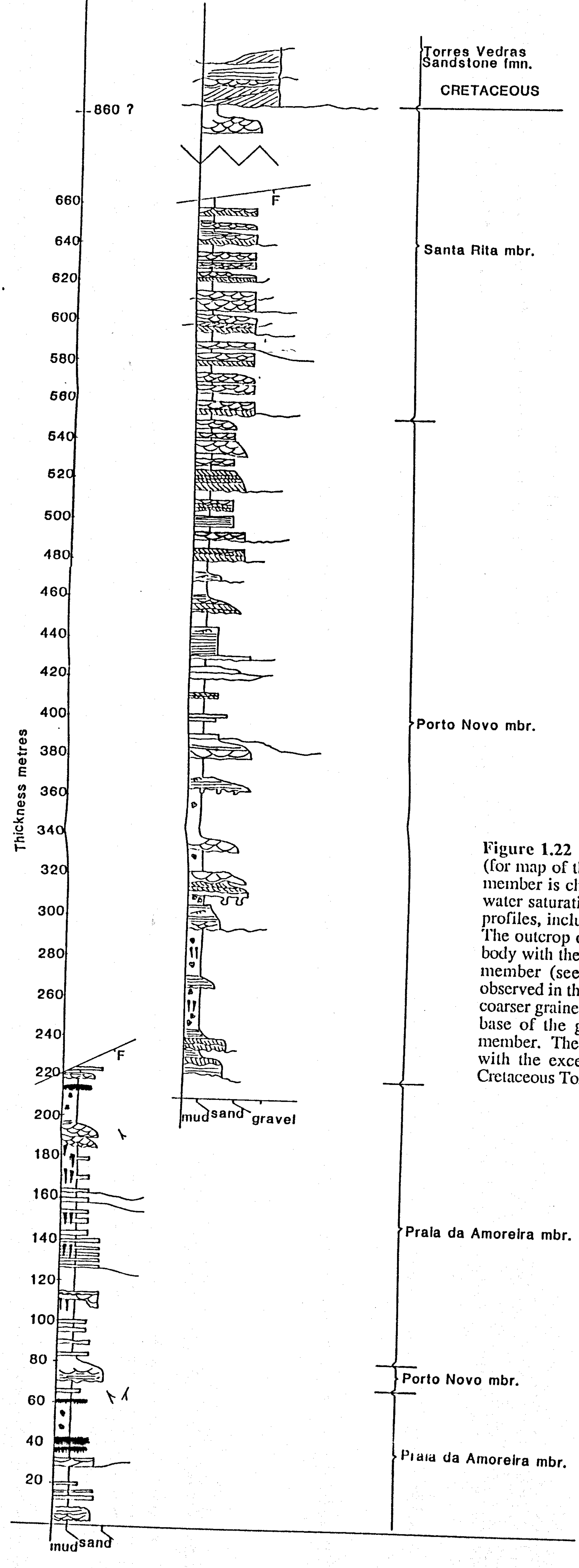


Figure 1.22 Summary graphic log for the Ferrel section (for map of the section see Chapter 6). Praia da Amoreira member is characterised by strata exhibiting signs of both water saturation and desiccation. There are numerous soil profiles, including some of the thickest calcretes observed. The outcrop of this member is punctuated by a sheet sand body with the lithological characteristics of the Porto Novo member (see Chapter's 2&3). Some coarsening-up is observed in the Porto Novo member section as a whole with coarser grained more closely spaced sand bodies close to the base of the gravelly fluvial deposits of the Santa Rita member. The top part of the Santa Rita member is covered with the exception of the unit immediately below the Cretaceous Torres Vedras Sandstone formation.

spines. A number of erosive based, lenticular, cross-bedded channel sands cut the sequence. Bioturbation, commonly of the thalassinoides-type indicating the presence of crustacea, is present in the fine lithologies in particular. There are a few reddened mudrock horizons in the section, associated with diffuse accumulations of pedogenic carbonate nodules.

Boundaries of the Praia Azul member.

The sheet sand which marks the base of the member crops out at beach level towards the northern extreme of Praia Azul. It consists of fine, shelly, locally faintly laminated sands. Elsewhere in the basin the first indication of marine conditions above the Porto Novo member is the presence of an argillaceous bio-micritic packstone, interpreted as a shell bank.

The upper boundary and thickness of the Praia Azul member is problematical. Fürsich (1981b) measured 89 m of section at Praia Azul and a further 50m may be covered by Quaternary deposits. To the south of Praia Azul and the Rio Sizandro (G.R. 6555 2845) is the base of the section which includes the youngest (Assenta) member of the Lourinhã formation. This member contains a similar fauna in places. However, the Assenta member is characterised by predominatly continental fluviatile sediments with occasional marine incursions. Channel sand bodies tend to have a sheet-like character and often contain re-worked caliche. Therefore, it is the dominance of marine indicators and an absence of sheet-like channel sand bodies which defines the Praia Azul member. The boundary thus defined, is in many ways gradational.

The thickness of the Praia Azul member is inversely related to that of the Porto Novo member below (Figure 1.10). As the Porto Novo member thickens northwards the Praia Azul member thins and parts, forming two thin tongues the last outcrops of which are observed below the Forte de Pai Mogo (Figure 1.8). The occurrence of tongues of brackish/marine strata within the Porto Novo member poses the problem of distinction between these outcrops and those of the Assenta member to the south. Fortunately, a further characteristic of the Assenta member is a widespread laminated facies interpreted as lacustrine deposits (see Chapter 5). These are largely absent in the Porto Novo member.

Reference Section

The Praia Azul–Assenta member boundary is exposed at another locality. Immediately south of Praia da Santa Rita (G.R. 6910 3385, Figure 1.23) is a 180m-thick section below an unconformity above which lie Santa Rita member sediments. The basal 60m is predominantly marine and very similar to that at Praia Azul. The last shell bed containing *I. lusitanicum* occurs at 61m (Santa Rita log No. 2) just below a thick sheet-like channel sandbody. Above this unit, the section is dominated by a number of sheet-like fluvial channel sandstones, up to 8m thick. A single very minor marine intercalation occurs, at 101m.

The Assenta member

Location and thickness

The type section for this member stretches from the Rio Sizandro G.R. 6655 2845 south to Porto da Calada G.R. 6400 2075, where it is unconformably overlain by Cretaceous fluvial sandstones (see section 1.6.5). The section is broken into two parts by a fault at G.R. 6405 2630 (Figure 1.24). Its throw is unknown and the section cannot be correlated across the fault. R. Hiscott has logged the top 140m of the Assenta member and the author has logged 140m of section below the fault, thus making the Assenta member at least 280m thick.

Description of Section

This member is characterised by erosive-based channel sandstones with a sheet-like geometry (Friend et al. 1979). These tend to fine-upwards into massive silts, often red, in which caliche soil profiles are developed. Lateral accretion surfaces are evident in some channel bodies (see Chapter 5).

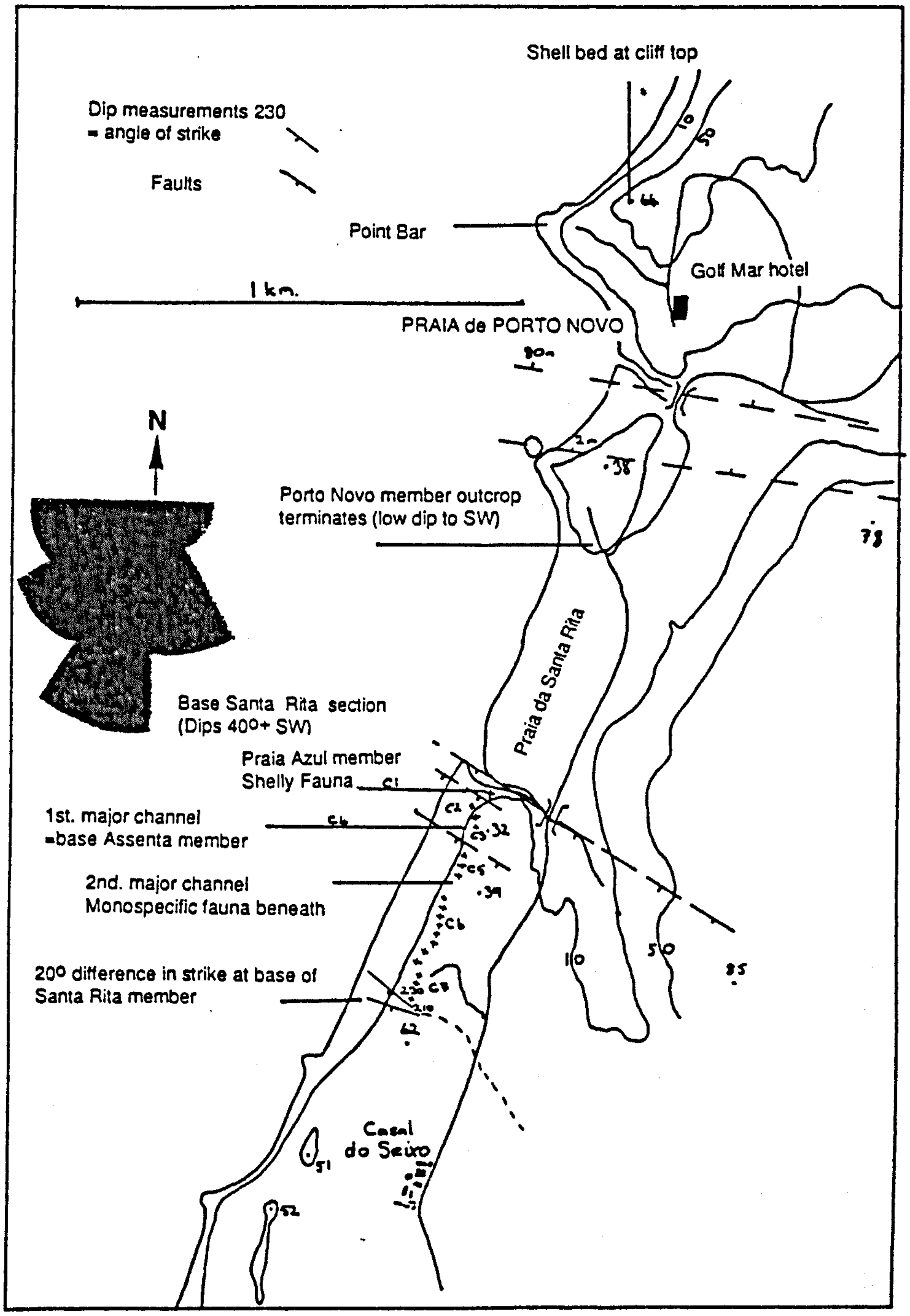


Figure 1.23 Sketch map of the Praia da Santa Rita section. Positions of logged section and important features indicated. Dip symbols show the difference in strike between the top of the Assenta mbr. and the base of the Santa Rita mbr. Palaeocurrent data from the P.Azul and Assenta members indicate relatively low sinuosity despite the presence of point bars, typical of fluvially dominated delta distributaries. Heights in metres.

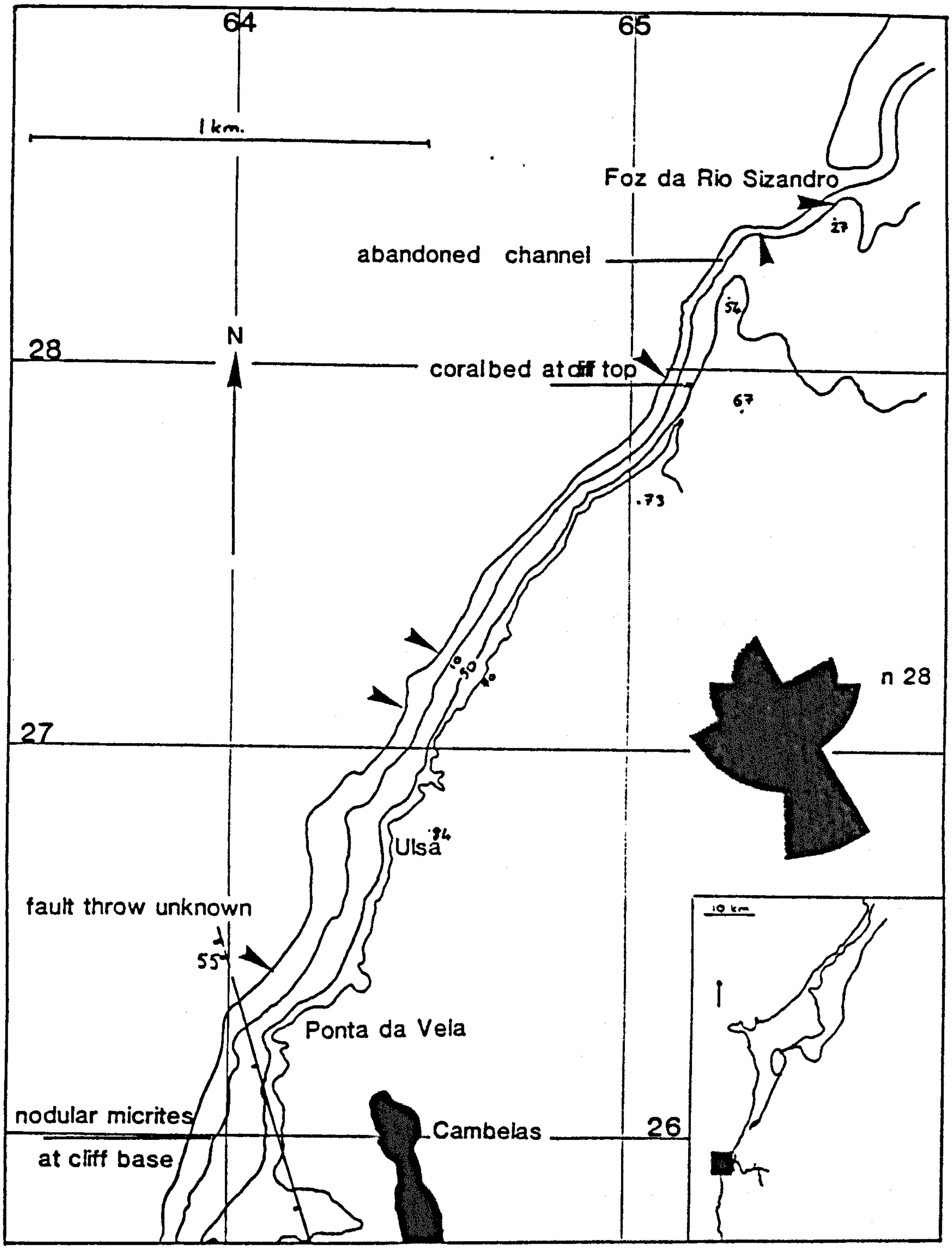


Figure 1.24 Sketch map of the Rio Sizandro section, showing the position of the major fault which breaks the section plus a number of important localities. Arrows mark the position of logged sections. Inset locates map area.

The sequence is periodically punctuated by thin marine intercalations, notably a 0.5m thick coral biostrome growing on a 0.5m oyster rich packstone (Figure 1.25). In the upper half of the member are a number of nodular micrites and ferroan dolomites (pers. obs. and R. Hiscott pers. comm.). Certain of the marine intercalations are very similar to those found within the Praia Azul member.

Figure 1.25 Coral biostrome with encrusting oysters above a 0.5m thick oyster shell packstone. Cliff top south of the Rio Sizandro (see Figure 1.24).

A further, quite distinctive facies, exposed at the very base of the section are units of finely laminated silts and thin sands, a few metres thick and often of hundreds of metres lateral extent. These are interpreted as lacustrine sediments and distinguish the Assenta member from the Porto Novo member in particular.

The Santa Rita member

Location

The outcrop area of this member is largely restricted to north of the Montejunto anticline and a fault which extends westwards from the south western limb of the anticline, intersecting the coast north of Santa Cruz (see Chapters 5&6, Figure 1.8). The type section, logged by R.Hiscott extends from below Casal do Seixo (Figure 1.23) at the southern end of Praia da Santa Rita. The base of the member is locally unconformable above the Assenta member due to deformation on the flanks of the Vimeiro diapir. A difference in strike of 20° can be demonstrated between the base of the first channel and the tops of thin sands in the Assenta member. When the upper unit is rotated on a stereonet to horizontal the Assenta member retains a dip of 4° to the NW (approximately at right angles to the axis of the Vimeiro diapir, R Hiscott pers comm.).

The southern limit of the type section is determined by the presence of a fault which downthrows white Cretaceous clastic sediments of the Torres Vedras formation to the south.

Other coastal outcrops of the Santa Rita member appear to the north of Ferrel (see section 1.6.5) and north of Foz do Arelho. The top of the member is not seen in the latter. Inland, the Santa Rita member crops out extensively to the south and east of Bombarral and also to the east of the Serra dos Candeiros where some of the coarsest examples with cobble grade clasts appear.

Description of the section

The Santa Rita member sediments are particularly distnictive being the coarsest of all the members of the Lourinhã formation. Lithologically they consist of buff or yellow, matrix supported pebble conglomerates. The matrix consists of coarse and very coarse, often angular sand. The sediments are organised in erosive based sheet sand bodies dominated by large scale trough and planar cross beds. Sand bodies are

1-11m thick and certain examples exhibit a fining up. These units comprise about 60% of the succession the remaining third consisting of locally brightly coloured mudrocks. Plant debris is quite rare.

1.7 CONCLUSIONS AND LAYOUT OF THIS THESIS

The lithostratigraphic scheme outlined in section 1.6 has been tested and modified in the field during the final field season. It is felt that it is a workable scheme meeting the criteria outlined in Hedberg (1976).

The following five chapters deal largely with the environmental interpretations of each of the members of the Lourinhã formation and how these interrelate. The final chapter draws these threads together, with a series of palaeogeographic maps and also looks at the broader implications in the light of the evolution of the North Atlantic Ocean.

CHAPTER 2

DISTAL ALLUVIAL FAN FACIES: THE PRAIA DA AMOREIRA MEMBER

NOTE: The bulk of this chapter forms the text of a paper entitled 'Distal Alluvial Fan Sediments from the Upper Jurassic of Portugal: controls on their cyclicity and channel formation' submitted to the Geological Society of London for publication.

2.1 INTRODUCTION

Only a few recent papers have concentrated on alluvial fans with a significant mud component in distal fan, sand and mudflat environments (e.g. Hardie *et al.* 1978; Hubert and Hyde 1982; Tunbridge 1984; Flint 1985). The majority of such fan environments pass laterally into playa lakes with features such as evaporites, mudcracks and adhesion ripples (Glennie 1970) indicating aridity and desiccation. This chapter describes an ancient distal fan sand/mudflat environment in a sub-tropical semi-arid climate where the active area of sedimentation was often water saturated so that features such as desiccation cracks are uncommon, except in soil profiles. The fan environment passed laterally into a fluvial meander belt. The author believes that fan apices were probably some 15 km north-west of the present-day coastline, along a line of horsts which presently mark the western boundary of the basin.

Particular attention is paid to the initiation of channels in the 'flat' environment. This is because channelised flow was apparently a secondary feature in the development of the sedimentary cycles observed. There is evidence for both autocyclic and tectonic control on the development of the sediment pile.

The sediments are spectacularly exposed in cliffs up to 50m high at three locations on the coast between Sintra and Peniche (Figure 1.3). Most of the data presented originates from the section north of Porto Novo (Figure 1.5). Though only the upper 90m of the Praia da Amoreira member (Figure 1.9) can be seen here, the strata are near horizontal and well exposed. In contrast, the type section south of Santa Cruz (Figure 1.12), though complete, dips steeply off the flanks of a diapir (Figure 1.21) and is heavily jointed and iron stained. The section to the north of Ferrel is rather small. The type section can be correlated with that at Porto Novo (Figure 2.1). A further outcrop of the Praia da Amoreira member occurs at São Bernadino (see Chapter 1). Lithologically the sediments exposed are the same as those which appear at other outcrops of the member. However, the sedimentary facies developed are rather different, with the development of lateral accretion surfaces and far more abundant plant debris. These sediments and their significance are discussed towards the end of this chapter.

On the coast, the Praia da Amoreira member is succeeded vertically by deposits of a fluvial meander belt which characterise the Porto Novo member (see Chapter 3). This is characterised by finer grained channel facies, significantly more fossilised vegetation, less well developed carbonate soils and larger scale palaeochannels than those of the fan facies. The Porto Novo member dominates areas to the east of the coastal outcrop. It is likely therefore, that the distal fan environment described in this chapter was deposited contemporaneously with a fluvial meander belt.

The Praia da Amoreia member has yielded no biostratigraphically useful palynomorphs, and its age can only be inferred from its stratigraphic position above the Abadia formation which is of late Oxfordian to mid-Kimmeridgian age (Ellwood 1987). The younger deltaic sediments of the Praia Azul member have been dated as basal Tithonian by Leinfelder (1987). Thus it is probable that the Praia da Amoreira member is of mid-to-late Kimmeridgian age. This is in broad agreement with biostratigraphic maps produced by the Portuguese Geological Survey (Camarate França *et al.* 1960, 1961).

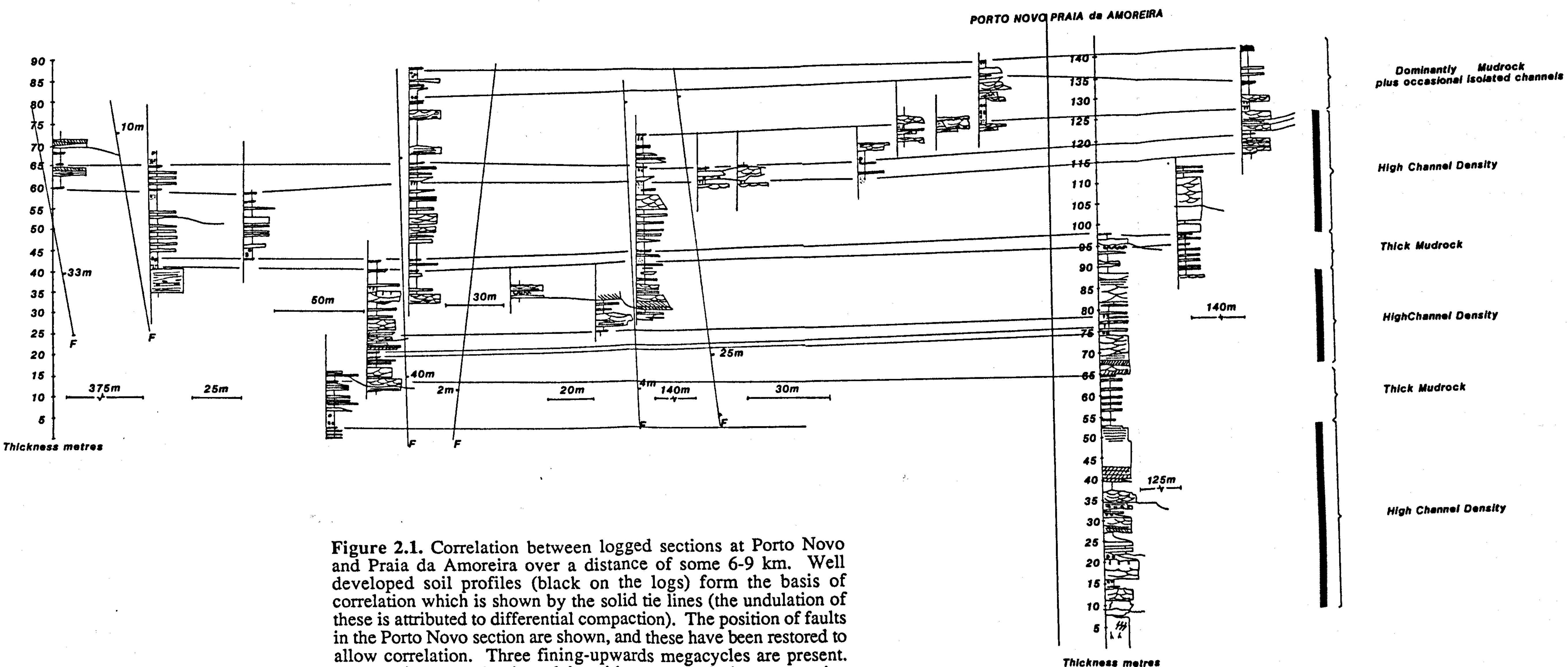


Figure 2.1. Correlation between logged sections at Porto Novo and Praia da Amoreira over a distance of some 6-9 km. Well developed soil profiles (black on the logs) form the basis of correlation which is shown by the solid tie lines (the undulation of these is attributed to differential compaction). The position of faults in the Porto Novo section are shown, and these have been restored to allow correlation. Three fining-upwards megacycles are present. The sections are dominated by either coarse facies or massive mudrock.

Table 2.1. Praia da Amoreira member, summary of the characteristics of Facies and Subfacies and their interpretation

FACIES *Sub Facies*	CHARACTERISTICS	INTERPRETATION
HETEROLITHIC FACIES	**cm-dm interbeds, wide range of grainsize and structures. Approx 50% mudrock.**	**deposits of sheetfloods with high suspended load; incipient development of channels; dominantly deposition under lower flow regime conditions.**
Subfacies (i): laminated sand and mudrock	ripple and parallel laminated sands thinly interbedded with mudrock.	low velocity flood deposits; which may at time have been deposited in shallow lakes.
Subfacies (ii): ripple laminated sandstone	dm beds of ripple and climbing ripple laminated sandstone; multiple internal erosion surfaces.	low velocity flood deposits; rapid deposition from high suspended loads; some falling stage modification as flood wane.
Subfacies (iii): brown, burrowed, fine sandstone	fine - medium sandstone intensely bioturbated; isolated dm beds; *Anchorichnus*; often at top of coarse part of cycles.	possibly widespread, shallow lake associated with low rates of sedimentation as coarse sedimentary deposition wanes.
Subfacies (iv): isolated cross beds	single dm-scale cross beds often with modified tops and mud-draped foresets.	isolated bars produced during large magnitude floods; mud drapes may represent velocity fluctuations during a single flood and/or several floods.
Subfacies (v): graded and parallel laminated beds	wide spectrum of grainsizes and cm-dm beds, graded, inversely graded or simply parallel laminated and ungraded.	upper flow regime conditions, graded = deposition from waning current inversely graded = rising stage deposition.
Subfacies (vi): vertically amalgamated sand beds	normally coarse sands in a series of vertically stacked beds, up to 1m thick - later beds frequently incised into those beneath.	incipient channels.
Subfacies (viii): deformed and structureless sand beds	variable lithologies; load casts on bases and laminae internally disrupted or not discernible; occasionally highly convoluted stringers of sand.	variety of causes; deposition over a wet substrate or in shallow lakes; pedogenesis, bioturbation, water escape due to rapid deposition and tectonically induced liquefaction.
MASSIVE SANDSTONE FACIES	**dominantly coarse sands; 1-5m thick lenticular 'winged' and occasionally sheet-like sand bodies scours at bases, cross beds dominant channel fill; parallel laminae more common in sheet-like bodies; 'abandonment' fills virtually absent; shallow scours in channel tops common mud drapes also; facies does not necessarily occur at the base of 'cycles'.**	**shallow ribbon channels with deposition dominantly under turbulent lower flow regime conditions; high suspended loads and frequent flooding; at late stages in development, consist of a network of small channels < 1m deep and a few metres wide; probably not an early feature of cycle development; channel initiation facilitated stacking-up of sand beds of heterolithic facies, abandonment gradual as thickness of sediment pile increases.**
MASSIVE MUDROCK FACIES	**siltstones and silty claystones with occasional sand beds, carbonate soil profiles frequently developed towards top of mudrock units which lie above heterolithic and massive sandstone facies.**	**deposition of fines from suspension during late stages of cycle development as increasing sediment thickness causes the switching of the active area of fan sedimentation; at a late stage rarely inundated and soil processes dominant.**
Subfacies (i): red-brown homogeneous mudrock subfacies	structureless, red-brown micaceous silts no organic debris.	more often subaerially exposed, little or no vegetation.
Subfacies (ii): dark green mudrock subfacies	occasionally with ripple and parallel laminae preserved; dark green micaceous and with much plant debris, associated with this is elemental sulphur.	may simply reflect preservational bias, or slightly more humid conditions, leading to lusher vegetation and consequent reduction of sediment; laminations suggest standing bodies of water.

2.2 DESCRIPTION AND INTERPRETATION OF FACIES

Three broad lithofacies are present within the Praia da Amoreira member (Table 2.1). These are:

1) Heterolithic facies: decimetre interbedded sand and mudrock with highly variable grain size, structures and lateral extent. Seven subfacies occur.

2) Massive sandstone facies: generally 1-5m thick, though rarely more than 3m, sand bodies have a lenticular 'winged' cross-section.

3) Massive mudrock facies: with associated soil profiles. Two subfacies are developed.

The three facies groups are arranged in a series of crude fining-upwards cycles (Figures 2.2 and 2.3a) with either massive sandstone or heterolithic facies at their bases passing up into massive mudrock facies with variable degrees of soil profile development. Cycle thicknesses vary between one and fifteen metres and they are strongly aggradational, forming discrete packets of sediment which normally have only mildly erosive bases. Soil profiles in older cycles are rarely cut by overlying cycles, and no re-worked caliche was observed.

Cycles have a considerable lateral extent, though the scale of this is difficult to ascertain due to the limitations of the outcrop, but a kilometre to several kilometres is likely.

2.2.1 Heterolithic facies

Description

This is an extremely variable unit of centimetre to decimetre bedded sands and mudrock. Although a full range of sand grain sizes occur, fine to medium sands dominate. Sorting is generally poor. Sands are normally light grey, rather kaolinitic, rich in mica and poorly cemented. The kaolinite is present both as the weathered residue of K - feldspars and as coarse 'books' in pore spaces. Mudrocks are dominantly brown micaceous siltstones and silty claystones which are completely structureless.

Figure 2.2. Crude fining-upwards cyclicity developed in the Praira da Amoreira member. The base of a cycle lies at the bottom of the prominent packet of heterolithic facies at the centre right. The proportion of sand decreases upwards and massive mudrock becomes dominant. A prominent 'stage III' (Gile et al. 1966) calcrete occurs at the top of the cycle which is 15m thick. This is directly overlain by a further much thinner cycle (3-4 m thick). A calcrete occurs at the cliff top. The individual cycles form discrete packets of sediment in a strongly aggradational sequence. Soil profiles in older cycles are rarely eroded or cut-into by subsequent deposits. The scale bar is approx. 15m. G.R. 6985 3830, north of Porto Novo.

Sandy facies usually have slightly erosive bases and occasionally contain mud clast pebbles (Figure 2.4). Coarse sand facies may in addition contain lithic pebble lags (see the section below on *massive sandstone facies* for a description of clast types). Individual beds often have internal scoured surfaces. The tops of sand beds also show evidence of re-working and beds may be truncated by an erosion surface.

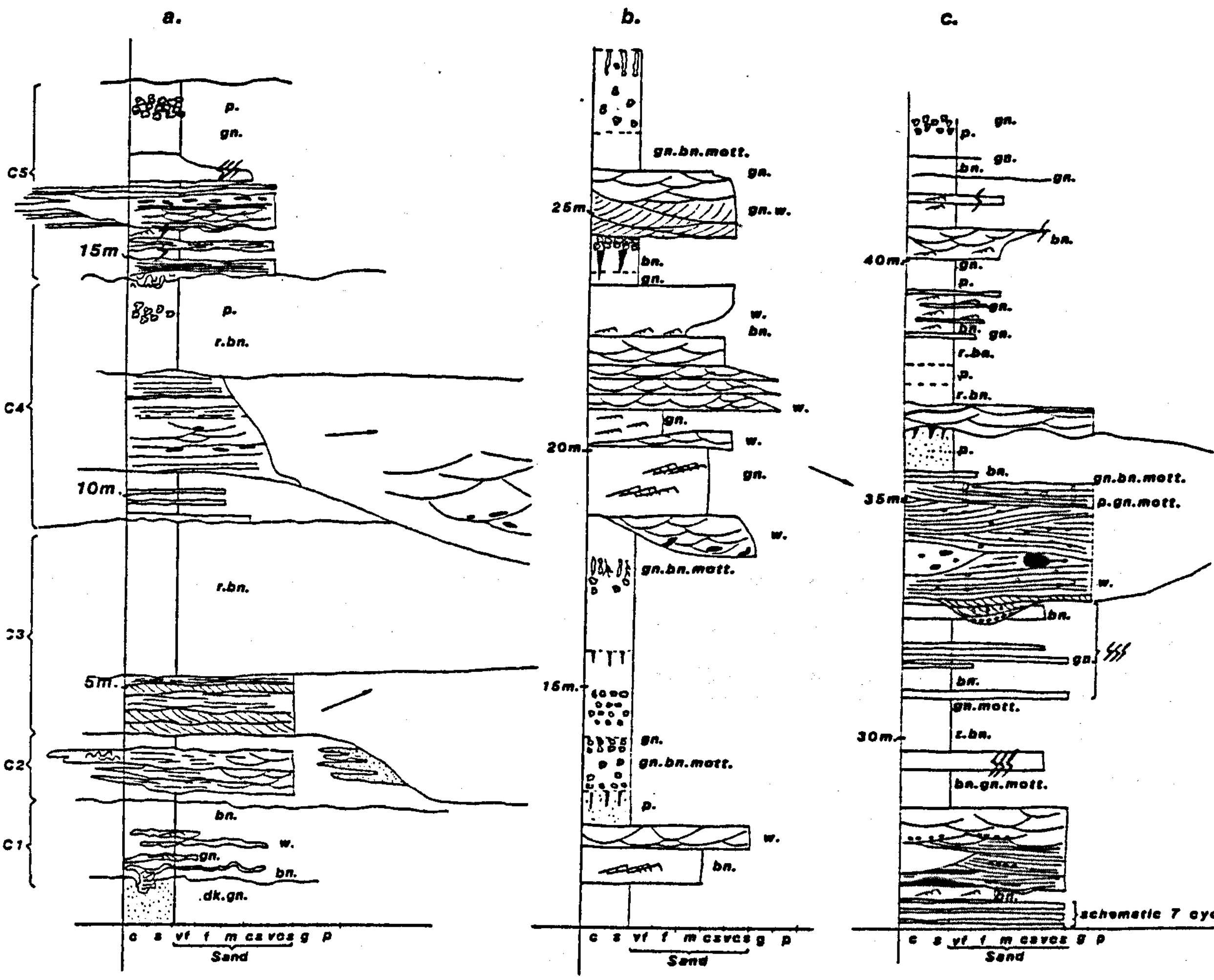


Figure 2.3. Three graphic logs illustrating features of the vertical sequence at Porto Novo. The position of (b) and (c) are shown on Figure 2.1.

a. Fining-upwards cyclicity (5 cycles C1-5). Note the variable scale of the cycles and of the relative importance of mudrock. Also, the abundance of soft sediment deformation and structureless beds and that channels often do not occur at the bases of the cycles. The smaller 'channels' in C2 and C5 are the vertically amalgamated heterolithic sub-facies. Only the larger channels are incised into cycles beneath. The lower three cycles are illustrated in Figure 2.5.

b. Top of the Praia da Amoreira member. Channel between 19-24m exhibits considerable grain-size and structure variations, consistent with a strongly fluctuating hydraulic regime. There is no development of fining-upwards. A scour with cross-bedded fill sits at the base. Climbing ripples in medium sands indicate high rates of sediment fallout. Graded beds are the product of waning floods.

c. In contrast to b, channel between 33-35m exhibits no grainsize variation. Its fill is dominated by planar and low angle laminations. Note the soil profile developed directly above the channel and the rhizocretions in the channels upper part.

Soft sediment deformation is an extremely common feature, with micro- and macro load casts beneath coarser lithologies. Mud flames sometimes penetrate deeply into sand beds, and whole beds may have undulose bases and tops with internal structures deformed accordingly. The latter feature could be compactional rather than synsedimentary in origin. Water-escape and liquefaction structures occur in both coarse and fine lithologies. Isolated zones of convoluted laminae are common. Locally deformation is severe and widespread, affecting up to a metre of sediment over several metres laterally. Fine sand facies frequently have upturned ruptured laminae.

Figure 2.4. Severe soft sediment deformation in heterolithic facies with highly convoluted stringers of coarse sand in mudrock rich in sand grains. It is probable that such deformation was triggered by seismic activity. Also note the rip-up clasts in the sand bed below. Scale bars on hammer are 10cm wide.

Bioturbation is a common feature of variable intensity. Normally, burrows are simple, lined, 5-12mm diameter tubes randomly orientated with either a sand or muddy fill. A few vertical examples have flared necks at sediment interfaces suggesting caving of a loose substrate (Fürsich 1981b). Faint curved meniscae are rare features indicating backfilling. Branching of burrows is not common. Examples of *Anchorichnus* (Frey *et al.*1984) have been found. These ichnofossils are considered to be the feeding structure of a worm. They consist of horizontal, slightly sinuous burrows (approx. 10mm diameter) with curved meniscae. Good examples have only been found in one subfacies (see below). The presence of *Anchorichnus* and other terrestrial indicators (such as soils) indicate that the traces belong to the Scoyenia Ichnofacies (Frey *et al.* 1984).

A number of subfacies are developed within the heterolithic facies and are described below.

Laminated sandstone and mudstone facies consist of thinly interbedded/laminated ripple and parallel laminated fine sands and silty clays. Individual beds are a few millimetres to 10cm thick. Sandy packets are usually slightly erosively based and contain multiple internal erosion surfaces. Laminae are commonly picked out by concentrations of mica and are often cross-cut by burrows, green/brown colour mottling is frequently associated with these. Minor soft sediment deformation is common. Desiccation cracks are absent. Locally, several metres of section are dominated by this sub-facies to the exclusion of all others, except for rare, structureless stringers of coarse sand with loaded bases. Elsewhere, thin packets of this facies occur interbedded with others.

Ripple laminated sandstone facies consist of ripple and climbing ripple laminated fine to medium sandstone in 10 - 40cm thick beds. These are erosively based and commonly form amalgamated units with many internal scours. They are often burrowed, though normally at low densities. Climbing ripple laminae occur as small sets within a ripple laminated bed or in thicker sets on their own. In the latter case climbing sets are truncated near bed surfaces, ripple laminae appearing at their tops. There are rare occurrences of 'type B' (stoss side preserved) climbing ripple sets

(Ashley *et al.* 1982). Isolated zones of ruptured or convoluted laminae are common. Beds may have loaded bases.

Brown, burrowed, fine sandstone facies are a relatively rare sub-facies, but very distinctive, comprising well cemented, intensely burrowed, 10-20 cm beds of fine brown micaceous sandstone. Faint ripple laminae are periodically discernible. Burrows are dominantly vertical, cylindrical and mud-lined, often with annular ribs. It is in this facies that specimens of *Anchorichnus* occur. The sub-facies is exceptional because of its colour, the intensity and preservation of burrows and also its lateral persistence, normally traceable over tens of metres. It often appears at or near the top of packets of heterolithic facies. Cement is usually poikilotopic calcite.

Isolated single cross bed sets are up to 0.4m thick, and are developed in medium to very coarse sandstone often with mud-draped foresets. Their bases are erosive, planar or undulose with angular toesets, and their upper surfaces are eroded and may be overlain by structureless stringers of coarse sand interbedded with silts or ripple laminated fine sands. This is a relatively rare sub-facies.

Graded and parallel laminated beds are a common and varied sub-facies with a wide spectrum of sand grades and bed thickness from centimetre to decimetre scale. Thicker beds tend to be associated with coarser sand grades. Grading is very variable: it may be uniform or confined to the top few centimetres of a 20 - 30cm bed. Siltstone drapes are common. Thinner, centimetre thickness beds often occur in packets separated by mud drapes. Laterally these may amalgamate into a single unit. Variations include coarse ungraded beds of parallel laminated sandstone, scours/gutter casts with parallel laminated fill, inversely graded beds and beds of parallel laminations with grain size oscillations. Structureless beds with normal or inverse grading are common.

Vertically amalgamated sand bed facies is a particularly distinctive development which results in small channel-like units (Figure 2.5) with lenticular 'winged' cross-sections. They have well defined erosive bases and are of limited lateral extent (up to 8m wide). Thicknesses are variable but can be up to one metre. Internally, they consist of a stack of several sand beds, often planar laminated, with the upper units often incised into those beneath. Towards their margins, individual beds are separated by mud drapes which thicken laterally as the sands thin and taper off.

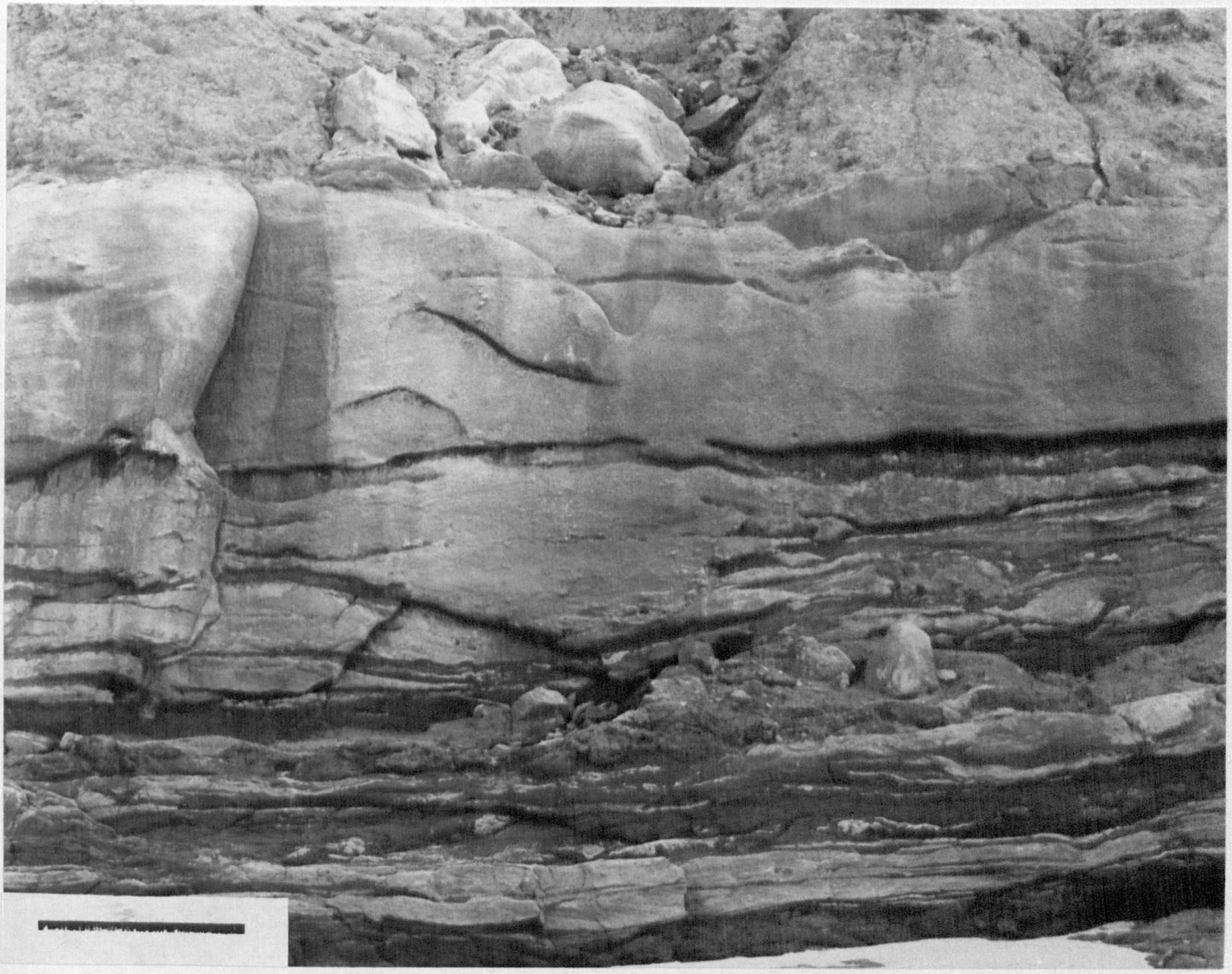


Figure 2.5. Lenticular 'winged' channel-like unit in the centre of the figure is the most important feature. This is an example of the vertically - amalgamated sub-facies. A mud drape almost completely divides this example. All three facies of the Praia da Amoreira member are illustrated . Heterolithic facies above a mildly erosive base, and sheet-like channel facies overlain by massive mudrock facies in the upper half of the figure. There are in fact three 'cycles' exposed in this figure - compare with lower half of figure 5a. Scale bar = 1m. Base of cliff some 1700m north of Porto Novo.

Deformed and structureless sand beds are both coarse and fine in grade. They are commonly interbedded with brown silts and silty claystones. Burrows are occasionally discernible, particularly when they have a muddy fill. Lower surfaces are often erosive though highly irregular. Upper surfaces are characteristically undulose and may be rather diffuse. Soft sediment deformation is also very common. In a few examples this is exceedingly severe with highly convoluted strings of coarse sand in a matrix of mudrock rich in sand grains (Figure 2.4).

Interpretation of heterolithic facies

The environment of deposition of the heterolithic facies was clearly highly variable. Features such as scouring, lateral and vertical variation, diversity of structures, grain-size oscillations and mud drapes indicate that considerable fluctuations in discharge occurred.

Deposition is likely to have been from an unconfined sheet flood flowing over a rather wet substrate, the latter indicated by the abundance of load structures. Each sand bed may be the product of a single flood event. However, the abundance of internal erosion surfaces suggests that local velocity fluctuations occurred during individual flood events and resulted in repeated episodes of scouring and deposition.

The range of grain sizes and structures which exist indicate that both sub- and super-critical flow conditions existed with velocities ranging between 0.2 and 1.5m sec^{-1} (Harms *et al.* 1982). The predominance of finer lithologies and rippled or graded beds suggests that turbulent sub-critical flow conditions were more common. The lack of extensive sheets of planar bedded sands often associated with high velocity sheetflood environments (Tunbridge 1981a, b) supports this conclusion.

A problem in determining the relative importance of sub- and super-critical flow is the abundance of structureless sand beds. As a number of these are clearly burrowed it is possible that bioturbation caused the absence of structures. In addition, evidence for the rapid deposition of these sediments plus dewatering structures suggests that fluidisation and liquefaction may also have been significant processes.

Weathering and pedogenic processes are also probable important causes of a lack of structures. The abundance of kaolinite is often attributed to weathering in a humid sub-tropical climate (Nami and Leeder 1978; Hallam 1984). In addition, the relative absence of plant debris and red colouration of the mudrocks suggest that the sedimentary surface was normally bare and exposed to weathering and oxidation. However, a sub-aerial interpretation is slightly problematic due to the general lack of features associated with such exposure, (e.g. desiccation cracks, footprints, aeolian deposits). Indeed, as pointed out earlier, there is much soft sediment deformation. Exceptions to this are found in the section to the north of Ferrel. Here, sediments which clearly show features of soft sediment deformation are also cut by vertical, rectilinear, sand-filled mudcracks (Figure 2.6). Therefore, in some areas, the sediments were alternately subject to both water saturation and desiccation. This suggests a seasonal rainfall pattern. At times is seems likely that standing bodies of water existed. The laminated sand and mudrock facies and the bioturbation tend to support this. However, mudrocks generally show little sign of the fine laminations commonly associated with lacustrine sediments (Hardie *et al.* 1978).

The whole sequence is strongly aggradational, as indicated by the generally mildly erosive contacts at the bases of sand beds. High rates of deposition from high suspended sediment loads are indicated by a number of structures, e.g. climbing ripple laminations (Ashley *et al.* 1982), graded and structureless beds and dewatering structures. Graded beds are the product of waning flows (Hardie *et al.* 1978). Those which only fine in their top few centimetres probably represent a rapidly waning flood. Water escape is a likely consequence of rapid deposition as such sediments are initially loosely consolidated and highly porous (Lowe 1975; Collinson and Thompson 1982). High suspended sediment loads are also indicated by mud drapes and the significant proportion of mudrocks overall. As mudrock units can be comparatively thick (locally tens of centimetres) it is probable that they often represent a number of low magnitude flood 'events', the lack of structures being due to a combination of lack of grain size contrast plus bioturbation and pedogenesis.

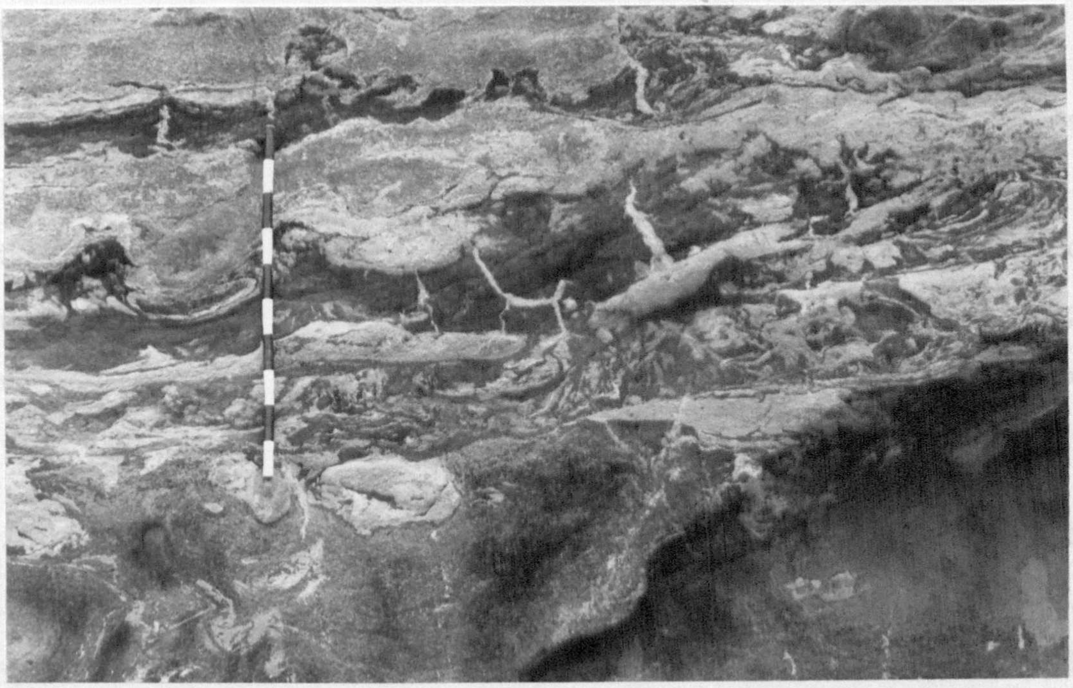


Figure 2.6 Subaqueously deposited heterolithic facies with soft sediment deformation structures; load casts, mud flames etc., cross-cut by sand-filled rectilinear desiccation cracks. Scale 1m.

The presence of inversely graded beds is somewhat problematic, but Allen (1981) attributes such organization (in conglomeratic facies) to the influence of high sediment loads on rapidly rising floods in 'flashy' regimes.

The more severe forms of sediment liquefaction seem hard to explain simply in terms of rapid deposition, loading and water escape. It is possible that some disturbance, probably seismic, produced the structures. In view of the location at the margin of the basin and the active tectonic regime envisaged, this is not unlikely.

The modification of bed surfaces (e.g. the ripple laminae above truncated climbing ripple laminae) was probably due to re-working of sediment at falling stages.

No palaeocurrent data is presented from the heterolithic facies because that collected is too sparse to be significant; trends generally mirror those found in the channel facies (see next section) with a downslope trend to the southeast.

2.2.2 Massive sandstone facies

Description

Massive sand bodies form prominent features in the cliff section (Figure 2.7). Most have a lenticular cross sectional geometry with width/thickness ratios of <15 forming *ribbon* sand bodies (Friend *et al.* 1979). These are invariably 'winged' (Bersier 1958) with a multi-storey fill (Figure 2.8) and are 1 to 5m thick, though rarely exceeding 3m. Similar complex 'winged' sandbodies were described by Friend *et al.* (1979) from the Ebro Basin and Stear (1983) from the Karoo.

Sheet-like sandbodies also exist, although these are often composite with major internal erosion surfaces cutting to their base. One example, though having a dominantly coarse grained fill, has the margin of the scour draped with fine sands and silts.

Palaeocurrent data (Figure 2.14a) from cross-bedding within channels indicate moderate sinuosities and a source to the northwest. Sand body margins have shallow inclinations, steep cut-banks not being a significant feature. Often there is a stack of 'wings' at the margin, interdigitating with mudrocks of the heterolithic facies (Figure 2.8). The mudrocks may extend into the body as a fine sediment drape over coarse facies. Erosion surfaces associated with the younger 'wings' often cross cut these drapes and the coarse sediment of older 'wings'. Fine sediment drapes, a few centimetres thick and only partially preserved, are a common feature.

In axial sections, the bases of sand bodies often have a relief of up to 0.5m with isolated scours having a cross-bedded fill. Blocky and platey mudclast pebbles and cobbles form lags at their bases and also occur on the foresets of scour fills and

Figure 2.7. Cliff section north of Porto Novo, 50m high trending approximately north-south. A number of lenticular channel sandbodies are clearly distinguishable, particularly above thick mudrock in mid-section. The lateral persistence of the 'cycles' is also apparent in this figure.

cross-beds and in association with partially eroded fine sediment drapes. Occasionally, bed bases are highly irregular and 'remnants' of heterolithic facies may be preserved as a pillar projecting a metre into the base of a sand body.

Lithologically, sediments are dominated by poorly sorted coarse and very coarse arkosic sandstones and granulestones. Buff or light grey, poorly cemented and kaolinitic, the sands are rich in pink feldspars, micaceous (dominantly biotite) and often pebbly. Pebble types are dominated by granites and phyllites with subordinate psammite, shale and milky quartz. Lithic clasts are sub-angular to well-rounded depending on their lithology. Kaolinite occurs both as the degradation products of feldspars and as coarse 'books' in pore spaces.

Figure 2.8. A stack of 'wings' at the margin of a channel. The interdigitating fines indicate that at no time was the palaeochannel very deep. Cross beds occur in the channel centre whilst planar beds are concentrated in the 'wings'. The base of the channel coincides with a prominent, undulose erosion surface marking the base of a 'cycle'. Note the well developed 'stage III' calcrete (Gile et al. 1966) in silts above the channel. Hammer at base of channel 0.4m long.

Much less important are medium to fine sands and silts. These tend to be green or mottled green-brown and are also micaceous. Structures associated with these sediments are planar lamination, ripples and climbing ripple laminae. In one example an upward transition from ripple laminae to planar laminae was observed. Fine sediments often show signs of deformation associated with loading by overlying coarser lithologies. No examples of entirely fine grained massive sandstones occur and only one example is known where the top one metre consists of burrowed ripple and parallel laminated silts and fine sands.

In coarse lithologies the dominant structures are trough cross beds 5-20cm thick. Individual sets are up to one metre wide and two to three metres long. A number of sand bodies are completely dominated by this structure. Larger scale trough cross beds and planar cross beds are less common. An exceptional example is 1.5m thick. Planar cross beds occur in 20-50 cm sets, occasionally in cosets and are generally confined to the lower half of the sand bodies. The axial extent of the thicker sets may be up to 10m. These larger cross beds occasionally have slumps on foreset planes and laminae are ruptured by water escape structures.

A striking feature, found at sand body margins, is climbing tabular cross-bed sets. Angles of climb may be steep. Individual sets are 10-20cm thick and have tangential bases. There is a pronounced grain size segregation, with brown medium sand toesets interdigitating with buff coarse foresets (Figure 2.9).

Parallel laminae are common, with two modes of occurrence; as 5-30cm graded beds of coarse to fine sand and silts and as ungraded beds of variable thickness, often exhibiting grain size oscillations between coarse and very coarse sand. Such oscillations are also a feature of cross bed foresets. There are examples of coarse grained sand bodies being dominated by parallel or low angle laminations. Where this occurs, packets of sediment 20-40cm thick are separated by low angle erosion surfaces. A small number of beds exhibit inverse grading.

The tops of a number of sand bodies have scours 0.5 - 1m deep and 3-4m wide cut into them. Their fill is usually coarse grained and consists of concentric laminae which thicken towards the centre of the scours and lap out onto their flanks.

The vertical and lateral arrangement of massive sandstone facies

The vertical arrangement of facies is variable and fining-upwards cycles are not necessarily developed (Figure 2.3b). The entire thickness of many sand bodies consists of coarse sand, which is directly overlain by massive mudrock facies (Figure 2.3c). A number do show fining in their top half metre to fine/medium sands and these are commonly rather structureless or deformed. Others have repeated grain size variations. Often the major vertical change is a reduction in the *scale* of structures from perhaps half metre thick cross beds, to sets ten centimetres thick. Laterally a certain degree of fining is often detectable, normally by only one or at most two sand grades.

Figure 2.9. Steeply climbing bedforms at a channel margin. Note the grainsize segregation between toe and foresets. This is attributed to fallout of fines in front of the bedform which interdigitate with coarse avalanching foresets. The tangential nature of toesets supports the proposed high rates of fallout from suspension. Scale bar = 0.4m.

Graded, parallel laminated sands are often found in the 'wings' of sand bodies. Cross beds are more common in the core of the sand body. Climbing cross-beds are an exception, always being found at sand body margins with foresets directed obliquely outwards. In addition, as 'wings' are traced laterally into the heterolithic facies, isolated cross sets directed away from the sand body may be found. However the 'wings' are more commonly structureless.

Interpretation of massive sandstone facies

The data presented indicates that the sand bodies are the result of deposition in channels which were rarely very deep and, like the heterolithic facies, experienced large fluctuations in discharge. There was some re-working of sediment and rates of sedimentation were high. It appears that the majority were filled by vertical accretion of bedload which ultimately choked the channel with sediment. The fact that there is only one example of a deposit that might be considered a fine-grained abandonment fill (see below) indicates this to be the case. The small scours in the tops of channels suggest that at a late stage in their development they consisted of one or more small channels, possibly forming a network rather like that of a braided channel at low stages (Collinson 1986). Abandonment was therefore probably a gradual process. The abundance of mudrock outside the channels, plus the fine drapes within, indicate the presence of high suspended loads (Picard and High 1973).

Shallow depths are indicated by the multi-storey nature of many channel fills. Erosion surfaces associated with the 'wings' at the channel margins rarely actually cut down to the base of the channels. The scale of bedforms, usually ten to thirty centimetres, may also suggest shallow depths though the thickness of a cross-set cannot be directly related to flow depth (Harms *et al.* 1982).

A number of the sheet-like sand bodies suggest that channels of greater width existed, though their depths still rarely exceeded 3m. These are often dominated by planar laminated coarse sands, perhaps a reflection of greater and more consistent stream velocities of about 1.5m sec^{-1} (Harms *et al.* 1982).

The dominance of cross beds suggests that subcritical turbulent flow was the norm. Parallel laminations indicate that locally, particularly in the shallows at channel margins, velocities may have been higher. Alternatively this may be a reflection of laterally decreasing grain size. Graded beds, indicating deposition from a waning flow and climbing bedforms are indicators of high sediment load. These structures are common in ephemeral streams (Picard and High 1973). Inversely graded beds are relatively uncommon, but along with the ripple-to-planar laminated sands are taken to indicate high rates of fallout during rising stage.

The distinctive climbing bedforms with grain-size segregation between toe and foresets are a special feature. The finer toesets are taken to be fallout from suspension in front of a bar so that the finer sediments interdigitate with the coarser avalanch foresets. The tangential toesets are consistent with the proposed high fallout from suspension. The foresets are always oriented towards the channel margins, suggesting that the bedforms migrated outwards up the shallowly inclined banks during streamflood.

The deep scours and rip-up clasts found at channel bases are probably the result of erosion by turbulent eddies during rising stages, with the cross-bed fills resulting from the transport of coarse bed load. Mudclasts were probably not transported great distances as they tend not to remain cohesive during transport (Smith 1972) and 'armoured' mud balls (Picard and High 1973) are not found. The fact that some scours are lined with fine sands suggests that scour-and-fill did not always occur during the same flood event. This may be due to a localised temporary lack of coarse sediment or rapidly falling stage resulting in fallout of suspended load.

The presence of syn-depositional deformation structures such as slumps on foresets could also reflect rapid discharge fluctuations. Examples from recent fluvial sediments are documented by Coleman (1969) and Levey (1978). Water-escape structures are probably indicators of rapid sedimentation as outlined earlier.

Deformation of fine sands by loading beneath coarser sediments may also be due to rapid deposition trapping water.

Lithologically, channel sediments are clearly immature first cycle sediments with a fairly local source, indicated by the abundance of pink feldspars and fissile rock fragments (mostly phyllite). Lithic clasts point to a basement source, comprised predominantly of metamorphosed sediments and granites.

Cross-sectional geometries and structures suggest that the channels were ribbons (Friend *et al.* 1979) and did not migrate laterally, for epsilon cross beds (Allen 1963) are largely absent. Palaeocurrent data supports a low to moderate sinuosity system with a source to the north-west. As the Berlengas and Farilhões Islands consist of exposed granites and schists, and are situated some 15km to the northwest of Porto Novo, it is reasonable to suppose that weathering of these or associated horsts provided the sediment source during the Kimmeridgian.

2.2.3 Massive mudrock facies

Description

There are two major sub-facies within the massive mudrock facies: *red-brown homogeneous* and *dark green mudrocks.* The latter occasionally contains preserved laminations and is also rich in plant debris, including logs.

Both subfacies consist of coarse blocky micaceous siltstones and silty claystones, the former often rich in sand and granule-grade clasts. Occasionally, isolated centimetre to decimetre, structureless sand beds are found. Thicknesses of mudrock units are highly variable, ranging from tens of centimetres to eight metres.

Palaeosols are widely developed in these facies. The simplest consist of completely homogeneous purple silts (within both red-brown and green silts) rich in coarse sand grains. Desiccation cracks are often developed, one to two centimetres wide, penetrating up to a metre with a sandy or silty fill. Some examples of soils of this type are developed in thin layers (ca 0.5m) of silts directly above channel sandbodies, the upper parts of which are often structureless and mottled purple, green and brown.

Mudrock units frequently contain diffuse-to-dense concentrations of carbonate nodules a few millimetres to centimetres in diameter and 0.5-3m thick. The more advanced examples of this facies consist of a dense packing of < 5 cm diameter glaebules (Brewer 1964), 0.3-1m thick. These may have a diffuse lower boundary, becoming more densely packed upwards. Massive and laminated horizons (Goudie 1973) are not found.

Silts associated with accumulations of carbonate nodules exhibit either a brown/green mottling, or dark green or purple colouration, the latter resembling the gritty silts with no carbonate accumulation.

A thin section from one of the more major carbonate accumulations is illustrated in Figure 2.10. The carbonate glaebules exhibit a crystic plasmic fabric (Brewer 1964), a characteristic of modern caliches. Coarser radially oriented calcite crystals, border skeletal grains whilst a randomly oriented microspar occurs elsewhere. Often skeletal grains are fractured and exhibit corroded margins and are similar to features in a modern caliche illustrated by Brewer (1964).

At a few localities there are vertical downward branching carbonate pipes up to 5 cm in diameter and 0.5m long, which normally have a hollow core. One example occurs within coarse channel fill sediment overlain by desiccated purple gritty silts. Very occasionally vertical carbonaceous traces are found in the silts, and in one instance a fossil tree root was observed *in situ* above an accumulation of carbonate glaebules.

The principle difference between the two mudrock subfacies is their oxidation state and, associated with this the presence or absence of plant debris. Red brown silts are structureless and plant debris is completely absent, whereas dark green silts occasionally have sandy ripple and parallel laminae preserved and often contain abundant lignified plant debris. Bright yellow elemental sulphur is frequently found associated with the organic debris.

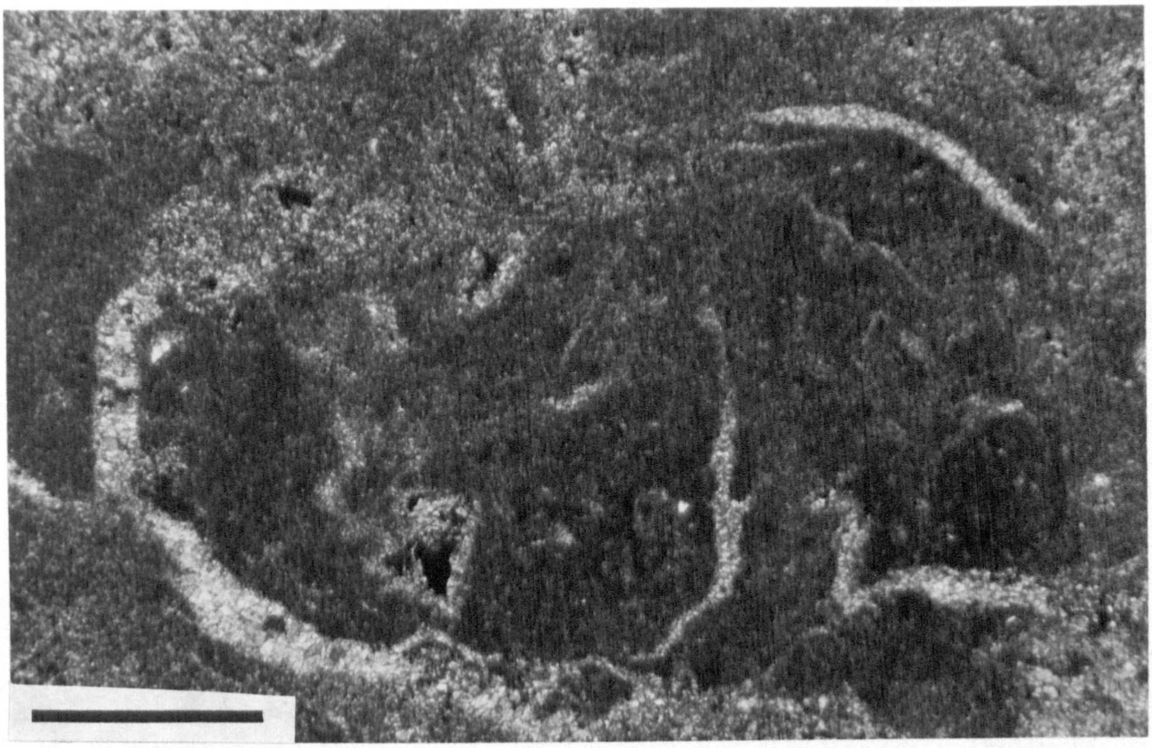


Figure. 2.10. Thin section photomicrograph of carbonate glaebules exhibiting Crystic Plasmic fabric (Brewer 1976). Scale Bar 1mm. Crossed polars. Early nodule rimmed by curved fractures with coarser calcite fill encloses a corroded sand grain with a radial spar rim. The fracture systems observed indicate that the glaebules were subject to in-situ brecciation, probably due to stresses induced by the displacive growth of carbonate, or by vertic 'churning' (Soil Survey Staff 1975).

Dark green silts occur at a limited number of levels in the section, but are volumetrically important as they are often thick (up to 6m) and of considerable lateral extent (2 km minimum). When overlain by coarser facies, there are usually large load casts with a relief of ten centimetres or more. In addition, channel sandbodies are often incised into these mudrocks.

Interpretation of massive mudrock facies

The presence of significant thicknesses of mudrock indicates the existence of considerable periods of low-energy sedimentation dominated by fallout from suspension. The purple gritty silts and various accumulations of carbonate are all interpreted as the products of pedogenesis. Variations between them were probably largely due to the length of time they were exposed to pedogenic processes. The purple silts are interpreted as fluvent entisols (Soil Survey Staff 1975); i.e. 'weakly developed soils' developed on alluvial deposits that are not confined to any particular climate zone. Their desiccation and homogenous nature suggests prolonged sub-aerial exposure and the purple colouration, which is very distinctive, is similar to that found in other soil profiles in the section, where carbonate has accumulated. An alternative classification would be as an Aridisol (Soil Survey Staff 1975); i.e. a semi-desert soil, usually dry and (in this case) with no horizonation developed. The gritty texture is probably the result of mixing of sand and silt in the solum.

The micro- and macroscopic features of the major glaebular calcretes confirm these as ancient equivalents of modern calcareous soils (Brewer 1964; Goudie 1973; Reeves 1976; Young 1976). Their classification according to the scheme of Soil Survey Staff (1975) would be as either Aridisols or Vertisols. The main problem with the latter term is that it implies a significant percentage (>30%) of expanding clays in the solumn. No XRD analyses have been carried out so this cannot be quantified. Though there is some evidence of shrink - swell behaviour in the occasional presence of desiccation cracks, other features such as pseudoanticlinal joint planes (Allen 1974, 1986), similar to features found in some modern calcretes (Reeves 1976), are absent. Classification of the glaebular calcretes as Aridisols therefore seems more appropriate

In the light of the petrographic characteristics of the massive caliches, which are equivalent to the 'stage III' calcic soils of Machette (1985) (after Gile et al. 1966) and their interpretation as palaeosols, it is probable that the less dense accumulations of nodules represent immature 'stage I or II' calcic soils (Machette op cit.). These seem to form an intermediate member between the desiccated silts with no carbonate accumulations and the better developed stage III profiles. As the various soil types appear repetitively, climatic conditions were probably relatively stable. Therefore, the control upon the degree of soil development is likely to have been the time available for pedogenesis i.e. the duration of periods of non deposition (see below).

The downward branching 'pipes' of carbonate are interpreted as rhizocretions, 'pedodiagenetic mineral accumulations around plant roots' (Klappa 1980). Their gross similarity to glaebular carbonate in hand specimen and tendency to occur at discrete horizons at outcrop supports such an interpretation. Rhizocretions are a relatively uncommon development in the Praia da Amoreira member. They are much more common in the fluvial meander belt deposits of the overlying Porto Novo member, perhaps reflecting the greater organic productivity in this environment, or simple preservational bias.

The time taken for caliche formation can be highly variable, depending upon factors such as the availability of carbonate and amount of rainfall (Gile et al. 1966; Goudie 1973; Machette 1985). However low-density accumulations of nodules develop in 10^2 years and well developed profiles form in 10^3 years (Reeves 1976). In less than optimal conditions this may take up to 10^6 years (Allen 1986). This indicates that there were lengthy periods of non-deposition, for an actively aggrading sequence would not allow the formation of discrete soil profiles (Kraus and Bown 1986; Bown and Kraus 1987; Kraus 1987).

The general lack of visible sedimentary structures in mudrocks could simply be due to little grain-size contrast, but bioturbation and pedogenic processes, such as vertic churning (Soil Survey Staff 1975), may have obliterated such features.

The dark green mudrocks are rich in plant material, which is probably responsible for their reduction. They may represent intervals where the climate was slightly more humid so that organic productivity was greater. Alternatively, the preservation of plant material may be due simply to preservational bias. Locally, isolated ribbon channels traversed this environment, with their 'wings' representing levee and/or crevasse deposits. More commonly the environment was oxidising with the development of carbonate soil profiles occurring, normally at the top of mudrock units. The general absence of mudcrack fills, except in soil profiles, indicates either that desiccation did not occur, or that cracks were filled with lithologically indistinguishable material.

2.3 VERTICAL AND LATERAL FACIES RELATIONSHIPS

As described earlier, the three facies are arranged in a series of crude fining-up cycles between one and fifteen metres thick, with a probable lateral extent of several kilometres. Though it is not possible to correlate individual cycles between the Praia da Amoreira and Porto Novo sections (a distance of some 6-9 km), it is possible to correlate major soil profiles (Figure 2.1). In addition, gross features such as dominance of coarse sediment or mudrock can also be correlated. As a result, three fining-up megacycles (*sensu* Heward 1978a, b; Stear 1983) can be distinguished. This correlation indicates that the two sections must be generically linked and were deposited contemporaneously. The megacycles do not show well defined thinning- and fining-up as detailed by Heward (1978a, b) from Stephanian sequences in northern Spain and Hubert and Hyde (1982) from the Triassic Blomidon Beds of Nova Scotia. The Portuguese examples are rather simpler and less regular, for the cycles are defined by either a dominance of coarse sediment or of massive mudrock.

Channels may occur at the bases of the cycles, and in a few examples are incised into the soil profiles beneath. More commonly the erosive base of a cycle has only moderate relief and is overlain by heterolithic facies. In places, mudrock was deposited above the erosion surface. Channels are generally developed some distance above the base of cycles. This is also a feature of the vertically amalgamating heterolithic sub-facies, which is never found incised below the relatively horizontal base of a cycle. Thus it seems that channels were a later development, occurring when a reasonable thickness of sediment had accumulated.

The relative proportions of heterolithic and channel facies to massive mudrock in cycles is extremely variable, with the latter representing between 10 and 75 per cent of the cycle thickness. Greater percentages are generally associated with the thickest of the cycles. Examples are ten centimetres of red claystone above one metre of heterolithic facies and six and a half metres of dominantly silty claystone above eight and a half metres of heterolithic and channel facies.

2.4 DIFFERENT FACIES DEVELOPMENTS

2.4.1 Introduction

No two of the four coastal outcrops of the Praia da Amoreira member are identical. In a number of instances there are significant differences. The purpose of this section is to briefly outline and where appropriate interpret some of the major differences which occur. Further interpretations will be made in section 2.5.

2.4.2 Praia da Amoreira

The lower 50m of the Praia da Amoreira member exposed at Santa Cruz is dominated by sheet-like channel sand bodies usually 2-4m, though up to 11m thick (Figure 2.1). The thicker sand bodies are clearly multi-storey, having a number of major internal erosion surfaces. Channel bases locally exhibit a relief of 1-4m. The sand/mud ratio for this part of the section is as high as 3:1 compared to 1:1.5 for the upper part of the member in the Porto Novo section. This lower part of the Santa Cruz section was studied by Ellwood (1987). He concluded that the sand bodies were probably the deposits of coarse grained meandering rivers. This model was largely based on the fact that palaeocurrent data indicated a rather dispersed pattern. His data was also strongly bi-modal. It consisted of only 16 measurements however. A somewhat larger sample collected by the author (Figure 1.12) indicates no such pattern and is similar to that collected to the north of Porto Novo (Figure 2.14a).

Figure 2.11. Shallow 'winged' margin of channel in the Santa Cruz section. Geometry is similar to those observed in the Porto Novo section and indicates a common origin for the sheet sand bodies near the base of the member at Sant Cruz (see section 2.5). Hammer for scale at centre 0.4m long.

Though steep cut banks (characteristic of meandering systems ?) have been observed, there are also examples of the shallow 'winged' geometries observed to the north of Porto Novo (Figure 2.11). This, is believed to have significance in terms of channel development (see section 2.5). The complete absence of any indication of point bar development is also a problem with Ellwood's model, as he himself noted.

2.4.3 The third megacycle

The uppermost of the three megacycles documented at both Praia da Amoreira and Porto Novo is particularly mud-dominated. Within this unit the regular, repetitive development of individual fining-up cycles breaks down. In their place mudrock is punctuated by packets of coarse sediment which occasionally only have a lateral extent of tens to hundreds of metres, as opposed to thousands. A number of these coarse units appear to stack-up locally (e.g. the most southerly logs at both localities Figure 2.1).

Pedogenic processes seem to have been particularly prevalent with a number of regionally significant caliches developed. Mudrock units frequently exhibit vertically continuous profiles of mottling, desiccation and carbonate nodule development.

At Porto Novo there are two examples of channel sand bodies with large scale lateral accretion surfaces. These both crop out in the top 10m of the member. The best of these appears at the cliff top just south of log A6 (see Figure 1.5) at GR 6963 3799 (Figure 2.12). In this example, the channel fill consists of a laterally stacked series of beds locally separated by thin inclined clay drapes. The direction of migration was from south to north. Sediments at the sand body's northern margin consist of a series of progressively thinner inclined sand beds. The erosive bases of these later units are cut at progressively shallower levels. This arrangement of structures suggests that, in addition to lateral accretion, there was a large vertical accretion component to the channel fill. The channel was apparently progressively abandoned with a gradual reduction in scale. Identical deposits were described by Cuevas Gozalo (1985) from the southern Pyrenees.

2.4.4 São Bernadino

As indicated in section 2.1 the section at São Bernadino, though lithostratigraphically part of the Praia da Amoreira member, does exhibit a rather different facies pattern. There is no evidence for the repetitive fining-upwards cyclicity found in the other three sections. Mudrock dominates the outcrop, accounting for 70% of the sediment pile and plant debris is more common within sand bodies.

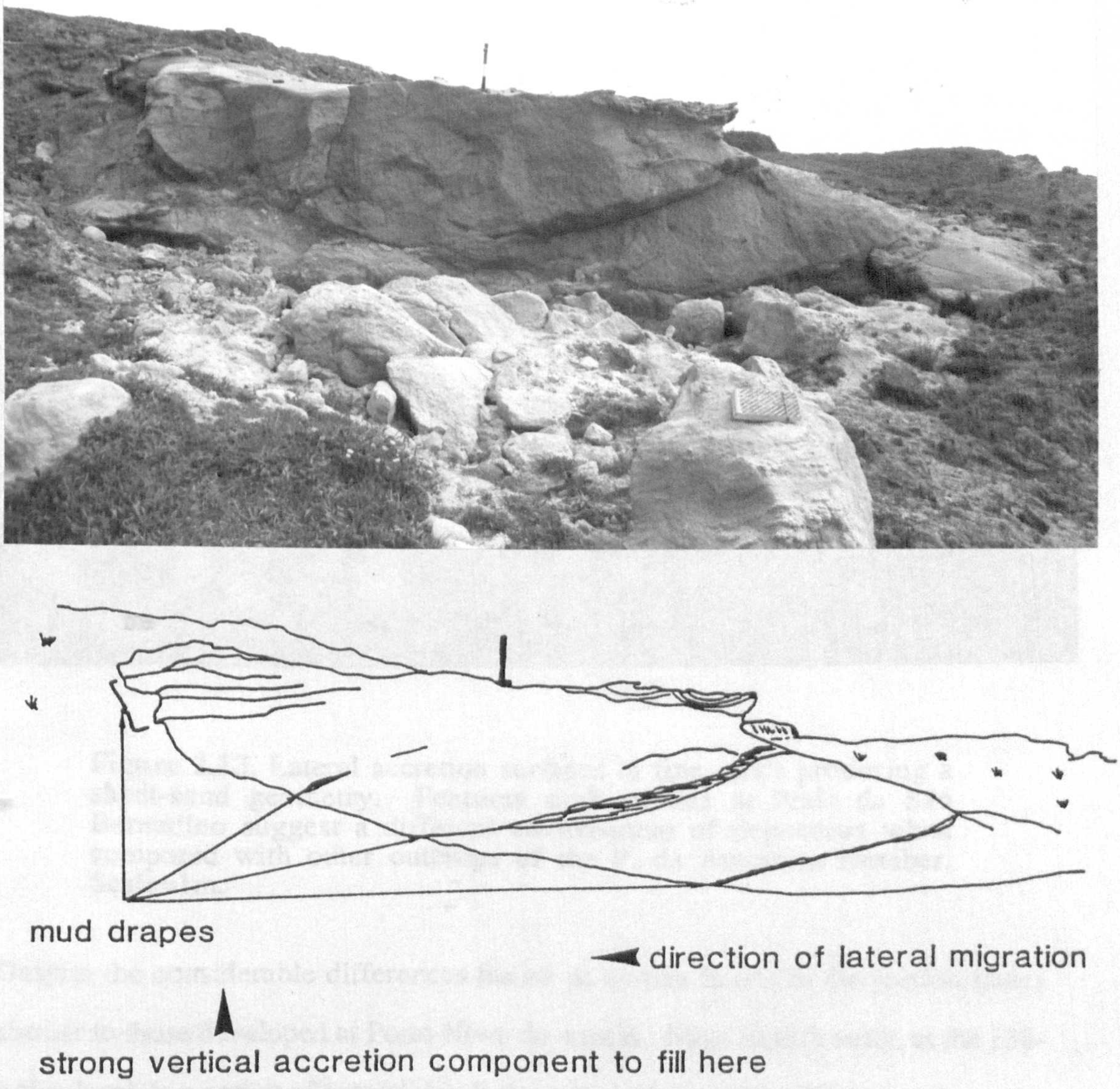


Figure 2.12. Photograph and sketch of lenticular, laterally accreted sand body near the cliff top north of Porto Novo. The channel fill also had a strong vertical component particularly in its abandonment phase. Scale=1m.

The mudrocks characteristically exhibit continuous profiles of mottling, and carbonate rhizocretions are very common (see logs 'SB series'). Perhaps the most significant development is the appearance of clear lateral accretion structures within a number of sand bodies (Figure 2.13). These appear at several levels within the section and are the clearest indicators of a rather different environment of deposition. Palaeocurrent data gathered from the section (Figure 1.11) has a dispersed pattern, characteristic of sinuous fluvial systems.

Figure 2.13. Lateral accretion surfaces in fine sands producing a sheet-sand geometry. Features such as this at Praia da São Bernadino suggest a different environment of deposition when compared with other outcrops of the P. da Amoreira member. Scale=1m.

Despite the considerable differences found at certain levels in the section, facies similar to those developed at Porto Novo do appear. Most significantly, at the 138-145m level, is a packet of heterolithic facies which persists laterally across the entire width of outcrop. Channel sand bodies in the vicinity of this unit tend to have the lenticular 'winged' geometries found elsewhere.

2.4.5 Ferrel

The Praia da Amoreira member crops out at the base of the Ferrel section and exhibits many of the facies found at both Porto Novo and the type section. As indicated earlier an additional feature is the widespread presence of dessication structures at certain horizons. The member is relatively thick at this locality (ca. 200m exposed and the base of the member not exposed). The most unusual feature is that the section is divided into two parts by an 8m-thick, sheet sand body with the lithological characteristics of the Porto Novo member (Figure 1.10, see Chapter 3)

2.5 ENVIRONMENTAL MODEL

In view of the strongly aggradational character of the sequence, the petrography of the coarse sediment, the volume of mudrock, the palaeocurrent data and proximity to the postulated basin margin (some 15 km to the northwest), it seems likely that the environment of deposition of the Praia da Amoreira member was a distal fan sand or mudflat (Hardie *et al.* 1978; Hubert and Hyde 1982; Flint 1985) which passed laterally into a fluvial meander belt (Figure 2.14b). As flash floods flow off the >4^{o} slope of the fan and spread out on the <1^{o} slope of the 'flat' they experience a rapid deceleration (Hardie *et al.* 1978, Hubert and Hyde 1982) resulting, in this case, in deposition under lower flow regime conditions. Fan gradients are based on analogy with Hardie *et al.* (1978) and the presence of caliches in the flat environment suggest gradients of 2 - 4 m km $^{-1}$ (Goudie 1973).

2.5.1 Cycle development

Given that channel sandbodies do not always occur at the base of cycles, they were not necessarily an early feature of a cycle's development. This may have been due to suppression of channel formation by high suspended loads (McGee 1897, Van der Meulen 1986). Therefore it is likely that cycles were initially the deposit of an

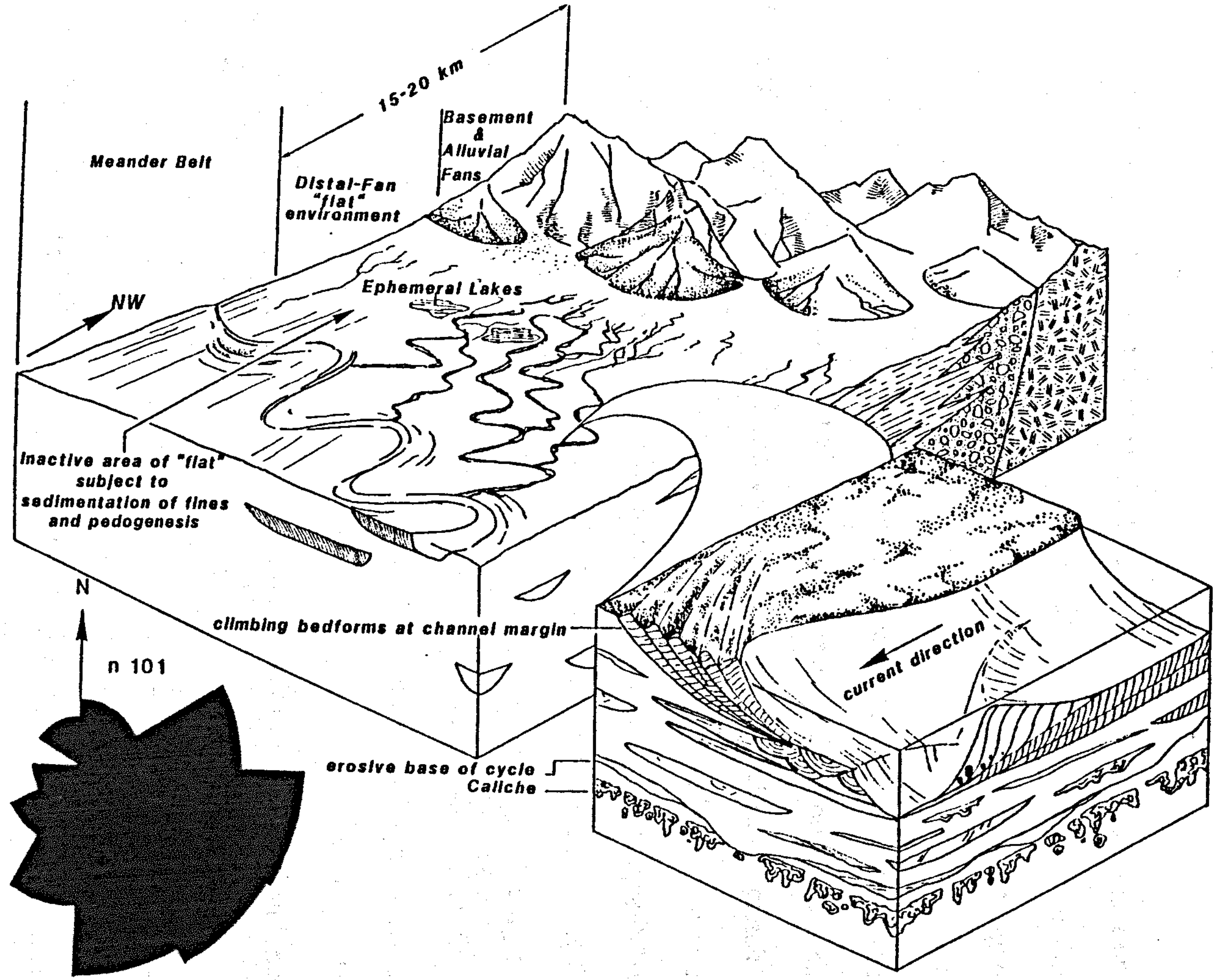


Figure 2.14a. Equal-area palaeocurrent rose from cross-beds within channels. The data indicates low to moderate sinuosities with a source to the northwest. Measurements from within individual sand bodies normally show relatively little dispersion, but occasional exceptions include different levels within a multi-storey sand body, and locally at channel margins where cross beds may be directed obliquely outwards. Data from the latter feature is not included.

b. Distal fan flat environmental model for the Praia da Amoreira member. Cyclicity was controlled by switching of the active area of fan sedimentation. This environment passed laterally into a fluvial meander belt characterised by larger scale palaeochannels, finer sediments and abundant vegetation (not shown). The "flat" environment illustrated is at an evolved stage with channelised flow well established.

unconfined, moderately erosive sheetflood with a high sediment load which deposited fine sands and silts over a broad area. It is probable that individual cycles were initiated during the switching of the active area of sedimentation on the fan surface. Such a process has been advocated for similar sequences by Heward (1978a, b) and Hubert and Hyde (1982).

If channels were not necessarily a feature of the early development of the cycles, then what controlled their initiation? The key seems to be in the vertically amalgamating (heterolithic) sub-facies (Figure 2.5). Small 'winged' channels probably developed at a point where, through chance, two or three sand beds were stacked on top of each other. During subsequent flood events, the sand beds would have been less cohesive and therefore more easily eroded than surrounding mudrocks. This allowed the initiation of a small channel which tended to concentrate flow during later flood events, leading to enlargement and upstream migration of the channel head (Schumm 1961, 1968). The subsequent development of a channel proper would be dependent upon the channel incising itself, or on the contemporaneous build-up of the surrounding sediments producing the stacked 'winged' margins illustrated in Figure 2.8. It is possible that flow became progressively more channelised as the sediment pile of each cycle developed. The winged morphology indicates that channels were rarely very deep. They filled rapidly by vertical accretion of bedload and were gradually abandoned so that eventually activity switched to another area of the fan. Channel and heterolithic facies deposition were succeeded by massive mudrock deposited from suspension from low magnitude floods. At a late stage, sufficient sediment had accumulated so that the area was rarely inundated and soil formation and desiccation became the dominant processes.

The calcretes present are related to present day Aridisol, Entisol, or Vertisol types (Soil Survey Staff 1975; Young 1976; Bridges 1979) and indicate a semi-arid climate with seasonal rainfall of probably less than 500mm and mean temperatures of 16-19°C annually (Goudie 1973). Considering the approximate timescales required for the formation of calcretes (Reeves 1976; Machette 1985), periods of negligible sedimentation (i.e. between cycles) were probably in the order of 10^2 to 10^4 years. This fits at the lower end of the time scale estimated for periods of inactivity on modern fan systems using similar criteria (Bull 1977). In view of the generally less mature 'stage I to III' profiles developed (Gile et al. 1966; Machette 1985), this is probably not unreasonable, despite the considerable variations possible in rates of calcrete development (Reeves 1976).

2.5.2 Megacycles

It is very likely that the duration of sedimentation and resultant thickness of individual cycles had a dominantly autocyclic control. However the megacycles may well have had an external control, being triggered by fault activity at the basin margin which led to the rapid progradation of the fans. Subsequent fining-up would be related to erosion that lowered relief in the highlands and the consequent progressive retreat of the fan head. A number of authors have attributed similar ancient sequences to this process (Heward 1978 a, b; Hubert and Hyde 1982).

The absence of clear *thinning* and fining-up sequences, so frequently observed in fan sequences needs explanation. The occasional features indicating sediment liquefaction in the section, suggest seismic activity. This could have overprinted the development of simple megacycles, and provides another possible explanation for the considerable variability in the scale of individual cycles, in addition to the autocyclic mechanism discussed above.

The Praia da Amoreira and Porto Novo sections can only be correlated on gross features, plus the more significant soil profiles. It is possible that the two sections represent the deposits of separate fans. If this is the case then the fans would only be linked by the effects of tectonism producing the large-scale cyclicity. Thus during quiescent periods, more advanced development of soil profiles would occur over several fans.

The dominance of channel sediments found at the base of the type section has two possible explanations. As the first megacycle is the thickest of the three developed, it may be the product of early, greater magnitude faulting as the fans began to develop at the basin margin. Such activity would most likely, result in the production and transport of significant quantities of coarse detritus to fan-distal areas.

An alternative explanation, is that the Santa Cruz locality lay in a distal position to a breach in the footwall between two en-echelon faults at the basin margin. According to Leeder and Gawthorpe (1987) large alluvial fans are sourced at such sites. The other features of the sand bodies at the base of the section, i.e. their multi-storey nature and the relief observed at their bases may all be a function of the dominance of coarse sediment, this contributing to a less cohesive, more easily erodable substrate.

2.5.3 Associated Environments

The top of the Praia da Amoreira member is marked by a thick (1-2m) 'stage III' calcrete (Gile et al. 1966; Machette 1985), representing a significant break in sedimentation. This is overlain by the deposits of the fluvial meander belt which must have progressively onlapped the shallowly inclined basin margin sequences. Further evidence supporting this conclusion comes from the outcrops at Ferrel and São Bernadino. At the former, the punctuation of typical Praia da Amoreira member facies by the presence of thick fluvial channel deposits typical of the Porto Novo member indicates that at the toes of the fans, fluvial channels onlapped during periods of low or inactive alluvial sedimentation. At São Bernadino, the facies present are typical of a sinuous fluvial system interspersed with the distal fan facies. As petrographically, the coarse facies are indistinguishable from those of the Praia da Amoreira member, this feature supports the interpretation of a distal fluvial transition beyond the toes of the active fans. In the light of this the more mud-rich third megacycle, which includes the early development of laterally migrating channels, probably also represents a transitional environment.

2.5.4 Climate

Finally, why was the substrate so wet and desiccation features so rare despite the occurrence of well developed calcretes? During the Late Jurassic, western Iberia lay at about 30° N. Upper Jurassic climates, particularly in western Pangea, are generally acknowledged to have been much warmer and drier than those at similar latitudes today (Drewry *et al.* 1974; Smith and Briden 1977; Frakes 1979; Hallam 1982, 1984). Hallam (1982) suggested that this was due to easterly wind systems blowing over land. This interpretation is supported by a study of atmospheric circulation by Parrish and Curtis (1982), whose maps suggest that the continental wind systems were most dominant during the winter months. During the summer months, easterly and south-westerly systems, blew onshore from Tethys, or the newly opened Central Atlantic respectively. If this was the case, then the summer months would have been more humid than the winter months. This may explain the low incidence of desiccation structures, as the time of maximum evaporation coincided with that of maximum precipitation. Obvious desiccation cracks are largely confined to soil profiles which developed upon inactive areas of the fan flat and so would only have been wetted directly by seasonal precipitation and not run-off.

2.6 SUMMARY AND CONCLUSIONS

During the late Kimmeridgian, a system of small low-relief alluvial fans distributed coarse, basement-derived clastic sediments south-eastward from upfaulted blocks at the western margin of the Lusitanian Basin. Distal-fan sediments comprise heterolithic, massive sandstone and mudrock facies deposited in a series of crude fining-upward cycles 1-15m thick. These sediments indicate a very 'flashy' regime with floods carrying large volumes of sediment, both in suspension and as bedload.

Deposition was often rapid as floods waned and sedimentary structures indicate that turbulent sub-critical flow conditions were dominant.

An autocyclic control on facies sequences is probable, as the active area of sedimentation on the fan surface switched positions. The fining-up cycles and lack of fine grained 'abandonment fills' in channels, support the idea of avulsive initiation, followed by gradual abandonment. The thickness of cycles may have been controlled simply by the length of time a particular segment of the fan was active, or by short-term tectonic factors. The variable character of mudrock units is probably the result of the presence or absence of significant volumes of plant debris. Frequently, although the sediment surface was often wet and small bodies of standing water probably existed, the sediments were oxidized and floral remains are consequently absent. Soil profiles represent breaks in sedimentation of between 10^2 and 10^4 years. At any one time most of the fan surface was probably inactive. Therefore individual cycles probably represent shorter periods of time, perhaps as little as 10^2 to 10^3 years.

The correlation of sections 6-9 km apart allows the delineation of three fining-upwards megacycles (*sensu* Stear 1983; Heward 1978 a, b). These are likely to have been initiated by faulting at the basin margin followed by denudation of the headlands and consequent fan retreat. The cycles do not show regular thinning and fining-up trends so common in alluvial fan sequences. This could be due to overprinting of the simple cyclic pattern by numerous minor earthquakes which also triggered the periodic liquefaction of sediments. The thick mudrock units at the tops of megacycles probably indicate tectonically quiescent periods.

Close examination of channel deposits indicate that they were not often an early feature of cycle development. They tended to form subsequently as a reasonable thickness of poorly cohesive heterolithic facies had accumulated. This allowed the excavation of scours and the initiation of channels. It is therefore possible that flow became progressively more channelised as the sedimentary pile of each cycle developed.

The distal fan environment passed laterally into a fluvial meander belt, which ultimately onlapped the fan system as its sedimentary activity waned.

CHAPTER 3

MEANDERING RIVERS: THE PORTO NOVO MEMBER

3.1 INTRODUCTION

The Porto Novo member has the most widespread outcrop area of all the members of the Lourinhã formation. Extensive cliff exposures appear between Porto Novo and São Bernadino, north of Ferrel and from 8km south-west of San Martino do Porto to Foz do Arelho. Inland, as with the other members of the Lourinhã formation, outcrop is poor, being largely restricted to overgrown roadside cuttings.

The majority of the data presented have come from the Porto Novo to São Bernadino sections. As indicated in Chapter 1, the more northern sections were studied in less detail. This was partly due to time constraints as the quantity of exposure is enormous. The data gathered was used to contrast and compare the sections to illuminate any regional variations. Environmentally, the facies indicate a meandering fluvial system, with sedimentation occurring on point bars and, during floods, on levees and muddy floodplains. The fluvial system was characterised by high suspended loads and significant discharge fluctuations. The presence of carbonate soil profiles, though these are not as well developed as those in the Praia da Amoreira member, indicate a semi arid to sub-tropical climate with marked wet and dry seasons. This is consistent with the evidence for discharge fluctuations found within the channel sediments.

Overbank sediments are generally less brightly coloured than those found in the P.da Amoreira member. In addition, considerable quantities of plant debris are preserved, largely in channel facies. Rhizocretions are common in soil profiles. These factors indicate that the central river valleys were more humid and supported a diverse flora when compared to the marginal fans on which the Praia da Amoreira member was deposited. The most notable fauna are the bones of large reptiles, commonly found in channel lags. In particular the area around Porto Dinheiro (Figure 1.3) is well known as a source of fossil reptile teeth (Thulborn 1974).

There is direct evidence for an active extensional tectonic regime during the deposition of the Porto Novo member (section 3.5). Strong subsidence is also required to accomodate the sediment pile.

3.2 DESCRIPTION OF THE STUDIED EXPOSURES

The main exposure between São Bernadino and Santa Cruz can be divided into four main segments. In each, the strata dip at low angles of only a few degrees. They will be described briefly from north to south. Data on the sections logged appear in Figure 1.10. Factors which affect thickness estimates are listed in Appendix 1.

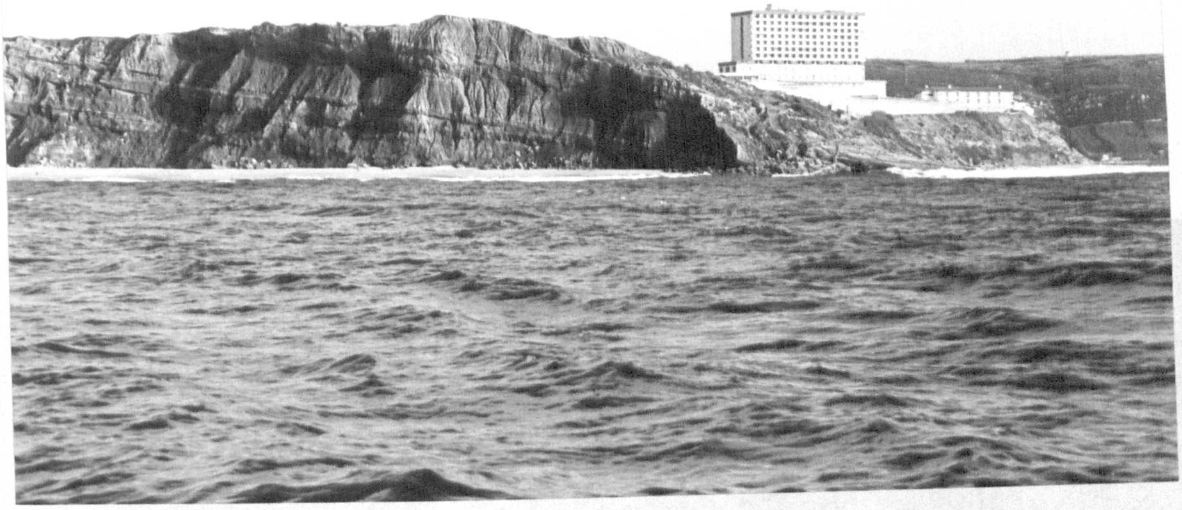
Each section consists of a series of sheet-like sandbodies, 3-8m thick, frequently with a lateral extent of hundreds of metres. Often, sand bodies extend across the full width of outcrop. As a result, cut banks and abandonment fills are rarely seen. Low angle (point bar) surfaces are often seen in cross section, their visibility

Figure 3.1. (over), Three photographs of cliff sections of the Porto Novo member. All illustrate the sheet-like sand body geometries.

a) The type section north of Porto Novo. Cliff trends NE-SW. The member is 67m thick at this locality extending from the base of the first sheet sand body at centre left to the 'plateau' at the cliff top which is a marine horizon. The 'point bar' outcrop (see 3.4.1) sits on the headland at lower right. Steep hillside behind the hotel is the west-facing flank of the Vimeiro diapir. Figure 3.15 is of the cliff base at the centre.

b) 600m north of Porto de Barcas, cliff trending N-S. Thick sand body at cliff base is 7.5m thick. Lateral accretion surfaces are apparent in two of the sand bodies. Thin sheet sand near the cliff top right of centre is a marine horizon.

c) 150m north of the Forte de Pai Mogo, cliff trending NE-SW. Sheet sand body at cliff base is the subject of Figure 3.32 and is ca. 4m thick. Lateral accretion surfaces are apparent, dipping from left to right. The sand body at the cliff top is a multi-storey unit. Hillsides behind exhibit the typical rolling relief characteristic of the Lourinhã formation inland.



enhanced by the regular occurrence of packets of muddy sediment between inclined beds of sandstone. The sandbodies are separated by massive mudrock, dull grey to red in colour which appear in approximately equal volumes to the sands (Figure 3.1).

Figure 3.2. The Porto Novo member downfaulted to the south (right) against the P. da Amoreira member. Southern end of Praia dos Frades (see Figure 1.11). The sand bodies in the foreground form part of a multi storey unit in the cliff section to the south of the fault (see Figure 3.39). Contrast this with the section to the left of the fault which contains a single sand body above a heterolithic unit. Cliff height ca. 45m.

The base of the member is not exposed to the south of São Bernadino. At this locality, a stack of fluvial channel sandbodies is downfaulted against similar, but more muddy facies of the Praia da Amoreira member.(Figures 1.11 and 3.2). Over 300m of the Porto Novo member are exposed in sections dipping shallowly to the south. These extend for some 5km, as far as the north side of Praia da Areia Branca where the section is broken by a fault and outcrop is lost for nearly a kilometre. The Porto Novo member is punctuated by thin tongues of the brackish to marine Praia

Azul member, between the Forte de Pai Mogo and GR 7120 4767 just south of Ponta de Vale de Frades (between 200 and 280m in the vertical section).

A further long section extends for 4km northwards from GR 7050 4105 just south of Porto Dinheiro to GR 7085 4502 on the south side of Praia da Areia Branca. The base of the member is exposed above the P. da Amoreira member but a probable fault at Porto Dinheiro prevents determination of the true thickness of the member. Neverthless some 170m of fluviatile sediments, punctuated by <7m-thick tongues of the Praia Azul member (above the 100m level), are exposed. The remaining sections, at Porto Novo and between Praia da Amoreira and Praia Azul are both complete and record the continued progressive southward thinning of the member from >300m in the north to 67m at Porto Novo and 47m thick to the south.

In outcrops to the north, at Ferrel and south of San Martino do Porto the Porto Novo member is also significantly thicker than at its type locality, being 330m and 800m respectively. The estimate made for the section to the south of San Martino has room for significant error due to the difficult nature of the outcrop, particularly the possibility of significant undetected fault displacements on the flanks of the Caldas diapir. The Porto Novo member is overlain by the gravelly Santa Rita member in these more northerly sections. Immediately below the Santa Rita member, the Porto Novo member frequently exhibits significantly coarser grain sizes than are found in the type section. This, particularly in the Foz do Arelho/San Martino section, makes the identification of a distinct member boundary difficult.

3.3 FACIES DESCRIPTION AND INTERPRETATION

Five main facies have been identified. These are:

1) Caliche pebble conglomerates

2) Sand facies

3) Mudrock within the sand facies

4) Heterolithic facies

5) Mudrock facies

Each type will be described and interpreted in turn, with a general lithologic description and where necessary, a sub-division based on the structures found.

3.3.1 Caliche pebble conglomerates.

Irregular pebble-grade clasts of light brown to white carbonate are the most obvious component of this facies. The conglomerates are normally dull grey to buff due to the presence of other components. These include mudflakes, sandstone pebbles, and fragments of wood, plant debris or logs (Figure 3.3), in a siliciclastic matrix of variable, but generally coarse sand grade. Mudclasts sometimes reach cobble grade but clasts larger than this are extremely rare. Amounts of lignitic debris present are particularly variable.

Both matrix- and grain-supported conglomerates are developed, though the latter are more common. Grading is locally present, with grain-supported orthoconglomerates passing up to paraconglomerates and coarse sands. Where there is good exposure of large-scale lateral accretion units, grading can also be traced laterally up the face of the bar from pebble conglomerates into sands.

This facies most often appears at the base of massive sand bodies above an erosion surface. As such it may form a thin veneer or exist in beds up to a metre or more thick. Structures are extremely variable. Beds may be massive, exhibit a rather crude horizontal to low angle stratification, or be planar to trough cross bedded. Cross beds are normally about 0.2m thick, appearing as isolated sets or cosets, but may be up to a metre thick. Isolated cross-bedded scour fills are occasionally found. Cross bed cosets may include interbedded sandy cross beds or be separated by drapes of ripple or parallel laminated, grey, fine sand to silt. Where good three dimensional exposure exists (see section 3.4.1) crudely stratified or massive conglomerates can be seen to pass laterally, within a lateral accretion unit, into trough cross bedded conglomerates and sands.

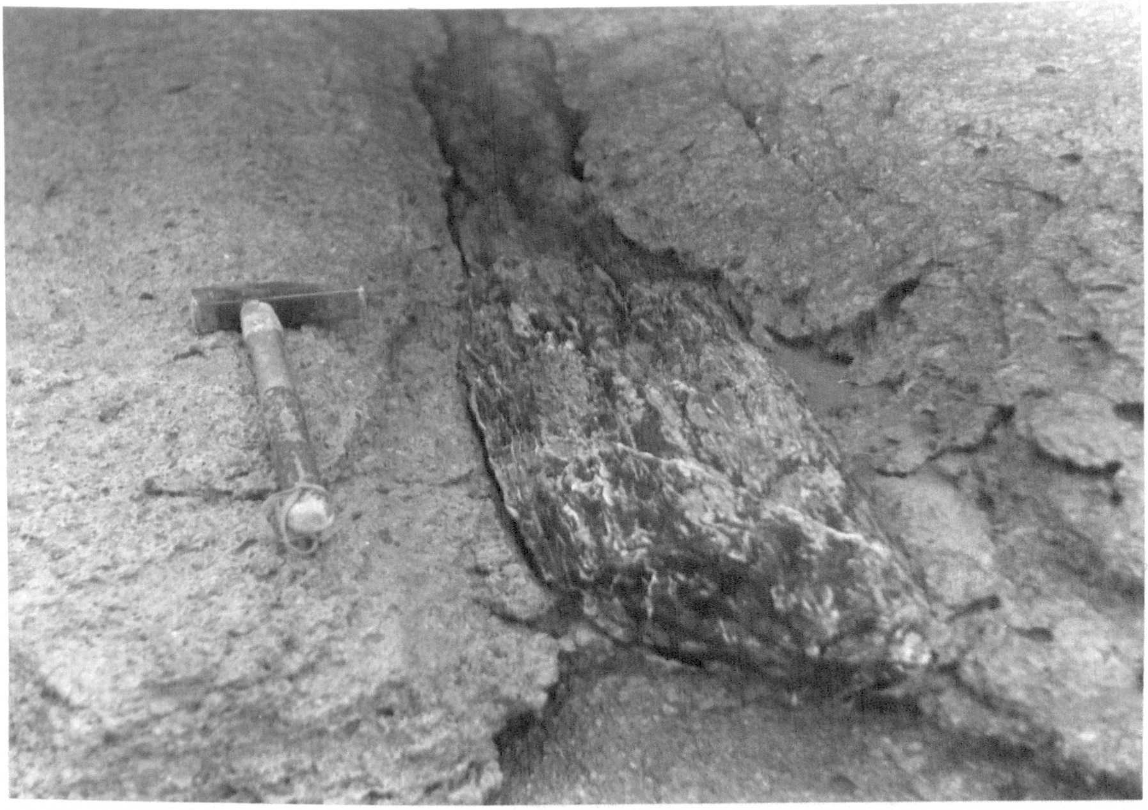


Figure 3.3. Silicified log enclosed within caliche and mudclast pebble conglomerates near the base of the 'point bar' outcrop. Hammer 0.4m long.

Locally, beds of this facies may be found elsewhere within massive sand bodies, exceptionally in their upper parts. In one example of the latter the sediment consists almost entirely of low-density components such as carbonate pebbles and mudclasts with very little sandy matrix. Deformation at the bed base is so severe that the bed varies in thickness from 0.1 to 0.4m. Detached balls of conglomerate have deformed enveloping silts, and mud flames project up into the bed base.

Interpretation of the pebble conglomerates:

This facies clearly represents channel lag deposits derived largely from the re-working of older floodplain and channel deposits. The major process was the disaggregation of slumped bank material, this providing the carbonate pebbles and much of the mudrock. Similar nodules of carbonate, of probable pedogenic origin appear within the mudrocks bounding the sand bodies. The presence of the pebbles within the channel sand bodies indicate that the soil profiles were an early feature of floodplain development and that channels may have swept across the floodplain over 10^3-10^4 years (this is the sort of timescale envisaged for the development of such relatively immature soil profiles according to Allen 1986).

The presence of rather poorly stratified to chaotically organised beds suggests that the conglomerates were deposited during floods. Smith (1972) carried out tests on the durability of mudclasts and concluded that they do not resist disaggregation if transported. In the light of this and their abundance and close association with caliche nodules in lag deposits, it seems likley that most of the material in the conglomerates was derived locally. Where crudely stratified, the conglomerates probably formed broad bars with shallow crests on the channel floor. The subdued relief of these features could be associated with high stream flows during flood.

The lateral transitions observed in good three dimensional exposures (see 3.4.1) indicate that these 'bars' had marginal dunes which in turn, passed laterally up-dip into sandy dunes on the faces of the major lateral accretion units. The few examples where this facies is found in the upper parts of the sand bodies probably represent exceptionally high energy flood events. The derived material could have been swept out of the channel or derived locally from eroded floodplain deposits. The severe soft sediment deformation observed indicates that the depositional surface was water saturated. In the one example mentioned above, with its limited lateral extent the carbonate clasts may have been delivered via a crevasse channel. The conglomerates are over- and underlain by mudrocks some 0.5m above channel deposits.

3.3.2 Sand facies

This is the most diverse facies grouping found within the Porto Novo member and volumetrically is second only to the massive mudrock facies. The full range of sand grades is present but most common are coarse to fine grades. The facies is sub-divided in terms of the grades and structures found:

Granulestones and very coarse sands are relatively uncommon, normally only appearing in the lower parts of sand bodies. Their mineralogy is dominated by sub-angular to sub-rounded quartz and feldspar with occasional pebbles of pink orthoclase feldspar, granite, phyllite, mudclasts and carbonate, plus minor quantities of lignite. The few examples seen are either cross bedded with decimetre-scale cosets, crudely parallel laminated, or form scour fills up to 0.7m-thick at the base of sand bodies. This facies becomes increasingly common in the upper part of the Porto Novo member in the northern sections.

Cross-bedded sands consist of coarse to fine sands exhibiting a full spectrum of cross beds from planar, tabular cosets through to trough cross beds. Individual sets are 5cm to over a metre thick, most often about 0.1-0.3m thick. Toesets of both angular and tangential type are found, the latter more common in medium to fine sands. Caliche and mudclast pebbles locally form lags at the base of sets or lie on foresets.

Cross beds usually appear in cosets a metre or more in thickness. In a few examples, the entire channel fill is dominated by decimetre-scale tabular cross beds. Generally the size of sets decreases up through a sand body. In other examples cross beds are rhythmically interbedded with sheets of caliche pebble conglomerate or planar laminated sands. Locally in these associations, cross beds form rather thin isolated sets. In the absence of a gravel lag, cross bedded sands appear as scour fills at sand body bases or as isolated large-scale sets up to a metre thick.

Primary current lineated, parallel laminated sands of medium and fine grades are very common. Bedding surfaces often exhibit high concentrations of coarse muscovite and finely comminuted plant debris. Laminae commonly appear in bundles up to 0.5m thick, separated by low angle erosion surfaces. In a number of examples, this facies comprises the dominant element of, or in an extreme example, the entire sand body thickness (Figure 3.4).

Figure 3.4. Margin of sand body dominated by parallel laminated sandstones. Note the rather abrupt termination at the top of the sand body. It is overlain by brightly coloured silts containing carbonate nodules. Beneath the erosive base of the sand body, at lower right, a series of inclined bedding planes of an older unit crop out. Location; cliff top on downthrown side of fault 1500m north of Port Novo (see Figure 1.5). Scale bars on hammer 0.1m.

Parallel laminated sands often pass up into facies exhibiting structures such as trough cross beds (Figure3.5) or, in graded beds in the upper parts of sand bodies, to very-fine sands and silts which are dominantly ripple laminated. These graded beds commonly repeat in a cyclic manner in decimetre-scale beds which also exhibit red-brown colour mottling and bioturbation. Burrows are commonly simple, cylindrical and unlined, in a variety of orientations.

Laminated fine sand facies: light grey to buff, or occasionally red, and ripple laminated sands are a common constituent of the upper parts of sand bodies. Concentrations of muscovite and finely comminuted plant debris often accentuate laminae. Laminated sands also form centimetre- to decimetre-scale drapes over coarser facies at lower levels in sand bodies draping, for example, a pebble lag. The upper surfaces of cross bedded facies are locally ripple laminated. Climbing ripple laminae are relatively uncommon, but examples do occur and, exceptionally, appear in thick units dominating a sand body. The best examples of this facies occur between São Bernadino and the Forte de Pai Mogo. Decimetre-scale ripple laminated sand beds are often cyclically interbedded or form graded units with ripple and parallel laminated, organic rich silts in the upper parts of sand bodies. Sand beds above silty units, frequently either contain small pebble grade mudclasts, or have micro load casts on their bases.

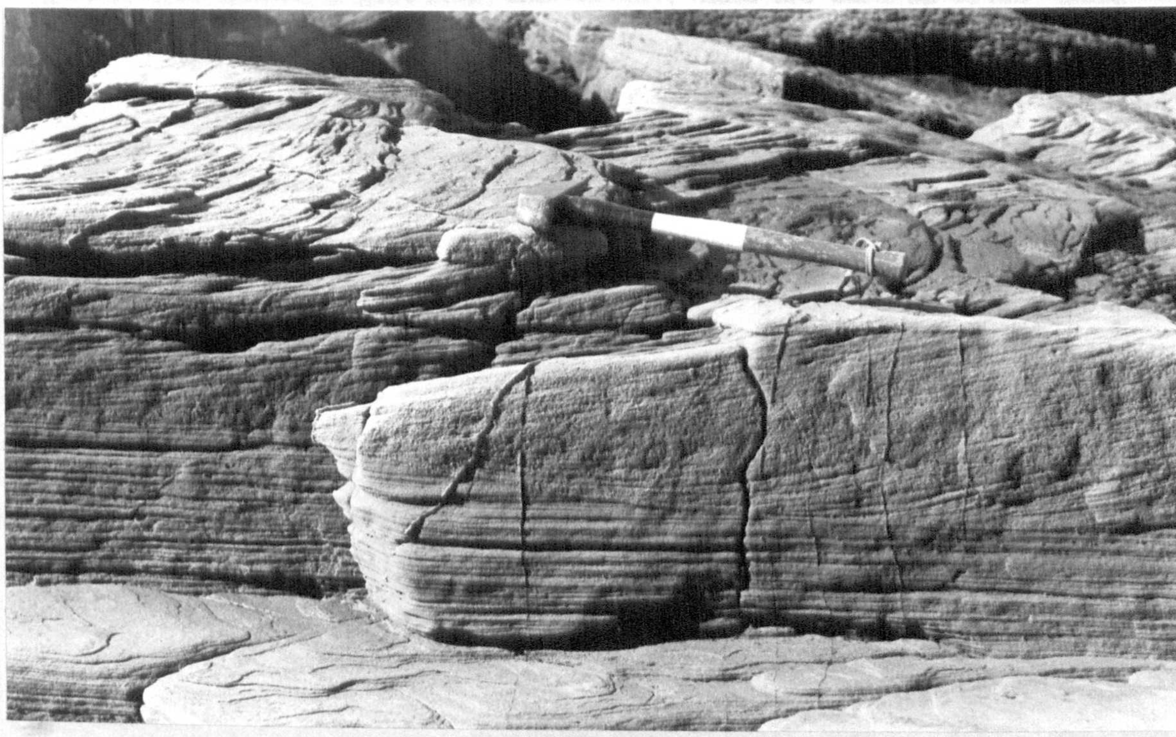


Figure 3.5. Vertical transition from current lineated to trough cross bedded sandstone. A possible indicator of waning flow (Harms et al. 1982). Outcrop immediately below the erosive base of the 'point bar' channel (see 3.4.1) Porto Novo. Scale bars on hammer 0.1m.

Simple ripple laminated sands may form a significant proportion of sand body thickness elsewhere in the Porto Novo member, exceptionally in beds up to a metre thick. Such beds are commonly normally or occasionally inversely graded. They may also appear in composite beds of ripple, climbing ripple and parallel laminated sands with decimetre thick sets of structures bounded by low angle, intraclast lined, erosion surfaces. Within these units, isolated cross beds of fine to medium sand, up to 0.25m thick and with several metres lateral extent may appear. Simple burrows frequently cross cut the laminated facies.

Channel sand macroforms form a rather varied facies of relatively uncommon, but striking bedforms, partly because of their scale. A few individual examples will be described to illustrate the range of bedforms that occur. To avoid confusion, each feature will be interpreted individually rather than at the end of this section.

1) On the south side of the bay at Porto Novo are a series of unusually thick, multi-storey sand bodies (Figure 3.6). Individually, these are each up to 8m - thick.

Figure 3.6. Lateral accretion surfaces and thick multi-storey sand bodies in cliff on the south side of the bay at Porto Novo. Sand bodies sit close to the base of the Porto Novo member. Thin sheet sand at cliff base belongs to the P. da Amoreira member. Camera facing SE.

On the west face of the outcrop is an axial cross section of one of these units which exhibits two interesting sandy macroforms. The first of these is a 1m-thick cross bed of medium sandstone the foresets of which have decimetre spaced, rather sigmoidal, bounding surfaces (Figure 3.7).

Figure 3.7. Bedform with sigmoidal foresets, 1m thick, close to the base of a storey within a multi-storey sand body. Location 50m SW of Figure 3.6. Camera facing due east.

Interpretation : The subdued bar crest resulting from the sigmoidal geometry strongly suggests that current velocities were high, close to the transition between cross and planar bedding. This bedform resembles the hump-back bars of Allen (1983) who likewise suggested close proximity to upper flow regime conditions.

2) At the same locality as the hump-back bar is a series of descending decimetre-scale planar cross-bed cosets, the bounding surfaces of which are inclined up to 10° down-current. A better exposed example of this facies appears to the south of Porto Dinhero. At this locality, individual tabular cross sets or packets of ripple laminated fine sand, 0.1 to 0.3m thick, bounded by laterally extensive planar erosion surfaces, are inclined at 10° from the palaeohorizontal (Figure 3.8). These facies are together over 3m thick and appear at the top of a sand body. They are directly overlain by a

graded unit of laminated fine sands and silts which pass up into massive red-brown mudrocks.

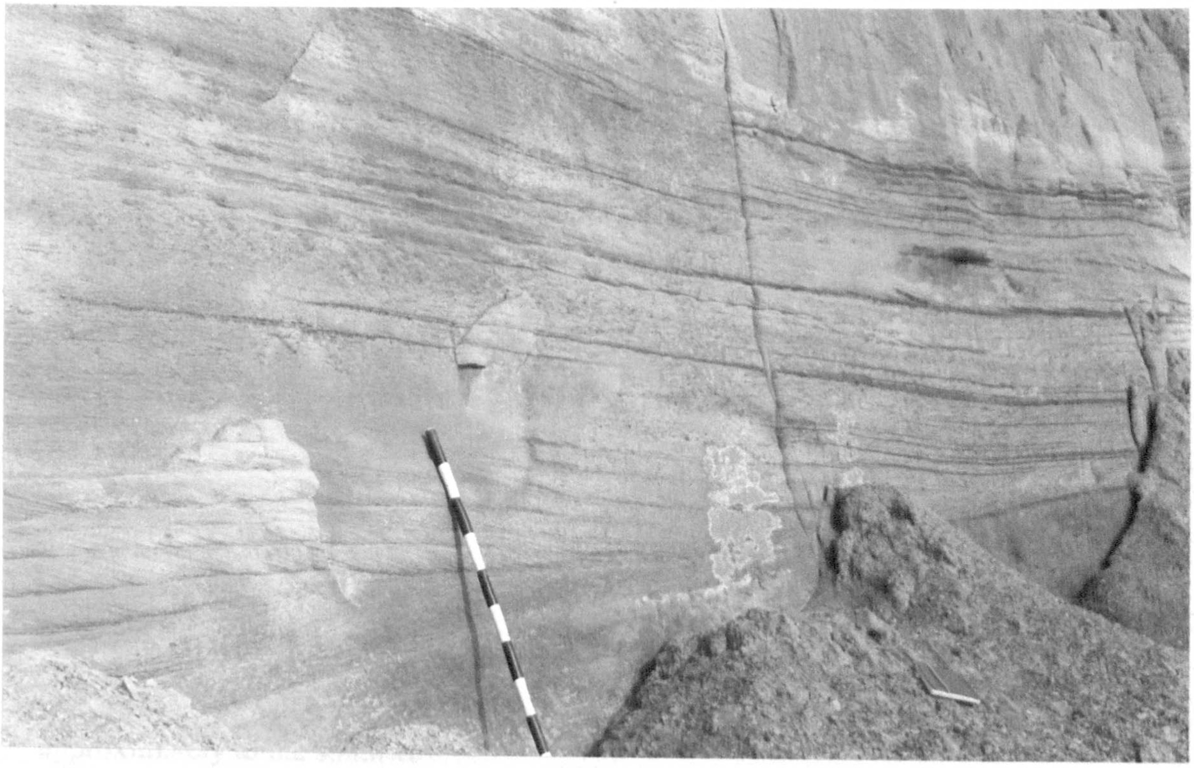


Figure 3.8. Descending tabular cross beds punctuated by horizons of ripple laminae. Dark bands are fine lignitic debris. Tectonic dip is from right to left, marked by horizon 60cm down from top of scale (Scale bars 10cm). These bedforms probably migrated down the lee slope of a much larger, slower moving bedform. Location south of Porto Dinheiro GR 7035 4025.

Interpretation : As this unit sits at the top of a sand body it was probably only shallowly submerged. It has a strong resemblance to the descending tabular cross beds described by Hazeldine (1983). He interpreted this facies as the result of small dunes descending the shallow lee slope of a larger, slower moving macroform. A similar interpretation is favoured for these Portugese examples. The additional packets of ripple laminae are probably the result of low stage re-working of the larger dune bedforms.

3) An unusual bedform which appears at the cliff top to the south of Porto Dinhero is an isolated cross bed of well sorted, medium sand, within what are otherwise interpreted as levee deposits (Figure 3.9a). The cross bed sits above a silt-draped

scour, in the top of a multi-storey channel sandbody. The cross bed thickens from 1 to 2m over its (axial) length of some 20m and in front of the bedform is a concave-up scour, approximately 5m wide at its base. This scour has a horizontally bedded fill, of grey silts and sharp-based 1-10cm beds of ripple laminated buff fine sandstone. In the upper part of the scour fill, the silts are red-brown and contain carbonate nodules. The margin of the scour is defined by truncated parallel laminated sands at its base and interbedded sands and silts in its upper parts. The sequence overlying the bedform is illustrated in Figure 3.9b. Palaeocurrent data with which to examine facies relationships is somewhat sparse and also enigmatic (Figure3.9c). The bar foresets (1) dip toward 036° whilst palaeocurrents, in what are interpreted as levee deposits above the bar, trend in completely the opposite direction (2). There are no data available from the (channel) sand body directly beneath the bar, but about 70m to the north palaeocurrents in the same unit trend between SSE and ESE (3). Features interpreted as point bar surfaces dip approximately to the southwest (4).

Interpretation : If the cross bed is related to the channel deposits below, then a possible scenario for its development is outlined in Figure 3.9d. It seems that the bedform may have occupied a chute channel, which migrated away from the direction of point bar growth and was itself a lateral accretion deposit, with its foresets oriented obliquely to the axis of its associated channel.

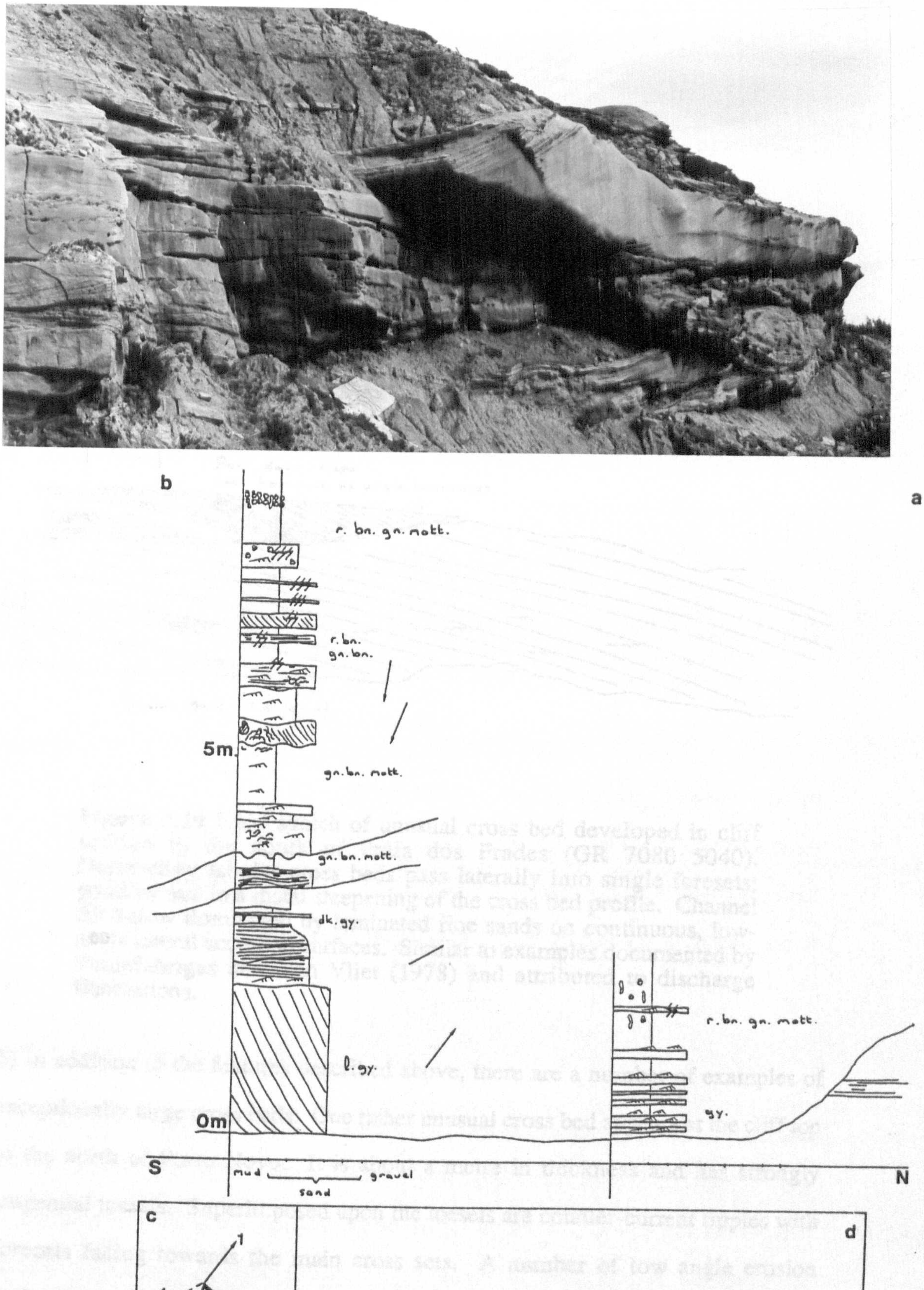
Figure 3.9. (over), Unusual 'bar' at the top of a multi-storey sand body at GR 7047 4015, south of Porto Dinheiro:

a) Photograph of the section (metre scale at centre). 'Bar' terminates in a scour which truncates the sand beds at the top of the sand body (at top left). Note also; the irregular erosive base of the main sand body; possible lateral accretion surfaces and fine (abandonment ?) fill in older sand body below the scale; and the single cross bed (at centre-left margin) indicating palaeoflow into the cliff face (i.e. at right-angles to the orientation of the large bar's foresets).

b) Graphic log of the 'bar' and the facies developed above it; interpreted as a levee deposit.

c) Palaeocurrent data from the main sand body, the 'levee' facies and the 'bar'.

d) Possible scenario for the field relationships observed.


a
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gy.
0m
S
N
mud
sand
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c
1
2
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4
d

4) Similar examples of the facies described in '3' above occur elsewhere: between São Bernadino and the Forte de Pai Mogo and to the north of Foz do Arelho. Both have foresets oriented approximately 90° to the associated channel's palaeocurrent direction. The former example is interesting because the bedform consists of a series of descending tabular cross beds, which pass laterally into a single large foreset, possibly because of a distal steepening of its profile (Figure 3.10).

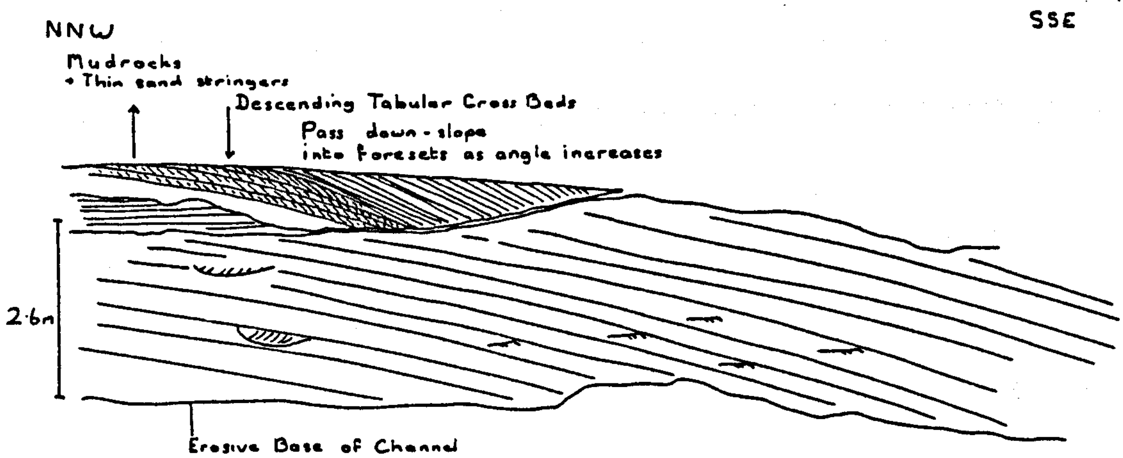


Figure 3.10 Field sketch of unusual cross bed developed in cliff section to the south of Praia dos Frades (GR 7080 5040). Descending tabular cross beds pass laterally into single foresets; possibly due to a distal steepening of the cross bed profile. Channel fill below dominated by laminated fine sands on continuous, low-angle lateral accretion surfaces. Similar to examples documented by Puidefabrigas and Van Vliet (1978) and attributed to discharge fluctuations.

5) In addition to the features described above, there are a number of examples of exceptionally large cross beds. One rather unusual cross bed appears at the cliff top to the north of Porto Novo. It is about a metre in thickness and has strongly tangential toesets. Superimposed upon the toesets are counter-current ripples with foresets facing towards the main cross sets. A number of low angle erosion surfaces truncate the foresets, above which are about 10cm of ripple laminated sand. These are not re-activation surfaces (vis. Collinson 1970), because the foresets above the laminated sands have a different inclination (and by inference, orientation) to those beneath. It is thought however, that these do have an origin similar to that

proposed by Collinson (op cit.) for reactivation surfaces, ie. low stage re-working of 'bar' foresets, followed by continued 'bar' accretion. What makes this structure unusual is that it appears to be an exceptionally large scour fill. The cross bed terminates in a concave-up scour which the foresets fill completely.

Structureless sands consist of massive lenticular beds of coarse to fine sandstone which occasionally appear stacked up on inclined bedding surfaces. Thicknesses vary from about 0.1 to 2.0m. Mud drapes usually separate individual beds, the bases of which may be deeply erosive with pebble to boulder grade mudclast lags. Alternatively, some bed bases are not deeply erosive but exhibit load casts with a relief of up to 0.3m.

A number of beds are structureless at their bases but have visible structures in their upper parts. One notable example has type 'A' climbing ripple laminae in its upper part. Grading is common in structureless sands

Deformed sediment is common, particularly at loaded contacts between coarse and fine facies. Within sandy facies there appear a variety of deformation structures from ruptured laminae to slump structures on cross-bed foresets. Also, isolated 'pockets' of deformed sediment are frequently found associated with clumps of lignitic debris.

Interpretation of the sand facies

The sand facies is interpreted as the deposits of fluvial channels. This interpretation is based on the presence of erosive bounding surfaces at the base of the sand bodies, the suites of structures present and the terrestrial indicators, notably palaeosols (see section 3.3.5), in associated facies. The variety of sand grades and suites of structures present, indicate that the channel environment was extremely varied experiencing wide fluctuations in discharge. Also, it seems that there were variations in regime between channels. This aspect is examined further in section 3.4.

High stream velocities, above 1m/sec were achieved during flood as the presence of significant thicknesses of current lineated sandstones testifies. The presence of thick bundles of this facies suggests that periodically, high velocities were maintained for significant periods of time, perhaps due to a series of seasonal rain storms. The occasional thin cross beds, which appear within parallel laminated sets and also upward transitions to cross beds or ripple laminae record the waning of floods (Harms et al. 1982).

Cross bedding, being the most common structure found, indicates that migrating straight to sinuous crested dunes were the most common bedforms, particularly in lower and middle parts of the channels. Current velocities of 0.6 to 1m/sec are implied (Harms et al. 1982). At the other end of the spectrum, the presence of considerable thicknesses of ripple laminated sands occasionally at low levels in a channel sand body, indicate that at low stages much lower current velocities were prevalent.

The limited number of macroforms observed are generally confined to the thicker channel sandbodies, rarely being found in those thinner than 3-4m. The larger channels in the system may have developed these large bedforms during large magnitude floods when significant quantities of bedload sediment were transported. The presence of probable 'humpback bars' adds weight to this proposal as it infers high velocities, close to upper flow regime conditions. Elsewhere, or perhaps at other times, channels carried little coarser bedload. This resulted in the development of isolated sandwaves enveloped in laminated sediment. Such features may be due to a localised deficiency of coarse sediment or to relatively low stream velocities moving only small quantities of bedload. The macroforms which appear at the top of channel sandbodies were probably the result of quite major floods when bypassing of a meander loop was initiated. At such times, the stream velocity was sufficient to allow the transport to an elevated position, of the significant quantities of coarse sediment required for their formation.

Thick beds of structureless sediment are possibly the result of rapid deposition during a waning flood or alternatively local bank collapse. The deeply scoured bases of a number of these units suggest that deposition was preceded by a strongly erosive flow or alternatively may represent failure planes. Both mechanisms would have entrained much sediment which rapidly settled out during waning stages. The presence of deformed sediments also suggest rapid deposition. The trapping of considerable quantities of water would subsequently result in the rupturing of laminae due to water escape, or slumping on bar surfaces, both due to high pore pressures developing as sands consolidated during falling river stages. Where deformation is associated with clusters of plant debris, it is possible that this was due to excessive compaction of vegetation.

3.3.3 Mudrock within the sand facies

This facies is closely associated with the sand facies, being interbedded with sands in the channel sand bodies. Dark, organic rich silts frequently form centimetre- to decimetre-scale drapes between sand beds. Such drapes, particularly in mid to upper parts of the channel sand bodies may exhibit sand filled desiccation cracks (Figure 3.11). This mudrock facies often contains significant volumes of plant debris. Parallel and ripple laminae are the dominant structures encountered. Climbing ripple laminae are relatively uncommon.

The amount of muddy sediment found within the sand bodies varies considerably. It may exist only as thin drapes between metre thick sand beds, or may form part of a rather heterolithic channel fill of decimetre interbedded lenticular sands (see section 3.4.1) where mudrocks represent up to 50% of the channel fill. As described earlier, mudrocks in association with fine sands dominate the upper parts of channel sandbodies.

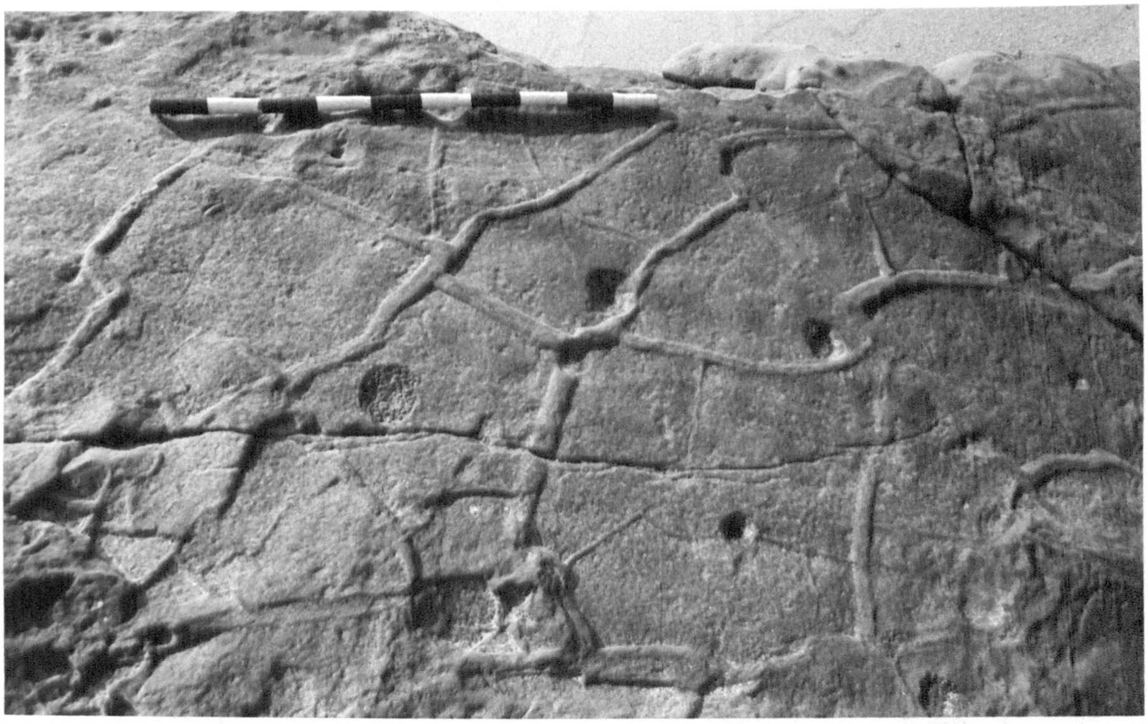


Figure 3.11 Desiccation crack fills on the base of an upturned block from a channel sand body. These indicate that at low stages a significant proportion of the channel bed was sub-aerially exposed. Cliff base north of Porto Novo. Scale = 1m.

Interpretation of the mudrock within the sand facies

The presence of significant volumes of mudrock within channel sand bodies, testifies to the high suspended loads that the streams carried, particularly during flood. Wide fluctuations in discharge are also indicated by the coexistence of sand and mudrock. This is supported by the desiccation of mudrocks within channels indicating that subaerial exposure of much of the channel bed occurred periodically. Variability in the proportion of mudrock present could reflect a number of things; e.g. relative availability of sediment, preservation of deposited sediment, or differing conditions within channels of different scales. In addition external controls affecting base levels such as tectonism or eustacy, may increase aggradation rates and therefore, the preservation potential of fine channel sediments. It certainly seems to be the case that the larger channel sand bodies, such as that exposed at Porto Novo (see section 3.4.1), contain a rather more heterolithic fill than the smaller channels which generally seem to be sand dominated. This could reflect a greater erosive power in the larger streams during flood, resulting in the incorporaton of much locally derived bank material.

3.3.4 Heterolithic facies

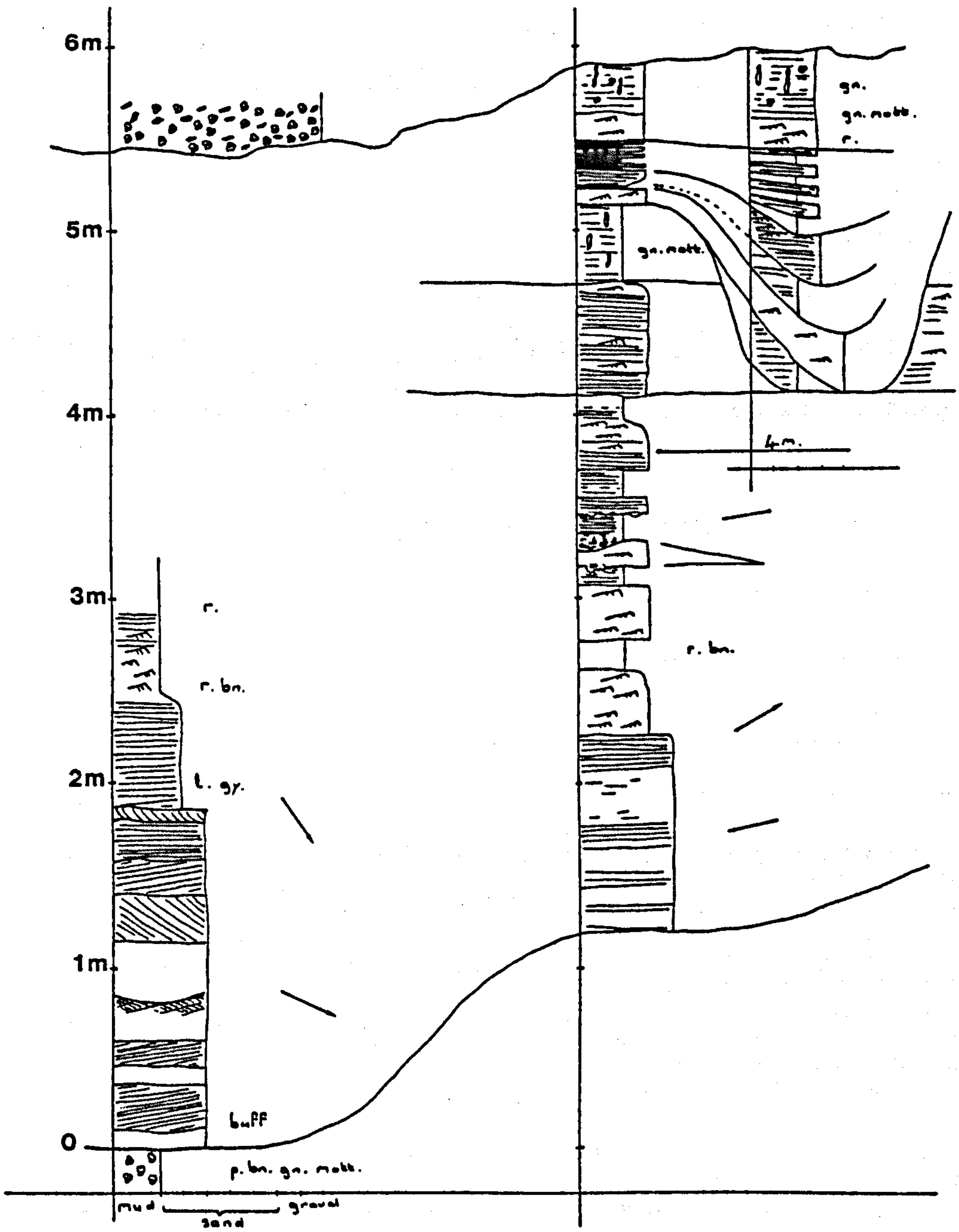


Figure 3.12 Graphic log through a small channel sand body and the heterolithic facies developed at its margin (above ca. 3m); GR 7040 4977 north of the Forte de Pai Mogo.

This facies is found in the upper parts of the channel sandbodies. It is characterised by decimetre bedded silts and very fine sandstones. Bedding is often irregular with thicknesses varying laterally. Thin biscuits of fine sand, rarely more than 0.1m thick and a few metres wide are also interbedded. Their bases may contain mudclast pebbles but the sands are otherwise often structureless or exhibit a few simple burrow casts. Sand beds are commonly normally graded. Occasionally, coarser deposits up to coarse sand grade are found. These frequently contain entrained pedogenic nodules and plant debris.

The sediments as a whole are rather structureless, due to bioturbation. They may be mottled, red-brown and even contain rootlets or small carbonate nodules. Mudrocks are often desiccated. Alternatively they may be dull grey and contain significant quantities of plant debris, which is occasionally pyritised. This facies is rarely thicker than about 1m, often rather less. Where primary depositional structures are preserved in the sands, these usually take the form of ripple or rather undulose parallel laminae. Exceptionally, an erosive based, single tabular cross bed draped by mudrocks is preserved.

A graphic log through a particularly well preserved, though somewhat atypical example of this facies is illustrated in Figure 3.12. The channel sediments below thicken northwards as the heterolithic facies lie close to the channel margin. Features of this deposit include laterally persistent bedding and well preserved structures dominated by parallel and ripple laminae. Beds thinner than 0.1m tend to pinch out laterally. Desiccation structures are rare, but the sediments are coloured a bright red-brown and the upper parts frequently contain small nodules and vertical pipes of carbonate. Near the top of the succession is a scour about 1m deep and 8m wide, with a fill in which individual beds thicken towards its centre and lap out onto its margins. Sands tend to be coarser in the axis of the scour than on its flanks. Palaeocurrent data and the axis of the scour are orientated approximately perpendicular to measurements recorded within the channel sand body beneath. Such scours are not a common feature of the heterolithic facies.

Interpretation of the heterolithic facies

From their position at the top of channel sand bodies, plus the frequent signs of sub-aerial exposure such as desiccation cracks, rootlet horizons and the carbonate nodules (interpreted as pedogenic accumulations; see section 3.3.5) the heterolithic facies is interpreted as being the deposits of levèes and crevasse splays. The structureless nature of much of this facies is attributed to bioturbation and pedogenic processes. Rapid deposition on the banks of channels by floodwaters was probably often preceded by minor amounts of erosion resulting in the mudclasts and (pedogenic) carbonate nodules often found within sand beds. The larger scours are probably crevasse channels. Isolated cross beds closely resemble modern crevasse splay deposits (e.g., O'Brien and Wells 1986).

It is often not possible to make a simple distinction between channel fill and levee facies, even in Holocene sediments (Nanson 1980). In the Porto Novo member this distinction is made difficult because, due to widely fluctuating flows channel fill sediments were often exposed and subject to weathering and pedogenesis. Hence there is a grey area between those sediments deposited within and without the channel. Generally, only where cut banks are observed, or exposure is particularly good, is a clear distinction possible.

The generally restricted or thin development of sandy levee facies suggests that most large magnitude floods did not carry large volumes of coarse sediment onto the floodplain. Significant quantities of suspended sediment were carried over channel banks however, as the thick mudrocks in the sections testify (see section 3.3.5). This distinction suggests that flooding was usually characterised by a widespread overtopping of channel margins rather than localised crevassing. The development of reddening, dessication cracks and pedogenic carbonate all suggest a semi-arid sub-aerial environment (Allen 1974; Leeder 1973).

3.3.5 Massive mudrock facies

Figure 3.13 Sinuous vertical pipe of carbonate in brightly coloured mudrocks. These features are interpreted as rhizocretions (sensu Klappa 1980). Scale bars = 0.1m.

Generally, mudrocks associated with the Porto Novo member are not as brightly coloured as those of the Praia da Amoreira member described in Chapter 2. Texturally they are very similar, consisting of coarse siltstones to claystones which are structureless and frequently micaceous. Bright red and purple mudrocks do appear at certain horizons. More commonly they exhibit continuous ochrous green-brown mottling and contain diffuse accumulations of carbonate nodules up to 10mm diameter. Dull grey or green silts, commonly containing plant debris or even logs are also common. Within the mudrocks are a wide range of carbonate accumulations, which include vertical pipes up to 0.5m long and 0.1m diameter (Figure 3.13). Locally these are associated with rootlets surrounded by reduced zones in otherwise oxidised sediment. Horizons of amalgamated carbonate nodules 0.1 to 2.0m thick, and interlocking, blocky, prismatic carbonate horizons generally only 0.1-0.3m thick are also present. Composite units with a combination of the above features occur locally.

Figure 3.14 Cliff section 500m SE of the fault illustrated in Figures 1.11 and 3.2. Profile dominated by highly coloured mudrocks. Log SB8 was measured just to the right of the outcrop illustrated. Horizontal colour stripes relate to the variable oxidation states of the mudrocks plus the concentration of carbonate nodules/rhizocretions (see log). Cliff height ca. 40m.

The thicknesses of mudrocks vary enormously along the section, not suprisingly being an inverse function of the density of channel sand bodies. However, up to 5m of mudrock between channel sands is common and more than 10m may be present. The maximum development in the absence of a channel is some 20m (Figure 3.14). This equates with almost half of the entire thickness of the Porto Novo member at its southern-most outcrop south of Santa Cruz. Thick mudrocks are particularly common in the parts of the section where thin tongues of marine strata are found; e.g., to the south of Areia Branca, or south of the Forte de Pai Mogo.

Interpretation of the massive mudrock facies

The thick units of mudrock interbedded with fluvial channel sandstones are interpreted as fine-grained overbank deposits. The numerous accumulations of carbonate and highly coloured horizons within the massive mudrock facies are all interpreted as the product of pedogenic processes on basically the same grounds discussed in Chapter 2; i.e. their close resemblance to recent carbonate soil profiles (Goudie 1973; Reeves 1976). As the vast majority of mudrocks persistently exhibit ochrous mottling and diffuse accumulations of pedogenic carbonate nodules it is clear that most of the floodplain mudrocks were subject to pedogenic processes almost continuously. Sediment accumulation rates were sufficiently high to prevent horizonation of the soils for much of the time.

The variability found in the better developed soil profiles is probably controlled by locally dependent factors, such as water saturation, plant growth and availability of carbonate. The degree of development of soil profiles in most alluvial suites is usually attributed to autocyclic controls; i.e., the proximity of a channel, and amount of sediment supplied during flood (Leeder 1975; Allen 1978; Kraus and Bown 1986). Allocyclic controls affecting base levels, such as tectonism and eustacy, are less likely to have a detectable effect in such a relatively thin succession which apparently aggraded rapidly. Both types of control operate continuously on all systems, but determination of their relative importance is difficult and may rely upon other geological clues (see section 3.5.2.).

Kraus and Bown (1986) suggest that temporal variations in base level may only be detectable through variable soil profile development in relatively thick succesions (ie. several hundreds, rather than tens of metres). However, Allen and Williams (1982) conducted an architectural analysis, involving discussion of the degree of soil profile development, on a sequence only a few tens of metres thick. A critical factor in their analysis was exceptionally good outcrop. Where this is available and a particular soil horizon can be correlated with confidence, regional considerations of lateral variability can be made (e.g., Atkinson 1986). A soil horizon which varies in character laterally, due to geomorphic influences, is termed a 'catena' (Bridges 1976; Atkinson 1986).

The most important aspect of the study by Allen and Williams (1982) was the tight stratigraphic control provided by tuff marker beds. As indicated in Chapter 1, stratigraphic control is particularly poor in the Lourinhã formation and though soil profiles proved very effective correlative tools in the analysis of the P.da Amoreira member, this is not the case in the Porto Novo member. In view of the lack of 'control provided by closely spaced and rigourous time markers of regional significance' (Allen and Williams 1982) in the Porto Novo member, an in-depth architectural analysis was not attempted.

The deposition of ca.10m of mudrock without the development of a caliche horizon, could take 17 000-25 000 years, assuming typical rates of accumulation of 0.6-0.4mm per year of fully compacted sediment. Where well developed soil profiles occur, a much reduced rate of sedimentation, perhaps as little as 0.1mm per year is implied: Thus it could take up to 0.1 my. for 10m of sediment to accumulate (rates after Leeder 1975).

3.4 FACIES ASSOCIATIONS

This section deals with the organisation of the facies described above, by examining the distribution of different facies types, using a number of case studies. These reflect the wide range of associations found within the Porto Novo member. In particular, some of the more unusual features alluded to in section 3.3 are dealt with.

Generally the facies exhibit a number of features usually attributed to meandering fluvial systems. Inclined epsilon cross beds (Allen 1963), interpreted as point bar surfaces, can be identified in a number of cases. Above an erosive base, caliche and mudclast pebble conglomerates are often found, indicating that the channel systems migrated laterally and in so doing reworked significant volumes of floodplain sediment. The succession above is often, though not always, observed to fine upwards, eventually passing into the heterolithic facies thought to represent channel levees (Figure 3.15). The levee facies pass into a sequence of homogeneous floodplain mudrocks in which carbonate soils are characteristically developed.

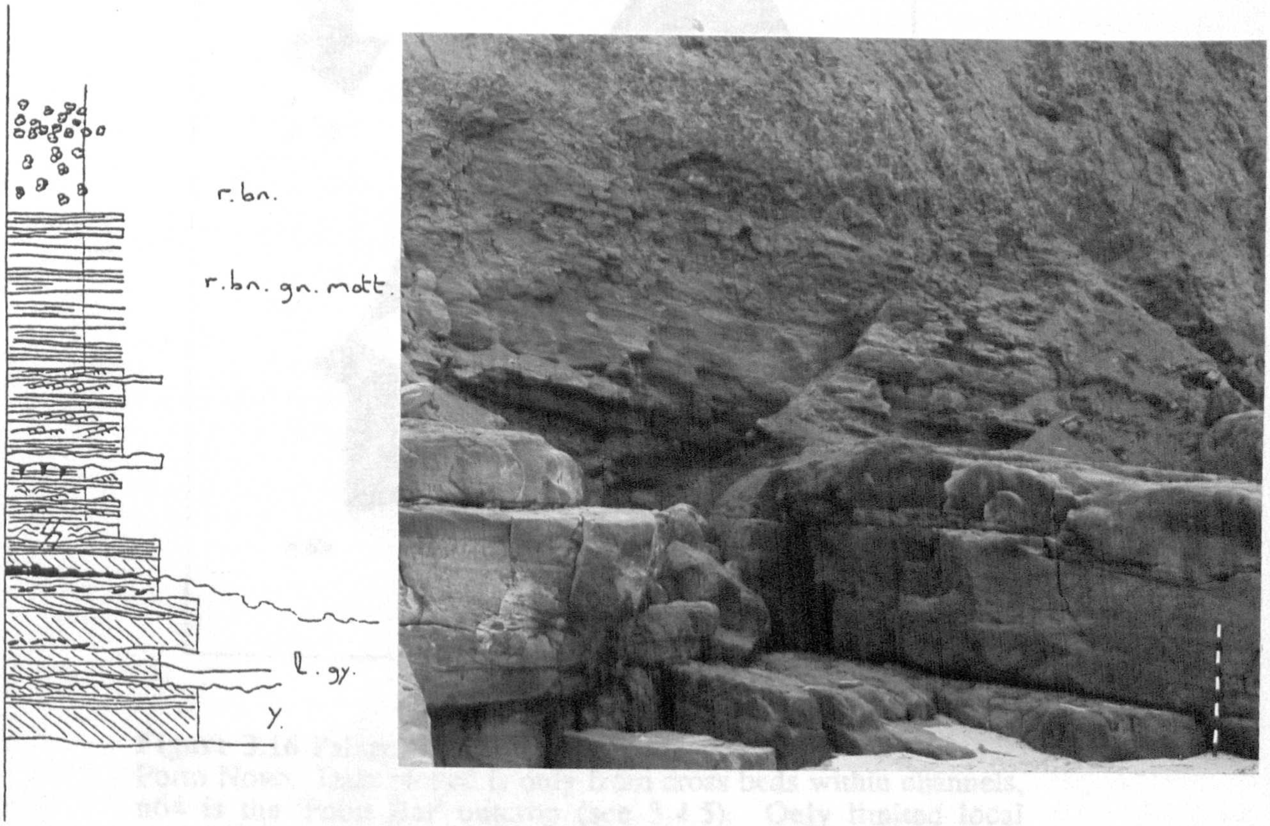


Figure 3.15 Massive sand, heterolithic and massive mudrock facies association. Facies are arranged in a fining-up sequence. Note: the desiccation cracks in the mudrocks (on log), these have a sandy fill; the irregular distribution of carbonate (pale areas in the mudrocks at the top of the photograph). A small fault cuts the outcrop. Location: cliff base 200m north of Porto Novo (on Figure 1a). Scale bar = 1m.

Palaeocurrent patterns usually exhibit only limited local variability within any single channel deposit, but significant dispersion between channels (Figure 3.16). The grouped data for each of the sections at the Forte de Pai Mogo, Porto Dinhero and Porto de Barcas are illustrated in Figure 3.17a. They suggest that,though the channels were sinuous they were not highly so,and that there was a strong south to south-easterly trend; this impression is reinforced by the grouping of all data recorded from São Bernadino to Porto Novo (Figure 3.17b).

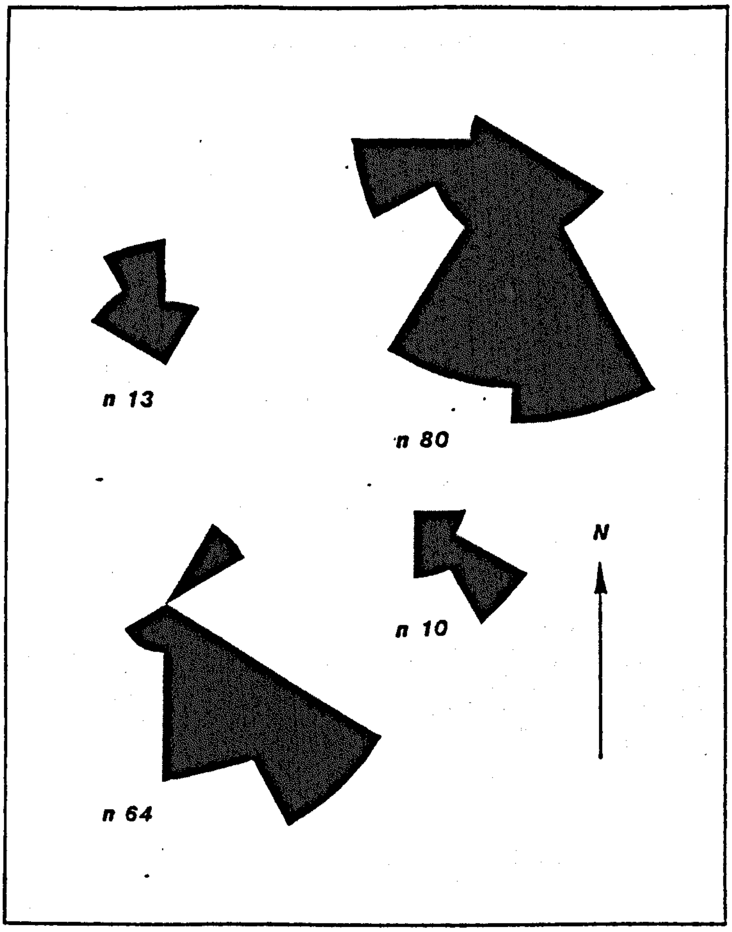


Figure 3.16 Palaeocurrent data from four channel sand bodies at Porto Novo. Data plotted is only from cross beds within channels, n64 is the 'Point Bar' outcrop (see 3.4.5). Only limited local variability is observed within any single channel deposit, but significant dispersion between channels suggests a sinuous system.

Some unusual featues of the succession include a low frequency of identifiable 'abandonment fills'. There seem to be two main reasons for this. Firstly, there is a strong lack of contrast between mudrocks within channels and those deposited on the floodplain. This makes the identification of cut banks very difficult, even at close quarters after excavation of the section with a shovel. This is because:

1) the sediments are very poorly lithified and tend to weather rapidly.

2) the sheer nature of many of the cliffs, which makes close inspection of the section difficult.

3) many channel sandbodies have a markedly sheet-like geometry. This provides continuous exposure of sands and no obvious development of abandonment facies. Data may be slightly biased because of the basin-axial nature of much of the north-south trending coastal outcrop.

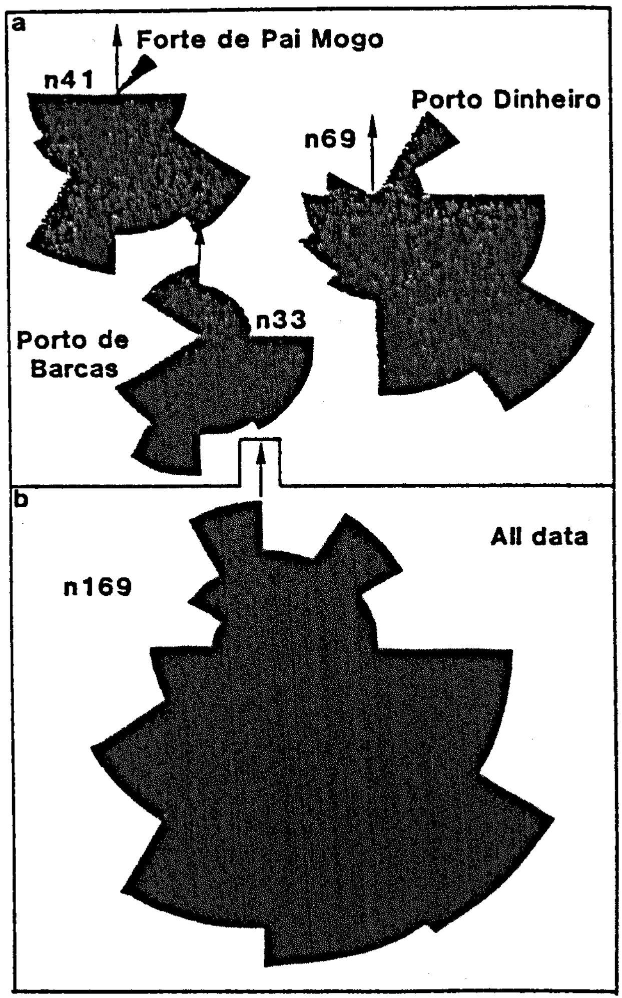


Figure 3.17 Palaeocurrent data for the Forte de Pai Mogo, Porto Dinheiro and Porto de Barcos sections (Porto Novo member): (a) **Individual,(b) Grouped data.** The data suggest a sinuous system with a strong south to south easterly trend. The relatively low incidence of measurements between 210-030° suggests that the system was only moderately sinuous.

Even taking account of the above problems with their identification, abandonment fills are not that common in the outcrops examined. This suggests that avulsion did not occur with a high frequency. In the absence of clay plugs provided by abandoned channel courses, which are anticipated in the classical models of meandering fluvial systems (e.g. Allen 1965), channels were probably not restricted in their lateral migration. Presumably this was due either to the nature of the bank material or more likely (as these are often clay rich), external controls such as the gradients and rates of of basin subsidence .

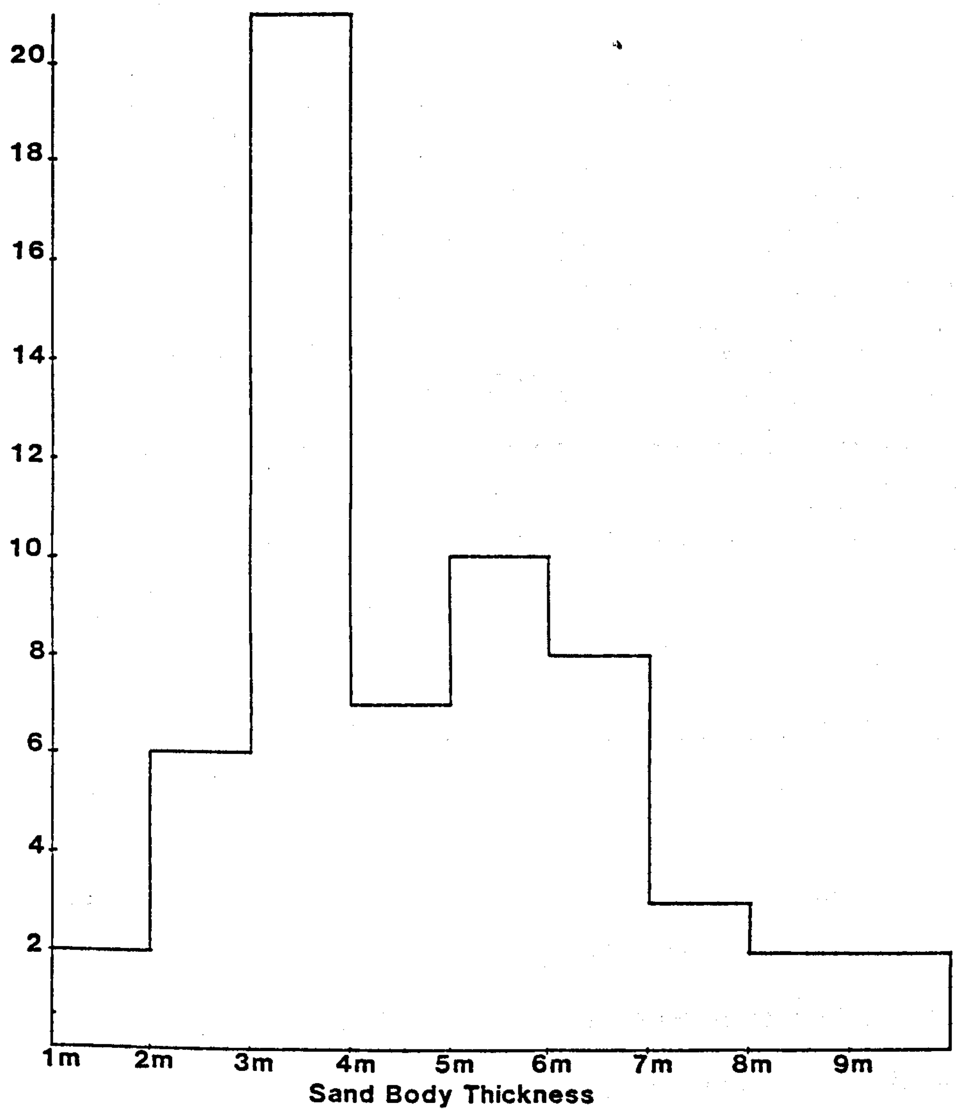


Figure 3.18 Histogram indicating the frequency of channel sand bodies of different thicknesses. Multi storey sand bodies excluded. The bi-modality probably indicates the presence of a trunk river with a number of tributaries.

Other unusual featues are largely associated with the existence of a clearly identifiable spectrum of channel 'types', these being defined by their scale and the nature of their fill. Figure3 .18 is a histogram of sand body thickness at intervals of 1m. The graph indicates that there is a degree of bimodality in sand body thickness and by inference palaeochannel depths. Such a pattern might be expected if the fluvial deposits exposed represented not just a single channel sweeping a floodplain but a major trunk river with a number of tributaries. The smaller scale sandbodies (<3m) are probably the deposits of tributaries, and the thicker units the trunk river. The channel 'types' are listed below:

1) Large-scale channel sandbodies, generally greater than 5m thick. often with well defined point bars, fining upward sequences and a heterolithic fill (e.g. the Porto Novo point bar Figure 3.19).

2) Large to medium scale, 3-7m thick, sand dominated, sheet - like sandbodies, point bar surfaces identifiable but of low angle and more difficult to discern (examples from south of Porto Dinhero).

3) Medium scale, fine-sand dominated sandbodies, often with identifiable point bar surfaces and the unusual prevalence of parallel and/or ripple laminations. Sediments may be rather reddened (e.g. from north of the Forte de Pai Mogo).

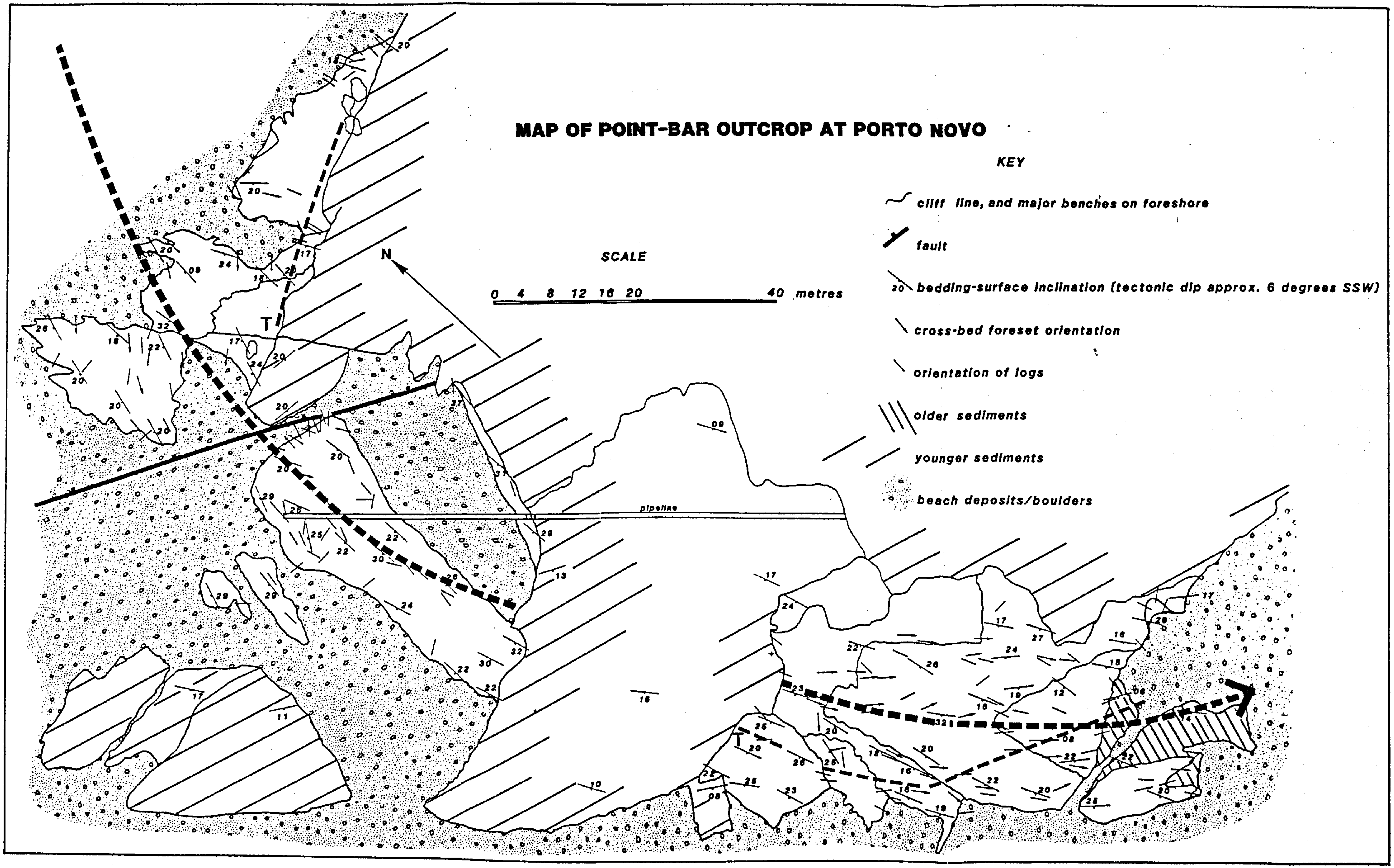
4) Mud dominated channels (e.g. from the Porto Novo section).

In addition to the above, section 3.4.5 will describe a channel abandonment fill, found to the north of Porto Dinheiro.

3.4.1 Large-scale sandbodies with a heterolithic fill; the PortoNovo Point-Bar

Despite the superb coastal exposure in cliffs up to 70m high, much outcrop is still rather two dimensional. Exceptions are found on headlands that separate the small bays and long narrow beaches which characterise most of the coastline. One of the most notable examples is a large channel sandbody some 6m thick which forms the headland immediately north of Porto Novo.

Figure 3.19 (over) Map of the 'Point Bar' outcrop at Porto Novo with superimposed palaeocurrent data. The outlined areas are the major wave-cut benches. Large arrow marks the radius of curvature of the palaeomeander, drawn parallel to the palaeocurrent indicators and strike symbols. Note the variable dips, exaggerated by 6-10° by tectonic dip and the paths of the graphic logs illustrated in Figures 3.23 and 3.24.


MAP OF POINT-BAR OUTCROP AT PORTO NOVO
KEY
cliff line, and major benches on foreshore
fault
bedding-surface inclination (tectonic dip approx. 6 degrees SSW)
cross-bed foreset orientation
orientation of logs
older sediments
younger sediments
beach deposits/boulders
SCALE
0 4 8 12 16 20 40 metres
N
pipeline
T

The outcrop consists of a series of seaward dipping, wave-cut benches which are successively younger point-bar surfaces (Figure 3.19). The shape of the benches was sketched in using compass and, where appropriate, measured distance. Approximately 150 palaeocurrent measurements are plotted on the map, two thirds of these being orientations of fossil logs, the remainder being cross bed orientations. Flume studies on the preferred orientation of logs and wood fragments by Macdonald and Jefferson (1985), indicated that these have a strong tendency to be orientated parallel to palaeoflow unless rotated and rolled when they will be orientated across the palaeoflow direction. Dip measurements indicate that palaeocurrents are approximately at right angles to the inclination of the point bar surfaces, with a slight tendency to be inclined obliquely up-face on the downstream part of the outcrop. It is also possible to trace an approximate arc through the mean palaeocurrent trends, thus obtaining an estimate for the radius of the palaeomeander. In this case the radius is approximately 160m and suggests a meander wavelength of over 300m.

The dips shown on the map are somewhat oversteepened, as the section dips at an increasing angle to the southwest. At its maximum this structural dip is about 6-10°. The increasing dip may be associated with a fault which passes down the centre of the bay at Porto Novo and is responsible for the gorge in the flanks of the Vimeiro structure, through which the River Alcabrichel flows (Figure 1.5). Dip measurements do show considerable variation over short distances however, and this is attributed to a combination of terracing and slumps on the face of the point bar, plus differential compaction of the heterolithic channel fill.

Measurements made at outcrop suggest that the channel was at least 5m and up to 7m deep (over much of the area the deposits of a younger channel lie directly above) and point bar surfaces extended for about 75m from the top of the channel to the base. Using the simple estimate of Allen (1965b) (point bar width = 2/3 channel width) suggests that the channel was approximately 110m wide. Figure 3.20 shows the southern part of the outcrop.

Figure 3.20 Southern part of the 'Point Bar' outcrop. The erosive base of the channel is arrowed. The upper boundary, in the background, is the erosive base of a younger sand body. Between these two erosion surfaces are a series of inclined lenses of sediment dipping from right to left. These are interpreted as major laterally accreted point bar units. The prominent stack of lenticular sand beds at centre right is ca. 5m thick. Camera facing NW.

The innermost cut bank, which is terraced, is exposed in the cliff face to the north of the headland. Parallel laminated medium and fine sands are the dominant structures developed. These either onlap, or drape successive intraclast-lined erosion surfaces (Figure 3.21). Sediments near the covered SW margin of the channel consist of parallel laminated dark grey silts and cross-bedded, buff, fine sands. Invertebrate burrows are common, and in the sands may obliterate all depositional structures (Figure 3.22).

A number of graphic logs were constructed, two of which are illustrated in Figures 3.23 and 3.24 . The lines of section are plotted on the map. The log of the southern part of the outcrop does show a general fining upward trend. This is not so apparent in the northern log, which has a notably heterolithic fill in its lower part

Figure 3.21 (top) Northern margin of the 'Point Bar' outcrop. The laterally migrated erosive margin of the channel is apparent from the two intraclast lined erosion surfaces which cut an older sand body. The tops of both units are in erosive contact with a younger sand body. Hammer 0.4m long.

Figure 3.22 Interbedded sands and dark organic rich silts pervaded by invertebrate burrows. Close to the SW margin of the 'Point Bar' outcrop. Scale bars = 0.1m.

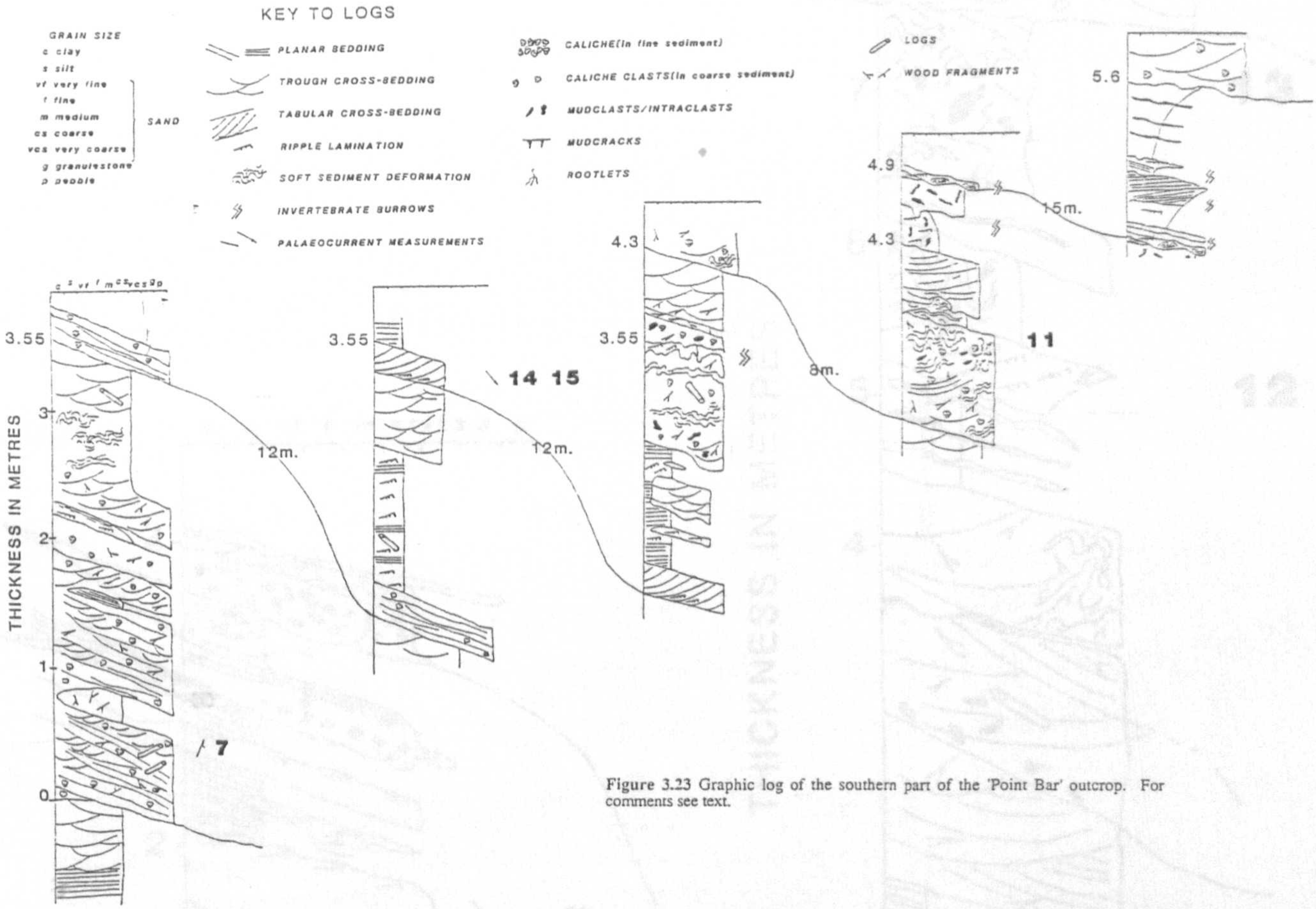
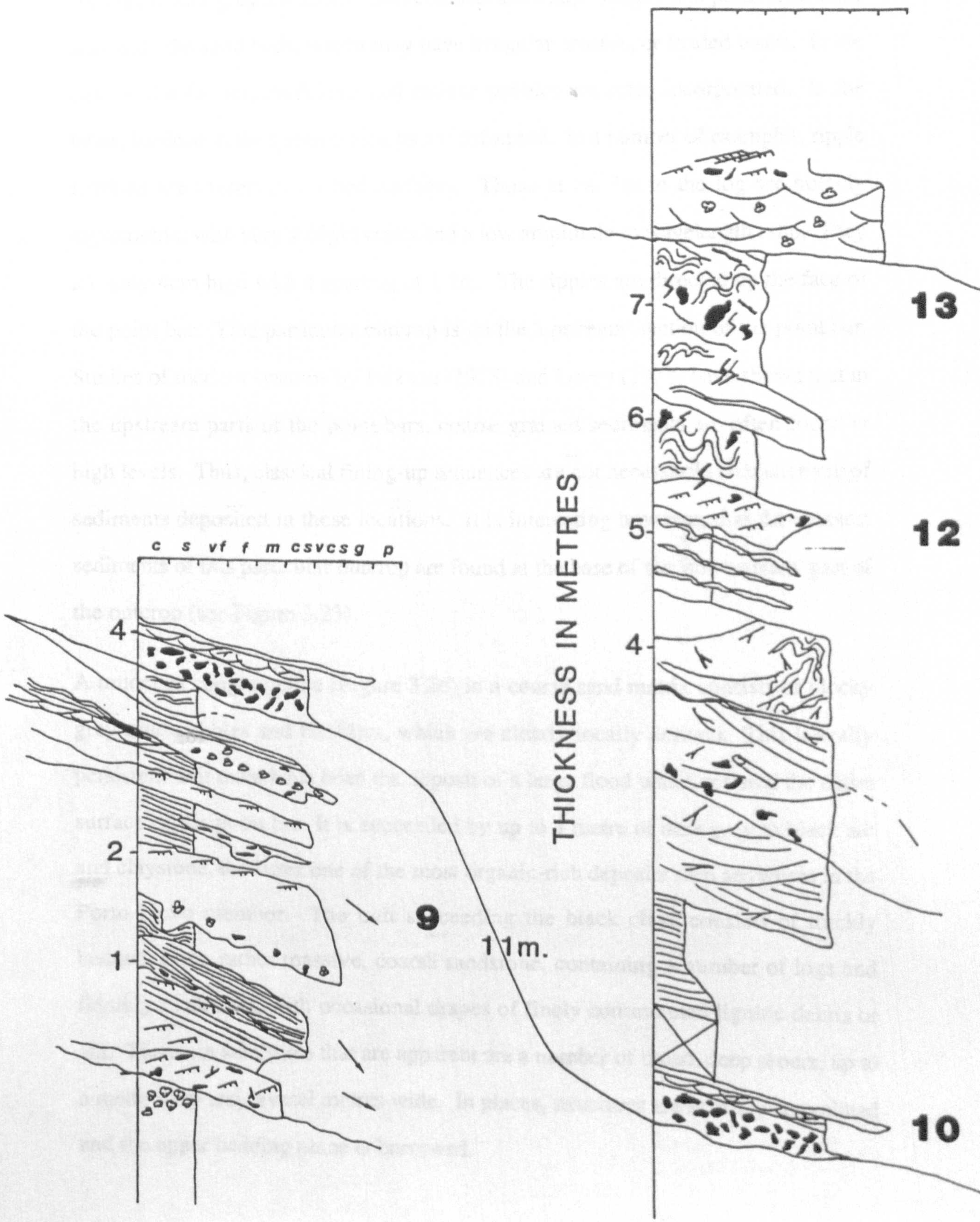


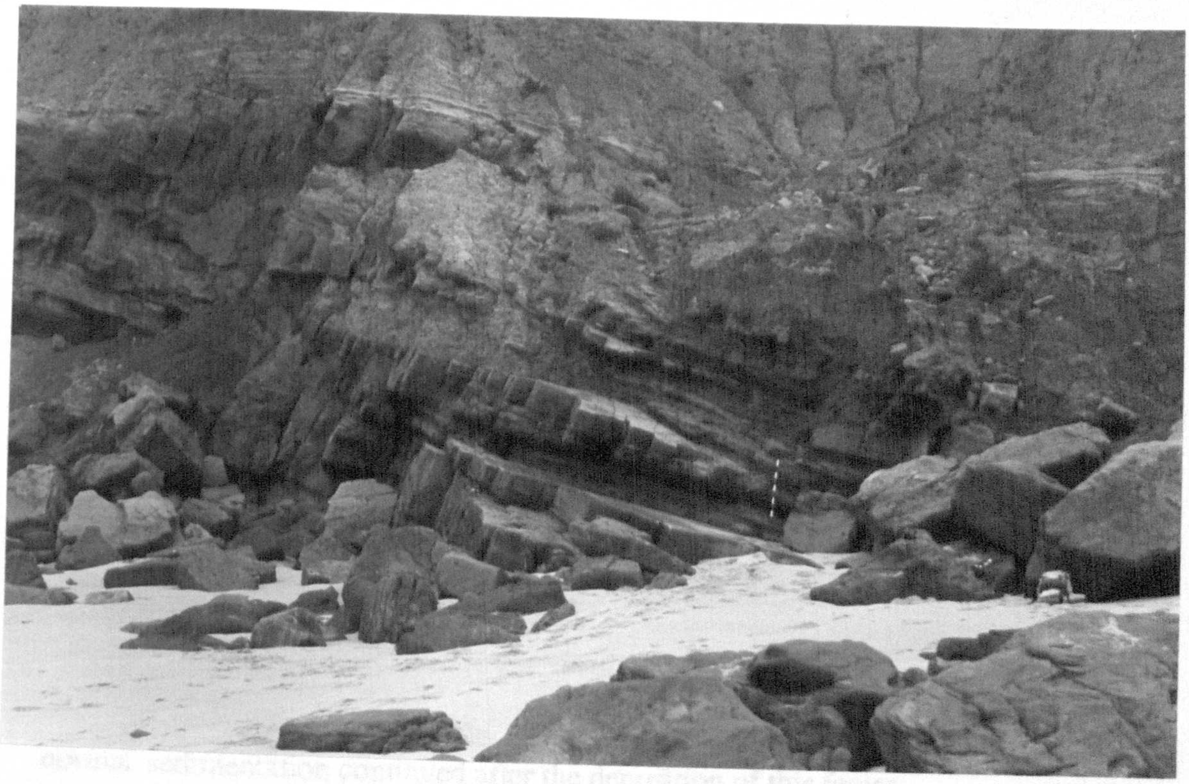
Figure 3.23 Graphic log of the southern part of the 'Point Bar' outcrop. For comments see text.

Figure 3.24 Graphic log of the northern part of the 'Point Bar' outcrop. For comments see text.



(Figure 3.25), dominated by lenticular, graded, medium and coarse sand bodies, up to 0.5m thick and with several metres lateral extent. These are enveloped within laminated dark grey siltstones. Bed contacts are sharp. Ripple and parallel laminae dominate the sand beds, which may have irregular erosive, or loaded bases. In the case of the former, mudclasts and caliche pebbles are often incorporated. In the latter, laminae in the enveloping silts are deformed. In a number of examples, ripple laminae are preserved on bed surfaces. Those at ca. 3m in the log are notably asymmetric, with very straight crests and a low amplitude to wavelength ratio. They are only 4cm high with a spacing of 1.2m. The ripples are directed up the face of the point bar. This particular outcrop is on the 'upstream' section of the point bar. Studies of modern systems by Jackson (1978) and Levey (1978) have shown that in the upstream parts of the point bars, coarse grained sediments are often found at high levels. Thus, classical fining-up sequences are not necessarily characteristic of sediments deposited in these locations. It is interesting however, that the coarsest sediments of this particular outcrop are found at the base of the 'downstream' part of the outcrop (see Figure 3.23).

A mudclast conglomerate (Figure 3.26) in a coarse sand matrix consists of blocky grey mud cobbles and boulders, which are clearly locally derived. This laterally persistent unit must have been the deposit of a large flood which scoured the entire surface of the point bar. It is succeeded by up to a metre of dark grey to black silt and claystone, the latter one of the most organic-rich deposits seen anywhere in the Porto Novo member. The unit succeeding the black clays consists of thickly bedded, often rather massive, coarse sandstone, containing a number of logs and floating mudclasts, with occasional drapes of finely comminuted lignitic debris or silt. The main structures that are apparent are a number of broad, deep scours, up to a metre deep and several metres wide. In places, structures are severely convoluted and the upper bedding plane is burrowed.



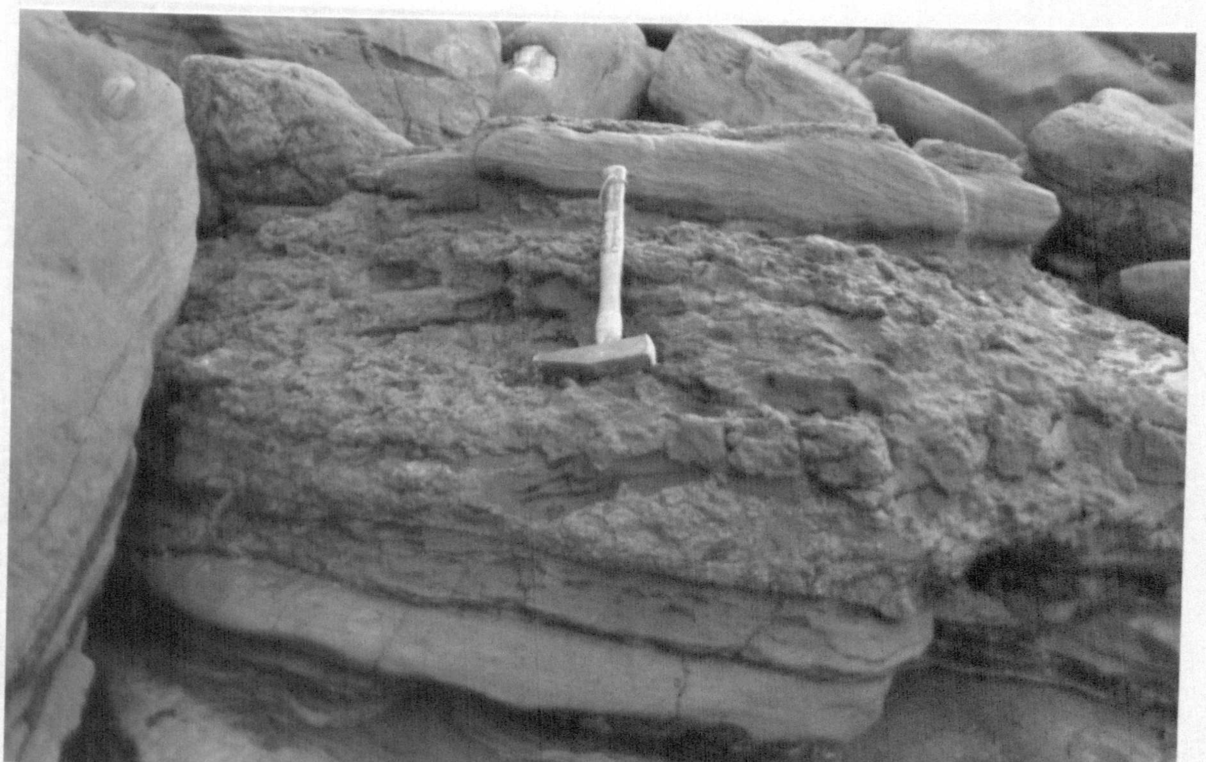


Figure 3.25 (top) Northern part of the 'Point Bar' outcrop. Note the heterolithic nature of the channel fill. This feature suggests that the palaeochannel was subject to significant discharge fluctuations and also carried a large suspended sediment load. Scale bar = 1m.

Figure 3.26 Blocky mudclast conglomerate with a coarse sand matrix. Northern part of 'Point Bar' outcrop. The mudclasts were clearly not transported far as if this was the case they would have disaggregated (Smith 1972). Hammer for scale 0.4m long. The thickness of this unit suggests that it was the result of erosion and deposition during a single, large magnitude flood.

Below the erosive base of an overlying channel sand body, the uppermost deposits of this channel consist of severely disrupted fine sands and silts. Large blocks of mudrock up to 0.5m diameter (Figure 3.27) are enclosed within convoluted silts with sandy stringers. A number of the blocks exhibit climbing ripple laminae. In the midst of this chaotic deformation are a few completely undisturbed, lenticular beds of ripple laminated and rather undulose, parallel laminated fine sands. The chaotically deformed beds pass down dip into undeformed laminated silt and fine sands. Laterally, it can be demonstrated that younger heterolithic sediments, like those described above, were deposited on the lower face of the point bar at least. This indicates that the disturbed sediments were not part of an abandonment fill as 'normal' sedimentation continued after the deposition of this facies.

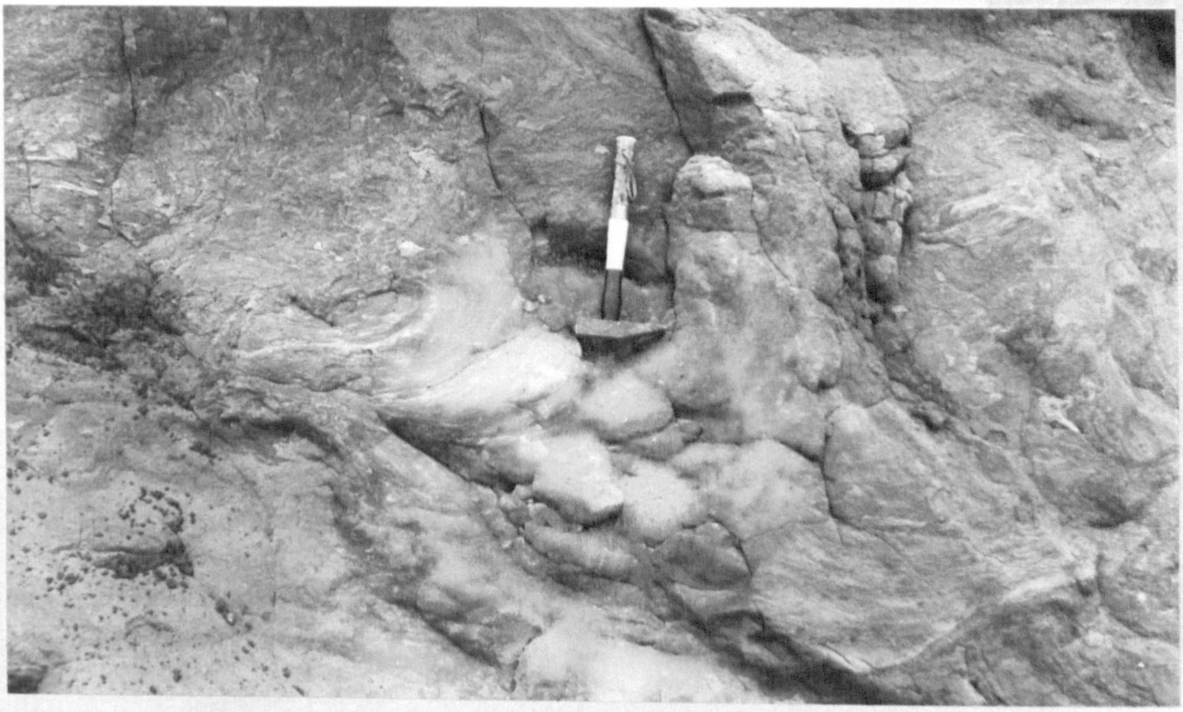


Figure 3.27 Chaotically deformed silts and fine sands in the upper part of the 'Point Bar' sand body. Discrete blocks of sediment up to 0.5m long are apparent at lower right and to the left of the hammer head. Scale bars on hammer = 0.1m. It is possible that such severe deformation was due to seismicity or slumping of the upper bar surfaces at low stages.

Deformed sediment is also a feature of the exposed benches on the southern side of the headland, particularly in the mid point-bar region (Figure 3.23). Over large areas, coarse sands containing lignitic debris, logs and caliche pebbles are contorted, with few apparent sedimentary structures and highly irregular bedding surfaces. Silty drapes pick out the deformation clearly. A particularly spectacular feature, in the mid-point bar region (at 3.5m on the log), is an exhumed, perfectly preserved series of straight crested sandwaves (Figure 3.28). Linguoid ripples mantle these bedforms which seem analogous to the 'transverse bars' recorded by Levey (1978) in the Congaree River, though they are the scale of the smaller megaripples described by the same author. On the surface of the bed, beneath the foremost of the bars, is a series of straight crested ripples. Their foresets dip at an angle opposed and highly oblique, to that of the bar foresets (Figure 3.29).

Figure 3.28 Series of straight crested megaripples on the southern part of the 'Point Bar' outcrop. The bedforms are advancing towards the camera; hammer (0.4m long) is resting against the foresets of the foremost of four bars with a spacing of 3-6m. The crests of the succeeding bedforms are the prominent breaks running across the field of view. The bedforms are clearly orientated approximately at right angles to the dip of the point bar surfaces clearly visible in the background. The surfaces of these bedforms are mantled by poorly preserved linguoid ripples.

In the uppermost sediments of the point bar a strong cyclicity is observed. Pyritic, light grey, medium to very fine sandstones oscillate with dark grey to black silts, on a scale of 2-20cm. Ripple and parallel laminae, often slightly deformed, are the dominant structures. Simple burrows and occasional sand-filled mudcracks are also present. Grading is often observed, both normal and inverse, in the dominantly parallel laminated silts and ripple laminated sands. The dominant presence of inverse grading in modern, fine grained fluvial systems has been documented by Taylor and Woodyer (1978).

Figure 3.29 Cross sectional view of one of the megaripples illustrated in Figure 3.28. The bedding plane beneath is covered with straight crested current ripples which are oriented highly obliquely to the megaripples and up-dip. These may have been due to wind driven currents at low stages when the bed surface was only shallowly submerged.

A relatively rare feature within a thick cross-bedded sand near the top of the sand body is a fossil tree root which penetrates up to a metre of sediment. The top of the root is covered, indicating that the tree had been growing during the active accretion of the point bar.

3.4.2 Sand dominated channels

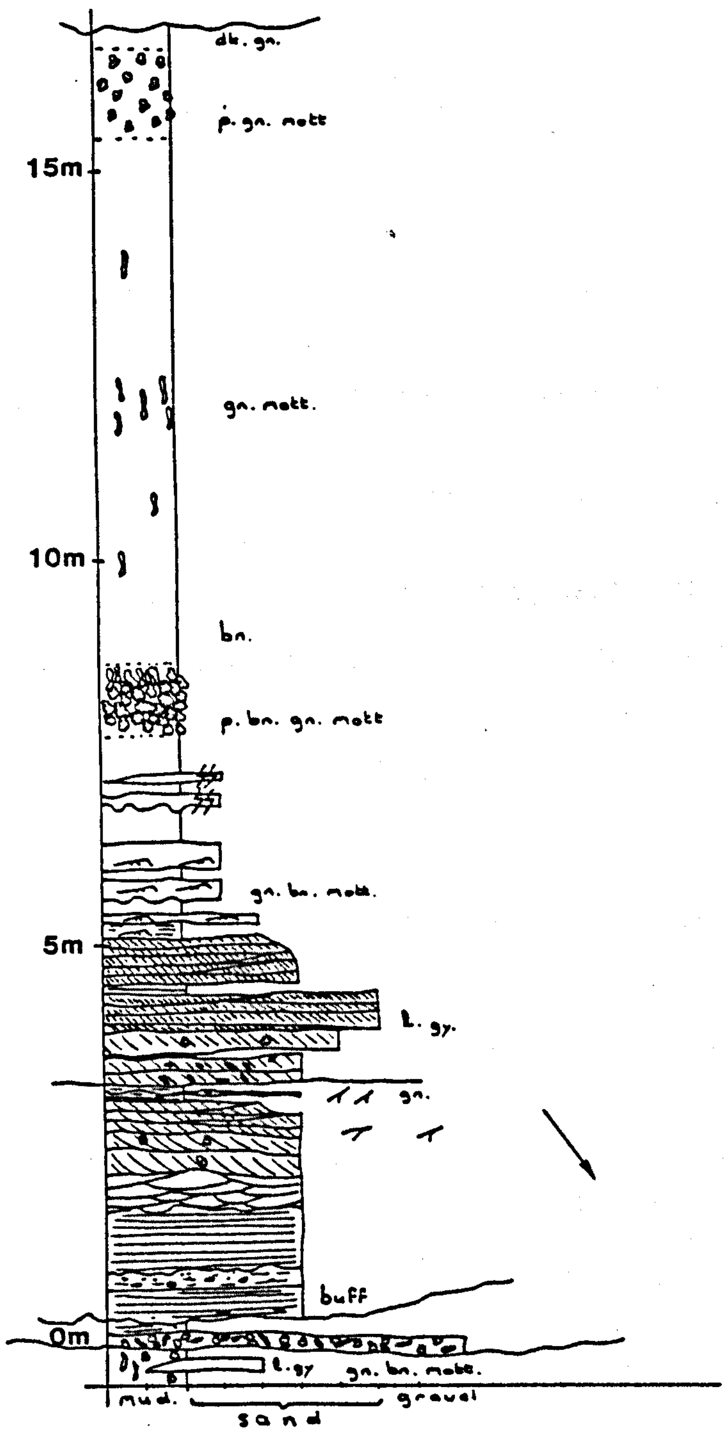


Figure 3.30 Graphic log of a sand-dominated channel deposit. For details see text. GR 7035 4020 south of Porto Dinheiro.

Figure 3.30 shows a single log through a channel which is exposed to the south of Porto Dinheiro. It is largely sand dominated and is a good example of the vertical succession most commonly encountered within the majority of channel sand bodies. Points to note are:

1) the lag at the base of the channel, with a parallel laminated silt drape;

2) the otherwise generally low proporton of mudrock within the channel;

3) the general lack of an obvious fining-up sequence except in the top metre;

4) excluding the lag, the coarsest sands are found at a high level within the channel;

5) cross-beds that show some decrease in size upwards dominate the sand body;

6) levee deposits consist of a few erosive based, laterally impersistent lenses of ripple laminated or burrowed, very fine sandstone, enveloped in structureless silts;

7) the channel is overlain by about 10m of mudrock, much of which shows evidence of pedogenic processes;

The low frequency and thin nature of drapes makes identification of point bar surfaces much more difficult. These usually take the form of major low-angle erosion surfaces the dip angles of which are significantly lower than those of 'type 1' channels.

Figure 3.31 illustrates a chain of three logs through a sand body from a slightly higher level in the section to the south of Porto Dinheiro (Figure 1.5). These all exhibit a fining-up sequence punctuated by mud drapes of laterally variable thickness. Major erosion surfaces are inclined from left to right and, as these are approximately at right-angles to palaeocurrent indicators, are taken to be lateral accretion surfaces. These surfaces generally have a shallowly inclined, occasionally stepped profile. A variable suite of structures is present, the scale of which generally decreases upwards. Laterally, sediment grades vary considerably between packets of sediment bounded by the major erosion surfaces. There is very little heterolithic facies developed at the top of the sand body. Instead a relatively rapid transition between sand beds with well preserved structures and structureless massive mudrocks exhibiting pedogenic features is observed.

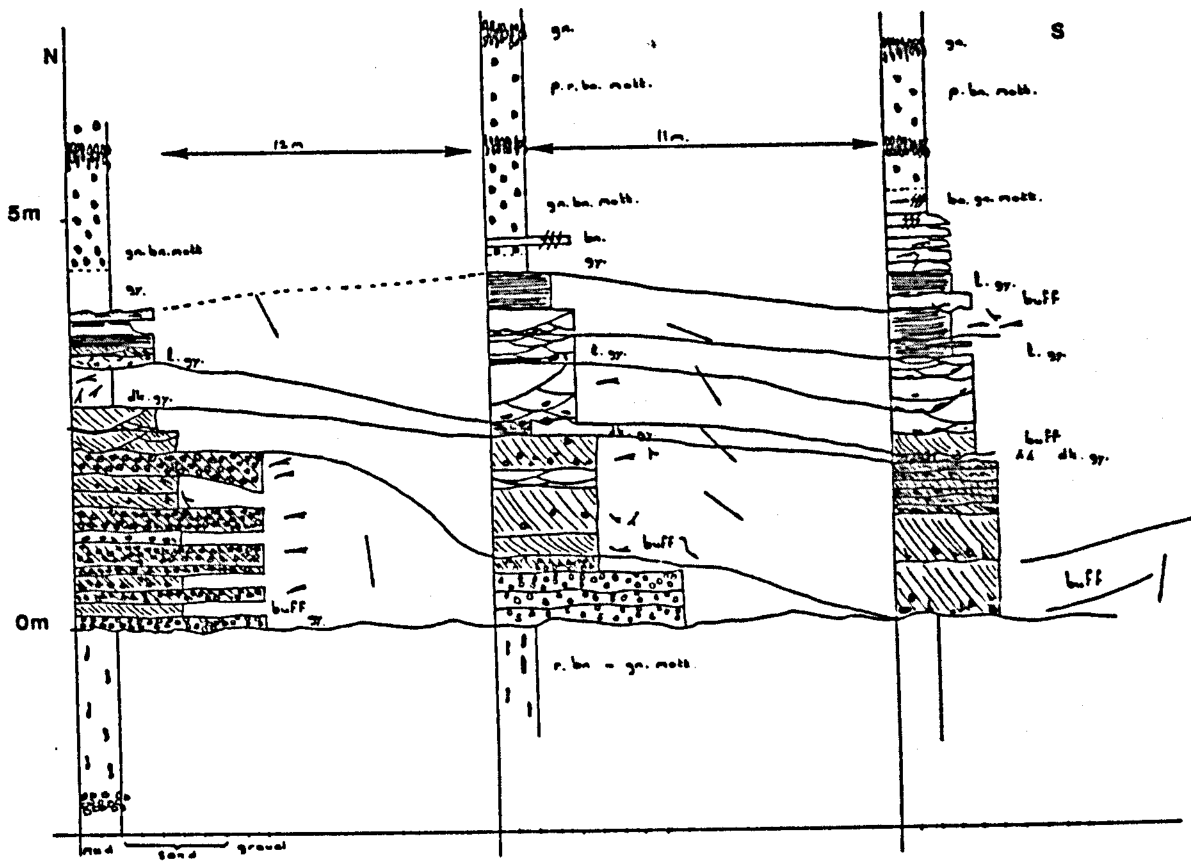


Figure 3.31 Chain of three logs through a sand-dominated channel deposit. For details see text.

3.4.3 Fine-sand channel fills, dominated by parallel and ripple laminations

Figure 3.32 shows a linked sequence of seven logs through a channel sand body exposed to the north of the Forte de Pai Mogo. This sand body, also shown in Figure 3.1c, has a high proportion of laminated fine sands in its fill. This characteristic is found in channel sand bodies along all the coastal sections, including that to the north of Foz do Arelho.

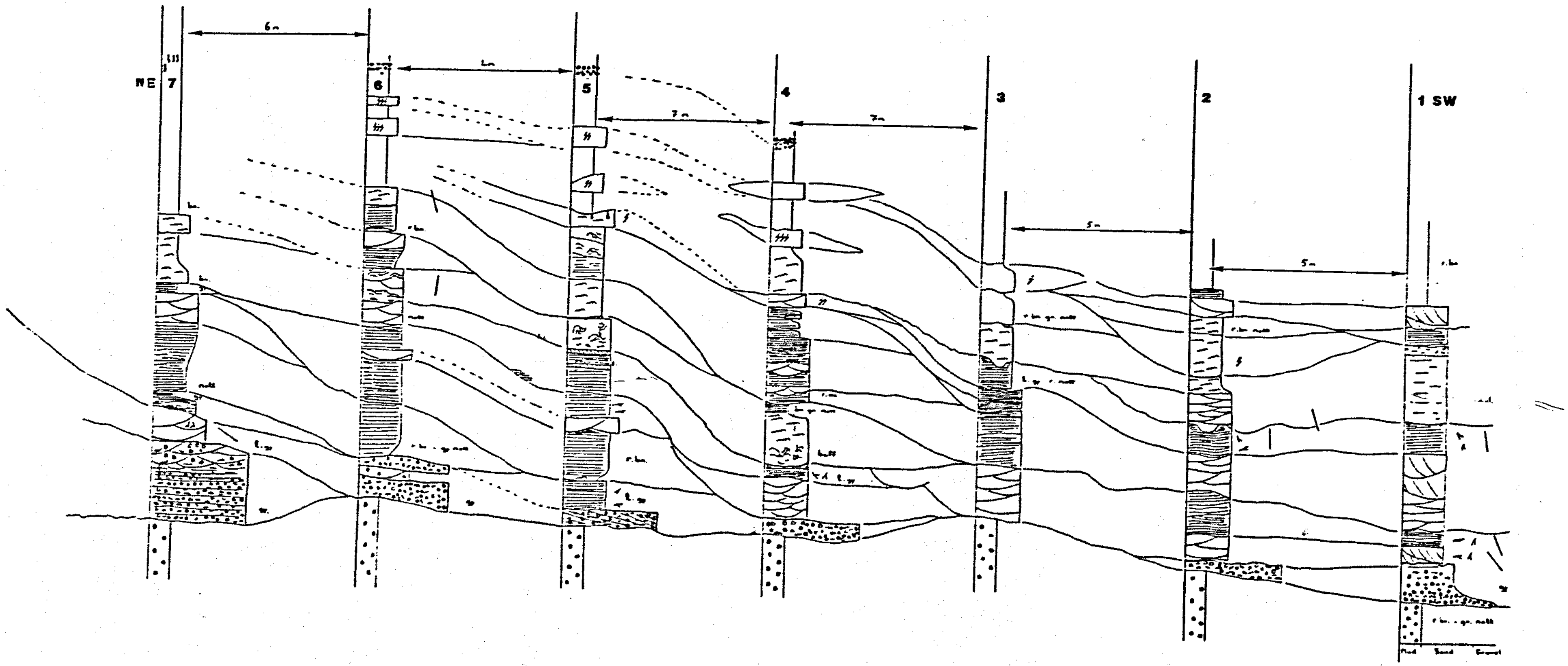


Figure 3.32 Chain of seven graphic logs through a sandy channel fill dominated by planar bedding. For details see text.

Where the base of the channel contains caliche and mud pebble conglomerates, it also tends to be more deeply eroded into the underlying mudrocks. This produces a rather scalloped basal erosion surface with significant relief. Figure 3.33 illustrates this feature. Sediments in the lowest parts of the channel are rather dull greys, but in the upper 2/3 are mottled red-brown.

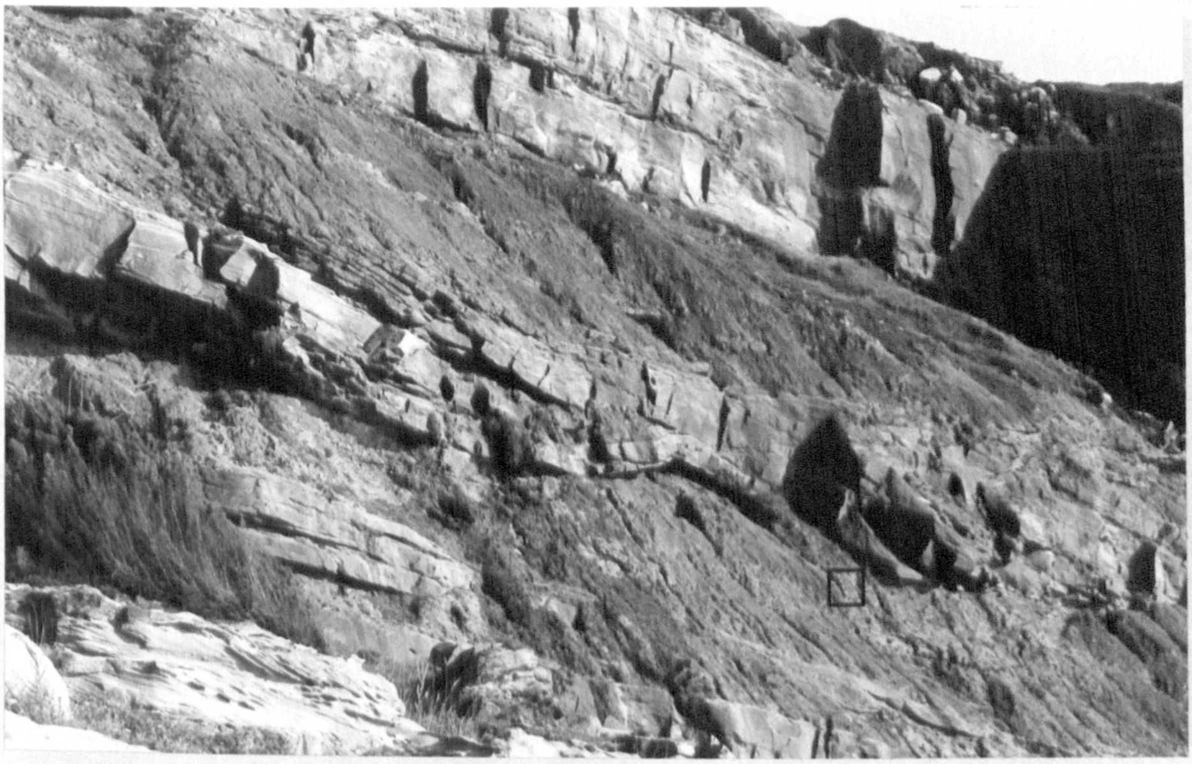


Figure 3.33 Scalloped basal erosion surface to laterally accreted sand body which is dominated by laminated sediment. Such features suggest a highly episodic discharge, with short lived, high velocity flows. The channel migrated from left to right. Metre pole for scale (circled). GR 8170 6760 2km north of Foz do Arelho.

The sand body depicted in Figure 3.32 is 3.25-4.4m thick. Point bar surfaces generally dip at low angles, averaging 4-7° over the 50m length of a single point bar surface. The erosion surfaces are often deeply incised into, and cross cut older units. They are often terraced and extremes of dip from 0-30° appear, though normally dips are in the region of 5-12°. Structures, particularly parallel laminae, lap out onto the margins of the inclined erosion surfaces. Silt drapes are uncommon but do appear and in places these are desiccated.

Figure 3.34 is located on Log 6. Parallel laminae in fine sands pass laterally into ripple laminae, indicating a decelerating flow. This is probably due to friction in the shallows at the channel margin. In slightly coarser medium sands, parallel laminae often pass vertically into cross beds. In the upper parts of the channel are a number of shallow scours with cross bedded or ripple laminated fills. These are probably the deposits of small chute channels, developed during flood stages.

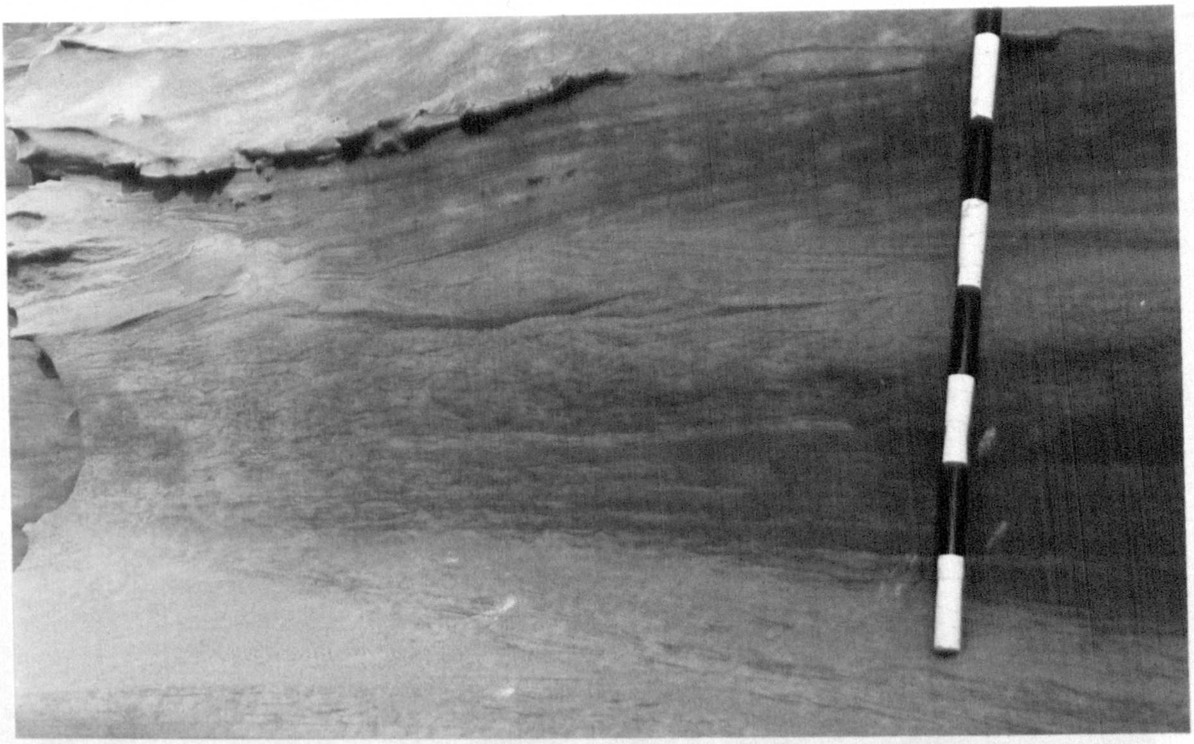


Figure 3.34 Fine sandstones exhibiting planar bedding passing laterally into ripple laminations from right to left. Note how both these, and the erosive based packet of parallel laminae in the upper part of the figure, lap out onto the margins of scours. The transition from parallel to ripple laminations represent a flow regime transition from upper to lower regime (Harms et al. 1982). This may have been due to friction in the shallows close to the channels inner bank. Scale bars = 0.1m.

3.4.4 Channels with a dominantly muddy fill

Figure 3.35 shows the margin of a small channel near the cliff top north of Porto Novo which was about 4m deep. Its fill is almost entirely muddy sediment. This is not thought to be an abandonment fill as it has no associated coarse fill at all. Also, the preserved channel margin is thought to be the inner bank; i.e., the point bar surface. This surface dips steeply, at up to 26°, but this is within the limits outlined by Taylor and Woodyer (1978) from modern suspended load streams in eastern Australia. Had the dips been in excess of 30°, much disturbance due to slumping would be expected. The point bar surfaces extend for some 17m from the top of the channel to its base. This gives a mean slope of 18-19°.

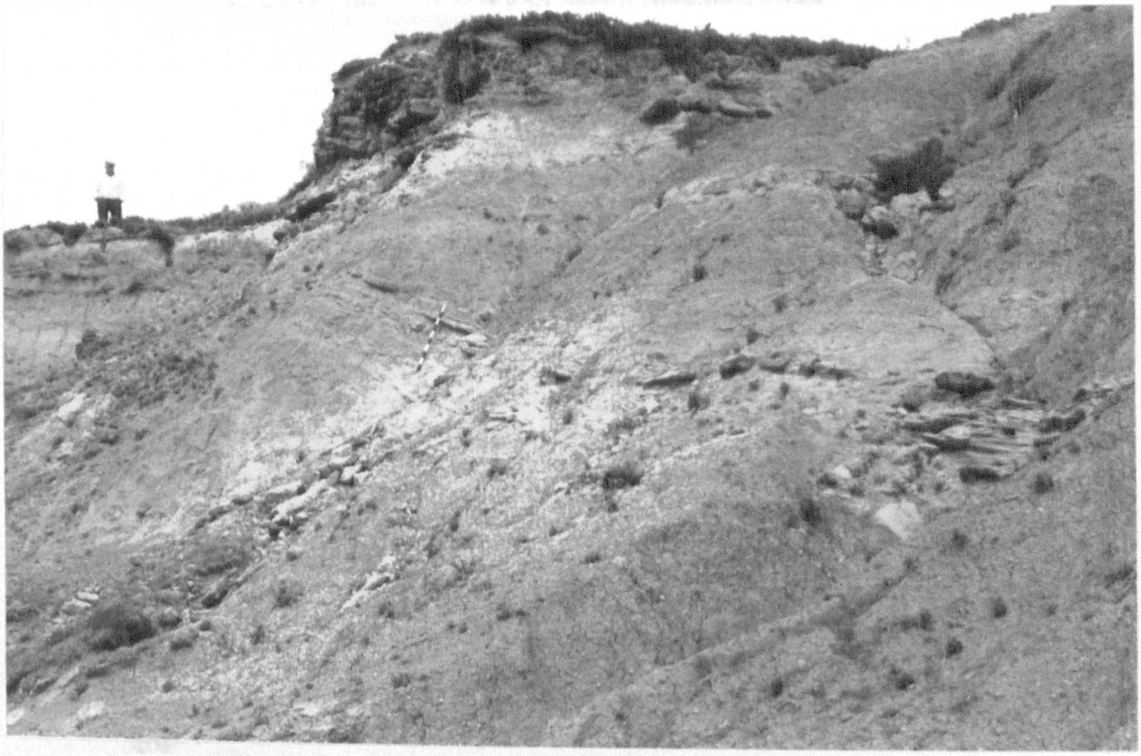


Figure 3.35 The inclined margin of a dominantly mud-filled channel. Channel base extends from foot of 1m scale, to the base of the packet of sand beds at centre right. This is thought to be the inner bank of a laterally accreted sediment body. The upper parts of the channel fill are colour mottled and exhibit desiccation cracks. Location: cliff top 900m north of Porto Novo.

The basal 0.6m of the fill consists of parallel and occasionally ripple laminated, light grey very fine sandstones with numerous irregular scoured surfaces. There are also rare 0.5-2cm planar-based beds of ripple laminated, fine buff sandstone and isolated 'balls' of fine sand up to 0.1m thick and 0.4m wide. These have undulose upper surfaces and very irregular, deformed bases which also disrupt the underlying sands. Higher up the point bar surface, the sands are graded, passing into silts. They are also burrowed in places. Mudrocks are cut by sand-filled desiccation cracks.

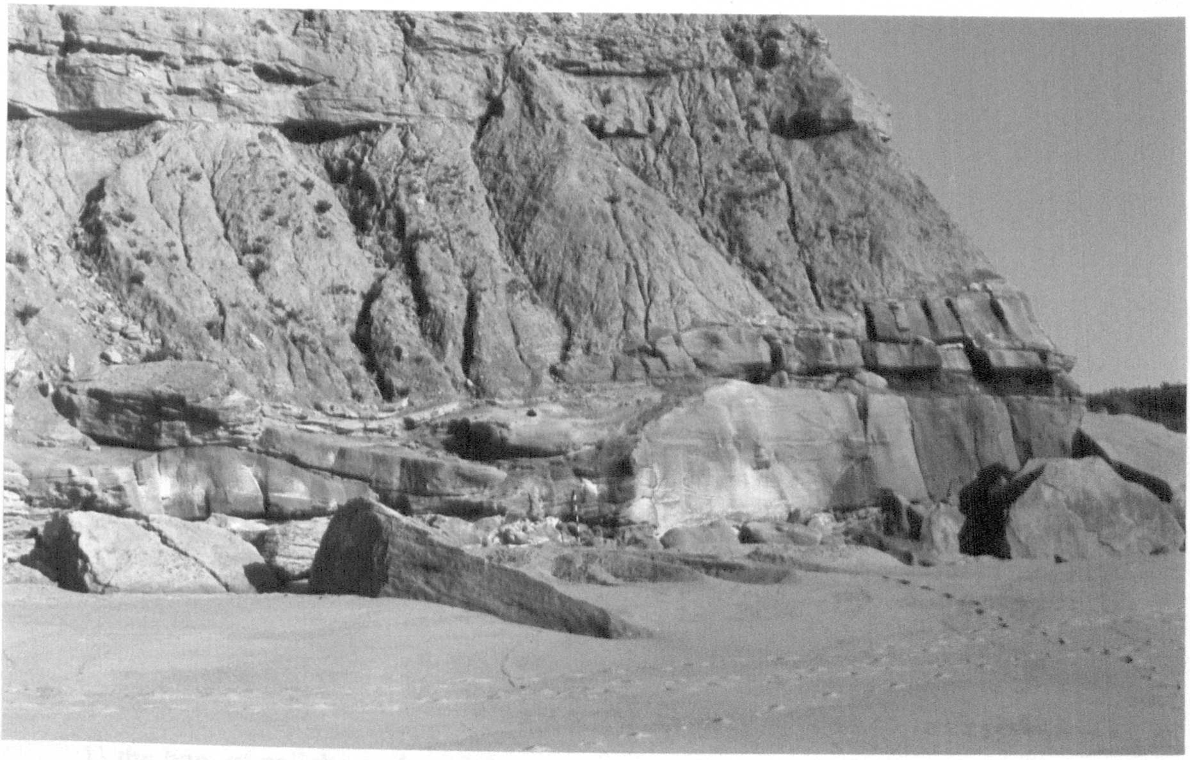
The remainder of the channel fill consists of faintly laminated to massive, dark green silts with rare lenses of very fine ripple laminated, light grey sandstone, up to 5cm thick. Silts in the upper 2m of the channel are mottled and contain 2-3mm diameter carbonate nodules which are still soft and easily disaggregated.

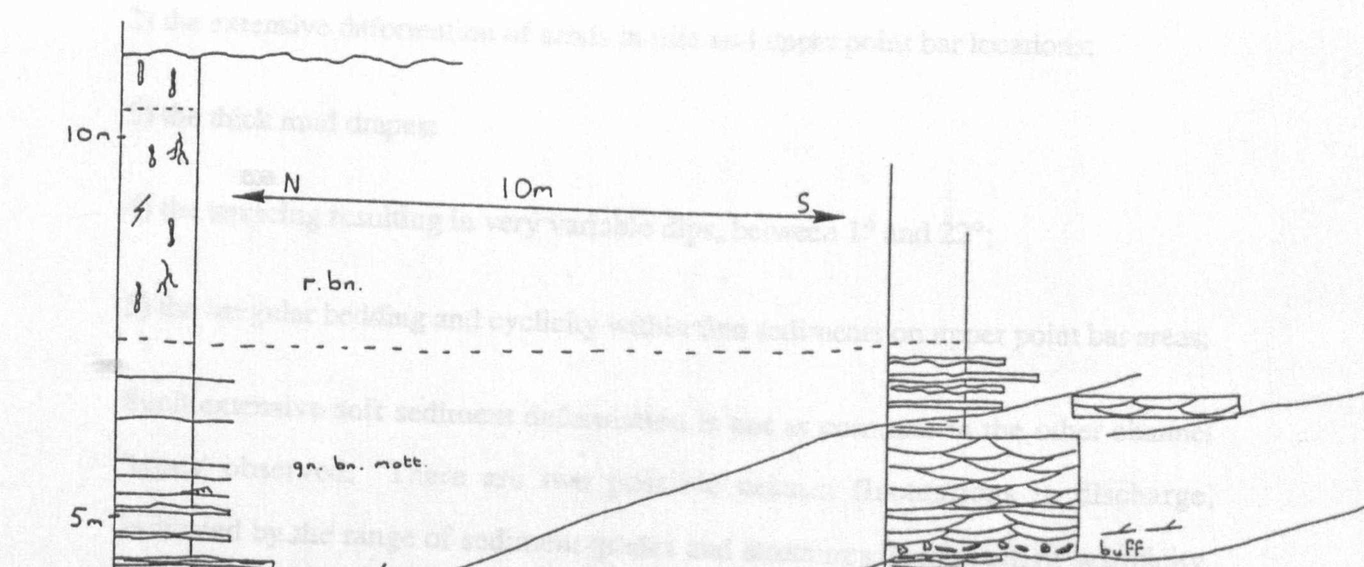
Most examples of this channel 'type' appear in the lower 15m of the type section at Porto Novo. However, a further example crops out to the south of São Bernadino (GR 7060 5020). The coarse facies at the base of this channel are dominated by caliche and mudclast conglomerates. Generally, channels of this type appear to have a limited lateral extent, suggesting that they were active for short periods of time and were abandoned rapidly.

3.4.5 Channel abandonment facies

Figure 3.36 shows a log and a photograph of a channel abandonment fill. The right hand log shows the normal, sand-dominated channel fill. This is abruptly truncated and replaced laterally by a horizontal fill, dominated by mudrocks and decimetre to centimetre bedded fine sands. The thicknesses and preservation of primary structures in the sands decreases upwards. Mudrocks are initially dull greys and become red mottled towards the top of this abandonment fill. Ten metres to the north of the left hand log, the entire channel fill consists of mudrocks and thin sands. The outer cut bank is obscured by modern debris.

Figure 3.36 (over) Photograph and graphic logs of a channel abandonment fill. Sand body is truncated by an erosion surface, above which is a bioturbated dominantly fine grained fill. Rootlets and carbonate rhizocretions appear in the overlying silts. Scale bar in photograph = 1m.





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buff
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buff

3.5 CONCLUSIONS

3.5.1 Interpretation of facies associations

This section will first briefly summarise the various aspects of the different channel fills, and will then discuss the likely controls on their development.

Large-scale channel sandbodies

Figure 3.37 is an idealised interpretation of the facies distributions observed in the Porto Novo 'Point Bar', based largely on the better exposed southern part of the outcrop. Points to note are:

1) the bars of caliche and mudclast conglomerate with low angle foresets flanked by sinuous crested dunes of the same lithology and passing laterally into straight crested megaripples in the mid point bar region;

2) the extensive deformation of sands in mid and upper point bar locations;

3) the thick mud drapes;

4) the terracing resulting in very variable dips, between 1° and 22°;

5) the irregular bedding and cyclicity within fine sediments on upper point bar areas;

Such extensive soft sediment deformation is not as common in the other channel 'types' observed. There are two possible causes; fluctuations in discharge, indicated by the range of sediment grades and structures developed, or seismicity. Terracing on the face of the point bar would have favoured instability on the higher slopes. In addition, the presence of thick mudrocks between sand beds would have provided planes of weakness to facilitate slumping. Fluctuations in discharge would result in excessive pore pressures in poorly consolidated sediments at falling stages. Seismicity is equally possible particularly as there is direct evidence for active

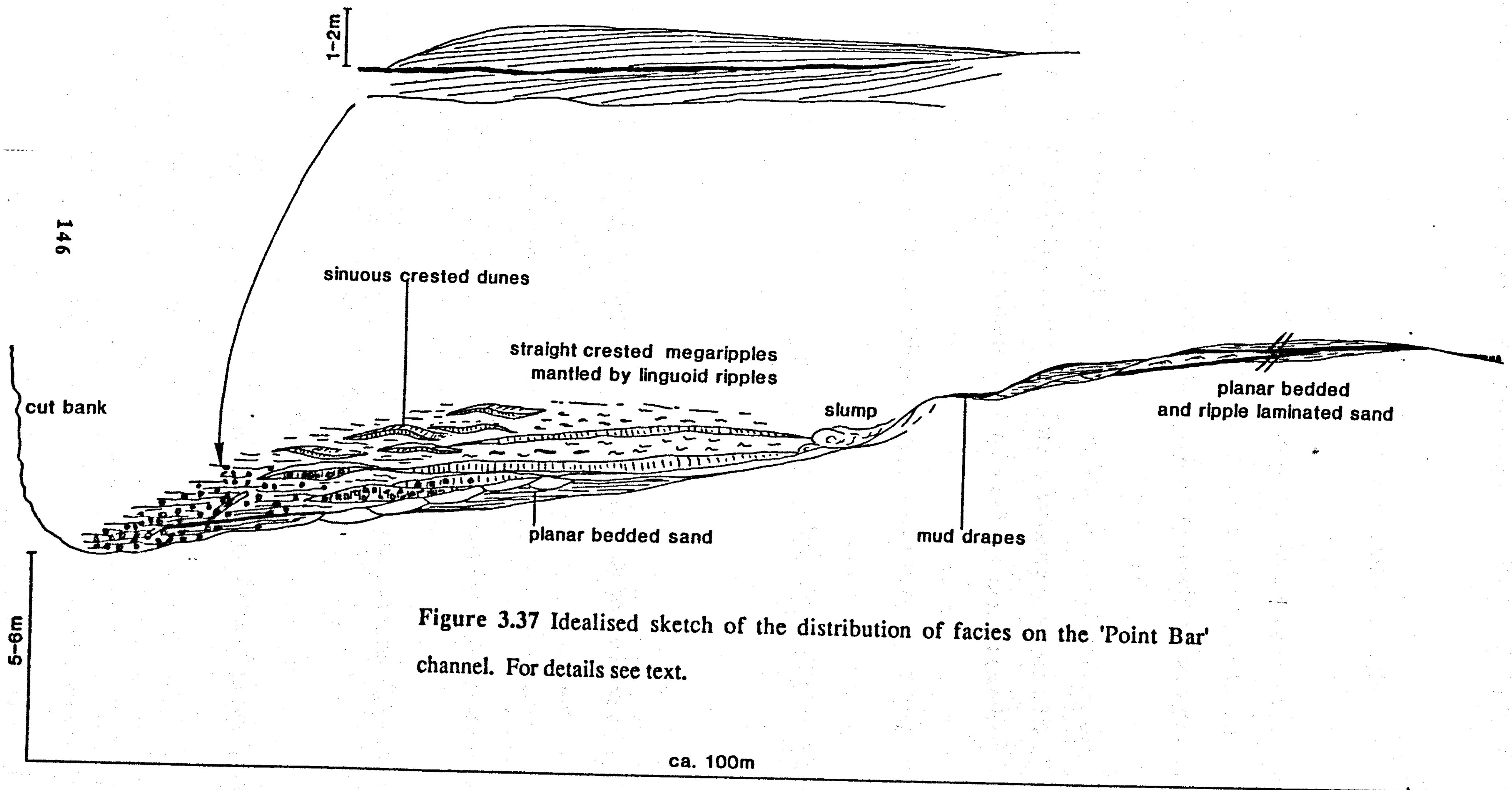


Figure 3.37 Idealised sketch of the distribution of facies on the 'Point Bar' channel. For details see text.

extensional faulting immediately before the incision of this particular channel (see section 3.5.2).

The presence of a heterolithic fill in this channel may, in addition to reflecting fluctuations in discharge, reflect the influence of tides on the river mouth (Smith 1987, 1988). In a review of work on the Orinoco delta, which is subject to a microtidal regime, Van Andel (1967) noted that the system was tidal over 150km from the river mouth. In the case of the Lusitanian Basin it is likely that an open marine basin was less than 80km to the south of Porto Novo (see Chapter 4). It is possible, therefore, that tides exerted some influence. The presence of small scale cyclicity in fines near the top of the sand body may be evidence of a mild tidal influence. However, when making comparisons it must be borne in mind that the basin and its fluvial systems were probably an order of magnitude smaller than the catchment of the present day Orinoco River which is some 950 000 km^2. An examination of the sections to the north of Peniche which are further removed from the palaeoshoreline, indicates that thick mud drapes and heterolithic fills are significantly less common than to the south.

Sand dominated channels

These are the most prevalent 'type' in the sections. Clearly, sandy bedload was a major component of total sediment load, though high suspended loads probably also existed because mud drapes do occur and associated overbank sequences remain substantial. Generally lower gradients on point bar surfaces would be expected in dominantly bedload steams (McGowen and Garner 1970; Allen 1970), largely due to the much lower cohesive strength compared to that of mudrocks. This is certainly the case with these sand bodies. In addition, lateral accretion surfaces in sand dominated channel deposits are not neccessarily visible as noted by Jackson (1978). Channel fills still exhibit evidence of fluctuating discharges, in the range of grain sizes displayed both vertically and laterally.

Channels dominated by parallel laminated fine sands

The channel fills dominated by fine parallel laminated sands, seem also to have been characterised by a widely fluctuating discharge, with high stream velocities during flood. The dominance of upper flow regime bedforms may be partly due to the finer grain sizes (Harms et al. 1982), but may also be characteristic of meandering streams which are not highly sinuous (Allen 1970). Such streams should allow higher velocity flows than those with tight meanders which tend to retard stream flows (Jackson 1975). Other examples of ancient meandering stream deposits dominated by parallel lamination are recorded by Nami and Leeder (1978), from the Jurassic of Yorkshire, and Turner (1986) from the Permo-Trias of the Karoo.

High sediment loads and rapidly decreasing discharges are suggested by the presence of thick units of climbing ripple laminae in other examples of this association. The occasional deeply scoured bases of channels, producing the scalloped basal erosion surfaces, must also be symptomatic of widely fluctating discharges. Prominent lateral accretion surfaces like those present in this section have been attributed to 'distinctly episodic deposition' by Puigdefabregas (1973) for Miocene deposits in the Ebro Basin.

Channels with a dominantly muddy fill

Channels dominated by a muddy fill once initiated, were rapidly abandoned and therefore accumulated a dominantly fine grained fill. This does not fit with the observation that avulsion does not appear to have been a frequently occurring process in this system. In addition it is strange that the vast majority of examples observed appear within a 15m-thick interval of one section. This incidentally, is also the interval in which the 'Point Bar' outcrop with the heterolithic facies occurs. It is possible that these reflect a rather unsteady base level, related to active tectonic movements. Section 3.5.1 deals with evidence for the tectonic environment prevaling during the deposition of the Porto Novo member and this will be discussed further in that section.

General

The channels which were characterised by short lived rapid flows, high rates of deposition and have an oxidised fill, contain very little organic material. It is possible that there was originally little organic material. These are the sort of characteristics that a channel fed intermittently by heavy rainfall in the comparatively arid basin margins might have. It is certainly true from studies inland, that this type of channel fill is uncommon in these more distal areas. Therefore, a basin margin source for these flashy high velocity stream deposits is a probability.

There is a lack of evidence for highly sinuous channels. Palaeocurrent data indicates a sinuous system with a strong SSE trend. The 'Point Bar' outcrop, though rather restricted, does not reveal a tight meander loop. Channels with a dominantly parallel laminated fill, are interpreted as not having tight bends, as these would have prevented the development of high current velocities. It is possible that flashy regimes themselves are characterised by moderate sinuosities. The lower incidence of abandonment fills, thought to be characteristic of highly sinuous fluvial systems, strongly suggests that such an environment did not exist. A direct consequence of this would be a lack of constraints to lateral movement, provided by clay plugs in the classical model (Allen 1965) and the development of laterally persistent sheet sands as observed.

Using the palaeochannel dimensions recorded from the 'Point Bar' outcrop it is possible, in a general sense, to give some estimate of the palaeo-drainage basin dimensions. Schumm (1977) discussed the discipline of palaeohydraulics and indicated that geologically similar basins would tend to have similar peak discharge and bedload sediment characteristics. These two variables have a dominant influence over channel shape and dimensions.

A comparison of the morphologic and palaeoflow characteristics of a number of ancient rivers by Ethridge and Schumm (1978) produced a number of palaeorivers with similar dimensions and width/depth ratios (of ca. 1:20) to those observed in the

Porto Novo member. All of these systems were thought to have had similar drainage basin areas of 15-38 000 km^2 (mostly 20-25 000km^2), and stream lengths of ca.500km. Assuming that the moderately sinuous Porto Novo river systems had a sinuosity of not more than 1.7 (just inside the lower boundary of meandering systems defined by Rust 1978) then the basin axis probably extended 300 to 500km north of the present day Porto Novo area. Such a figure would extend the basin's margin north of Porto (close to the present day northern limit of Mesozoic outcrop (Figure 1.2)) or possibly a further 100km or more to the north. The basin areas above fit well with present day basin parameters (23 000km^2 onshore area).

3.5.2 The Structural Setting of the Porto Novo member

There is only limited evidence for the structural regime existing during deposition of the Porto Novo member. Most of the available data comes from two localities, Porto Novo and the base of the member where it first appears to the south of São Bernadino. In addition, data from the logged thicknesses of section are presented in Figure 1.10.

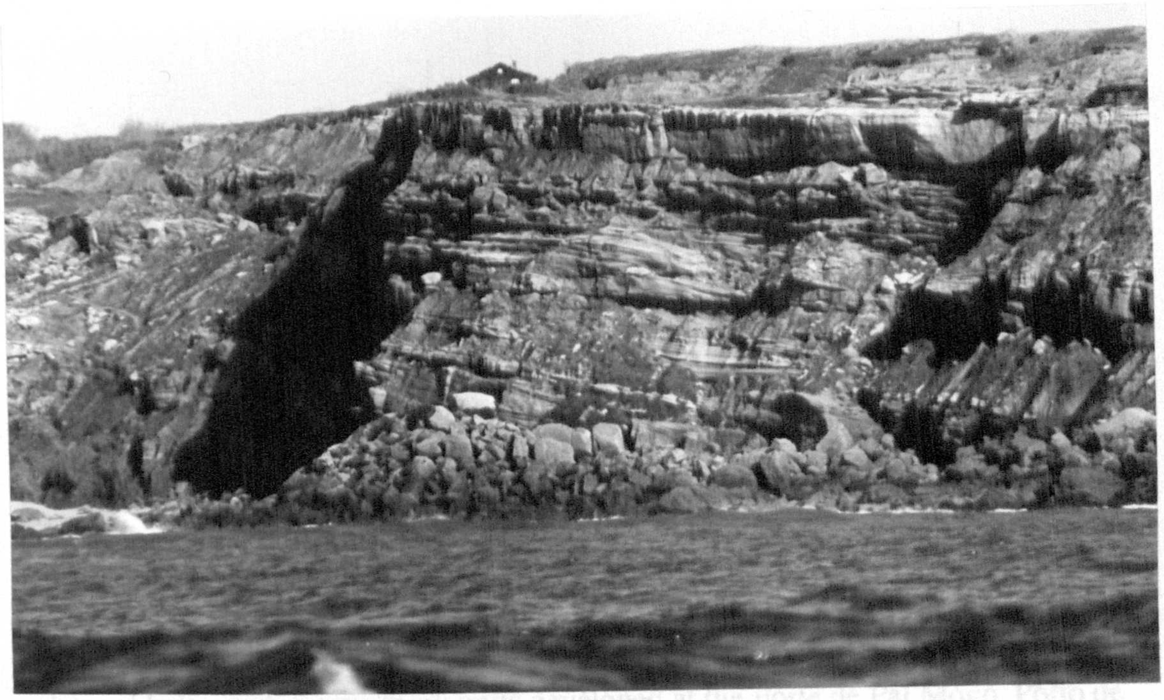
Below the erosive base of the 'Point Bar' channel, decimetre interbedded silts and fine sandstones are cut by a series of small-scale synthetic and antithetic normal faults. The channel fill sediments are clearly not affected by this deformation (Figure3.38) which is clearly contemporaneous with the deposition of at least the lower part of the Porto Novo member. As the shales exhibit brittle fractures, it is likely that the faults are the shallow sub-surface expression of a normal fault at depth. It is possible that the location of the channel above the faults illustrated in Figure 3.38 was influenced directly by the faulting, which presumably provided a topographic low. It may also be no coincidence that the major multi-storey channels on the south side of the bay at Porto Novo are in that location (cf.Alexander 1986).

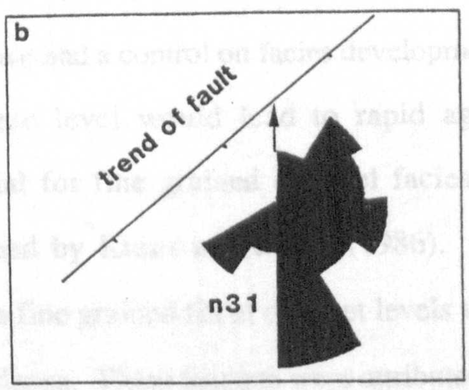
Figure 3.38 Small-scale brittle faulting of fine sandstones and shales below the 'Point Bar' sand body. The erosive base of the channel above is not affected. These structures parallel the larger scale faults observed in the section to the north (Figure 1.5). A few tens of metres to the south is a fault which downthrows ca. 80m to the north (see Figures 1.5 and 3.6). The scale of certain channel sand bodies is particularly large in the vicinity of Porto Novo. It may be that the small scale faults were the shallow sub-surface expression of faulting at depth, and may have produced a depression into which major channels avulsed repetitively. GR 6925 3710.

The minor faults illustrated, parallel the major NW-SE or WNW-ESE structural trend recorded in faults which are prominent features in the sections to north and south (Figures 1.5 and 1.22). Extensional faults with these trends are a notable feature of the Lusitanian Basin's structure (Willis 1988). The author has observed faults with identical orientations cutting the older San Martinho mudstone and Montejunto formations, immediately south of San Martinho do Porto (Figure 1.7). These structures are usually characterised by a single fault plane, which is occasionally mineralised and slickensided. Displacements are usually a few tens of metres. Synthetic and antithetic faults with displacements of only a few metres are often associated. Occasionally, more complex structures with extensive brittle deformation of the footwall rocks appear (Willis 1988). Willis dated this extensional phase as uppermost Jurassic/lowermost Cretaceous on the grounds that the white Cretaceous fluvial sands of the Torres Vedras formation, thought to be of Valanginian age (Rey 1972), are not cut by these structures. Radiometric Ar^{40} dates from a suite of intermediate dykes intruding some of these faults were 130-140my (Willis 1988).

Active extension was occurring during the deposition of the Porto Novo member. The cross section in Figure 1.10 shows that the thickness of the Porto Novo member, below the Praia Azul member, increases considerably as one moves northwards. It is possible that the increase in thickness is due, in part, to syn-sedimentary faulting and subsidence, though the base of the Lourinhã formation is almost certainly diachronous which would also account for thicknesses increasing northwards.

Particularly strong evidence for some syn-sedimentary fault activity and localised thickening was observed around the fault which separates the Praia da Amoreira and Porto Novo members at the southern end of Praia dos Frades (Figure 1.11). This fault zone contains a 0.5m thick gouge, perhaps suggesting prolonged activity (Figure 3.2). In the hanging wall there is a small synthetic fault with a displacement of about 1m and an antithetic shear zone of highly fractured rock, with little actual displacement. The most striking difference between the hangingwall and footwall is the ratio of channel sandbodies to overbank mudrocks. In the footwall there is a single channel, at the cliff top, cut by the fault (Figure 3.2). In contrast, the hangingwall is dominated by a series of, often multi-storey, channel sandbodies (Figure 3.39a). The limited amount of palaeocurrent data available indicates palaeoflows dominantly parallel to, or directed away from, the fault (Figure 3.39b). Sediments within one of the channels are cut by active growth faults (Figure 3.39c). The relatively thin mudrocks between these channels exhibit very limited pedogenic development, with the exception of a single, well developed profile below the channel at the cliff top. In addition, a notable feature of most of the channels in the section is the low incidence of caliche pebble lags.





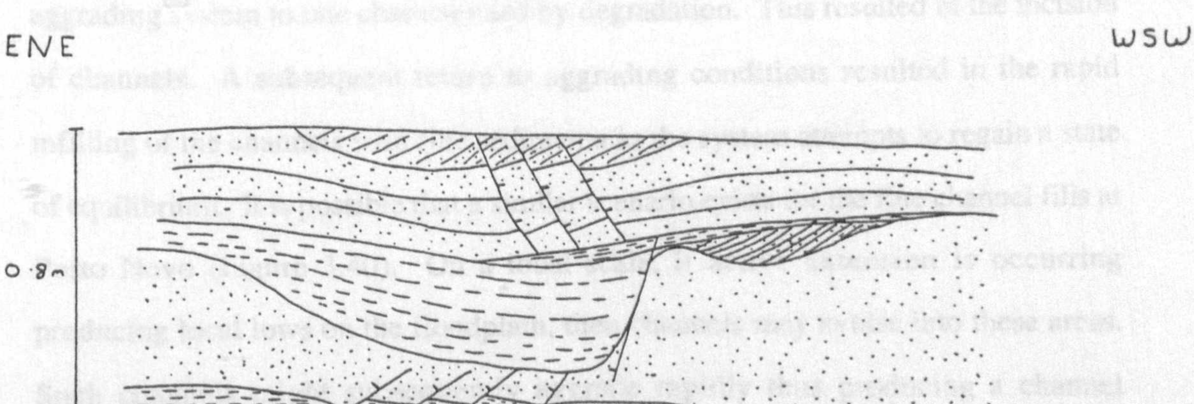


Figure 3.39 a) (top) Multistorey sand bodies in cliffs immediately south of Praia dos Frades and the fault which juxtaposes the Praia da Amoreira and Porto Novo members (see Figure 3.2). As these sand bodies are on the downthrown side of the fault, their presence may be due to the concentration of channels close to the fault above the hanging wall. **b)** Palaeocurrent data, gathered from channel sand bodies in the logged sections forming the hanging wall block. These exhibit a trend parallel to and away from the fault. **c)** Sketch illustrating growth faults observed in a channel fill from the logged section. Stipple indicates sandstone, dashed lines shale.

Computer models of alluvial stratigraphy, such as that of Bridge and Leeder (1979), predict that in asymmetrically subsiding half graben, channels will be preferentially concentrated close to the footwall during periods of active tectonic subsidence. Potter (1978) and Alexander (1986) suggested that channels would concentrate within, and flow parallel to, surface downwarps caused by faulting. The combination of the data strongly suggests that these and other faults were active during the deposition of the Porto Novo member and were responsible for the lateral variations in thickness observed. It is also possible that through temporal variations in relief and base level, faulting also exercised a control on the distribution of channel belts, with multi storey units developed at the Forte de Pai Mogo, Porto de Barcas and Porto Novo in particular.

Faulting may even have had a control on facies development and the type of channel fill; a raising of base level would lead to rapid aggradation and increased preservation potential for fine grained channel facies. A possibly analogous situation was outlined by Kraus and Bown (1986). They observed scours of variable depth with a fine grained fill at discreet levels within Eocene and Triassic fluvial systems in Arizona. These features were attributed to an initial reversal of an aggrading system to one characterised by degradation. This resulted in the incision of channels. A subsequent return to aggrading conditions resulted in the rapid infilling of the channels with fine sediments as the system attempts to regain a state of equilibrium. It is possible that a similar scenario exists for the fine channel fills at Porto Novo (Figure 3.40). On a local scale, if active extension is occurring producing local lows on the floodplain, then channels may avulse into these areas. Such channels might subsequently aggrade rapidly thus producing a channel dominated by fine grained fill because of the high suspended loads carried by the system.

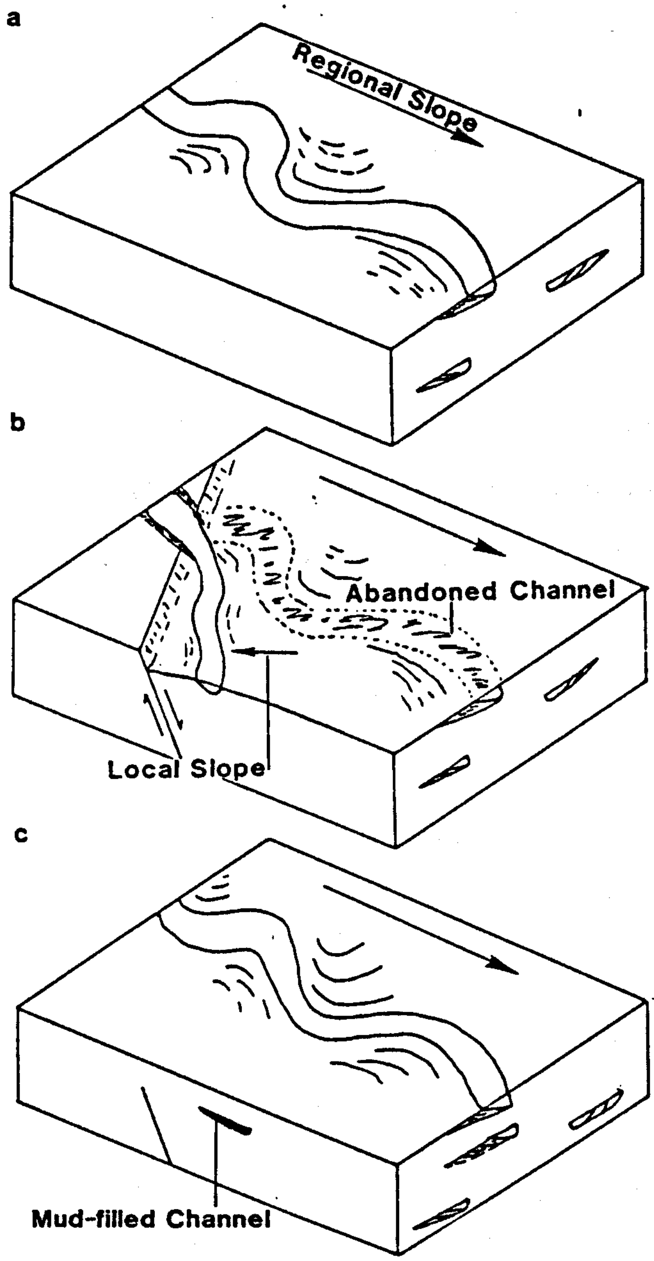


Figure 3.40 Series of sketches illustrating the possible effect of faulting on river channel courses, and the formation of the mud-filled channels observed at Porto Novo. **a)** 'Before' sketch of a river meandering across its floodplain. Downslope direction is to the right. **b)** Faulting occurs, resulting in a topographic low at the land surface (the fault need not break through to the surface). The channel avulses into this area and rapidly aggrades to re-establish equilibrium (floodplain mudrocks above the footwall may be rapidly re-worked providing much fine-grained sediment). **c)** The channel is plugged with mud and avulses again returning to its more usual course.

CHAPTER 4 MARINE TRANSGRESSION: THE PRAIA AZUL AND SANTA RITA SECTIONS

4.1 INTRODUCTION

The alluvial sediments of the Porto Novo member were innundated by a marine transgression from the south. The age of the transgression has, until recently, been in doubt. Leinfelder (1987) dated the base of the unit at the Kimmeridgian-Tithonian boundary on the basis of the ostracode *Cetacella armarta* (found at Praia Azul by Fürsich 1981b). Leinfelder (op cit.) linked the transgression to a peak on Vail et al's. (1984) sea level curves. This, the most widespread of all the transgressions recorded in the Lourinhã formation and presumably the most long lived, extended at least as far north as the Forte de Pai Mogo (Figure 1.7).

The sedimentary facies which developed characterise a shallow, predominantly muddy, brackish/marine environment, in places dominated by the influence of fluvial channels. These formed small lobate deltas which prograded into the shallow marine environment.

This chapter will primarily be concerned with the facies of the Praia Azul member. However, enclosing units that contribute to the assessment of the overall palaeoenvironment will also be considered. In particular, the basal 120m of the Assenta member which crops out at Praia da Santa Rita will be discussed in detail.

The two major coastal outcrops of the Praia Azul member are a) the proposed type locality between the southern end of Praia da Amoreira and the mouth of the Rio Sizandro, and b) to the north at Praia da Santa Rita (Figure 1.8). At the former, the member is some 70-120m thick. It thins progressively northward where smaller outcrops at Porto Dinheiro, Porto das Barcas and the Forte de Pai Mogo record a declining marine influence (Werner 1986), (Figure 1.10). The section at Praia da Santa Rita exposes some 60m of predominantly marine and brackish water facies, and 120m of the Assenta member, which is largely continental in character. At Praia da Santa Rita the Assenta member is locally unconformably overlain by the coarse grained Santa Rita member.

The succession at the type locality of the Praia Azul member was briefly described by Fürsich (1981b) who interpreted the succession as a delta distributary system discharging into a shallow sea with protected bays and lagoons. The most detailed aspect of his study concentrated on assemblages of macrofauna, their distribution and salinity tolerances. I broadly agree with Fürsich's interpretation but specifically disagree with the interpretation of individual facies.

Inland, outcrop is exceptionally poor and confined to isolated and relatively recent road cuttings. It seems that the palaeoshoreline extended northeast-southwest, as similar facies are recorded several kilometres inland by Leinfelder (1986). Leinfelder referred to these facies as the Santa Cruz member in his lithostratigraphic scheme. Towards the western part of the Arruda region, Leinfelder recorded a stronger marine influence with carbonate facies predominating. These are similar to facies recorded in the region of Cabo Espischel some 80km to the south (Fürsich and Schmidt Kittler 1980) (Figure 1.2). The Cabo Espischel sections are thought to be near the eastern margin of the basin, as tongues of continental clastics appear and thicken to the east. In view of the above, it seems likely that the axis of the marine part of the basin was narrow and elongate in a NE-SW direction (Leinfelder pers. comm.). No marine sediments have been found to outcrop inland north of the Torres Vedras - Santa Cruz line, though the existence of a northward thinning unit would be expected on the evidence provided by coastal outcrop.

4.2 DESCRIPTION OF THE TWO MAIN SECTIONS

4.2.1 The Praia Azul section

Figure 4.1 The type section of the Praia Azul member is dominated by mudrock punctuated by a number of thin, laterally persistent shellbeds (prominent beds in cliff section) and about half way up the cliff, a series of lenticular channel sand bodies. One of the sand bodies is cut by the small reverse fault near the end of the cliff. The persistence of shell beds suggests that the environment of deposition consisted of a simple series of facies 'bands'. Cliff height 50m. Camera facing due N; fault downthrows to the north.

Extending for some 1.5km from G.R.6635 3083 just south of Alto da Vela to G.R.6600 2935 mid-way along Praia Azul where outcrop is lost, the Praia Azul member is exposed in near vertical cliffs up to 70m high. The section dips at a shallow angle (ca 4°) to the southwest. An apparently conformable lower boundary is exposed, marked by a unit of heterolithic laminated sediment and medium to thick bedded sands with marine trace and body fossils above bright red terrestrial mudrocks. In all, some 70m of the Praia Azul member is exposed (Fürsich (1981b) logged 89m). The top does not outcrop due to a break in section formed by the valley of the Rio Sizandro. It is estimated that about 50m of section is missing between the top of the exposure at Praia Azul and the base of the outcrop south of the Rio Sizandro. This probably conceals the junction between the Praia Azul and Assenta members, as the character of the section to the south is dominantly continental (see Chapters 1&5). It is possible that the break in section formed by the Rio Sizandro valley marks a fault zone but no evidence to either confirm or deny this possibility was found.

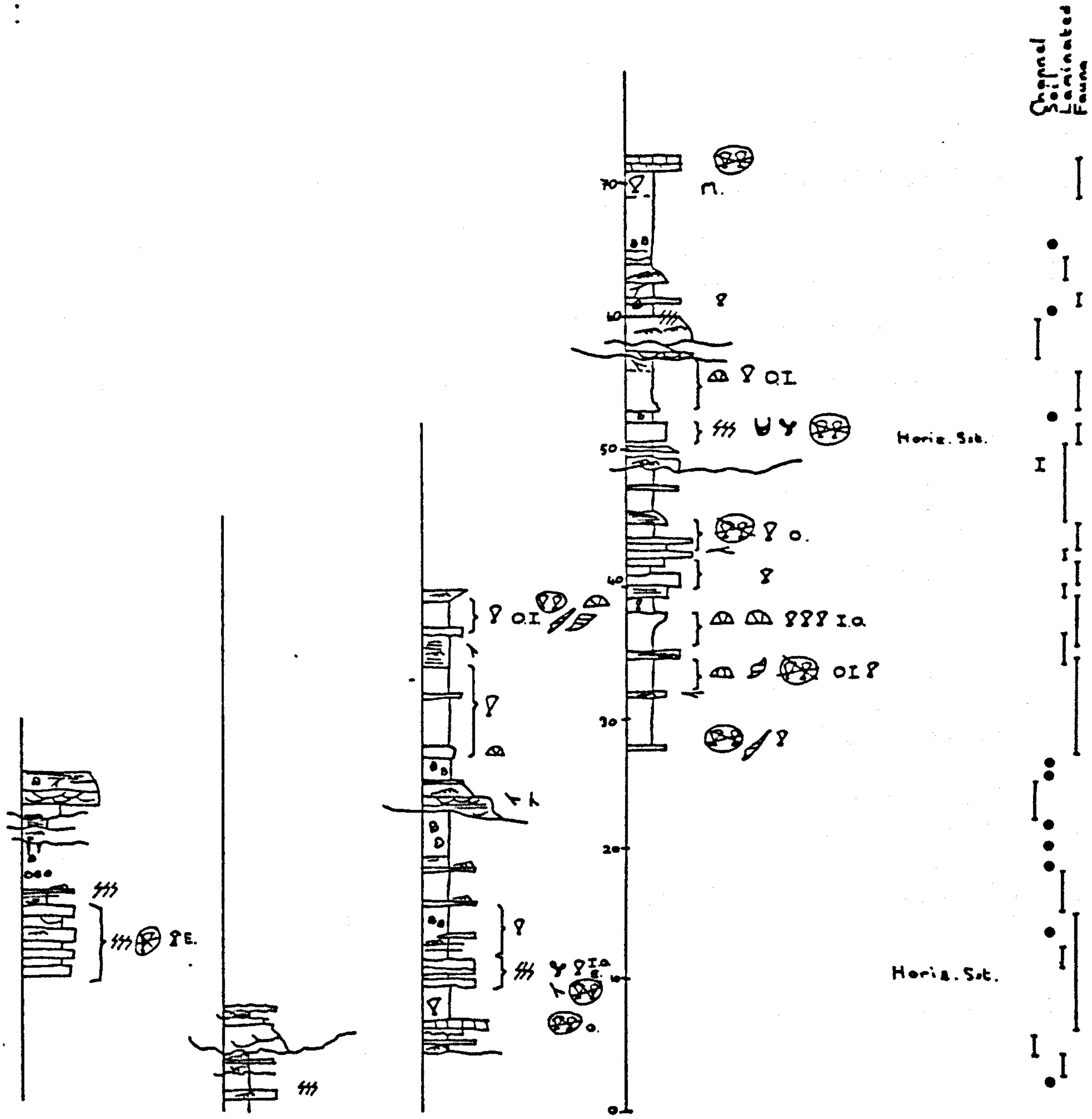


Figure 4.2 Log correlation diagram for the Praia Azul section. Logs and Key are in enclosures. Positions of channel, soil, laminated facies and fauna indicated at right hand. Note the lateral persistence of facies bands (compare with Figure 4.1). Vertical scale in metres.

About 100m of section was logged at five points along the cliffs, thus producing a limited degree of overlap. There are problems with access to much of the section because of its near vertical nature. The most striking feature of the outcrop is the lateral persitence of many beds (Figure 4.1). Different facies tend to occur at discrete levels. Even channel sandbodies which have a lenticular rather than sheet-like geometry all tend to occur at one of four levels in the section. This is reflected in the correlation diagram (Figure 4.2) with channel facies at the 25m level and the close correlation of fossil-rich and laminated sediment horizons at 30-40m. The lateral persistence of facies indicates that localised environmental variation was very limited: the environment of deposition was characterised by the development of a simple series of broad facies bands.

The other major feature of the section is the predominance of mudrock. Amounting to some 60% of the logged succession, it is indicative of the prevalence of low energy conditions in the marine environment. There is markedly little or no re-working of sediment and structures such as wave ripples have not been found.

A variety of lithofacies exist, many with the associated faunal assemblages documented by Fürsich (1981b) indicating abnormal, probably lowered salinities. The facies and faunal assemblages documented by Fürsich are outlined in Table 4.1. They are dominated by muddy facies with less common sandstones which crop out as isolated or grouped horizontal beds of fine sands, or lenticular channel sand bodies. Carbonate rocks are virtually absent, being confined to a few very impure silty beds.

4.2.2 The Santa Rita section

A photograph of the Santa Rita section appears in Chapter 6 (Figure 6.1). At Praia da Santa Rita the base of the section dips steeply at more than 40° to the south west. Dips decrease to about 20° at the base of the coarse grained Santa Rita member (Chapter 6). Below this the outcrop consists of 60m of the Praia Azul member overlain by 120m of the Assenta member. The base of the section is cut by a fault

1)Fluvial Channels:	lenticular sand bodies; fining-up; trough cross bedded coarse-medium sandstone; erosive bases
2)Flood Plain:	unfossiliferous red/green sandy marly silts with calcareous nodules and sandy intercalations
3)Levees:	intercalated between (1) and (2); fine and medium sandstone; ripple laminae and bioturbation; unfossiliferous
4)Marsh:	grey marl with calcareous nodules, plant debris, root horizons; uncommon facies
5)Delta front sands:	fine sands in units several metres thick; some ripples or trough cross beds; bioturbation (*Thallassinoides Planolites*); shell beds and layers of shell debris; low diversity fauna
6)Protected Bay /Lagoon:	grey fine sandy/silty marl; bioturbated; abundant low-diversity fauna, *I. lusitanicum* banks, oyster patch reefs and thin sandstones burrowed by *Thallassinoides*
7)Delta abandonment facies:	thin laterally persistent shelly micrites or shelly calcareous sandstones, diverse encrusted shelly fauna, *Rhizocorallien*

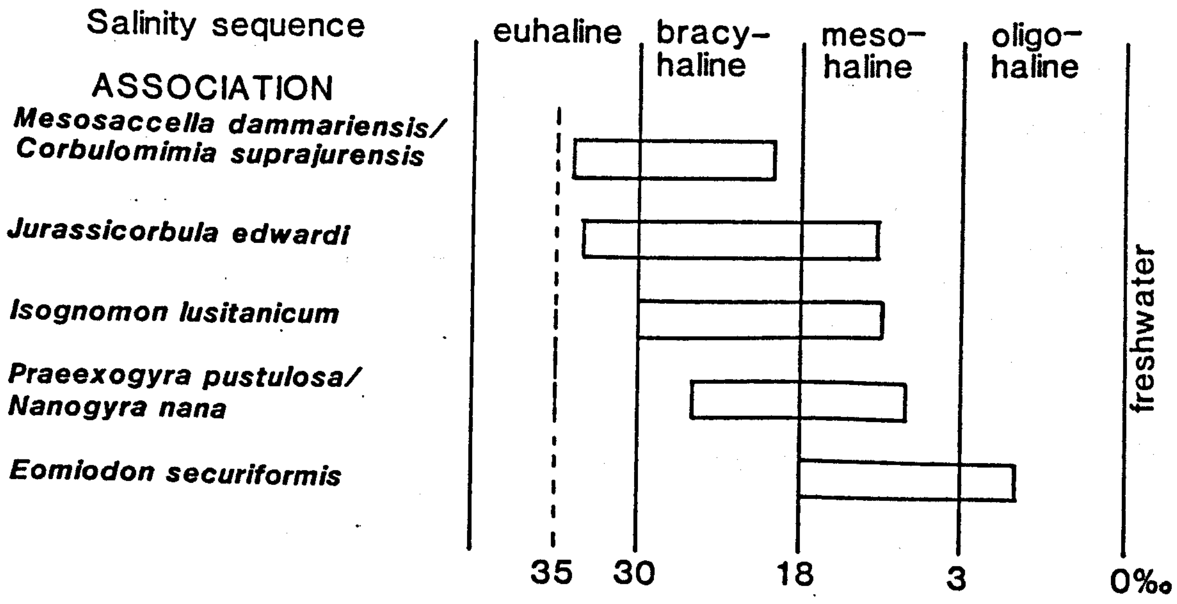
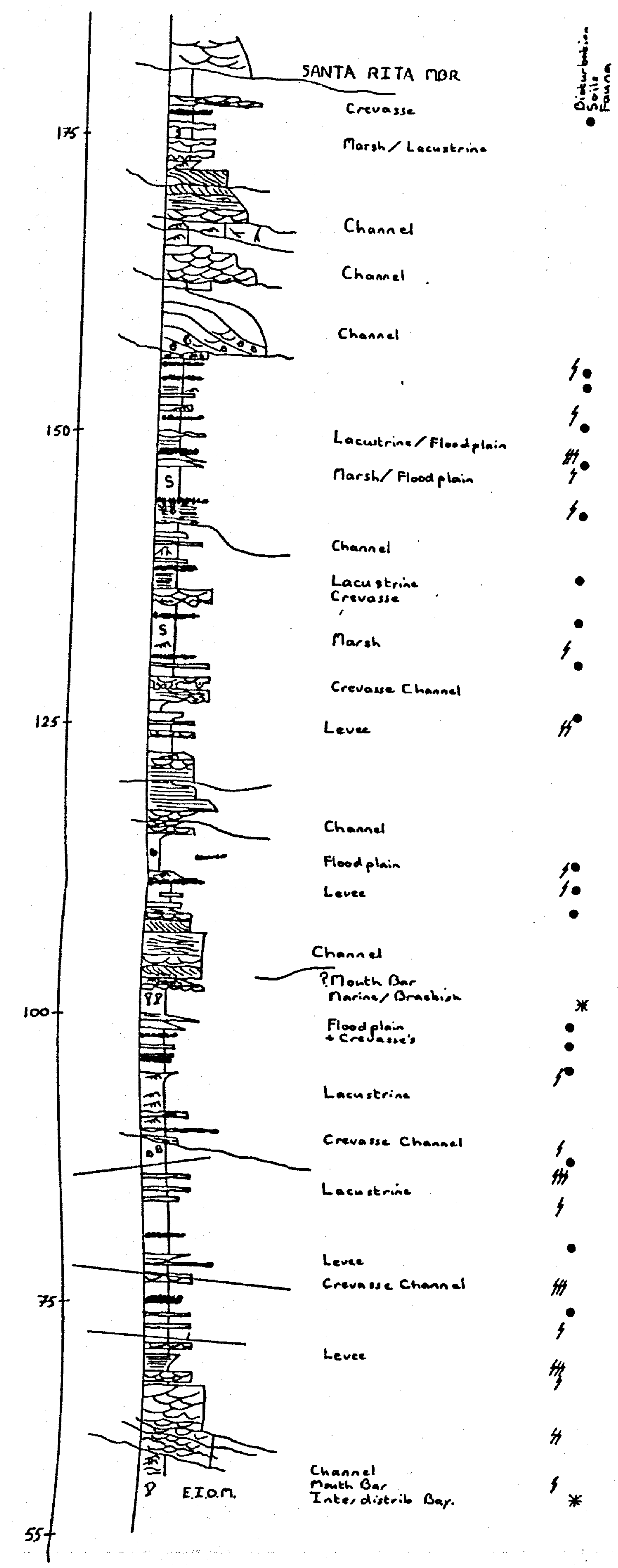


Table 4.1 Lithofacies and faunal assemblages, plus their presumed salinity ranges, documented by Fürsich (1981b).



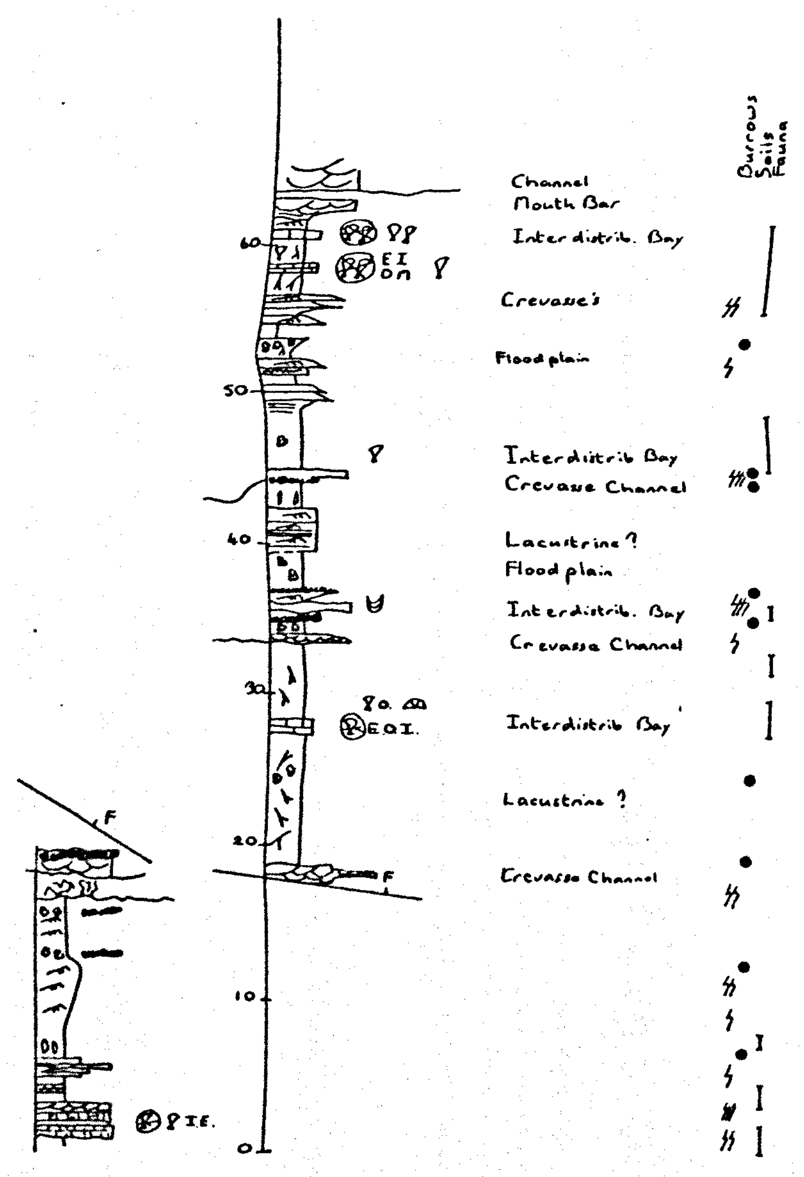


Figure 4.3 Log correlation diagram for the lower part of the Santa Rita section below the unconformity with the younger Santa Rita member. Logs and Key appear in the enclosures. The first 60m (left) are of the largely marginal marine Praia Azul member. The remainder (above) which is largely continental in character belongs to the Assenta member. The presence of bioturbation, soils and fauna are indicated to the right of the logs. Vertical scale in metres.

which cuts across the southern limb of the Vimeiro structure inland (Figure 1.23). On the coast this marks a break in outcrop for some 550m to the north where the Porto Novo member crops out. The boundary between the Porto Novo and Praia Azul members is not exposed here, but at the cliff top at Porto Novo there is a shell bed dominated by *E. securiformis* above several metres of red floodplain silts. A number of other faults, the throws of which are unknown, but not thought to be great, cut the lower part of the Santa Rita section.

The section can be divided into two distinct parts separated by the first of a number of major sheet sand bodies which crops out 62m above the base of the section (Figure 4.3). Beneath the sand body the sediments are dominantly marine (Praia Azul mbr.) and above it they are dominantly terrestrial (Assenta mbr.).

4.3 FACIES DESCRIPTION AND INTERPRETATION

The various facies exposed at the main outcrops are described below. The facies and brief interpretations appear in Table 4.2.

4.3.1 Fossil rich facies

This facies encompasses a number of sub-facies which are listed below. Many of the sub-facies were worked on by Fürsich (1981b). Fürsich's work concentrated on the faunal assemblages present and their salinity tolerances. He defined five assemblages (Table 4.1) reflecting a range of salinity tolerances from eu- to oligohaline and also noted the low incidence of stenohaline fauna. As my sub-facies are based largely on lithological criteria a number of Fürsich's faunal assemblages fall into the same sub-facies. The sub-facies are as follows:

1) Fossil rich siltstones and fine sandstones

2) Sandy biomicrites

3) Nerineid wackestones, mudstones and marls

Fossil rich facies:		
a)Fossil rich silt stones:	massive silts with a variety of fauna; patch reefs and shell banks	interdistributary bay environments
b)Sandy biomicrites:	similar to (a); laterally persistent beds often with diffuse bases; much fragmented and encrusted shell debris, *I. lusitanicum* and *M. lusitanicum* in life position; *Diplocraterion* burrows	abandonment facies, or interdistributary bay environment locally protected from clastic influx
c)Nerineid wackestones.......	*mudstones and marls*	restricted lagoon
Horizontally bedded sandstone facies	fine bioturbated sands, dm-m beds, in packets up to several metres thick *E. securiformis* and gastropod fauna; occasional ripple laminae, cross beds and low-angle laminae	shoreface and beach environments
Isolated sharp-based sand beds	fine-coarse unfossiliferous sandstones intercalated with mudrock, fossiliferous or laminated sediment; bioturbated or cross bedded	crevasse splays
Laminated silt and sand facies	laminated silts and fine sands; a variety of structures/associations	a variety of environments including mouth bar, crevasse and lacustrine see Table 4.3
Massive mudrock facies	red/green with pedogenic nodules or grey with plant debris and/or fossils	delta-top/floodplain or marsh/interdistrib. bay environments
Massive sandstone facies	large-scale sheet sand bodies up to 8m thick; Santa Rita only	trunk rivers in upper delta plain
	small-scale lenticular to sheet like sand bodies <3m thick P. Azul only; tend to occur as multiple sand bodies at certain horizons	distributary channels in the lowere delta plain
	isolated, small-scale, lenticular 'winged' sand bodies, <3m thick	semi-permanent crevasse channels

Table 4.2 Lithofacies of the Praia Azul and Assenta members

Fossil rich siltstones and fine sandstones

This is the most common sub-facies. It consists of massive dark green or grey silts or rarely fine sandstones, rich in bivalves which are frequently preserved as fragile impressions only. This is punctuated by horizons periodically packed with *Isognomon lusitanicum*, a large semi-infaunal filter feeding bivalve (Fürsich 1981b) and oysters, the most common faunal elements. Frequently the latter encrust *I. lusitanicum*. Less common are the trigonid bivalve *Myophorella lusitanicum, E. securiformis* and gastropods, particularly two species of spired ceretiid type, plus an occasional coiled trochus species. In addition, the club-like spines of the cidaroid echinoderm *Pseudocidaris* appear and are taken to be indicators of normal or at most brachyhaline conditions (Fürsich 1981b). At one horizon (45m level Santa Rita section) the serrated tooth of a bipedal carnivorous dinosaur (identified by N. Hollingworth) was found. Texturally the silts are often fine-grained, micaceous, soft and blocky, though occasionally they may be better lithified and rather fissile. Lignitic debris is often common. Significant thicknesses of this facies may occur (up to 5.5m) with little variation.

This sub-facies encompasses four of the faunal assemblages identified by Fürsich (1981b) (see Table 4.1). Faunal elements often appear quite abruptly within a sequence and may disappear equally quickly beneath interbeds of the laminated facies. As a result of the abrupt appearance/disappearance of fauna and variations in assemblages and amount of carbonate sediment present (always rather minimal), this facies exhibits considerable variability. The variability includes localised concentrations of oysters, often encrusting each other and occasional disarticulated specimens of *I. lusitanicum* (the *Praeexogyra pustulosa-Nanogyra nana* association of Fürsich op cit.) forming mounds up to 1m thick. Occasional *Pseudocidaris* spines have also been observed. This association often forms prominent features in the cliff section due to its coarse texture. Laterally the mounds thin over a few metres and pass into fossil rich silts.

Figure 4.4 Upper surface of one of the shell beds dominated by *I. lusitanicum*. These beds tend to have considerable lateral persistence (vis Figure 4.1). This facies is thought to have formed large 'shell banks' (Fürsich 1981b). Hammer 0.4m long.

In contrast to the mounds most other examples of this sub-facies tend to have considerable lateral persistence. Shell beds up to 0.8m thick and visually dominated by *I. lusitanicum* (though Fürsich (1981b) states that they only represent 5% of the biomass) appear at several levels in the section (Figure 4.4). These beds are examples of Fürsich's *I. lusitanicum* association. The fauna are often disarticulated and heavily encrusted with oysters. The beds pass laterally into thinner, persistent beds with fewer specimens in life position. Along strike, such beds can be traced for hundreds of metres (Figure 4.1). Their bases are rather diffuse with pockets and clusters of *I. lusitanicum* extending down into the silty substrate. The matrix consists of marly silts and fine sands. Elsewhere, there are thin units dominated by specimens of *I. lusitanicum* in life position, notably at 34m (Log 2 Figure 4.2) where unusually large specimens, most fully articulated with a dorso-ventral axial length of 15cm or more occur.

Locally, erosive-based, structureless very-fine sand beds, rich in shell debris with no intact specimens, occur. These are also laterally persistent, examples occurring at 30m and 42m (Figure 4.2). An unusual occurrence in silts containing sparse fragmental oyster debris is the appearance of a series of vertical tapering wedges of carbonate all lying at the same level in the sediment. Laminated sediments below are burrowed by *Diplocraterion* whilst sediments above are increasingly terrestrial.

Interpretation : Most of the fossil rich facies probably formed in interdistributary bay areas. As these facies most often form laterally continuous zones of fossil rich fine sediment without channel deposits dissecting them, the bays were probably quite extensive shallow regions between a series of small lobate deltas. There is little evidence of significant relief within the bays except for the mounds, interpreted as oyster patch reefs by Fürsich (1981b), and the shell beds which he interpreted as "banks". I agree with his interpretation. The patch reefs and banks probably had a positive relief of up to a metre.

The structureless nature of the silts is probably due to bioturbation, as burrows are certainly common in other lithologies in the succession. The silts indicate that in the bays fine grained sedimentation from suspension was the dominant process. This was punctuated by spells of more rapid and coarser grained sedimentation during floods, as indicated by the presence of laminated and coarser sediments. Flood influxes were probably also responsible for the presence of significant quantities of plant debris. Local concentrations of fauna may have acted as baffles trapping sediment which contributed to the build-up of the patch reefs and shell banks.

The presence of occasional erosive based shell beds containing only fragmental shell debris is strongly suggestive of rare storm events concentrating shell debris by winnowing of the fine sediments. Alternatively they could represent delta lobe abandonment facies (see 'sandy biomicrites'). The fact that these beds are often rather sandy supports either proposal. However, the former interpretation is favoured as carbonate-rich beds are thought to be the product of abandonment (see 'sandy biomicrites').

The rarity of carbonate beds in the sections strongly suggests that rates of clastic sedimentation were normally too rapid to allow the concentration of carbonate, despite conditions otherwise conducive to its deposition. This is supported by the presence of much more carbonate rich successions to the east in the Arruda region (Leinfelder 1986, 1987) and to the south on the coast near Cabo Espischel(Fürsich and Schmidt Kittler 1980). The presence of shallow marine carbonates only 80km away on what is thought to be the opposite margin of the basin strongly suggests that the entire basin was rather shallow at this time.

As the silts often pass into reddened sediment it is thought that the bays were always rather shallow and prone to silting-up, with consequent exposure to the atmosphere and oxidation. The fossil rich silts containing vertical pipes of carbonate may be the result of bay sediments passing laterally into marshland subject to infiltration of meteoric waters and, during dry spells, pedogenetic processes.

Sandy biomicrites

This sub-facies is very similar to 'shell banks' observed at Praia Azul and also to the north of Porto Dinheiro. It consists of grey to light brown well cemented, shelly, sandy, micritic packstones dominated by fragmented shell material. The proportion of intact specimens tends to increase upwards through a bed. These are usually 20-50cm thick with undulose upper and lower surfaces. They also have considerable lateral persistence though this cannot be quantified due to the orientation of the beds. A bivalve fauna dominates with both *E. securiformis* and *I. lusitanicum*, plus oysters. Sparse lignitic debris is often present. Two examples occur at the base of the Santa Rita section, interbedded with a poorly cemented grey marl also containing shell debris and fine lignitic debris. The upper example is interesting as it contains examples of *I. lusitanicum* in life position, at the top of the bed. This is sharply overlain by a cross bedded sandstone which contains no fossil material (Figure 4.5).

Figure 4.5 Cross bedded sandstone sharply overlying a shell bed near the base of the Santa Rita section. It is likely that the sand bed represents a single depositional event, probably a crevasse, following the breaching of a distributary channel levee during flooding. Scale bars on hammer = 0.1m.

Examples of this sub-facies have rather diffuse bases above silts. The amount of bioclastic material present can vary considerably between beds, lithologies varying between wackestones with sparse, mostly fragmentary, shell debris and packstones. In the latter, up to 95% of the fossil material present may be represented by fragmentary debris the remainder being made up of whole body fossils of the faunal elements already described including *M. lusitanicum* (Figure 4.6). These often appear as either pockets or a veneer of sandy bioclastic packstone on bed surfaces with a number of specimens in life position. Most are intensely burrowed. *Diplocraterion* is the most common identifiable trace fossil with 6mm diameter tubes commonly filled with a dark green silt. Spreite are dominantly protrusive.

There is a single example of a carbonate bed which appears to have a limited lateral extent. It is also, at one metre, the thickest example of this facies and lies just below one of the small channel sandbodies in the Santa Rita section (see massive sandstone facies). Laterally it passes into massive silts. The amount of fossil debris present decreases upward from a packstone at the base to a wackestone. Silts between this unit and the channel include specimens of *E. securiformis* in life position, plus *I. lusitanicum* and oysters. The bed base is rather diffuse, grading up from shelly siltstones to marls.

Figure 4.6 Large specimen of *I. lusitanicum* in life position (the cracked shell with prominent growth lines) at the top of a biomicrite of packed shell debris. Also prominent are a number of specimens of *M. lusitanicum* (the nobbly shells). Scale bar = 0.1m.

Interpretation : The most important feature of this sub-facies is the significant amount of carbonate. This contrasts with shell beds and banks and indicates that at times the environment was starved of clastic input. The gradational bases of a number of the beds suggest a gradual increase in the amount of carbonate present, or more likely a fall off in the supply of silt. The common retrusive spreite may be indicative of the winnowing of sediment by some limited current action. This is also supported by the concentration of shell debris and the common occurrence of sand grains in the carbonate beds.

The most likely location for the facies as a whole is in interdistributary bays; the carbonate present suggests a distal position. Another possiblity is that the deposits are the product of abandonment. Abandonment facies are characterised by lateral persistence and the re-working and winnowing of sediment as a delta lobe subsides and is encroached and re-worked by the sea (Elliott 1974b). Abandonment facies are not favoured for all of these beds; for example the unit overlain by a cross bedded sandstone which suggests proximality to a distributary channel and the isolated carbonate bed which does not have the required lateral persistence. It is possible that the latter may have formed in an area locally isolated from clastic input, possibly by perturbations of the bay floor.

Nereneiid wackestones, mudstones and marls

This sub-facies appears at only one location; near the base of the Santa Rita section. Only about 0.75m thick, it consists of decimetre bedded impure silty mudstones and wackestones with a monospecific fauna of small, turreted, nereneiid gastropods. Bed bases and tops are rather irregular. The gastropods have straight sided shells with an axial length of up to 15mm and a maximum diameter of about 4mm across the apetural whorl. Normally about six whorls are developed. The micrites are also burrowed by *Diplocraterion* and are interbedded with dark green, occasionally parallel laminated silts. Where unlaminated, the silts also contain the gastropod fauna.

Interpretation : The uniqueness of this sub-facies suggests some special conditions, at times associated with low rates of clastic sedimentation facilitating the concentration of precipitated carbonate. A low diversity and unusual fauna also suggest that environmental controls, such as salinity, were unusual. It is possible that the environment was isolated from others, perhaps in a saline lake or lagoon subject to rare influxes of clastic sediment during floods.

4.3.2 Horizontally bedded sandstone facies.

This facies occurs at a number of levels within the Praia Azul section, as isolated beds up to 0.75m thick, or as packets of sediment particularly near the base of the section and at the 52m level. In two places this sandy facies lies above channel sand bodies and below marine sediments. Single isolated beds repeatedly occur above reddened mudrocks and pass upwards into marine silts.

Figure 4.7 Northern part of the Praia Azul section. The base of the member occurs where outcrop is lost in the debris at the cliff base. The prominent sediment body just above the cone of debris at the end of the outcrop is a small lenticular sand body. This lies above, and is flanked by, laminated facies and overlain by the horizontally bedded sandstone facies (the prominent, laterally persistent beds). Several metres above the sand facies and separated from them by laminated sediment lie channel sand bodies, overlain by marine strata. Cliff height ca.45m. Camera facing due south.

At the base of the section this facies is up to 5m thick (Figure 4.7). It consists of very fine to medium sandstones usually 0.1-0.4m thick and often rich in shell debris (Figure 4.8). Normally upper and lower bed surfaces are well defined and horizontal, though locally shallow scours < 0.4m deep penetrate the top of the bed beneath. Commonly original depositional structures have been completely obliterated by bioturbation. *Diplocraterion* and *Thallassinoides* are the most common burrow types observed, the latter on bed bases. *Planolites* burrows, <5mm diameter appear on bed surfaces. Occasional remnants of ripple and parallel

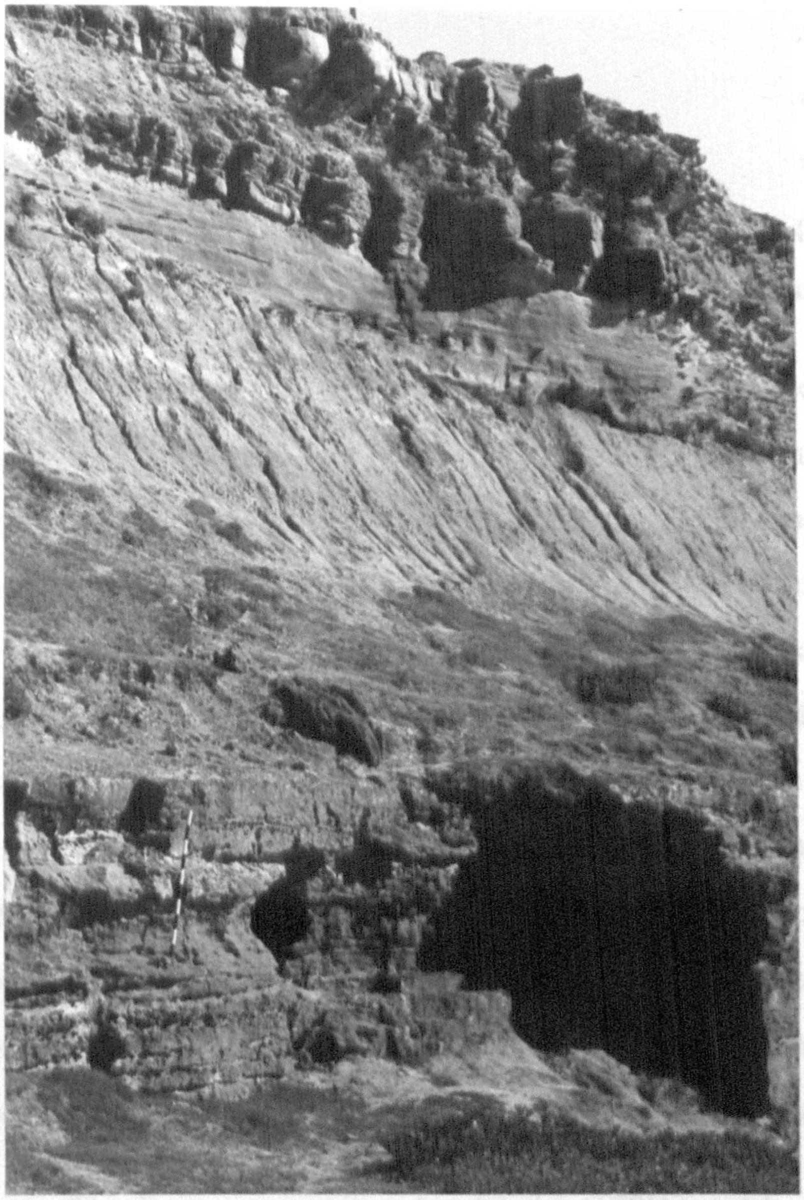


Figure 4.8 Close view of the upper part of the horizontally bedded sandstone facies and its relationship to (channel) sand bodies (one of which occupies the cliff top. The slightly rough texture in the mudrocks just above the grass line is a carbonate soil profile. (A more detailed description of this succession appears in section 4.3.6). The structureless nature of the sandy sediments is apparent from this figure. Metre scale.

laminae occur and small 0.1m deep, concave-up scours suggest the original presence of cross-bed sets. A single example of a medium, granule rich sandstone with low angle parallel laminae was observed.

Shell debris found within the sands seems dominantly derived from oysters, while larger generally intact specimens are of the shallow infaunal bivalve *Eomiodon securiformis*. These often appear as concentrations of disarticulated valves, particularly on bed surfaces. Articulated specimens in life position also appear. The only other body fossil found are rare occurrences of turreted cerithiid gastropods.

Laterally, units dominated by sand can be traced into areas where the sand beds may be separated by thin (< 0.1m) beds of grey siltstone. The siltstone is usually rather homogenous as it is intensely bioturbated, *Thallassinoides* often being the sole trace fossil present. Faunal diversity may also increase laterally with species such as *Isognomon lusitanicum* and various oysters appearing. Sand colours tend to be duller greys as opposed to the yellows of the thicker units. Such a transition can be seen in the correlation diagram (Figure 4.2), between Logs 4 and 2. The latter has a 0.65m patch reef dominated by *I. lusitanicum* interbedded with the sands and also has fossiliferous silts above and below the packet of sand beds. In contrast the thicker sand beds in Logs 4 and4A are bracketed by laminated sediments. A few hundred metres to the south a facies transition into fossil rich (shell bank) facies occurs. These units also lie only a few metres above terrestrial sediments.

Interpretation: This facies was interpreted by Fürsich (1981b) as delta distributary mouth bars. This seems unlikely as nowhere are they incised by channel sandbodies. On the contrary, they are overlain by either silts with a shelly fauna representing a low energy marine environment or by reddened (terrestrial) mudrocks suggesting the progradation of a muddy shoreline. Channel sand bodies often appear above the terrestrial mudrocks, but it is hard to see how this sandy facies could be related directly to the sand beds several metres below. In addition, the sand beds do not resemble laminated sediments found elswhere below channel sand bodies, which I interpret as mouth bar deposits (see 4.3.4 and 4.3.6).

The sands probably originated as beach or shoreface deposits subject to winnowing of fines by wave action. The coarsest unit with its low angled laminae is unfossiliferous and strongly suggestive of sands in the surf or swash zone (Elliott 1986b). The variable thickness of this facies may well be linked to rates of sediment supply, possibly a local variable dependant on a fluvial distributary; i.e. the extent of sand available for reworking.

The location of this facies between terrestrial and marine facies is strong evidence for its origin as a shoreline deposit. Planar bed bases above reddened terrestrial sediments are characteristic of transgressive shoreface sands (Elliott 1986b). It is therefore possible that some examples of this facies represent the reworking of sandy delta top sediments following abandonment of a delta lobe. The coquinoid horizons and scoured surfaces observed within this facies are most likely due to periodic storms, possibly related to Winter/Summer cycles.

4.3.3 Isolated sharp-based sand beds

This facies appears at a number of levels in both sections. It consists of single beds of cross bedded, bioturbated or structureless sandstone between 0.2 and 1.1m thick. Beds have sharp, often erosive bases and many contain lags of lignitic debris, mudclasts and caliche pebbles. The tops of a number of beds exhibit rather abrupt termination of coarse grained sedimentation. The lateral extent of beds is variable, ranging from a few tens to hundreds of metres. Lithologies consist mostly of medium or fine micaceous sandstone which may grade up into ripple or planar laminated silts. Rare examples contain coarse sand grains of pink alkali feldspar. A single bed at Praia Azul has its top colonised by small (< 2 cm dia.) bivalves. The sparse almost monospecific fauna is mostly preserved as casts on the upper bed surface. A few oysters are also present.

The thickest example of this facies crops out near the top of the Santa Rita section. It consists of a basal pebble conglomerate with quartz, feldspar and granitoid clasts which grades into trough cross bedded very coarse sands overlain by a thinner repetition of this cycle. The bed base is loaded into the enveloping grey silts.

In the lower, marine part of the Santa Rita section, a biomicrite with fauna in life position is overlain sharply by a 0.25m cross bedded medium sand which grades into ripple laminated fine buff sands and dark grey parallel laminated silts (Figure 4.5). Another example, with sharp upper and lower surfaces, consists of fine light grey sandstone with alternately parallel and ripple laminated, trough cross bedded and intensely burrowed horizons. Burrows are dominated by *Diplocraterion*.

In the upper part of the Santa Rita section, a bed 0.65m thick and of limited lateral extent consists of medium sands. It has a loaded erosive base with pebble grade mudclasts and is cross bedded with a single set at the base 0.4m thick.

Interpretation: The most likely origin of this facies is as crevasse-splay deposits. The grain size contrast with the surrounding mudrocks, the sharp erosive bases, locally limited lateral extent and the poor sorting are all indicative of rapid, short lived pulses of sedimentation, the sort commonly associated with discharge into an interdistributary area via a crevasse channel (Arndorfer 1973; Elliott 1974, 1986a). This would result in the deposition of a lobe of sand. The colonisation of a lobe in the Praia Azul section by a low diversity fauna probably indicates brackish conditions in the receiving bay area.

Enclosed carbonate pebbles, probably of pedogenic origin are probably either locally derived in the case of splays in continental parts of the section, or have been eroded and transported from subaerial levees into the interdistributary bays where sand beds are found within marine facies.

The thick coarse unit at the top of the Santa Rita section has limited lateral extent and was probably deposited close to the crevasse channel. The cyclicity observed suggests episodic sediment supply.

The bed overlying the biomicrite is quite striking, particularly as its deposition may have extinguished a variety of fauna. It is possible that this bed records the breaching of a levee following a period of time during which clastic input had not been great.

4.3.4 Laminated sand and siltstone facies.

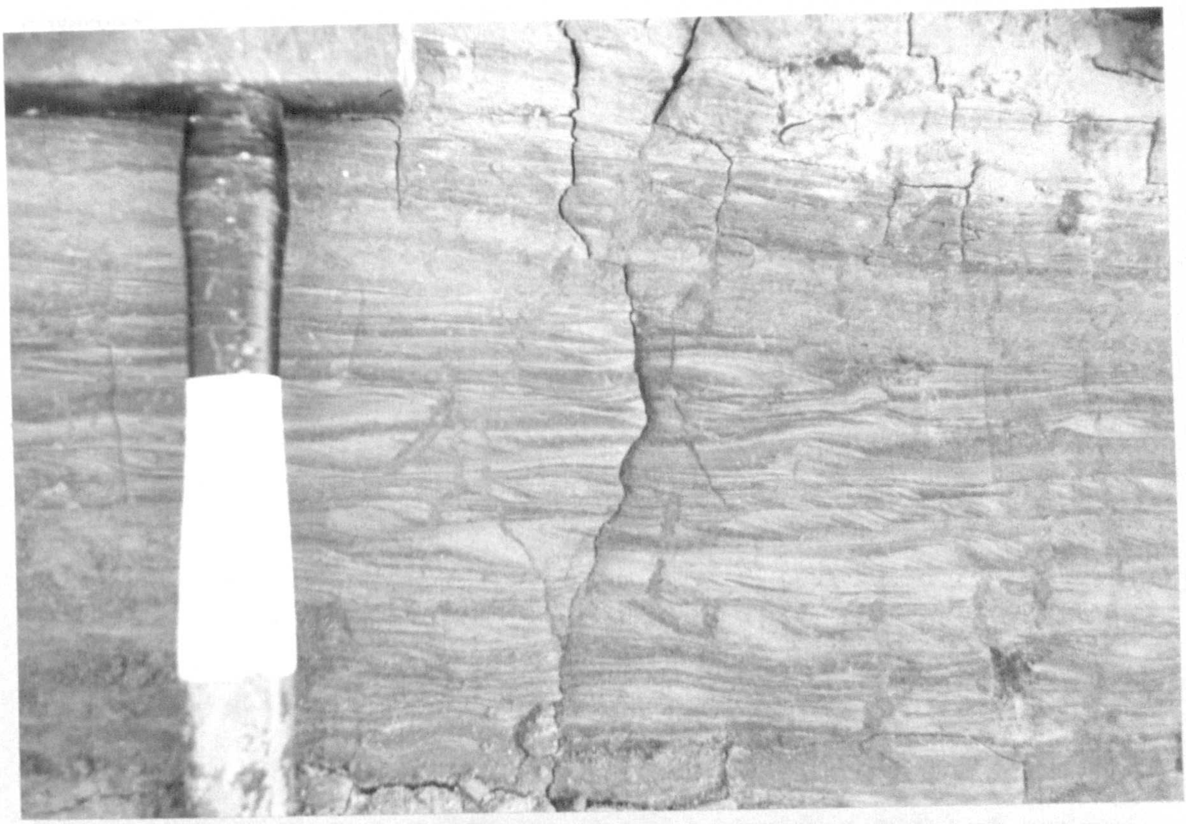


Figure 4.9 Laminated facies outcrop beneath the first major sand body in the Santa Rita section. High rates of sediment fallout are indicated by the presence of both type 'A' and 'B' climbing ripple laminae (Ashley et al. 1982). these are cross-cut by a few invertebrate burrows. This particular example of the laminated facies is interpreted as distributary mouth bar deposits. Scale bars = 0.1m.

This is a common and highly variable facies in terms of sediment type, structures present and also its position relative to other facies both marine and terrestrial. The presence of primary depositional structures is usually associated with an absence of body fossils. A single exception to this exists in the Praia Azul section where a sparse bivalve fauna appears in parallel laminated silts. This facies is best exposed in the Santa Rita section where it occurs in outcrops of both the Praia Azul and Assenta members.

Normally, this facies consists of finely laminated to decimetre interbedded fine sands and siltstones with a variety of structures. Ripple and parallel laminae are ubiquitous and climbing ripple laminae are also common (Figure 4.9). A number of examples are described in Table 4.3 to illustrate the range of structures and associations observed.

Laminated facies are often found interbedded with fossil rich silts. They consist of current ripple laminated very-fine sandstones and siltstones in packets about 0.2-1m thick, interbedded with units of structureless mudrock some with a fossil fauna and others with fauna absent. Often where fauna is absent mudrock is red-brown or purple coloured and locally contains small irregular carbonate nodules.

Graded beds are common, often occuring in packets of 10-30cm beds up to a metre thick. Examples of this include brown and green mottled decimetre scale beds exhibiting a transition from very fine parallel laminated sands up to climbing ripple laminated silts. Others exhibit predominantly ripple laminae and there is a degree of scouring between beds in places.

Inverse grading appears in the laminated façies at several levels in both sections. Rather than affecting a single bed, coarsening-up often expresses itself as an increase in the proportion of sandy laminae over silt. The setting of these units which are 1-3m thick vary, but they commonly represent a transition from marine facies to terrestrial (see Table 4.3). Parallel and ripple laminae are common, as are climbing ripple laminae. Laminae are often broken and have upturned ends resembling dish structures (Lowe 1975). Locally there are severely disrupted

LOCATION	ASSOCIATION	DESCRIPTION	INTERPRETATION
Base Praia Azul section	lies above terrestrial mudrocks containing carbonate nodules, and below small lenticular channel sand body or horizontally bedded sandstone facies	red silts and buff fine sandstones; current ripple and parallel laminated, enclosing pebble grade mudclasts; local horizons of convoluted laminae and in places load casts and mud flames with a relief of <0.3m	transgressive deposits; flood derived sediment deposited at the margins of interdistributary bays and in front of crevasse channels
Top Praia Azul member Santa Rita section	lies above marine facies and below erosive base of thick sheet sand body (interpreted as a fluvial distributary channel deposit)	grey silts and fine sandstones; overall inverse grading over 3m from dominantly silt to dominantly sand; current ripple, parallel and climbing ripple laminae present (Figure 4.10); minor bioturbation	distributary mouth bar
Various	bracketed by marine facies, or with marine facies below and terrestrial indicators, such as reddened silts containing (pedogenic) carbonate nodules, above	parallel and current ripple laminated silts and fine sandstones; unfossiliferous and often reddened	interdistributary bay and bay margin sediments, supplied with sediment by overbank and crevasse channel flooding
Santa Rita section	above small lenticular (channel) sand bodies	parallel and ripple laminated facies with associated decimetre-scale lenticular sand beds with concave-up bases	crevasse channel levee deposits
Santa Rita and Praia Azul section	below lenticular channel sand bodies	laminated facies with widely fluctuating grain sizes and structures; some carbonate nodules and severe soft sediment deformation (Figure 4.11)	interdistributary bay marginal sediments introduced via a crevasse channel

Table 4.3 Examples of the Laminated facies variability; Praia Azul and Santa Rita sections.

vertical pipes of sediment. In places wavy and lenticular bedding with isolated ripples or ripple trains draped with silt appear.

Laminated facies also lie above some of the small channel sand bodies particularly in the Santa Rita section and may be interbedded with thin lenticular beds of fine sandstone exhibiting similar structures. Commonly the sands have curved concave-up bases and planar tops. They occasionally have loaded bases and may be burrowed.

One small channel sand body has a particularly unusual laminated facies about 1m thick developed beneath it. The base of the laminated unit is sharp and highly irregular due to loading into massive dark green silts beneath. As a result the lower part is severely convoluted. It is also burrowed. Within this part of the unit there is also a single isolated 0.15m thick lenticle of cross bedded granulestone rich in pebbles of caliche. The upper parts of the unit are finer, consisting of very fine sand and silt with occasional granule rich laminae. Rather undulose parallel laminae are the dominant structure plus lenticles of fine yellow sandstone and wood fragments. This is disrupted by a 0.2m wide and 0.5 m deep pipe of disturbed sediment (Figure 4.10). Laminae at the margins of this structure are downbent.

Laminated facies in the upper, dominantly continental (Assenta mbr.) part of the Santa Rita section, include considerable thicknesses of red-brown silts and thin sands which are parallel, ripple and occasionally climbing ripple laminated. These are locally ruptured or convoluted and interbedded with significant thicknesses of massive structureless silts. Parallel laminae in particular dominate several metres of dark grey or mottled silt rich in plant debris, plus occasional logs. These are intercalated with thin structureless or ripple laminated fine sands which may be graded. Bed bases may be loaded or even convoluted. Carbonate soil profiles also punctuate the section. Burrows appear within this facies, generally simple unlined tubes either 1-2mm or about 10mm in diameter most often sub-vertical. The best

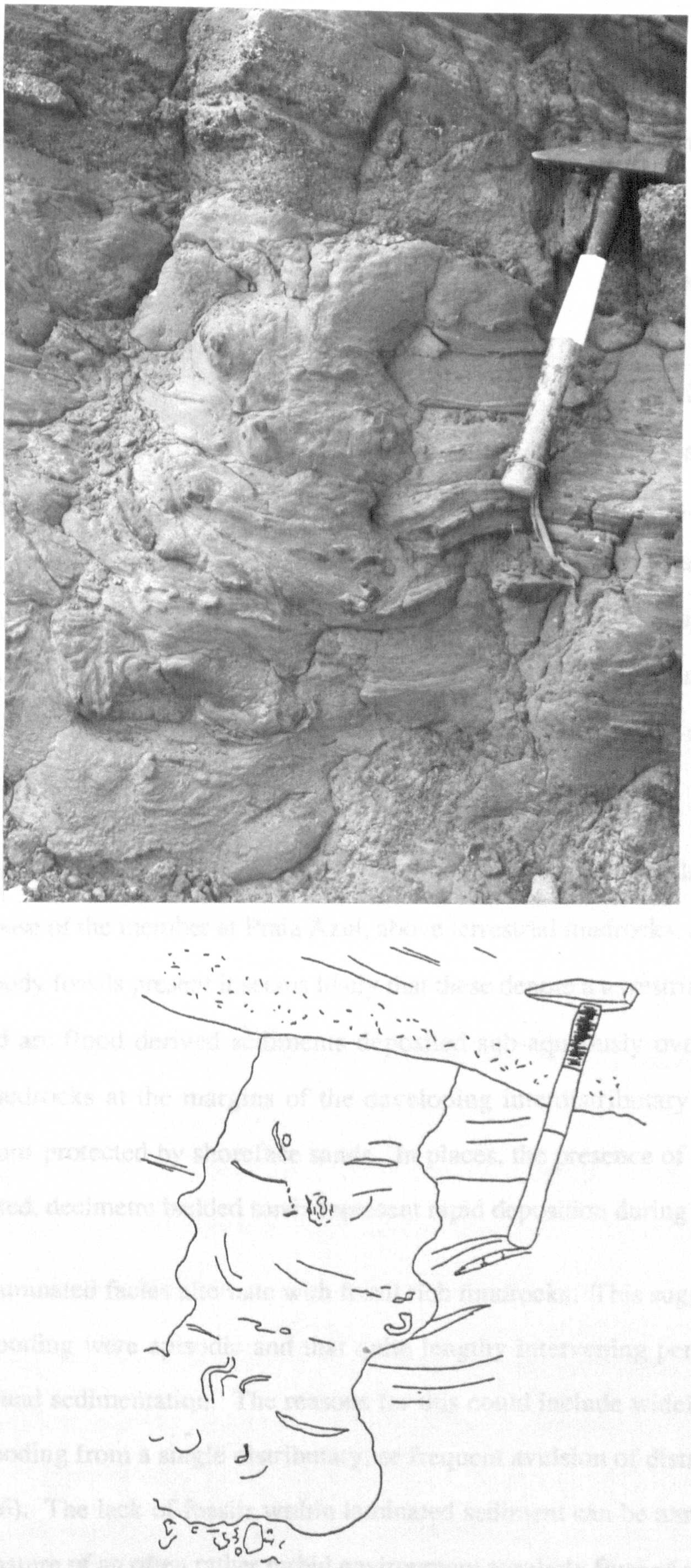


Figure 4.10 Photograph and sketch of a pipe of disturbed laminated sediment beneath a lenticular channel sand body; Santa Rita section. Such disruption was probably caused by water escape, perhaps a consequence of rapid sedimentation or compaction. Scale bars on hammer = 0.1m.

examples appear in the Santa Rita section. *Diplocraterion* is the most common trace fossil in the lower (P. Azul mbr.) part of the section.

Interpretation : The range of structures and associations observed suggests that the laminated facies represents more than one depositional setting. The most common origin of this facies is thought to be as flood deposits, either resulting from widespread overtopping of levees where these do not have great relief, resulting in the thinly laminated facies, or in the case of packets of graded beds, as crevasse splay deposits. Finely laminated facies are recorded in both modern and ancient deltaic successions in interdistributary bay settings (Coleman and Gagliano 1965; Elliott 1974a). The laminated facies seems to occur in four broad settings: a) at terrestrial-marine upward transitions; b) within marine facies; c) at marine-terrestrial upward transitions; and d) within terrestrial facies.

Terrestrial-marine transitions are represented by the development of the laminated facies at the base of the member at Praia Azul, above terrestrial mudrocks. Although there are no body fossils present it seems likely that these denote a terrestrial-marine transition and are flood derived sediments deposited sub-aqueously over former floodplain mudrocks at the margins of the developing interdistributary bays, or shallow lagoons protected by shoreface sands. In places, the presence of climbing ripple laminated, decimetre bedded sands represent rapid deposition during floods.

Quite often laminated facies alternate with fossil rich mudrocks. This suggests that periods of flooding were episodic and that quite lengthy intervening periods saw only background sedimentation. The reasons for this could include widely spaced periods of flooding from a single distributary, or frequent avulsion of distributaries (see sect.4.3.6). The lack of fossils within laminated sediment can be attributed to the stressful nature of an often rather turbid environment regularly flushed with fresh to brackish water. The laminated facies appearing in packets within the fossil rich silts represent pulses of flood generated sediment transported into the interdistributary bay environment. The widespread nature of these packets of sediment suggests that they do not always result from localised crevassing, but rather are overbank sheet floods. Such sequences were termed 'phase 1' by Elliott (1974a) and represent an early stage in the development of bay fills where distributary channels have not yet built up sufficient levees to contain flow at high stages. In Elliott's scheme these units form the base of coarsening up sequences which pass up into crevasse splay, subaerial levee and channel facies. In the Praia Azul section, transitions from laminated facies to crevasse splays to red mudrocks with pedogenic nodules represent localised emergence due to silting-up of the bay.

Laminated facies were often deposited in very shallow water less than one metre deep, as they appear at marine/terrestrial transitions passing up into purple or red-brown silts, the latter often with small (<10mm diameter) irregular carbonate nodules. Like the 'transgressive' laminated sediments these probably represent bay margin/ channel-levee deposits. Locally these form graded beds up to 1m thick and may be the result of deposition from a large flood causing rapid emergence of a shallow bay margin area.

A particular type of marine-terrestrial transition is envisaged for some of the inversely graded units. These often appear beneath the channel sand bodies and are probably channel mouth bar deposits. Significantly it is these units which show load casts and water escape features such as the disturbed pipes of sediment, indicating a water saturated substrate and common occurrences of climbing ripple laminae indicating periodically rapid rates of deposition (Figure 4.9).

The unusual example of this facies, in the Santa Rita section below a lenticular sand body, is thought to represent bay marginal sediments introduced via a crevasse channel, the latter represented by the overlying unit. This is supported by the widely fluctuating grain sizes observed, indicating considerable fluctuations in discharge and sediment supply and the presence of 'pipes' of deformed sediment which probably indicate liquefaction and water escape.

Laminated sediments found above some of the small channel sand bodies in the section are probably levee deposits, with the decimetre-scale lenticular sands representing small crevasses. In the dominantly continental (Assenta mbr.) part of the Santa Rita section, laminated facies are also probably the deposits of levees. Subaerial exposure has resulted in a deep red colouration and the destruction of structures by pedogenesis. Bioturbation may often be the result of plant roots. In contrast, extensive areas of parallel laminated sediment with occasional convolutions and water escape ruptures suggest a subaqueous origin. As the sequence is effectively wholly terrestrial with no fauna present, these units are thought to represent shallow lacustrine environments subject to periodic incursions during flooding. The presence of interbedded soil profiles indicate sub-aerial exposure, perhaps due to infilling of lacustrine depressions.

4.3.5 Massive mudrock facies

Mudrocks dominate both sections, forming significant proportions of both terrestrial and marine facies. They consist of silts to silty claystones and rare claystones which are often micaceous. Colours vary from dark greys and greens through to reds, browns and purples, the latter in particular associated with carbonate soil profiles. Lignitic debris is common and in the upper part of the Santa Rita section this is often associated with patches of bright yellow elemental sulphur. Gypsum also appears in the same mudrocks, consisting of small glassy acicular crystals in the joints that give the silts a blocky texture. Elsewhere in the section, fractures and joints associated with faults also contain gypsum in coarse fibrous bundles.

Massive silts are also interbedded with packets of laminated sediment. Where silts are dominantly red and brown in colour, decimetre scale, interbedded sands are commonly structureless with the exception of burrows and/or diffuse accumulations of carbonate nodules. At one level in the Santa Rita section, carbonate is preferentially concentrated within interbedded sands.

Interpretation : In the marine parts of the section, massive mudrocks probably represent fine grained sedimentation from suspension, occuring in the interdistributary bays during floods. In modern environments such as the Mississippi a lack of structures is attributed to either, bioturbation or a lack of visible textural contrast (Coleman et al. 1964; Coleman and Gagliano 1964; Donaldson et al. 1970).

In the terrestrial parts of the section massive mudrocks probably represent a number of depositional environments. Where red, brown and purple they are often interbedded with structureless or burrowed red sandstones. These are most likely levee or sub aerial, delta top / floodplain environments, and their structureless nature is probably attributable to bioturbation by rootlets and other pedogenic processes. The carbonate nodules in these silts are probably pedogenic accumulations equating with the 'stage III' calcretes of Gile et al. (1966) and Machette (1985).

The silts in the upper half of the Assenta member exposed at P. da Santa Rita are most often dark grey or green and rich in plant debris. The ochreous colouration and preserved plant debris indicate that reducing conditions were normal in the sediment. Therefore, these silts most likely represent delta top/floodplain marshland. They are quite often interbedded with packets of parallel laminated sediment, interpreted as lacustrine deposits, indicating that marsh and lacustrine environments were closely associated. The gypsum found within the grey silts is thought to be a late diagenetic phase because of its euhedral form and confinement to joints and fractures. The section is located on the flanks of the Vimeiro diapiric structure and remobilised Triassic evaporites may be the source of the gypsum.

4.3.6 Massive sandstone facies and facies associations

Large sand bodies of variable scale and geometry crop out at both of the major sections. All are characterised by erosive bases, which are often lined by pebble and cobble grade mudclasts. Many of the suites of structures are the same as those described in Chapter 3 for the Porto Novo member. In the light of this, the following description and interpretation of structures will be minimal, most attention being concentrated on the unusual features observed.

Figure 4.11 Fining-upward cyclicity, of variable scale, in major sand body at Praia da Santa Rita. For full details see text. One cycle begins about 20cm above the hammer (scale bars = 0.1m) and is 20cm thick. A second continues beyond the field of view.

In the Praia Azul member, sand bodies are 1-3m thick and commonly have a lenticular cross sectional geometry, with concave-up erosive bases. The lateral extent of these units is usually restricted to a few tens of metres. In contrast the

Assenta member outcrop at Praia da Santa Rita is characterised by sheet sand bodies up to 8m thick, plus a number of smaller lenticular sand bodies.

A wide range of sub-facies are developed within the sand bodies. Matrix supported mudclast and caliche pebble conglomerates, locally containing lignitic debris, often appear near sand body bases. Trough cross bedded and current lineated fine and medium sandstones are common, particularly in the Assenta member sand bodies. Cross bed sets are usually 0.1-0.2m thick, up to a metre wide and 2-3m axial length. Sand beds often exhibit rapid grading in their upper few centimetres and linguoid ripple crests or invertebrate burrows are occasionally preserved on bedding planes. Sand-bed bases often cut down into those below through the intervening silt drape.

Decimetre to metre scale graded beds of fine sands to silts, often ripple or planar laminated are also common (Figure 4.11). Locally, outcrop may be dominated by a marked fining-upward cyclicity in stacked-up laminated beds arranged in crude thinning-up sequences. The beds illustrated in Figure 4.11 have only mildly erosive bases. They consist of fine curent-ripple to 'type B' (stoss side preserved) climbing ripple laminated sandstones (Ashley et al. 1982), which pass up in their top few centimetres to grey parallel laminated silt drapes. Mud drapes are usually ripple laminated, though laminae are often convoluted due to loading by subsequently deposited sands. They have often been reworked, forming mud-pebble lags. Lignitic debris is locally concentrated along laminae. Water escape structures in places rupture thinner sand beds.

A number of sand bodies in the Santa Rita section have structureless upper parts due to the development of carbonate soil profiles. This commonly takes the form of a dense nodular crust of carbonate. In one example well preserved rhizocretions have weathered out (Figure 4.12). Sand body tops also frequently display invertebrate burrows, either as weathered out circular tubes approximately 10mm diameter or as straight to sinuous horizontal trails of similar diameter, a number of examples bearing spreite.

Sand bodies occur at four levels in the Praia Azul section, at its base and around 20m, 47m and 57m. Data on facies beneath and above channels are presented as this will better facilitate later discussion of their environmental context. This appears to be different for channels at different levels in the section.

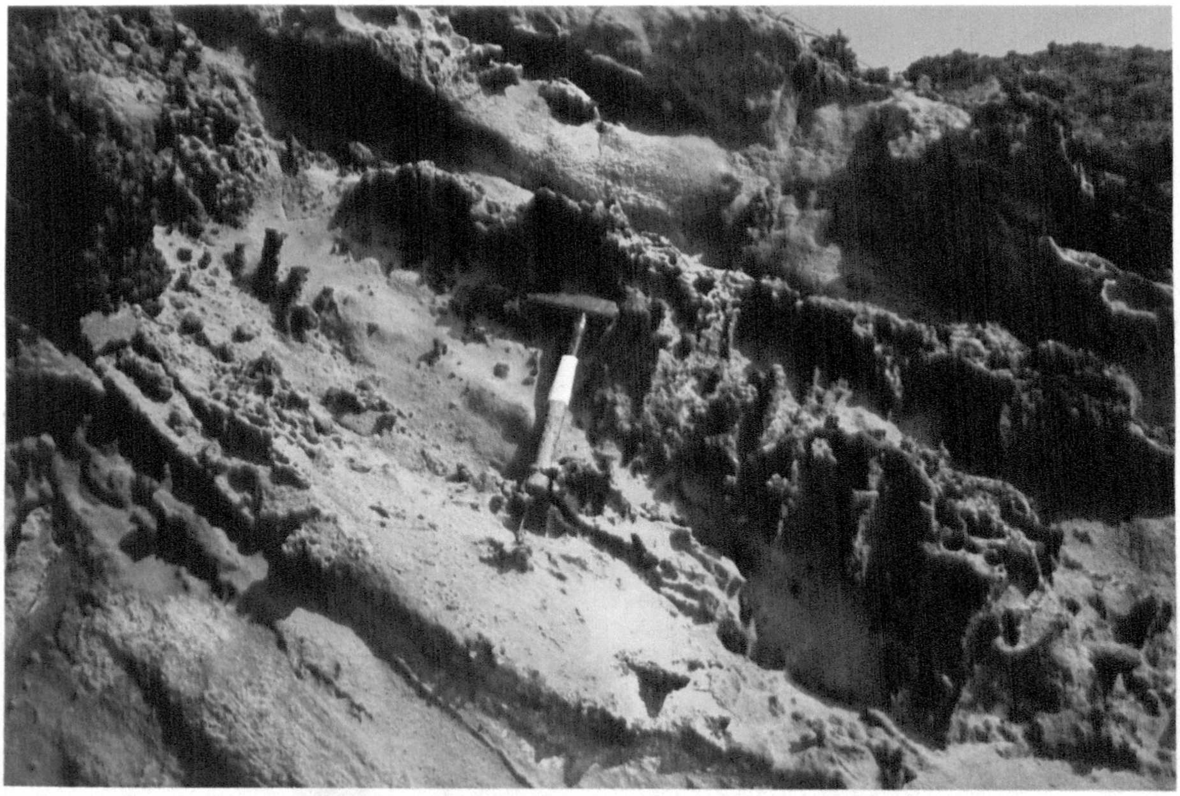


Figure 4.12 Carbonate rhizocretions in the upper part of a channel sand body; Praia da Santa Rita. Hammer 0.4m long.

The only sand body observed at the base of the section lies at the terrestrial - marine transition and has beneath it 1.5m of laminated facies. It has a lenticular "winged" cross section with a thickness of 1.75m and width of only 10m. Structures are poorly preserved. Bioturbated horizontally bedded sandstone facies, interbedded with grey silts, lie above the sand body. The wings, which extend for considerable distances are severely deformed, structureless fine sandstones with loaded bases. The sand body at 47m has a similar morphology and setting.

Sandbodies at the 20m-level in the section are the most prominent, appearing at the cliff base half-way along Praia Azul and extending northwards to the cliff top about

100m south of Alto da Vela (Figure 4.1). They appear above a horizon of red-brown mudrocks. At least seven sand bodies appear at this level, each having variable lateral extent and a lenticular, to sheet-like cross section. At the cliff top one example is interesting because of the sequence directly beneath it (Figures 4.8, 4.13).

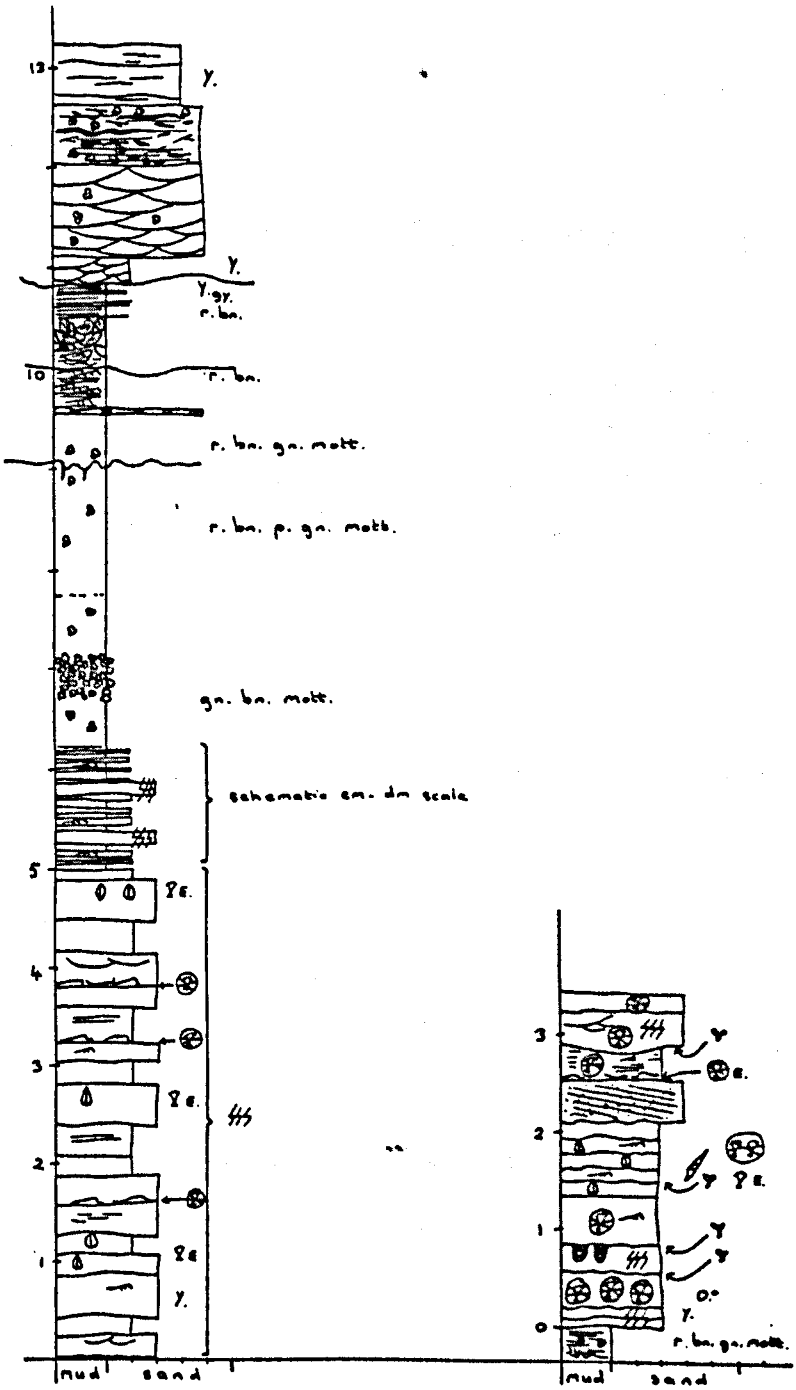


Figure 4.13 Graphic log of the succession illustrated in Figure 4.8. For details see text. Vertical scale in metres.

Horizontally bedded sands, with an *E. securiformis* fauna, near the base of the P. Azul member, pass up into 1.2m of centimetre bedded fine sands and silts with ripple laminae or intense bioturbation. These grade into massive mottled silts with 5mm diameter carbonate nodules and, just below an erosion surface, mudcracks with a sandy fill. Above this erosion surface lie 0.5m of calcareous silts containing a few small clasts of caliche. These are overlain by a thin coarse sand, rich in pebble grade clasts of mud and caliche, and 0.5m of parallel, ripple and climbing ripple laminated siltstones. The erosive base of the sand body cuts into this unit and the lower half metre of sediment consists of a siltstone breccia containing blocks up to 150mm square surrounded by smaller clasts and deformed stringers of sand. It is succeeded by 0.3m of parallel laminated silts and very-fine sands above which the fill is dominated by sand. The top of the outcrop is lost to erosion.

A further sand body at the 20m level crops out half-way up the cliff at the northern end of Praia Azul (Figures 4.1, 4.2). Its maximum thickness is some 3m and it has a lateral extent of some 50m. The geometry of the sand body therefore falls close to the lenticular/sheet boundary as defined by Friend et al. (1979). A fault cuts the southern limb of the sand body. This is the small reverse fault appearing in Figure 4.1. Unfortunately it was not possible to make a detailed study of the sand body due to problems with accessibility of outcrop. However, most sedimentary units within it lie on low-angle planes inclined to the south. This sand body also lies above red-brown silts containing irregular carbonate concretions up to 10mm in diameter, no laminated facies occurs directly beneath it. Also, fine sediment at the margin of the sand body are reddened, burrowed and contain carbonate nodules. Thin parallel laminated, or burrowed fine sands at the top of the sand body are overlain by 2m of red-brown silts with carbonate nodules, above which marine sediments reappear.

In the Santa Rita section, the smaller-scale sand bodies are largely restricted to the basal 60m (P. Azul mbr.) of the outcrop. They consist of trough cross bedded coarse to fine yellow sandstones with horizons of convoluted fine sands, ripple laminated fine sands and ripple and parallel laminated silts. In the dominantly marine part of the section most channels lie above two to three metres of the laminated facies, which may, or may not, exhibit a coarsening up. Frequently the laminated sediments show signs of scouring, load structures, convolutions and water escape structures. Lags of caliche pebbles and mudclasts are common and so too are intensely burrowed horizons within the sand body. Partially eroded mud drapes and lignitic debris also occur. Commonly, the sand body margin will be gently tapered with a concave upward base passing laterally into wings of rather structureless fine sands which are burrowed and exhibit the development of carbonate nodules. One example near the base of the section has a terraced margin with a series of scours cutting downward across older deposits. Another, in the upper part of the section, cuts into fine sands which appear to have an early fine carbonate cement. Blocks of the carbonate cemented lithology appear as clasts in the upper part of the sand body.

The large-scale sand bodies belonging to the Assenta member, comprise a variety of sheet-like sand bodies 2-8m thick. A few of these are likely to be multi-storey, though with the exception of those at 155-165m (Figure 4.3) it is difficult to be certain due to the limited lateral extent of the outcrop. Palaeocurrent data gathered from cross beds within these sand bodies are presented on the section map (Figure 1.23). There is little data available from any of the other facies developed.

A number of the more interesting features are described below. A more comprehensive description of Assenta member sand bodies, from the proposed type section south of Praia Azul, appears in Chapter 5.

The first of these major sand bodies appears between 62 and 70m, above 2m of laminated facies exhibiting coarsening-up. It marks the main marine-terrestrial transition and, therefore, the boundary between the Praia Azul and Assenta members. The sand body has a sheet-like geometry with a lateral extent of 80m exposed in the cliff section. In addition it can be traced to a wave cut platform at exceptionally low tide thus proving a minimum lateral extent of some 200m. In all probability the lateral extent is considerably in excess of this. It exhibits well defined, low-angle lateral accretion surfaces, which are inclined up the structural dip of the section (Figure 4.14). These are picked out by silt drapes which separate sand beds 0.5-1m thick. The sand body fines-up, both vertically through the section and laterally along the inclined accretion surfaces.

Figure 4.14 Eight metre thick channel sand body exhibiting lateral accretion surfaces which are inclined up-dip. Palaeohorizontal marked by the thin, laterally persistent beds below (to the left). these are marine shell beds. Laminated facies ca. 2m thick occupy the space between the uppermost of the marine beds and the sand body which is the first major unit in the Santa Rita section, and marks the base of the Assenta member.

In the upper parts of the sand body dips on the lateral accretion surfaces flatten out. Sediments are dominated by red-brown and green mottled silts and very fine sands which are dominantly horizontal, or low angle parallel laminated, forming thin coarsening up cycles. These are burrowed; both horizontal and vertical burrows are developed with the intensity of bioturbation increasing upwards. Two sizes of horizontal burrow appear, one 3mm and the other 6mm in diameter. Individual units occasionally have scoured bases filling small concave-up channels in the surface of the beds below.

The sand body is overlain by massive silts interbedded with centimetre to decimetre beds of ripple laminated fine sands. The silts are either rich in plant debris or have a diffuse development of carbonate nodules. The thicker sand beds often have loaded bases and enclose both nodules and plant debris. It is only this first major sand body in the Santa Rita section which has laminated facies beneath it. The laminated facies is absent beneath all sand bodies higher in the section.

The penultimate sand body in the Santa Rita section is the multi storey unit between 155 and 165m (Figure 4.3). The top 3.5m consists of a graded unit dominated by westerly directed palaeocurrents. The lower unit is also graded from a caliche pebble conglomerate up to parallel laminated silts over a thickness of some 5.9m. At its coarsest, pebbles of caliche reach 7cm in diameter at the channel base. The outcrop exhibits lateral accretion surfaces which consist of irregular, low angle erosion surfaces cutting across much of the sandbody. Fining grain size trends can be traced both laterally along these units and vertically across them.

The base of the sand body is underlain by brecciated siltstone in a fine sand matrix. This lies above massive and ripple laminated purple and green mottled silts with two thin blocky carbonate horizons. The lower part of the sand body consists of sigmoidal beds up to 0.5m thick at their centre and dominated by parallel, rather undulose laminae. Individual sets may exhibit fining upwards from caliche pebble conglomerate to coarse or medium yellow sandstone, others show little grainsize contrast. All have numerous internal erosion surfaces draped with finely comminuted plant debris. Plant debris is particularly common at the base of the sand body which in places is discoloured by bright yellow elemental sulphur. Palaeocurrents have a SSE orientation compared to a SW dip on the lateral accretion surfaces.

The last sand body in the Santa Rita section, below the unconformable contact with the Santa Rita member, lies about 1m above the multi storey unit. They are separated by poorly exposed dark grey, ripple laminated silts rich in fine plant debris and rather sulphurous. The erosive base is gently undulose and overlain by 0.8m of ripple laminated medium sands also rich in plant debris. The remainder of the sand body, above a further erosion surface, exhibits a fining upwards trend from very coarse to medium sandstone. Large scale trough cross beds and bundles of planar laminae separated by low angle erosion surfaces dominate the lower and middle parts of the sand body which is 5.5m thick in total.

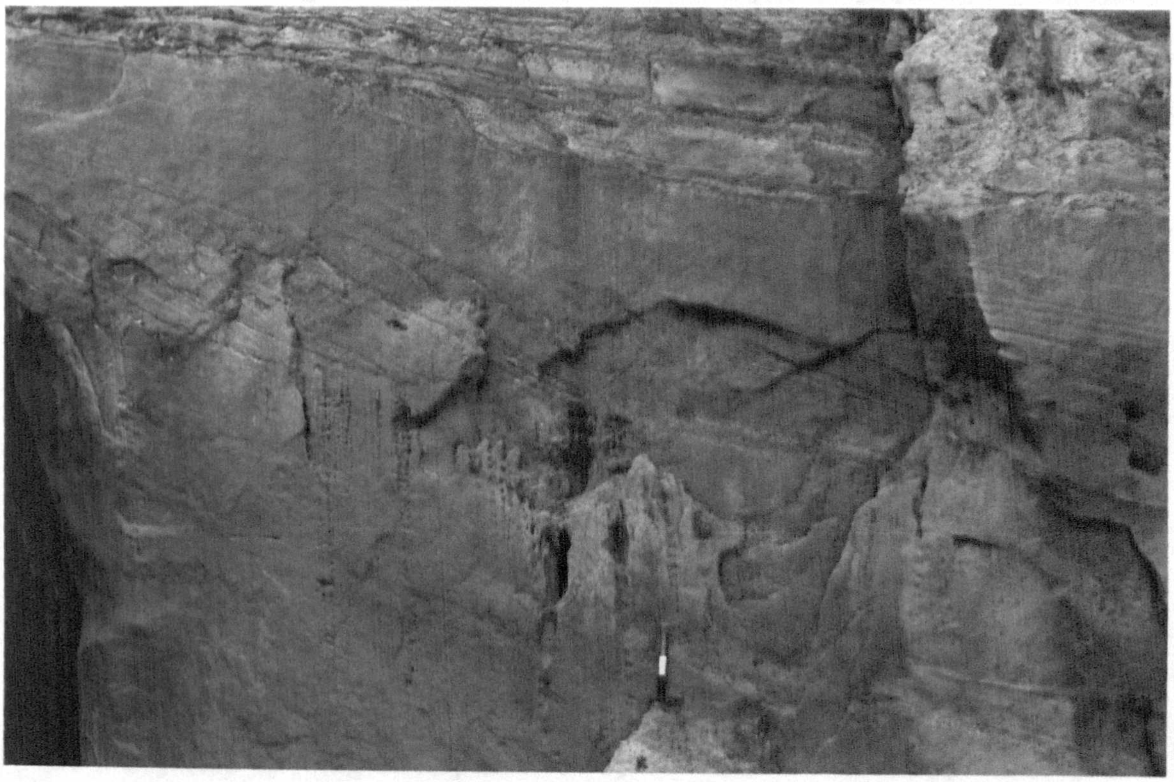


Figure 4.15 Sigmoidal cross bed at the top of the last sand body Assenta member (Santa Rita sectiôn). Thin drapes of mud and lignitic debris suggest discharge fluctuations. The bedform shape suggests high velocity flows, having similarities with the hump-back bars of Allen (1983). Its elevated position suggests that it may have been the deposit of a chute bar (see text). Hammer 0.4m long.

Towards the top of the sand body a number of scours have a varied fill including ripple laminae, inclined parallel laminae draping scour margins and planar cross beds. The top metre of the sand body is dominated by a single tabular cross bed with markedly sigmoidal foresets (Figure 4.15). The foresets are orientated towards the west and in this direction the sets progressively flatten out into planar laminae. Individual foreset laminae frequently have drapes of finely comminuted plant debris. There are also a few decimetre-scale blocky mudclasts. A rapid upwards transition to ripple laminated, sulphur rich, grey silts interbedded with occasional decimetre-scale structureless yellow fine sands with loaded bases occurs.

Interpretation : Massive sand bodies are all thought to be the deposits of channels. Features they have in common which are consistent with this interpretation are their erosive bases, coarse fill and unidirectional palaeocurrents. The specific depositional settings of different sand bodies are probably quite varied. The following interpretations have been based on criteria such as sand body scale, geometry and associated facies.

1) Large-scale sand bodies within the Assenta member. These have many characteristics in common with channel sand bodies in the Porto Novo member including lateral accretion surfaces (Figure 4.14), graded beds and mud drapes (Figure 4.11), plant and caliche debris, and structures indicating rapid deposition such as climbing ripple laminae. Overall the sandy facies are finer grained than those of the Porto Novo member.

The position of the continental Assenta member facies above dominantly marine Praia Azul member facies suggests a close relationship. The large scale sand bodies are interpreted as being the deposits of major trunk rivers in an upper delta plain setting. The presence of numerous graded beds and abundance of mud drapes are indicative of considerable fluctuations in discharge, a characteristic shared by most other members of the Lourinhã formation. This suggests that there is a strong overriding climatic control as discussed in Chapter 2. However, it must be borne in mind that the distal reaches of modern fluvial systems close to marine basins, characteristically display sand/mud couplets (Smith 1988). The major control postulated by Smith was still seasonal discharge fluctuations but with some influence from tidal fluctuations presumably arresting stream velocities in the fluvial system. Even in a microtidal regime some effect on the fluvial system can be expected over measurable distances, particularly in the light of significant discharge fluctuations. This is the case in the Orinoco Delta (Van Andel 1967), where a tidal range of about 2m contrasts with a range of river stage of about 6m. At low stages tides make themselves felt up to 150km inland from the river mouth. Conversely, at high stages the entire lower delta plain is inundated by fresh water.

During major floods, quite rapid rates of deposition were achieved. This is suggested by the significant thicknesses (<1.5m) of the graded units within a number of the channel sand bodies (Figures 4.11 and 4.14). The sheet-like geometry and the lateral accretion surfaces observed within sand bodies, indicates that they migrated laterally. However, palaeocurrent data suggests that the channels were not highly sinuous. Low or moderate sinuosity is a characteristic of channel systems in modern fluvially dominated deltas (Kanes 1970; Bagaz et al. 1975; Galloway 1975).

The sediments beneath the first channel in the Assenta member at P. da Santa Rita, suggests strongly that the sand body is the deposit of a major distributary. It lies only 2m above a biomicrite (Figure 4.14), the intervening space occupied largely by the coarsening-up mouth bar facies. It is interesting to note the relative thickness of channel and mouth bar (8m and <2m respectively). The channel is clearly deeply incised into the delta top, a characteristic of many recent shoal-water deltas (e.g. Mississippi (Gould 1970) and Guadalupe (Donaldson et al. 1970)). This also provides supportive evidence for the shallow nature of the receiving basin.

Most of the remaining sand bodies in the section above have purely continental affinities and were probably deposited by trunk rivers near the apex of the delta, or actually traversed the upper delta plain. The dominance of parallel or low angle lamination in a number of these is quite striking and is a feature of the Assenta member in its type locality (see Chapter 5). It suggests that at least periodically, sustained high velocity flows (of the order of 0.6-0.8m sec^{-1}, Harms et al. 1982) occurred.

Large volumes of detrital caliche in the multi-storey sand body are indicative of the reworking of pedogenically altered sediments. The last channel in the section also has features indicative of fluctuations in discharge, notably the scours with variable fill. The sigmoidal cross-bed at the top of the channel fill has numerous reactivation surfaces and drapes of lignitic debris. In view of its position and structures it is probably the deposit of a chute bar (e.g. Levey 1978) cutting across a meander loop. It would be subject to periodic high velocity flows and consequent episodic and rapid sedimentation. The sigmoidal profile and flattening out of foresets both laterally and vertically could be a consequence of conditions similar to those proposed for Allen's (1983) hump-back bars. The flattening of foresets could be the result of high rates of aggradation, and the sigmoidal profile the product of the combination of aggradation plus high velocities supressing the bar form. Levey (1978) suggested that there may be a link between flashy discharge patterns and the development of chute bars.

2) Small-scale sand bodies in the Santa Rita section: A rather different origin is anticipated for these sand bodies. They all have a lenticular 'winged' cross section, are less than 2m thick and few are wider than 10m. In the lower 60m of the section they all have beneath them laminated (mouth bar) sediments. The laminated sediments often exhibit soft sediment deformation and water escape structures which suggest sub-aqueous deposition (see 4.3.4). In contrast, the upper parts of the sand bodies exhibit signs of sub-aerial exposure and pedogenesis. In the light of the associations outlined, and the small scale of the sand bodies relative to the major distributary channel deposits, these units are thought to be the deposits of crevasse channels (e.g. Arndorfer 1973; Elliott 1974a). The laminated facies beneath are interpreted as minor mouth bars (sensu Coleman et al. 1964; Elliott 1974a; Fielding 1987), and the reddened laminated facies above as crevasse channel levees.

In modern settings, small crevasse channels form when the levee of a major distributary is breached during a flood (Arndorfer 1973). The resulting channel may be deeply incised into the levee, and permanent or semi-permanent discharge established. Large volumes of sediment are swept into interdistributary bays at high stages. The lenticular geometries observed are characteristic of crevasse channels (Visher et al. 1975).

3) Channel sand bodies in the Praia Azul section: Channel sand bodies in this section also fall into two groups. Those at the base of the section and at 47m (Figure 4.2) both have geometries and associations similar to the small crevasse channels in the Santa Rita section. The two examples in the Praia Azul section therefore are interpreted in the same manner.

The remaining sand bodies are interpreted in a different manner. Their thicknesses (ca. 3m) are intermediate between the crevasse channel sand bodies and those of the trunk distributaries of the Assenta member. The sand bodies have lenticular to sheet-like geometries of limited lateral extent (Figures 4.1 and 4.7), and there is evidence in the presence of lateral accretion units, for a limited tendency to migrate laterally. The characteristics of distributaries in lower delta plain settings are a tendency to bifurcate, and also avulse frequently (e.g. the Guadalupe delta, Donaldson et al. 1970). This results in a pattern of numerous channel sand bodies, smaller in scale and lateral extent compared to the trunk distributaries of the upper delta plain (Horne et al. 1978). This is the pattern observed at Praia Azul which, in addition, has a stronger marine signature than the Santa Rita section (compare Figures 4.2 and 4.3). In the light of this the channel sand bodies are interpreted as the deposits of a distributary network in a lower delta plain setting.

There is a problem with the above interpretation because not all the sand bodies lie above marine/mouth bar facies. One example described lay above red-brown mudrocks containing carbonate nodules thought to be of pedogenic origin. It is possible that this outcrop preserves the deposits of a channel, locally incised into delta top mudrocks, which passed laterally into marine strata. The other channel sand body described from this section had a thick mudclast conglomerate at its base. This may be the product of erosion of the channel's mouth bar, the deposits of which lie beneath.

4.4 THE NORTHERN SECTIONS

4.4.1 Introduction

Marine strata related to the transgression at the base of the P. Azul member reappear north of Porto Dinheiro (Figure 1.8). They consist of 3-4 thin, laterally persistent tongues of shelly sediment 0.4-7m thick. One example can be traced for 3.5km (Figure 4.16). They are interbedded with reddened or mottled mudrocks and fluvial channel sands.

Figure 4.16 Laterally persistent thin tongue of marine strata extends for 3.5km northwards from Porto Dinheiro. The marine unit begins ca. 2m below the lower of the two prominent beds, and ends ca.1m above the upper bed. It is about 7m thick at this point (see Figure 4.18). A further marine horizon occurs 29m above (arrowed on cliff face in middle distance). Channel sand bodies of the Porto Novo member crop out in the cliff face below, and also in strata intervening between the marine horizons.

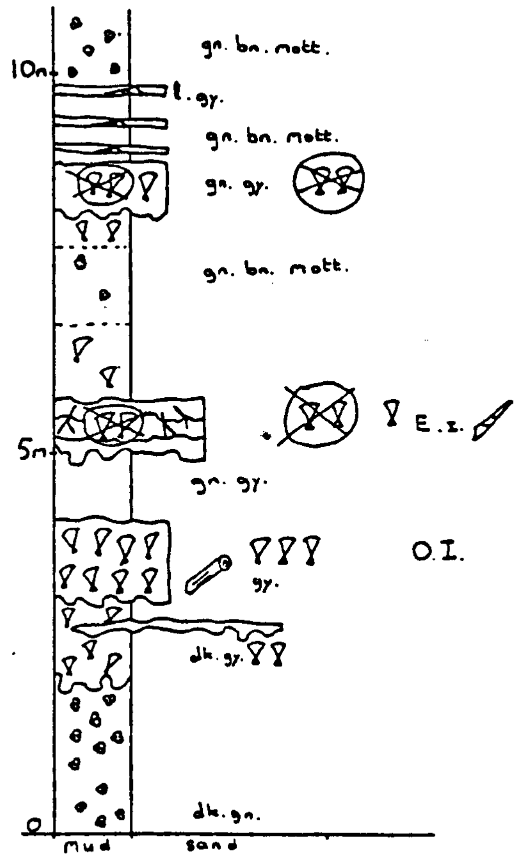


Figure 4.18 Graphic log of the sequence illustrated in Figure 4.16. For description see text.

On the beach at Mexilhoal (GR 7085 4505) near Areia Branca, and in the cliffs just south of Porto das Barcas at GR 7085 4120 (Figure 4.17), channel sandbodies are incised into the thin marine horizons. The former contains re-worked shell debris. North of Areia Branca as far as the Forte de Pai Mogo, appearances of shelly sediment become rather intermittent and appear to be localised in the mudrocks in the upper parts of the channels, at their margins.

4.4.2 Description and interpretation of facies

Figure 4.18 is a graphic log illustrating the succession to the north of Porto Dinheiro. It consists of a series of three prominent structureless sandy beds with sharp, but irregular bases, two of which are dominated by fragmentary shell debris. Plant debris is locally abundant. The uppermost two beds are separated by a horizon of barren silts which are mottled and contain sparsely distributed carbonate nodules. The lower shell bed is an *I. lusitanicum* shell 'bank'. Silts below this bed also

Figure 4.17 The margin of a channel sand body incised into fossiliferous facies ca. 0.5km south of Porto do Barcos. The slightly prominent bed 0.3m down from the top of the metre pole is the upper shell bed illustrated in Figures 4.16 and 4.18. The bench at the cliff base is the lower of the three shell beds and is rich in specimens of *I. lusitanicum*. The palaeochannel incised right through its thin laminated mouth bar deposits (sediment above the upper shell bed) into the bay sediments beneath . This is a characteristic of modern shoal,water deltas e.g.Guadalupe (Donaldson 1970); Colorado (Kanes 1970).

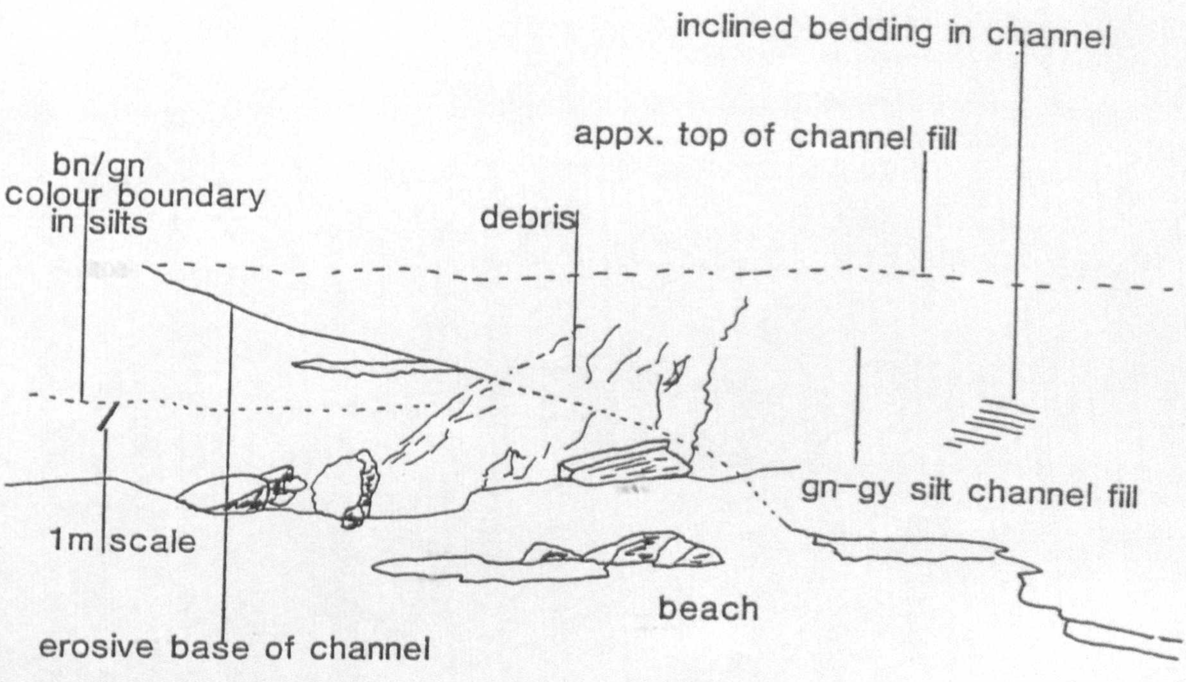


Figure 4.19 Sketch and photograph of a mud-dominated channel fill which contains a low-diversity, oyster dominated, fauna. In the most northerly outcrops of marine/brackish strata this association seems to be the most common encountered.

contain a shelly fauna and lie above an irregular interface with silts containing carbonate nodules. The fauna in the upper beds is dominated by *E. securiformis* at the expense of *I. lusitanicum*. Laminated facies lie above the topmost of the shell beds.

Some 6-700m to the north of the logged section, a thin (0.4m) shell bed appears in massive silts 29m stratigraphically above the top of the shell bed in Figure 4.18 (see Figure 4.16). A third horizon appears 13m above this at GR 7077 4440, near the top of the exposure south of Areia Branca. The 7m-thick unit thins progressively northwards, being 3.2m thick at Porto de Barcas (GR 7080 4278) and 2m thick at GR 7070 4417 above which outcrop is covered.

To the north of Areia Branca appearances of fossiliferous strata are laterally restricted, with the exception of one thin shelly horizon at GR 7115 4830, 600m south of the Forte de Pai Mogo (see enclosures Log FPM2). The fauna are generally sparse and dominated by oysters, suggesting brackish conditions (Werner 1986). More commonly they appear restricted to the mudrocks at the margins of channels such as that illustrated in Figure 4.19. A similar example appears below the Forte de Pai Mogo, the northern extreme of the marine incursion.

Interpretation : The sucession to the south of Areia Branca exhibits facies similar to those observed in the major sections to the south and attributed to interdistributary bay environments. A similar interpretation is implied here. The transition at the base, wholly in muddy facies, suggests that in these more northerly areas the shoreline was entirely muddy.

Occurrences of carbonate nodules are interpreted as pedogenic accumulations in line with earlier interpretations. The occurrence of nodules in silts between the fossiliferous strata probably indicate emergence of the substrate due to silting-up of the bay followed by renewed submergence, probably due to compactional subsidence. The distribution of fauna in the succession is interesting. The strongest marine signature in the succession is near the base. This interpretation is based on Fursich's (1981b) classification of faunal assemblages (Table 4.1), *I lusitanicum* being less tolerant of reduced salinity than *E. securiformis*, this species belonging to the assemblage most tolerant of reduced salinity. *E. securiformis* dominates the upper two sand beds at the expense of *I. lusitanicum* and this suggests a less saline, more stressful environment.

The thin nature of the succession and the alternation between marine and terrestrial strata suggest that the bay environment was very shallow indeed, probably no more than 2-3m. Similar shoal water depths are documented for the Holocene Colorado delta of the Gulf Coast by Kanes (1970). In such shallow conditions, the presence of the shell banks could act as barriers to the mixing of fresh and marine waters.

The accumulations of shell debris in the upper two sand beds, could be the result of periodic storms concentrating both shell debris and sand, or simply winnowing by currents.

A very shallow bay environment is also suggested by the channel sand body illustrated in Figure 4.17. The heterolithic sediments to the right of the scale bar are probably the channel's mouth bar deposits. These have been locally incised to below their base by the channel. Such a feature is very common in present-day shoal-water deltas where channels incise right through their mouth bar sediments into older bay sediments beneath (e.g. the Mississippi, Gould 1970; Colorado, Kanes 1970; and Guadalupe deltas, Donaldson et al. 1970).

The outcrops to the north of Areia Branca (Figure 1.8), where oysters appear at the margins of mud dominated channels, record a declining marine influence. The location of the fauna, suggests that either there was some marine ingress into channel mouths, or that the channels were inundated by marine waters once they had been abandoned.

4.5 CONCLUSIONS: ENVIRONMENTAL MODEL

Figure 4.20 is an environmental sketch which attempts to illustrate the various ideas outlined in the previous sections. The sedimentary facies developed indicate that a basal Tithonian transgression (dated by Leinfelder 1987) occurred, followed by the development of a shoal water lobate delta system with broad interdistributary bays on a shallow marine, mud-dominated shelf.

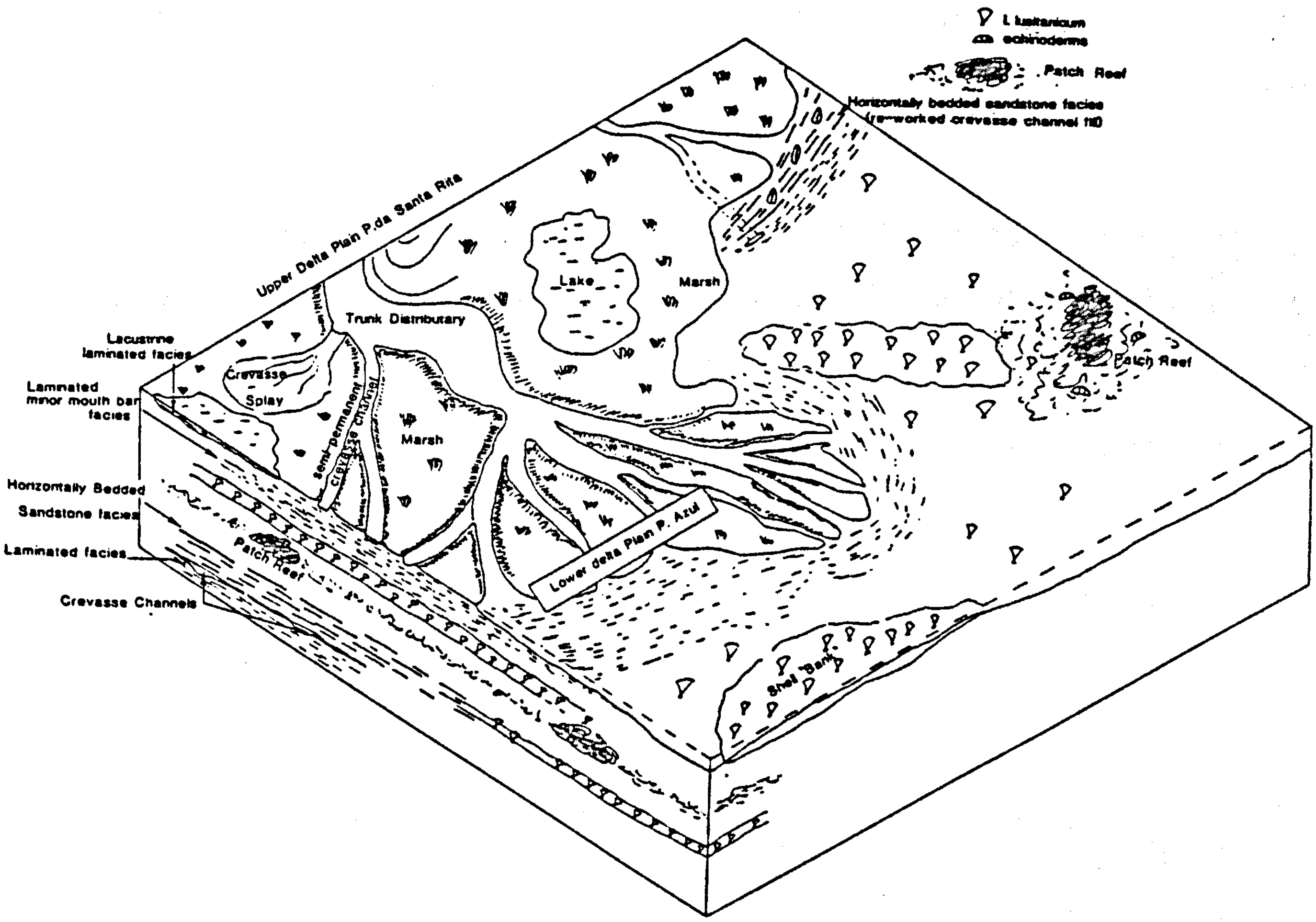


Figure 4.20 Environmental model for the depositional environments of the Praia Azul and Assenta members. The deposits of the former were laid down in a lower delta plain setting and the latter an upper delta plain setting. The deltas developed in a shoal water environment and probably contributed to a highly irregular dominantly muddy coastline. Locally sediment was re-worked, producing the horizontally bedded sandstone facies. For further details see text.

The nature of the transgressive deposit at the base of the Praia Azul member varies. In the type section, laminated sediments, associated with small lenticular sand bodies represent input from crevasse channels. Laterally extensive sand beds, are the result of local reworking and winnowing of sediment by waves on the shoreface. The almost complete bioturbation of the sandy facies suggests that reworking was a slow and prolonged process. To the north, laterally extensive, irregularly based beds with an encrusted fauna, record a retreating muddy shoreline. At the northern extreme of the incursion, around the Forte de Pai Mogo, the margins of flooded distributary channels were colonised by oysters.

The entire basin it seems, was very shallow at this time. Evidence for this comes from studies elsewhere in the basin (Fürsich and Schmidt Kittler 1980; Felber et al. 1982; Leinfelder 1986,1987), and from a number of features within the Praia Azul member. The thin nature of mouth bar sequences (ca. 2m), particularly when compared to the larger channel sand bodies, the laterally extensive shell banks, and the widespread but thin nature of the marine sequences as a whole, often oscillating with terrestrial deposits, all support the proposed shallow nature of the basin. The last feature also indicates that the lower delta plain had little relief.

The successions documented probably resulted from the rapid switching of distributary networks, characteristic of lobate, fluvially dominated delta systems (Donaldson et al. 1970; Horne et al. 1978). Such a system is also suggested by the clustering of channels noted at the 20m level in the Praia Azul section.

The switching of distributary networks was possibly linked to seasonal fluctuations in discharge and sediment load. Marine areas were prone to salinity fluctuations which in the presence of low levels of tidal current or wave induced mixing were controlled primarily by (seasonal ?) variations in fluvial discharge. It is likely that in inshore areas, or somewhat restricted bays, normal salinities were hardly ever attained. In view of this, the presence of echinoderms is somewhat problematic, as they are conventionally taken to be indicators of normal marine salinities (Clarkson

1983). Either sustained marine salinities existed periodically or *Pseudocidaris* possessed at least some tolerance to salinity fluctuations. The thicker channel fills observed in the upper part of the Santa Rita section, are thought to characterise an upper delta plain environment, compared with a lower delta plain setting for the channel sand bodies developed at Praia Azul. Major trunk rivers up to 8m deep (exposed at P. da Santa Rita), bifurcated on the delta top into a number of distributaries which were usually no more than 3m deep (exposed at P.Azul).

Broad interdistributary bays were sites of fine grained background sedimentation punctuated by spells of coarser grained sedimentation expressed as laterally extensive sheets of laminated sediment and decimetre-thick lobes of fine sand. The former, which are the most common, were probably due to extensive overbank flooding, itself a consequence of poorly developed levees associated with an immature channel system (Elliott 1974a). The latter are crevasse splay lobes, a consequence of the widespread breaching of a levee during flood (Elliott op cit.). Generally, it seems that levees were not significantly developed in distal regions as most bay sediments (at Praia Azul) lack significant crevasse splay deposits. More common are packets of laminated sediment, the product of widespread flooding of the lower delta plain at high river stages. Where levees were better developed, semi-permanent crevasse channels were initiated and their deposits are more common in the adjacent marine sequences. The localisation of sediment input may have led to there being areas that were relatively sediment starved. In such areas carbonates were able to accumulate.

Thick packets of laminated sediment below lenticular 'winged' channel sand bodies are probably the deposits of minor mouth bars (Arndorfer 1973; Elliott 1974b; Fielding 1987). The bays also contained isolated patch reefs and laterally extensive shell banks with reliefs of up to a metre. Similar features occur in bays surrounding the modern Gulf Coast Colorado River Delta (Kanes 1970). These features may have assisted in restricting circulation and preventing mixing of marine and brackish waters. There is no direct evidence for diapiric barriers forming lagoons as postulated by Fürsich (1981b).

Volumetrically, bay and marginal marsh sediments dominate the sections, generally with a sparse fauna punctuated by horizons rich in specimens of *I. lusitanicum* and occasional other large shelled bivalves. Periodic emergence, perhaps where sedimentation marginally outpaced subsidence, is suggested by reddening of the silts and the appearance of carbonate nodules. Bay fills frequently underwent more than one episode of emergence and re-submergence.

The remarkable absence of signs of reworking by waves or currents in most of the succession may also be providing evidence of an extensive shallow marine basin to the south. Heyward (1981) noted that epeiric sea coasts with extremely low bottom gradients are characterised by rapid and extensive transgressions and regressions, and that the shallow gradients may contribute to the complete dissipation of wave energy and damping of tidal effects.

The fluvial channels were subject to significant fluctuations in discharge, a consequence of a strongly seasonal climate (discussed in Chapter 2). The presence of carbonate-rich soils at a number of levels, in both mudrocks and sandy lithologies supports the general semi-arid to sub-tropical climatic regime envisaged. The upper delta was clearly vegetated as evidenced by rhizocretions in the soils and preserved plant debris. There were swampy and lacustrine areas on the delta top. Evidence for these environments appears within the Assenta member deposits at Praia da Santa Rita.

The existence of two major transgressive pulses is suggested by the presence of two levels in the Praia Azul section where thick shoreface sediments are developed (Figures 4.2 and 4.7). A discussion of the possible causes of transgression and fluctuations in relative sea level appears in Chapter 7.

CHAPTER 5

FLUCTUATING RELATIVE SEA-LEVEL: THE ASSENTA MEMBER

5.1 INTRODUCTION

The brackish/marine conditions prevailing over the southern part of the basin during the depositon of the Praia Azul member significantly declined in importance during the deposition of the Assenta member. The character of the sediments is predominantly terrestrial, with a number of thin, often distinctive, laterally extensive marine horizons. This arrangement of facies suggests that the environments of deposition consisted of a broad alluvial plain characterised by sand-bed rivers of moderate sinuosity, plus numerous lakes, which were subject to periodic, short-lived, but widespread marine incursions. During such events marine areas were often sediment starved. This allowed the accumulation of carbonates and even the establishment of coral bioherms.

There are two coastal outcrops of the Assenta member. One relatively short section at Praia da Santa Rita (described in Chapter 4) and the other extending 8.5km from south of the Rio Sizandro GR 6655 2845 to Porto da Calada GR 6400 2075 (Figure 5.1). The Santa Rita outcrop is 120m thick and lies conformably above the Praia Azul member. It is unconformably overlain by the Santa Rita member (see Chapters 1 & 4).

Exceptionally low southerly dips, often only ca.2°, characterise the Rio Sizandro section. A number of normal faults are present. The section can be correlated across all but one of these because of the presence of marine horizons. This fault, at GR 6405 2630 (Figure 5.1) marks the southern limit of the author's detailed investigations. To the south of this point, where dips are near horizontal, the section was measured by R Hiscott. His logs appear in the enclosures and certain of his

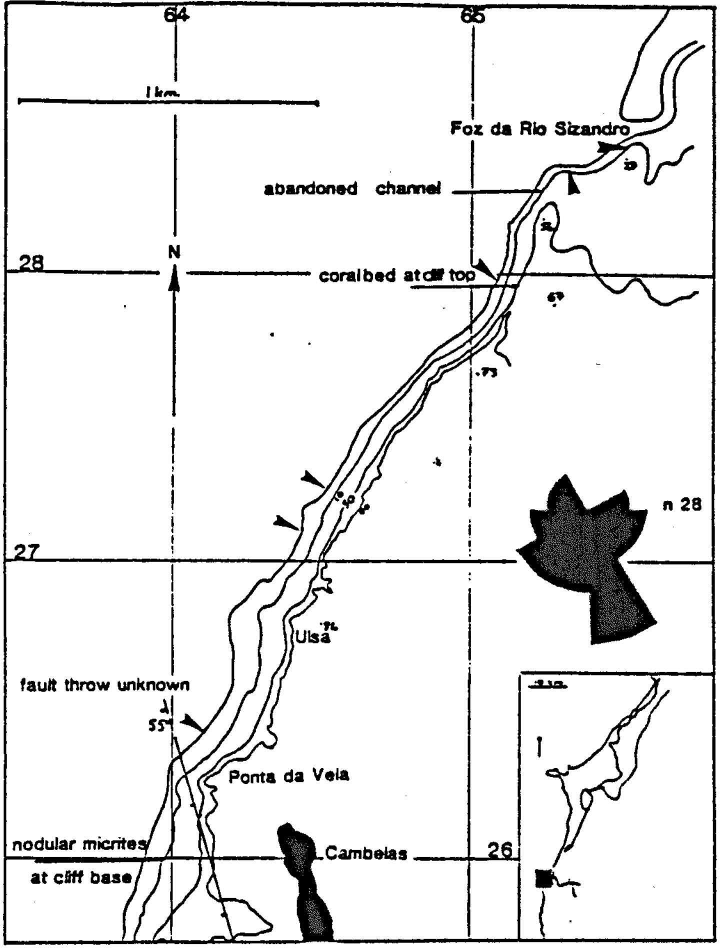


Figure 5.1 Sketch map of the Rio Sizandro section (location shown in inset). The fault shown on the map divides the type section of the Assenta member into two halves. The northern (basal) section illustrated and studied by the author, and the southern (upper) part studied by R. Hiscott (St. Johns Newfoundland). The two sections cannot be correlated across the fault. The position of logged sections (shown in figure 5.2) is indicated by the arrows. Palaeocurrent data from cross beds within channel sand bodies, indicates a low to moderately sinuous system. The 'coral bed' is illustrated in Figure 1.25 and the 'abandonment fill' in Figure 5.12. Heights in metres.

descriptions and interpretations appear in this chapter. These are identified in the text. One hundred and forty metres of section appear to the north of the fault, within which there are four marine horizons, most only 2m thick (Figure 5.2). Hiscott logged a further 140m of section above the fault (Figure 1.10). He recorded three thin marine horizons in the lower 85m. The top 55m has a stronger marine character and includes a number of dolomite beds, plus faunal assemblages not found elsewhere in the Lourinhã formation (Rey 1972). Coarse, fluviatile clastic sediments, of Cretaceous age, unconformably overlie the Assenta member at Porto da Calada.

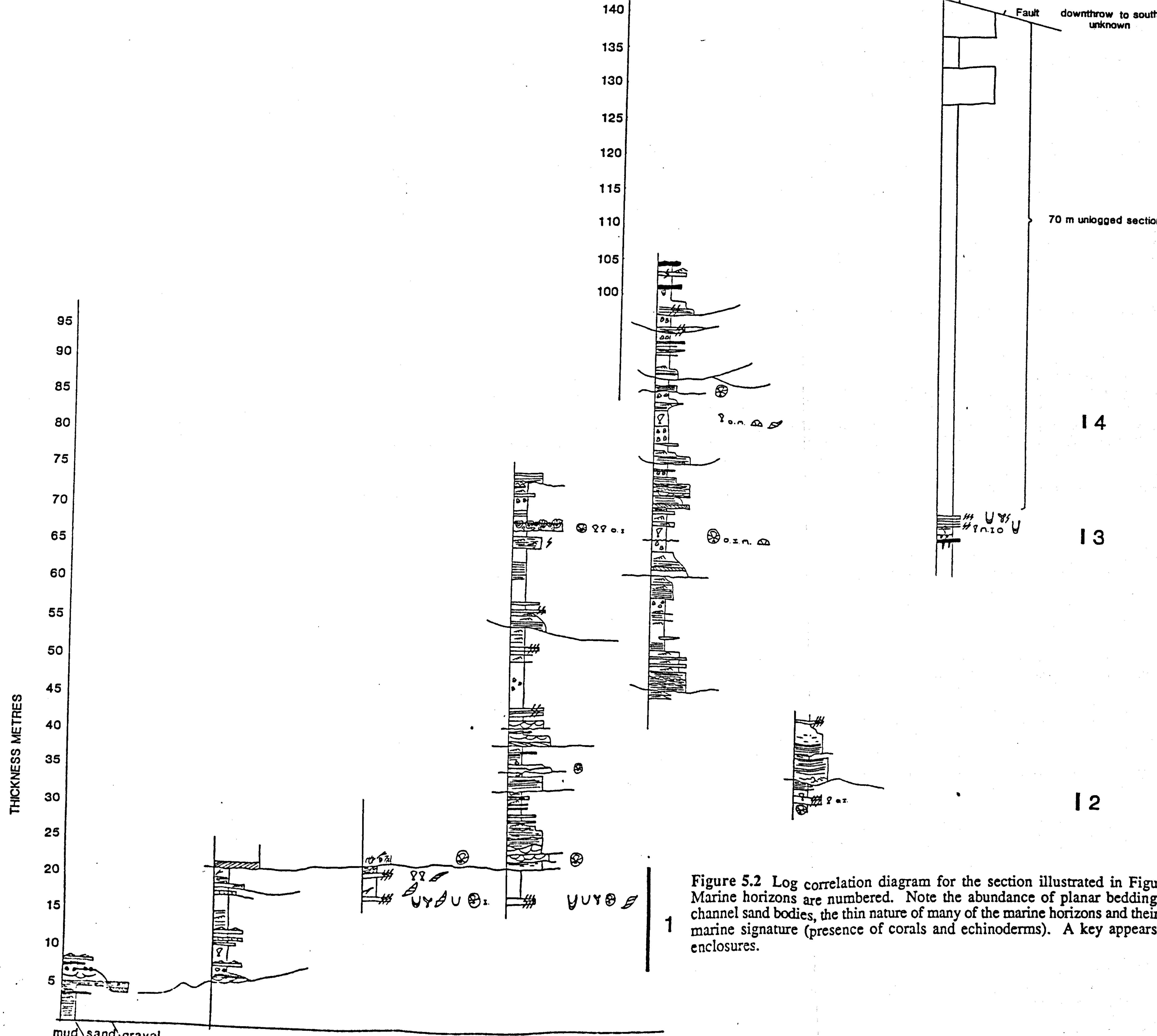


Figure 5.2 Log correlation diagram for the section illustrated in Figure 5.1. Marine horizons are numbered. Note the abundance of planar bedding in the channel sand bodies, the thin nature of many of the marine horizons and their strong marine signature (presence of corals and echinoderms). A key appears in the enclosures.

In addition to normal faults, the section is cut by a few high angle reverse faults which parallel the coast. Displacement across these is usually relatively small. There are also a number of basic dykes 1-2m in width which are often associated with faulting, particularly in the vicinity of Assenta. A volcanic breccia plug occurs to the south of Assenta at GR 6400 2275.

The areal extent of the Assenta member is effectively confined to south of the course of the Rio Sizandro (Figure 5.1). This river trends approximately east-west from the coast to Torres Vedras. Inland of the northernmost Santa Rita outcrop exposure is so poor that it has not proved possible to identify the Assenta member. Even south of the Rio Sizandro, the number of outcrops worth examination number less than ten.

5.2 DESCRIPTION OF FACIES

Many of the facies found within the Assenta member are similar to those found in either the Porto Novo or the Praia Azul members. As the Santa Rita section of the Assenta member has already been described in Chapter 4 this section is confined to relatively brief descriptions for the most part. However marine horizons will be described individually and in some detail. Sections on marine strata also include some data on their variable associations.

There are five main facies, some of which contain a number of sub-facies:

Massive sandstone facies; including Massive sandstones, Caliche pebble conglomerates and heterolithic interbeds, Cross bedded sandstones, Planar bedded sandstones, Ripple laminated fine sandstones and Ripple and parallel laminated mudrocks.

Heterolithic facies

Massive mudrock facies

Laminated facies; including Graded beds, Ungraded beds and Isolated cross beds.

Fossiliferous facies; including Massive siltstones, Silty biomicritic wackestones, Massive sandstones, Oyster packed silts, Coral biostromes and Nodular micrites.

5.2.1 Massive sandstone facies

This facies has many similarites to the Massive sandstone facies found within the Porto Novo member, but generally grain sizes are confined to medium sandstone or less. Sands are always micaceous.

Figure 5.3 Thick sheet sand bodies in the cliff section to the south of the Rio Sizandro. Cliff height to right 70m Log No. 2 passes up through the stack of sand bodies. Lateral accretion surfaces dipping shallowly from left to right are evident in the sand body at mid. section (37-42m on log). The coral bed (Figure 1.25) crops out just below the cliff top (arrowed). Abandoned channel is the sand body at beach level on the extreme left of the photograph.

Massive sand bodies occur on two scales: Either, large sheet sandbodies 2-8m thick (Figure 5.3), or small lenticular sand bodies 1-2m thick and a few metres to tens of metres wide. Both types exhibit a degree of fining-up. Larger sand bodies may have a basal relief of up to 1m. The following sub-facies are developed;

Massive sandstones appear at the bases of the larger sand bodies. They consist of structureless coarse to medium sandstones containing logs and lignitic debris. Where a sand body cuts into fossiliferous strata, minor amounts of shell debris also appear. Pebbles of irregular carbonate nodules (commonly found as pedogenic accumulations in mudrocks) may also be present.

Caliche pebble conglomerates and heterolithic interbeds are confined to the basal parts of sand bodies directly above the erosive contact. These consist of sandy, matrix-supported, mudclast and caliche pebble conglomerates. Most often horizontally stratified, they are also found as cross bedded scour fills incised up to 0.5m into underlying strata (Figure 5.4).

Figure 5.4 Cross bedded scour fill at the base of a channel sand body. Lithologically the sediments consist of caliche and mudclast pebbles in a sandy matrix. Also note the outcrop of 'laminated facies' directly beneath the sand body. Scale pole = 1m. This outcrop is at the base of the Rio Sizandro section.

This facies also appears in an unusual association at the base of one particular sand body. In this example, inclined decimetre-scale beds of caliche pebble conglomerate, usually horizontally stratified, are interbedded with a variety of other lithologies which include buff medium sandstones and dark grey mudrocks. Individual beds are up to 0.25m thick and the unit as a whole is 0.45m thick. All beds are inclined at about 5° to the south. This is the same orientation as the major breaks within the sandbody which mark lateral accretion surfaces (Figure 5.3, see section 5.3.1/2). Most individual beds have a lenticular cross section, tapering off both up and down dip. The upper bed boundary is an erosion surface.

Cross bedded sandstones consist of decimetre-scale predominantly trough cross beds up to 0.5m thick, which frequently occur in cosets up to a metre thick. Tabular cross beds also occur, generally as isolated sets near the base of channel sandbodies, or as thin cosets near their tops. Cross beds often have mudclast pebbles or cobbles, or fine lignitic debris on their foresets.

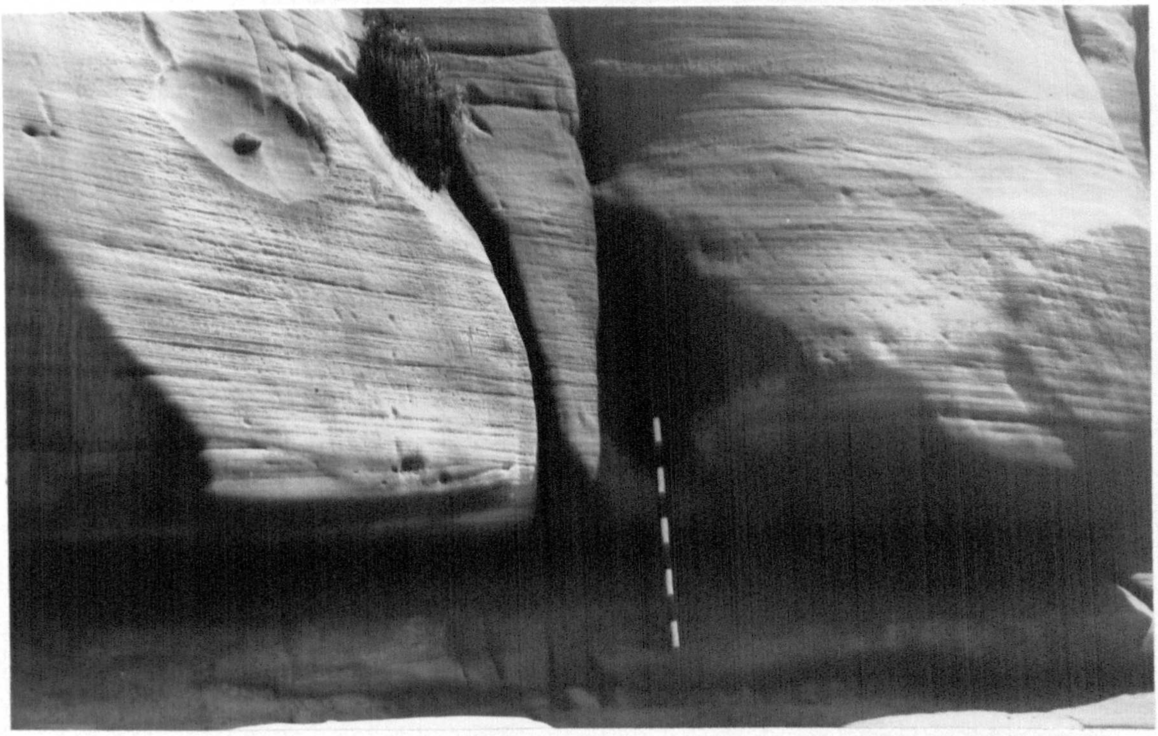


Figure 5.5 Channel sand body dominated by packets of planar bedded fine and medium sandstone. Scale pole = 1m. This locality is illustrated in Figure 5.3; the massive sand body at the cliff base (right of centre).

Planar bedded sandstones are by far the most common structure encountered within channel sand beds. Rarely coarser than medium or fine sand, these are commonly graded, passing into ripple laminated very-fine sands and silts in their upper few centimetres. Primary current lineations are found on bed surfaces. The sands are often red and green colour mottled in their upper parts. Coarse mica flakes and concentrations of plant debris are common components. Individual beds may be a few decimetres to several metres in thickness. Within the larger of these, packets of planar bedded sediment a few tens of centimetres thick are separated by irregular erosion surfaces (Figure 5.5). Horizons of convoluted laminae occasionally appear.

Ripple laminated fine sandstones generally form relatively thin units at the top of graded beds. Fine current ripple laminated sands may, in a few instances, form a significant component of channel fill. Individual beds may be up to a metre thick and include horizons of parallel laminated or convoluted sediment.

Ripple and parallel laminated mudrocks also appear in graded units and as thin drapes to sand beds. These may be found at almost any level within channel sandbodies, but are more common in their upper parts. Thick, dark grey, organic rich mudrocks occasionally appear between sand beds. Rarely greater than 0.5m thick, small 'biscuits' of buff sandstone up to 0.1m thick are often enveloped by the mudrocks. These and sand beds above the mudrocks often exhibit soft sediment loading structures. In addition to the load casts, sand beds above these muddy units also exhibit groove casts on their bases.

A number of the thick mudrock units are not horizontal but inclined, passing from the upper parts of the sandbody to lower levels and even the base. Such units are often cut out down-dip by the base of younger sand beds. The palaeocurrents indicated by groove casts were orientated approximately at right angles to the dip of the inclined mud beds.

5.2.2 Heterolithic facies

These are largely confined to the uppermost parts of channel sandbodies. Usually red in colour, this facies may also contain invertebrate burrows and carbonate nodules. Exceptionally up to 2m thick, it is commonly only 0.5-1m thick.

Decimetre or centimetre-scale medium to very fine sand beds are interbedded on a similar scale with red-brown mudrocks. Both are usually massive, but may exhibit fine ripple or parallel laminae. Sand beds have a variable lateral extent, from a metre to several metres, usually having a lenticular geometry and tapering off along strike. However, the orientation of the section relative to palaeoflow indicators is important. Where inclined accretion surfaces are evident, the sand beds often pass laterally into lower levels of the sandbody, thickening down dip. Invertebrate burrows are also prevalent and may increase in intensity up through a packet of this facies until eventually no other structures are apparent.

Carbonate nodules usually a few millimetres to centimetres in diameter also appear. These rarely form more than a diffuse accumulation. However, in one example these amalgamate to form an irregular crust up to 0.5m thick in the upper part of the sandbody. This is associated with green and buff colour mottling and the complete loss of primary depositional structures.

5.2.3 Massive mudrock facies

Much of this facies consists of red-brown or dark green, micaceous mudrock, much like that described from the Porto Novo member. Diffuse accumulations of carbonate nodules commonly appear within this facies. There are also horizons of purple gritty silts, much like those described in the Praia da Amoreira member and rare vertical pipes of carbonate. Dark, organic rich rootlet traces are occasionally present.

Rare examples of 'beds' of carbonate about 10cm thick also exist. These consist of closely spaced, irregular prismatic blocks of fine grained carbonate. The upper and lower contacts of these units with the surrounding mudrock are abrupt and well defined. Associated silts are mottled red, green and purple.

5.2.4 Laminated facies

These consist of packets of laminated sediment 1-5m thick. These units have a lateral extent of tens to hundreds of metres. Their bases, tops and margins are usually rather diffuse or gradational. Their colour is variable, from dull greys and greens, through to browns and red-green mottles. Parallel laminated silts or alternations of silts and fine sands dominate this facies with less common current ripple laminations. Horizons of convoluted laminae are very common and these are often truncated by erosion surfaces. The laminated facies generally contains multiple internal erosion surfaces. These usually have a planar, or only slightly irregular morphology.

Invertebrate burrows are very common, usually appearing as 5-7mm diameter, unlined sand-filled tubes, at a low intensity. Exceptionally, completely bioturbated horizons occur.

Mica is a common constituent of both the silts and associated sands. Periodically, concentrations of fine lignitic debris pick out laminae. A rare feature is the development of small carbonate nodules, which is usually associated with a purple colour of the sediment.

The laminated facies also contains a number of associated sub-facies:

Graded beds of buff fine sand to silt are common. Usually ripple or parallel laminated, with an irregular erosive base and up to 30cm thick. They often appear at 10-30cm intervals between packets of laminated silts. As the coarsest sediment present in this facies, bed bases may have load casts with a relief of several centimetres.

Ungraded beds of fine sand also appear. Usually ripple or parallel laminated, they exceptionally exhibit sets of 'type A' climbing ripple laminae (Ashley et al. 1982). These beds are usually up to 20cm thick, have basal erosive surfaces and exhibit deformation features much like those detailed for graded beds. At the base of the section, a number of ungraded beds also exhibit unusual deformation structures, consisting of paired 'humps' about 20cm apart (Figure 5.6). This deformation may penetrate the sediments below for up to 10cm, below which bedding is unaffected.

The tops of a number of these structures are planed off at a horizontal erosion surface above which the sequence is again parallel laminated.

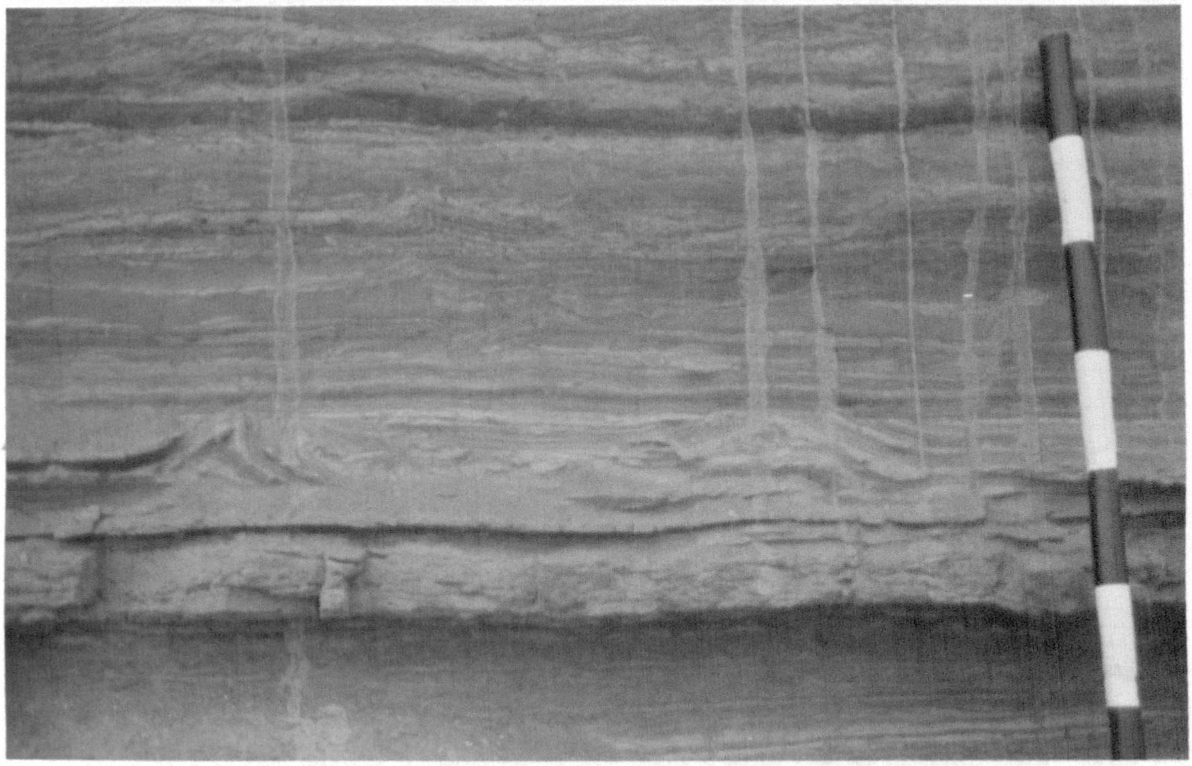


Figure 5.6 Outcrop of the laminated facies at the base of the section. Dominated by parallel laminated silts and fine sands plus thin sand beds. The sand bed at the centre of the photograph is deformed, having an unusual pair of 'humps'. Such deformation clearly ocurred whilst the sand was at the sediment-water interface, as it is truncated by an erosion surface. This may have been caused by loading by a large reptillian foot. Scale bars = 0.1m.

In certain associations (see section 5.3.4), the laminated facies contain centimetre beds of fine grey sandstone with straight crested ripples on bedding planes. Such units also contain more abundant climbing ripple laminae and structures such as lenticular and wavy bedding (Figure 5.7) (Reineck and Wunderlich 1968). Certain sand laminae in this association exhibit small scale load structures which project down into the underlying mud drapes. Invertebrate burrows in these beds include examples of *Diplocraterion* . Exceptionally, a small number of thin shelled bivalves have been found in one example of this facies.

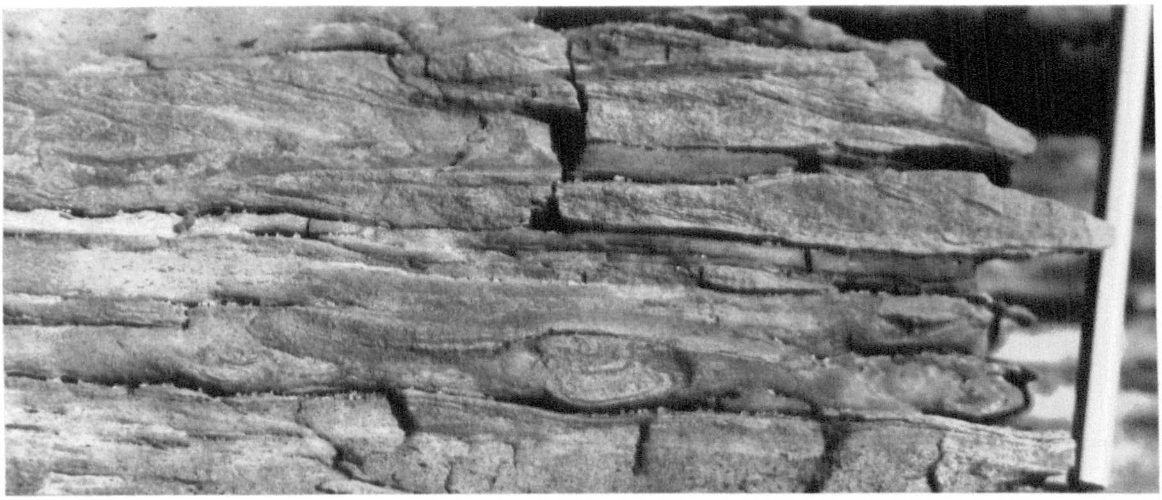


Figure 5.7 Lenticular bedding; solitary mud-draped ripples, within the laminated facies. Such structures form where current activity exists in a mud-rich environment starved of sand-grade sediment, resulting in isolated ripples migrating across a muddy substrate. Location: cliff base 100m north of fault shown on Figure 5.1. Note also the load structure below the ripples.

Isolated cross beds of buff, or light grey, fine to medium sand occasionally appear. These are rarely more than 0.1m thick, exceptionally up to 0.2m. They consist of a single planar tabular cross set, with a near planar erosive base. Mud drapes, or concentrations of fine plant debris occasionally appear on cross-bed foresets. In the context of the laminated facies this sub-facies is unusual in that it is often of only limited lateral extent.

5.2.5 Fossiliferous facies

This includes a wide variety of sediments associated with the thin marine horizons which punctuate the Assenta member. Because the nature of these is so variable, a number will be described individually after the more common facies have been illustrated.

Often only a metre or two of sediment can be unequivocably described as marine and may be bracketed by reddened mudrocks with carbonate nodules, or pass up into laminated facies incised into by the base of a channel sandbody. The thickest, almost consistently marine unit appears at the base of the section. It is up to 15m thick. The facies observed are as follows;

Massive silts are dark green or grey in colour and locally contain sparse accumulations of the delicate mould of a single bivalve species, probably an oyster. Elsewhere, fauna are more abundant, with a variety of small fragile bivalves, plus gastropods. The bivalve shells in this association may exhibit small borings.

Silty, biomicritic wackestones 10-20cm thick, containing fragments of oysters, *E securiformis* and *I. lusitanicum,* plus a few intact specimens of each and plant debris. Ichnofossils include *Diplocraterion* and *Arenicolites* both with relatively thin 5mm diameter tubes.

Massive sandstones are usually white, thoroughly bioturbated and fine grained. These often form prominent marker beds up to 0.35m thick. Fragmented shell debris is common and concentrated at the base of one example. The fauna is much the same as that found in the silts with the occasional addition of gastropods, most often turreted forms several centimetres in length plus, exceptionally, coiled varieties much like the recent common snail. Burrows identified include *Diplocraterion* and *Thalassinoides*, the latter particularly prevalent on bed bases.

Fossiliferous facies associations

Of the four marine horizons identified, the lower two have combinations of the facies described above. The upper two, particularly the third unit, exhibit rather different facies. These generally have a stronger marine signature, yet at the same time are very thin, being only about 2m or less in each case.

The uppermost unit lies above mottled silts with sparse carbonate nodules. A colour change to dark grey occurs and these silts contain an initially quite varied fauna of bivalves, including oysters, *M. lusitanicum* and a mussel-like bivalve, plus gastropods and cidaroid echinoderm spines. Within 20cm the diversity of fauna rapidly tails off to a single species of bivalve. The fauna is completely absent 2m above the base of the unit and is replaced by laminated facies. A small quantity of shell debris appears at the base of a small channel 2.5m above the disappearance of the fauna.

The third marine unit is the most varied, with lateral facies changes a common feature. Three logs pass through this horizon (Figure 5.2). The middle of these occurs above dark green silts containing sparsely distributed irregular carbonate nodules. There is a colour and texture change across an irregular boundary to dark grey organic rich silts containing abundant oysters plus echinoderm spines and specimens of *I. lusitanicum* and *M. lusitanicum*. This facies, some 2m thick, is identical to the 'Patch Reefs' found in the Praia Azul member. Fauna are absent when coarser laminated sediment appears about 2.5m below the erosive base of a channel.

Figure 5.8 A terrestrial-marine transition. A thin shell bed in dark green silts lies above green or reddened, desiccated silts containing pedogenic carbonate accumulations. Invertebrate burrows extend down into the terrestrial sediments at the boundary The marine horizon is less than a metre thick and passes up into coarsening-up laminated facies. Location as per Figure 5.7.

The southernmost outcrop is the best exposed. It forms a wave-cut platform about 50m north of the fault which divides the section. Marine fauna briefly appear 0.5m above a soil profile. The soil consists of reddened silts with sand filled desiccation cracks overlain by a 0.2m laminated blocky carbonate horizon and silts with sparsely distributed carbonate nodules. At the transition, short stumpy 5cm diameter curved invertebrate burrows penetrate the terrestrial sediment up to 0.2m (Figure 5.8). Burrow fills consist of dark grey coarse silts containing shell fragments. A small number of specimens of *M. lusitanicum* , *I. lusitanicum* and oysters lie in the sediments immediately above. These are rapidly replaced by a coarsening-up cycle of laminated facies including wavy and lenticular beds with *Thalassinoides* and *Diplocraterion* burrows. Traces of the latter are also found in a 0.5m fine sand bed which caps the sequence.

The character of this unit is completely different at its northernmost outcrop at the cliff top. Silts containing carbonate nodules are overlain by 1.5m of laminated facies. These are occasionally pervaded by vertical and horizontal invertebrate burrows up to 10mm dia., the latter particularly on bed bases. This unit is succeeded by 0.5m of grey siltstones packed with oysters plus a few specimens of *I. lusitanicum*. Upon this are a further 0.5m of large, in situ, colonial coral heads (Figure 1.25), which are encrusted with oysters. Laminated facies 2.5m thick which overlie the corals, are succeeded by purple silts containing carbonate nodules.

Hiscott recorded several more thin marine units in the upper parts of the section. The first three of these, with a spacing of ca. 20m, are all thin (ca. 2m) nodular micrites with thin seams of grey shale. Rey (1972) recorded the presence of foraminifera, gastropods, bivalves, echinoderms and ostracodes within these sediments.

The uppermost marine units in the top 55m of the section consist of ca. 0.5m beds of microcrystalline dolomite and contain a fauna of bivalves, gastropods and the large foraminifer *Anchispirocyclina lusitanica*. These are associated with marine siliciclastic and dolomitic sandstones which are intensely bioturbated. *Thallasinoides* and *Diplocraterion* are among the idntified ichnofossils.

5.3 FACIES ASSOCIATIONS AND INTERPRETATIONS

5.3.1 Massive sandstone facies associations.

Channel sand sub-facies are commonly found in a variety of associations in erosive based fining-upwards sand bodies. These tend to occur on two scales. The smaller of these are 1.5-2m thick and have a lenticular 'winged' cross section (Friend et al. 1979). A number are multi-storey with two or more sand bodies assymetrically stacked upon each other. Sandbody widths vary from 6-40m. Channel fills are rarely coarser than fine sand and primary depositional structures are dominated by packets of parallel, and occasional current ripple, laminae. These sediments are usually brown in colour, often with green mottles. Mud drapes are very rare. Horizons of convoluted laminae and invertebrate burrows are common. Bioturbation may be intense, almost obliterating all primary structures.

Sandy channel fills often pass vertically into heterolithic facies. Laterally these form 'wings' to the sandbody. Lateral accretion surfaces are apparent in only one example of this channel type. Silts, both above and below sand bodies, often contain small irregular carbonate concretions. These are interpreted as the product of sub-aerial exposure and pedogenesis. Convoluted laminae are also present and may be due to water escape. This is often a consequence of rapid sedimentation (Leeder 1982, Allen 1985).

The larger scale channel sand bodies have sheet-like geometries (Friend et al. 1979). Thicknesses vary from 1.5-9m and lateral persistence usually exceeds several hundred metres. Grain sizes and the suites of structures observed are far more variable though parallel lamination often dominates channel fills (Figure 5 5).

Mud-draped, low-angle, lateral accretion surfaces are apparent in a number of sand bodies. With the exception of the occasional multi-storey unit, channel bases are incised into one of two facies types: red-brown silts containing pedogenic nodules, or laminated facies. The latter may be associated with a marine fauna and possibly show coarsening-up, or with red or purple silts exhibiting pedogenic nodules and occasionally, mudcracks (see section 5.3.5).

The range of facies developed is illustrated in Figure 5.9 by four sedimentological logs through a single sand body. These have been correlated by linking, as far as is possible, prominent erosion surfaces. A number of important features are illustrated:

1) The erosive base of the channel is incised into laminated facies which include structures such as wavy and lenticular bedding (Reineck and Wunderlich 1968, Reineck and Singh 1980). These sediments lie above bioturbated fossiliferous sandstones and also contain, in places, a sparse fauna themselves.

2) Locally, sediments at the channel base are contorted and enclose numerous pebble and cobble grade mudclasts, considerable quantities of lignitic debris, plus logs, caliche pebbles and some fossil debris.

3) Channel-fill sediments do not exceed medium sand grade. Structures are dominated by packets of planar bedding.

4) Major erosion surfaces, which occasionally cross-cut earlier examples, all dip from left to right (north-south) and are often picked out by quite thick muddy units. Thin sand beds within these are commonly severely deformed. Though not apparent from the logs, these erosion surfaces generally have a smooth, gently curved profile and bound inclined lenticular sedimentary units (Figure 5.10).

5) Palaeocurrent data, though relatively sparse, indicates consistently a south eastwards palaeoflow obliquely into the cliff face and approximately at right angles to the orientation of the major erosion surfaces.

6) Log number 3 has, near its base, a thick mudrock unit which contains an internal erosion surface and fines up to claystone.

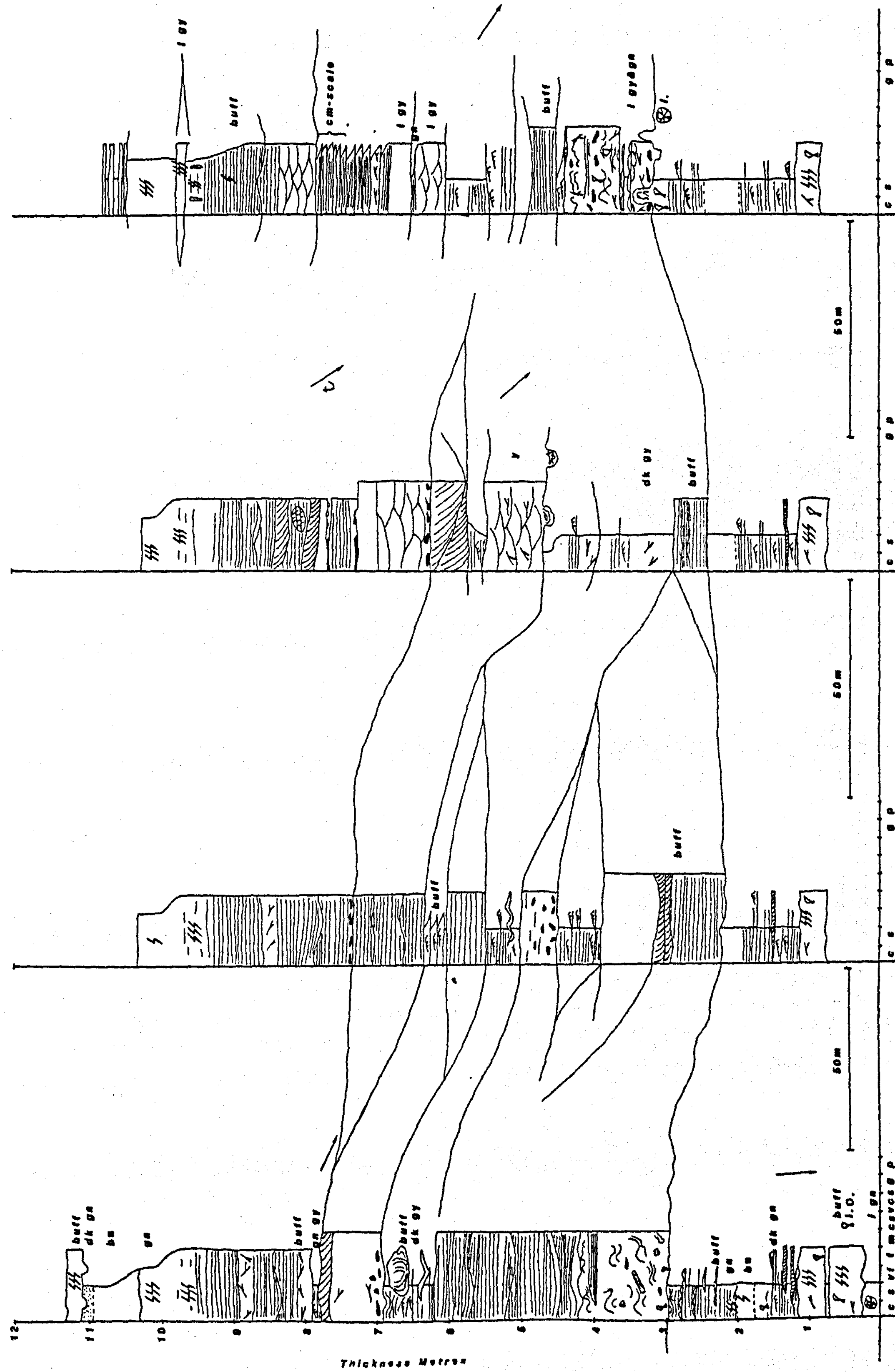
7) Sediments at the channel top have few primary structures preserved, are intensely bioturbated and are overlain by limited thicknesses of heterolithic facies.

5.3.2 Interpretation of Massive sandstone Facies

Both channel 'types' exhibit an extraordinary prevalence of planar bedding. In contrast, the larger scale sandbodies include thick muddy units. The significantly different flow conditions which the two sediment grades and structures record (Harms et al. 1982) indicate that the channels were subject to considerable fluctuations in discharge. There is no suggestion that they were ephemeral for the mudrocks are generally dark grey and organic rich and features such as desiccation cracks within channel fills are absent.

In the Porto Novo member, channels dominated by parallel laminated sands were intrepreted as the product of periodic, high discharge floods. Such an explanation would account for the abundance of upper flow regime bedforms (e.g. Laracy and Hiscott 1982, Turner 1986). A semi-arid, sub-tropical climate, with marked seasonality would account for such fluctuations in discharge. The development of caliches, albeit mostly poorly developed Stage I and II profiles (Gile et al. 1966), with diffuse accumulations of nodules, fits such a climatic interpretaton. This is also consistent with the climatic conditions envisaged during the deposition of the other members of the formation.

Figure 5.9 (over) Four graphic logs through a sheet sand body; illustrating the range of facies developed both vertically and also laterally along the major accretion surfaces. The sand body lies above fossiliferous and laminated facies, recording the progradation of a fluvial channel into a shallow marine environment. Note the entrained shells, lignite, logs and carbonate nodules in the basal units of the sand body. Also the dominance of planar laminae and the thick muddy unit. The muddy unit in the third log suggests that the channel may have been abandoned and subsequently re-initiated during its lifetime. Palaeocurrents indicate flow approximately parallel to the strike of the inclined lateral accretion surfaces. Part of this sand body is illustrated in Figure 5.10.


Thickness Metres
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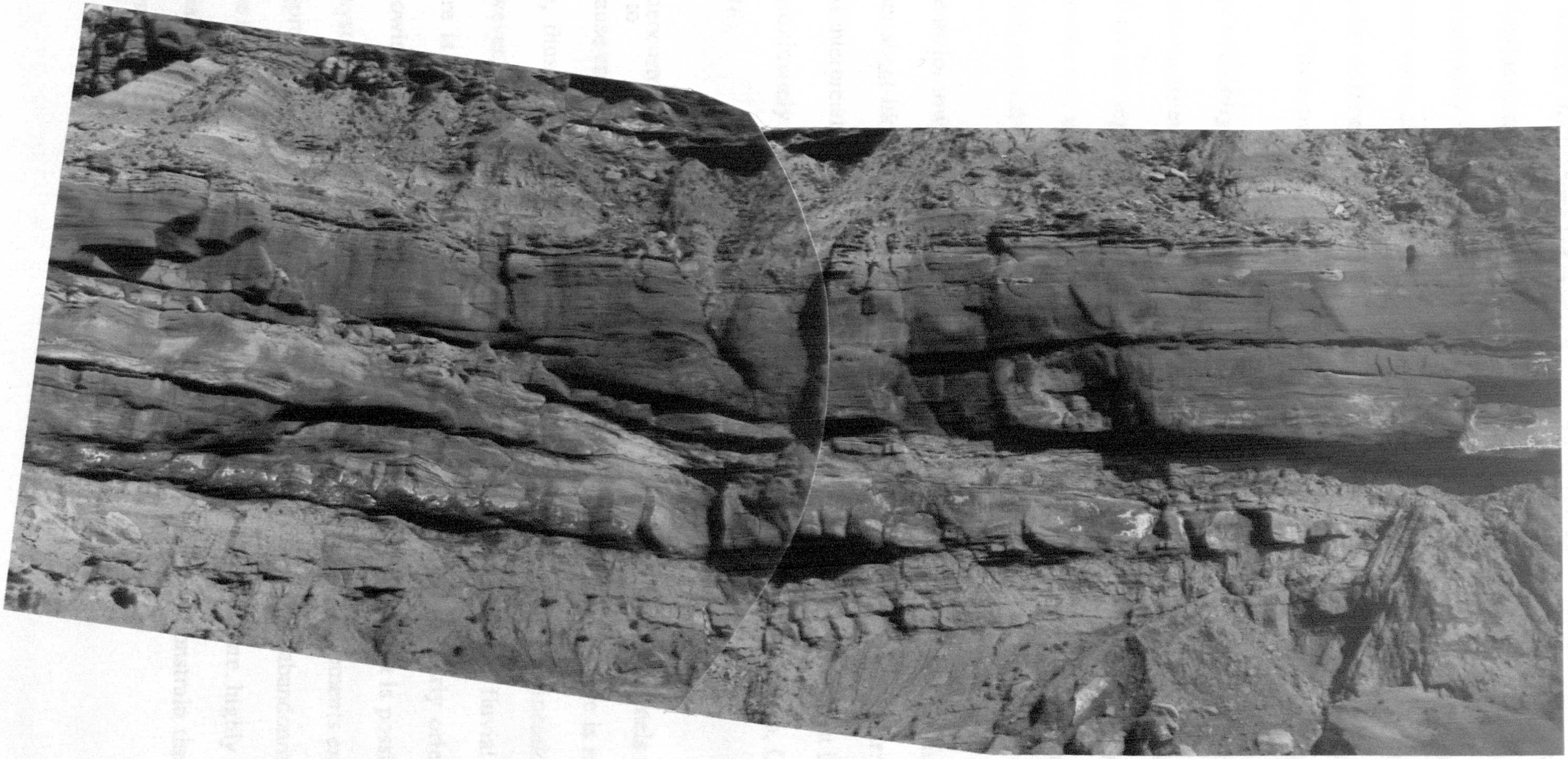


Figure 5.10 Photomontage of part of the sand body illustrated in Figure 5.9. Note: the two thin marine sands near the base of the section overlain by laminated facies; the low-angle lateral accretion surfaces which dip from left to right and truncate older sand beds; the thinly developed levee sediments at the top of the sand body. Scale =1m. Logs 2&3 (Figure 5.9) bracket this figure.

Discharge fluctuations led to much re-working of sediment during rising stages and subsequent rapid deposition may have been responsible for the chaotic units occasionally observed at the base of channel sandbodies. During low stages, the rivers were characterised by sluggish flows resulting in the deposition of thick muddy sediments. Minor increases in discharge moved relatively small quantities of sand. These subsequently deformed due to the soft nature of the muddy substrate.

The unusually thick mud unit between 3 and 4.75m in log 3 (Figure 5.9) requires some explanation. It is laterally coincident with a thick mud drape in logs 1 and 2 and clearly represents a major break in high stage flows and sand deposition. It is possible that this feature represents the partial abandonment of the channel course which was subsequently re-initiated by a large magnitude flood event.

The major low angle erosion surfaces in the channel illustrated, and in a number of other sheet-like sandbodies, clearly define lateral accretion or point bar surfaces. This interpretation is supported by the relative orientation of palaeocurrent indicators approximately at right angles to the dip of these surfaces (Allen 1970b, Collinson 1986).

Palaeocurrent data (Figure 5.1) collected from cross beds within channels is sparse because cross beds are not very common and bedding plane exposure is relatively rare, thus preventing the collection of cross bed and current lineation data. However, the limited data available is consistent with a meandering fluvial system. There is a considerable spread to the data with a strong southerly orientation. Following the interpretations made for the Porto Novo member, it is possible that the system was only moderately sinuous as there are few measurements extending the spread of data into the northern quadrants. There are also few abandonment fills in the section (see below). These are classically expected in mature, highly sinuous fluvial systems, where meander abandonment fills serve to constrain the lateral migration of the meander belt (see Collinson 1986).

In contrast to the sheet-like channels, the small-scale lenticular sandbodies rarely show any sign of lateral accretion and mud drapes are absent. However, it is not thought that these were subject to significantly different flow conditions. Their fill is dominated by parallel lamination, indicating predominantly upper flow regime conditions. Rapid deposition is also suggested by the horizons of disturbed laminae. This is probably due to the escape of trapped water during compaction (Leeder 1982, Allen 1985). Occasional horizons of ripple laminae indicate waning flows. Although mud-drapes are largely absent, it is still possible that these occurred but that they were re-worked during sustained flood conditions. Such conditions would also lead to the rapid disaggregation of any mudclasts present (Smith 1972). The presence of intensely bioturbated horizons, or exceptionally of whole channel fills, could be a consequence of extended breaks in sedimentation. This would allow invertebrates to extensively burrow a particular horizon. More diffuse burrow concentrations may therefore be a consequence of more persistent sedimentation.

The most likely explanation for the smaller scale channels is as relatively low sinuosity distributaries or crevasse channels branching off major trunk rivers represented by the large-scale sand bodies. Crevasse distributaries would have been characterised by considerable discharge fluctuations and high velocity flood discharges. They were probably relatively short lived features filling rapidly by vertical accretion of bedload.

An argument in favour of moderate to low sinuosity for both the sheet and lenticular sand bodies is the prevalence of upper flow regime bedforms. Significantly high stream velocities are not favoured by highly sinuous systems with tight meander loops (Turner 1986).

5.3.3 Abandonment Fills

As in the Porto Novo member, relatively few abandonment fills are seen. The author observed two in the 2.5km long section examined, one of which is described below. The other was inaccessable at the cliff top. Hiscott (pers. comm.), recorded two further examples in the 6km of section between the fault at GR 6405 2630 and Porto da Calada.

The abandonment fill which is well exposed crops out at the cliff base at GR 6525 3835. It may be a slightly unusual example inasmuch as it is bracketed by marine sediments. The base of the channel is incised into a bioturbated marine sand. Dark organic rich silts with *Diplocraterion* burrows, overlain by a fossiliferous marine sand, lie above. Channel sands contain coquinas of oyster shells. They are also burrowed, exhibiting sinuous epichnial traces very similar to recent gastropod browsing trails, (Figure 5.11). A number of straight, to sinuous crested, symmetrical ripple sets, exist on bedding surfaces. It is possible that these were produced by wind driven waves.

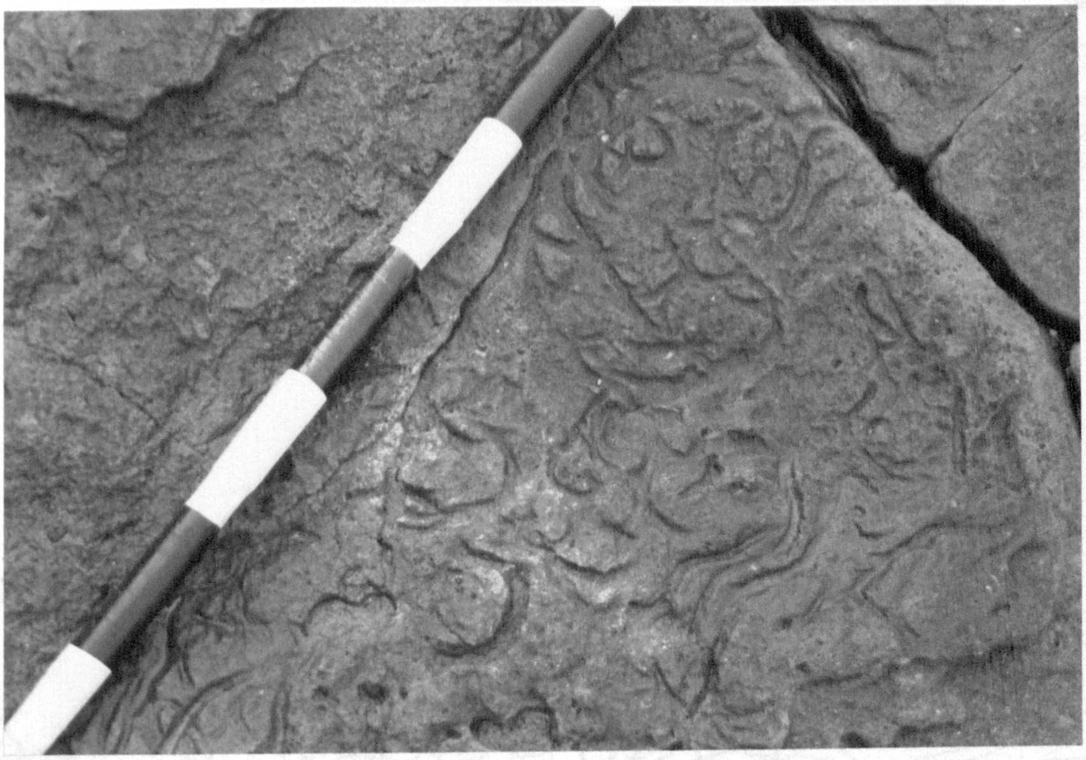


Figure 5.11 Gastropod ? browsing trails on a bedding plane within the abandoned channel's sandy fill. For location see Figure 5.12. Scale bars = 0.1m.

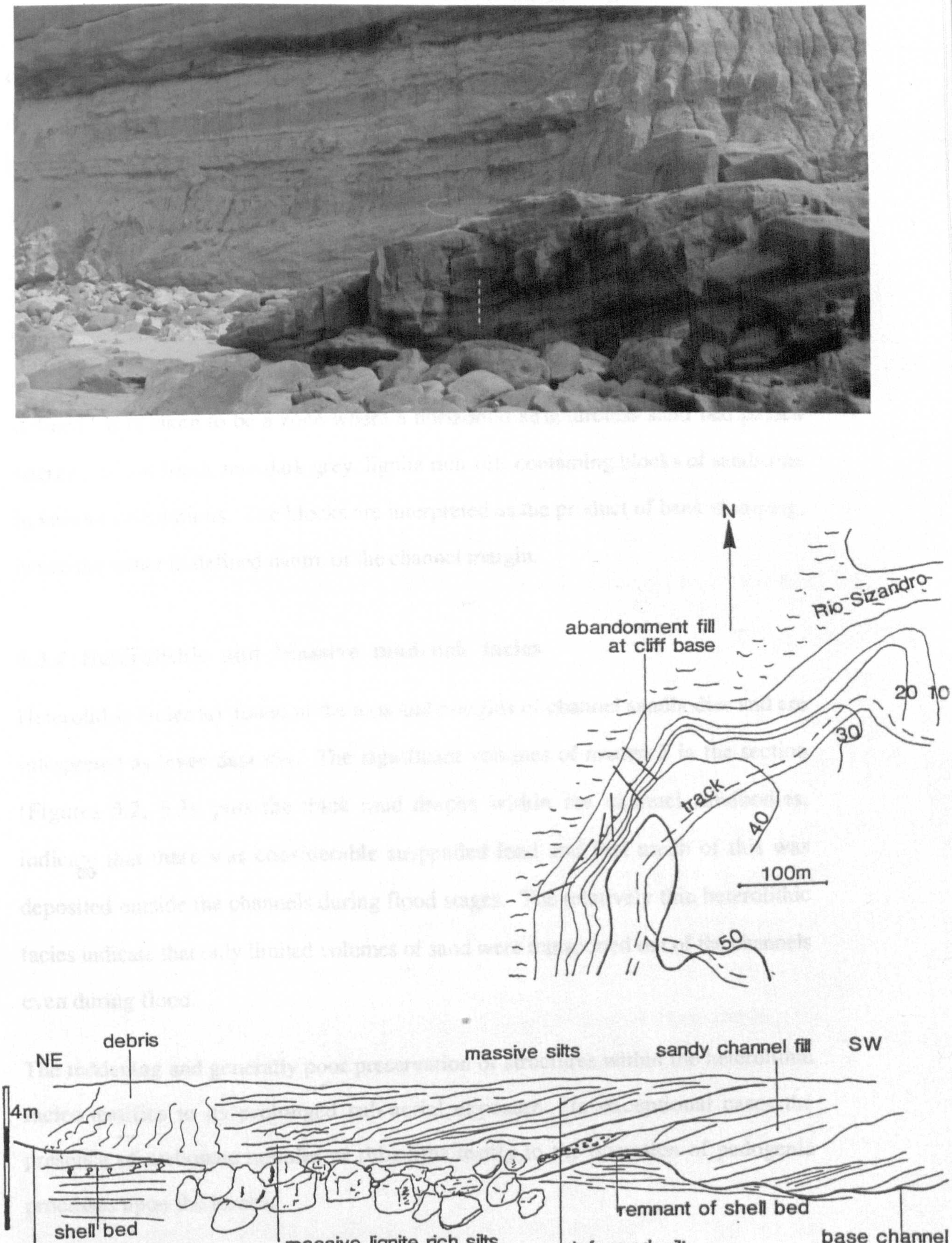


Figure 5.12 Abandoned channel at cliff base to the south of the Rio Sizandro: a) Location map with dip and strike measurements taken from the sandy fill, indicating the palaeochannel curvature and changing orientation of point-bar surfaces. b) photograph of the outcrop. c) labelled sketch of (b).

The 'normal' channel fill consists of a fining-up sequence from caliche pebble conglomerates to fine sands, some 4-4.5m thick. Well defined lateral accretion surfaces dip to the north at the northern-most extreme of the outcrop. The orientation of these surfaces changes southwards over some 175m, first into the cliff face (i.e. to the east) and then to the south (Figure 5.12a).

The abandonment fill (Figure 5.12b,c) consists of inclined beds of burrowed and deformed grey silts and fine sands which extend laterally for some 50m to the north of the sandy channel fill. An outer cut bank is not well exposed and rather ill defined. It is taken to be a zone where a horizontal structureless sand bed passes laterally, to the south, into dark grey, lignite rich silts containing blocks of sandstone in various orientations. The blocks are interpreted as the product of bank slumping, hence the rather ill defined nature of the channel margin.

5.3.4 Heterolithic and Massive mudrock facies

Heterolithic facies are found at the tops and margins of channel sandbodies and are interpreted as levee deposits. The significant volumes of mudrock in the section (Figures 5.2, 5.3), plus the thick mud drapes within the channel sandbodies, indicate that there was considerable suspended load and that much of this was deposited outside the channels during flood stages. The relatively thin heterolithic facies indicate that only limited volumes of sand were transported out of the channels even during flood.

The reddening and generally poor preservation of structures within the heterolithic facies testifies to its prolonged sub-aerial exposure. In exceptional cases the presence of carbonate nodules or rhizoliths testify to the operation of pedogenic processes upon the levees.

The development of pedogenic carbonate within the mudrocks is generally in the form of diffuse accumulations of carbonate nodules. This suggests that rates of sedimentation generally may have been relatively high (see discussion in Chapter 3).

Alternatively, the low elevation considered appropriate to the Assenta member environments of deposition, plus the presence of much lacustrine sediment (see 5.3.4), may indicate that floodplain sediments were water saturated much of the time, thus inhibiting the growth of pedogenic carbonate. The exceptional thin 'beds' of carbonate are unusual. However, Reeves (1976) has documented 'pseudobedding' in pedogenic carbonates where these have accumulated at discreet horizons in laminated lacustrine sediments. Though structures characteristic of lacustrine environments are absent in the associated silts, their presence elsewhere suggests that lakes did exist. Therefore it is possible that the operation of pedogenic processes have contributed to the destruction of primary depositional structures. Similarly, the development of a crust of caliche on the top of a channel deposit suggests that it remained either sub-aerially exposed, or only shallowly buried, for a significant period of time, perhaps hundreds to a thousand or more years (timescales after Reeves 1976; Machette 1985).

5.3.5 Laminated facies

As mentioned earlier, the laminated facies occurs in two facies associations. In the first of these structures observed include wavy and lenticular bedding. Units such as these are normally found above marine facies and below channel sediments or mudrocks exhibiting terrestrial indicators such as reddening or the presence of carbonate nodules (vis sections 5.2, 5.3.1). As this association occurs between marine and terrestrial facies it is interpreted either as mouth bars when overlain by channel sandstones (Fisk 1955; Coleman et al. 1964; Coleman and Gagliano 1965), or as muddy coastal margin sediments which are possibly tidally influenced where there is no evidence of an associated channel (Reineck and Singh 1980).

In the other facies association structures such as wavy and lenticular bedding are absent, with parallel laminae, reddening and simple invertebrate burrows more common. These units occur within wholly terrestrial parts of the succession and are often bounded by reddened mudrocks containing carbonate nodules thought to be of pedogenic origin (e.g.Allen 1974, 1986). A number of examples above such soil profiles are incised by the bases of channel sand bodies. Such a unit occurs at the base of the succession and includes the unusual deformation structures (Figure 5.6).

This second association, in which parallel laminae are particularly dominant, is thought to be of lacustrine origin. As the thickness of laminated sediment rarely exceeds 2m, the lakes were clearly rather shallow. Silty laminae are interpreted as the product of background sedimentation, punctuated by influxes of sand from rivers in flood. The occasional erosion surfaces observed could be the result of vigorous flood influxes subjecting the shallow lake floor to tractional currents.

The shallow lacustrine interpretation assists in finding an explanation of the deformation structures observed. They are consistent with loading from above causing deformation of several centimetres of sediment vertically and/or the lateral migration of sediment to form the 'humps' on either side. A reasonable explanation would be the footprint of a fairly large bi- or quadropedal animal, possibly a dinosaur.

5.3.6 Fossiliferous facies

The variable nature of these units seems to hold the key to the sedimentary environments existing, and possibly the external controls operating, during the deposition of the Assenta member. In the lower part of the section the facies have many similarities with those of the Praia Azul member. Fossil rich silts, silty biomicrites and bioturbated sandstones all have analogues described and interpreted in Chapter 4.

The marine horizons in the middle part of the section are all relatively thin, despite their strong marine character. Therefore, they all probably represent quite shallow marine environments which were starved of significant clastic input. Their widespread nature and consistent thickness suggests that both marine incursion and shoreline progradation were relatively rapid events

5.4 ENVIRONMENTAL INTERPRETATION AND DISCUSSION

Following the interpretations made in Chapter 4, the majority of the facies observed in the Assenta member outcrops are the deposits of delta top and alluvial plain environments. Major fluvial channels had a number of small-scale low-sinuosity crevasse channels or distributaries. These tended not to migrate laterally, were subject to a flashy, high velocity discharge and were filled largely by vertical accretion of bedload. They were probably relatively short lived features.

In contrast to the minor distributaries, the major channels migrated laterally, deposition usually occurring on point bars. However, the channels were not highly sinuous and meander abandonment or channel avulsion did not occur with great frequency. Following the interpretations given in Chapter 4 for the Praia Azul member as the deposits of a series of small, lobate, shoal - water deltas dominated by fluvial processes, the Assenta member systems, linked to this, formed the major fluvial distributaries on the delta top and on the alluvial plain behind (Figure 4.24).

The major channels were also subject to significant discharge fluctuations, possibly of a seasonal origin. High stages were often characterised by upper flow regime conditions, and low stages by sluggish flows and the deposition of thick units of fines from suspension and as low velocity tractional deposits.

Both delta top and alluvial plain were characterised by numerous shallow lakes. Parallel laminated silts and fine sands dominate the lacustrine fill. Sandy laminae and periodic, erosive-based thin sand beds reflect influxes of coarser sediment during flooding. The lakes were sufficiently shallow for their bottoms to be subject to mildly erosive tractional currents, associated with these turbid influxes.

The variable character of the marine units with time probably indicates a subtle change in the controls on base level. Initially, it seems likely that normal sedimentary processes such as delta lobe switching, were operating. The first marine unit in the section is relatively thick and though it rarely has a very strong marine character, is similar to marine units within the Praia Azul member. It probably represents the fill of an interdistributary bay which silted up, and was then cut through by a channel which was subsequently abandoned and finally re-submerged. The latter event probably reflects either the abandonment of the prograding delta lobe due to switching of the system to another position, or marine encroachment due to compactional subsidence locally exceeding rates of sedimentation.

Channels advancing into shallow marine areas carried significant volumes of fine sand and silt. These were deposited as thin mouth bars of laminated sediment. The presence of structures such as lenticular and wavy bedding in these and other units at marine-terrestrial transitions, suggests that some tidal currents may have been present (Reineck and Singh 1980). This contrasts with the situation existing during the deposition of the Praia Azul member (Chapter 4). The sediments of that member contain virtually no evidence of tidal activity.

The nature of marine units changes in the middle part of the section. They become consistently exceptionally thin and widespread. In particular the nodular carbonates can be traced for up to 6.5km along section. These are also interstratified with wholly continental deposits. There is no evidence for a deltaic origin for these associated clastic deposits (R. Hiscott pers comm.). A discussion of the potential causes of fluctuation in relative sea level appears in Chapter 7.

A return to deltaic deposits occurs in the top 55m of the section. Hiscott (pers. comm.) interprets the sequence containing dolomites and marine sandstones as the deposits of a delta system with associated brackish lagoons.

CHAPTER 6

A GRAVELLY FLUVIAL SYSTEM: THE SANTA RITA MEMBER

6.1 INTRODUCTION

The sandy fluvial/deltaic systems of the Assenta member, with their occasional short lived and widespread marine incursions, persisted for the remainder of the Jurassic in the southern part of the study area. According to Rey (1972), the units immediately beneath the unconformity at Porto da Calada, above which lie Cretaceous fluviatile sediments (see Chapter 1), are Berriasian in age.

In the north and northeast however, a completely different fluvial facies apparently existed at the same time. It seems to have extended no further south than an east-west trending 'high' between Santa Cruz and Torres Vedras. To the east of Torres Vedras Leinfelder (1986, 1987) recorded the presence of coarse pebbly strata in the NE part of the Arruda region (his 'Bombarral' formation; Figure 1.7).

The fluviatile sediments consist of gravelly sheet sandstones, which represent approximately 65% of the succession, plus maroon to grey, largely structureless, mudrocks (Figure 6.1). Large-scale cross beds, representing bars up to two metres or more in height are a prominent structure found within channel fills (Figure 6.2).

Inland, the Santa Rita member is the best exposed of all five members of the Lourinhã formation. Good outcrops appear in a number of quarries, where the associated red, clay-rich mudrock is excavated for brick clay. The area of outcrop inland lies dominantly to the south of Bombarral (Figure 1.7); only the older members of the Lourinhã formation which have survived erosion remain to the north. Some of the coarsest examples, with significant volumes of cobble-grade, matrix-supported clasts, appear to the east of the southern limb of the Serra de

Candeiros, in what Willis (1988) referred to as the 'Fraguas Monocline' (Figure 1.7).

Figure 6.1 Type section of the Santa Rita member south of Porto Novo. Base of the member is about 10m below the unusually thick sand body which is 14m thick. Sheet sand bodies account for 65% or more of the vertical profile mudrocks the remainder. The succession beneath includes the Praia Azul (below the first obvious sand body) and Assenta members. The section is located on the flanks of the Vimeiro diapir dips decreasing rapidly away from the structure. There is an unconformity across the Assenta/Santa Rita member boundary which is expressed as a difference in strike. The Santa Rita member dips into the cliff face whilst the section below clearly dips outwards; towards the sea.

There are three coastal outcrops: at Praia da Santa Rita, north of Ferrel, and north of Foz do Arelho (Figures 1.7, 1.22, 6.3). The 'type' section is incomplete, with 106m of sediment faulted against Cretaceous sediments by a major fault which also separates the Santa Rita member from the Assenta member in the western part of the basin. The succession has rather a low dip and a lateral extent of 1.8km. The base of the member is unconformable upon 120m of the Assenta member (see Chapters 1&5; Figure 6.1).

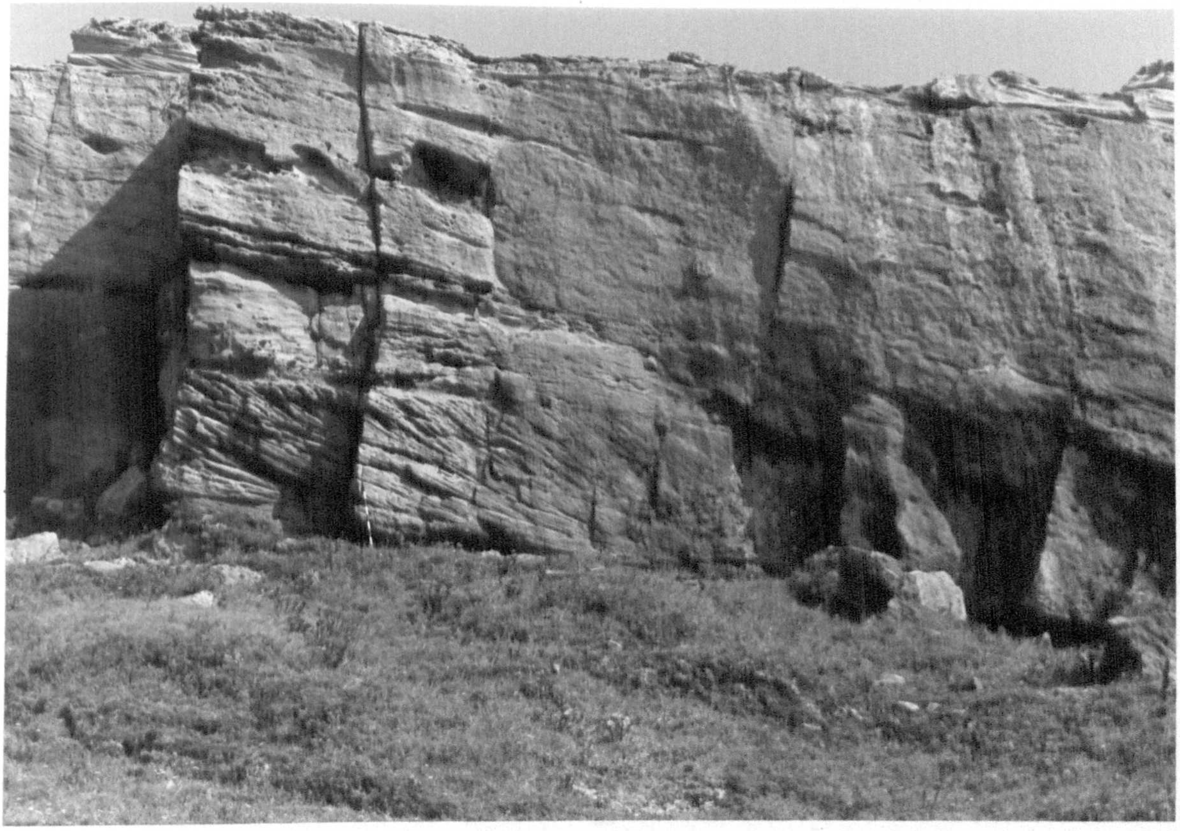


Figure 6.2 Two metre thick tabular cross bed at the base of a sheet sand body. Note the prominent low angle erosion surface 2m above the cross bed cutting down to the right. Palaeocurrents in the sand body above show little difference, suggesting that the erosion surface may be a lateral accretion surface. Section north of Foz do Arelho. Scale bar = 1m.

In the upper part of the section north of Ferrel, over 100m of the Santa Rita member crops out (Figure 1.22). It apparently rests conformably upon the Porto Novo member. A fault at GR 7360 5920 (Figure 6.3), across which correlation is not possible, breaks the section. Between GR's 7390 55950 and 7440 6000 the Santa Rita member is covered and additional faulting cannot be ruled out. However the top of the formation is exposed at this locality (see Chapter 1). Making no allowance for errors produced by faults (which, due to their probable normal character are likely to result in overestimates), the member is probably some 300m thick at Ferrel. This is the only coastal location where the top of the member is exposed.

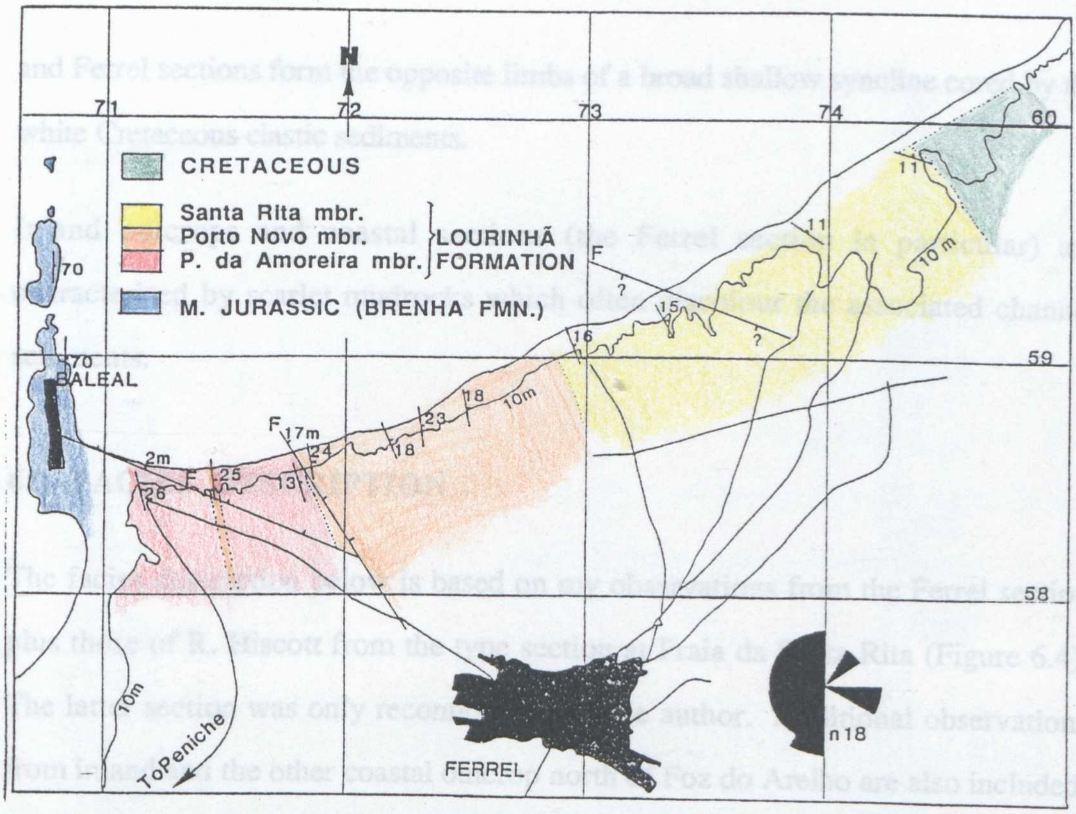


Figure 6.3 Map of the Ferrel section illustrating the distribution of the P.da Amoreira, Porto Novo and Santa Rita members of the Lourinhã formation, plus the Cretaceous (Torres Vedras formation) and Middle Jurassic (Brenha formation). Palaeocurrent data consists of a single measurement from each of the sand bodies in the Santa Rita member outcrop. It suggests a moderately sinuous system with a source to the west. The tongue of Porto Novo member strata within the P.da Amoreira member outcrop consists of a single 8m-thick sheet sand body. Grid lines are at 1km intervals.

To the north of Foz do Arelho, 300m or more of the Santa Rita member lie above Porto Novo member facies. The section is broken by the outlet of the Lagoa de Obidos, and Cretaceous clastics crop out to the south. A number of faults are suspected as the rocks are heavily jointed and fractured in places. This often is associated with poor outcrop, and consequently the faults cannot be proven. This section dips at about 20-30° off the flanks of the Caldas diapir, directly into the sea. Traversing the northern parts of the section is difficult and dangerous at all states of the tide. Because of its highly indented nature the section does, however, include some of the best three dimensional exposures. The southern-most parts of the section are accessible from Foz do Arelho except at high tide. The Foz do Arelho and Ferrel sections form the opposite limbs of a broad shallow syncline cored by the white Cretaceous clastic sediments.

Inland outcrops and coastal sections (the Ferrel section in particular) are characterised by scarlet mudrocks which often discolour the associated channel sediments.

6.2 FACIES DESCRIPTION

The facies description below is based on my observations from the Ferrel section plus those of R. Hiscott from the type section at Praia da Santa Rita (Figure 6.4). The latter section was only reconnoitered by the author. Additional observations from inland and the other coastal outcrop north of Foz do Arelho are also included. Where Hiscott's observations are included these wil be clearly identified in the text.

There are two main facies:

1) Sand and gravel facies;

2) Mudrock and sand facies.

The former is the most varied and has a number of sub-facies, namely;

- planar bedded pebble conglomerates;

- planar cross bedded gravels ;

- trough cross bedded gravels;

-planar bedded fine sands;

- structureless sands and gravels.

The facies analysis which follows is not as extensive as those given in earlier chapters largely due to the generally much simpler range of facies developed. It is also symptomatic of the less detailed analysis undertaken in the field.

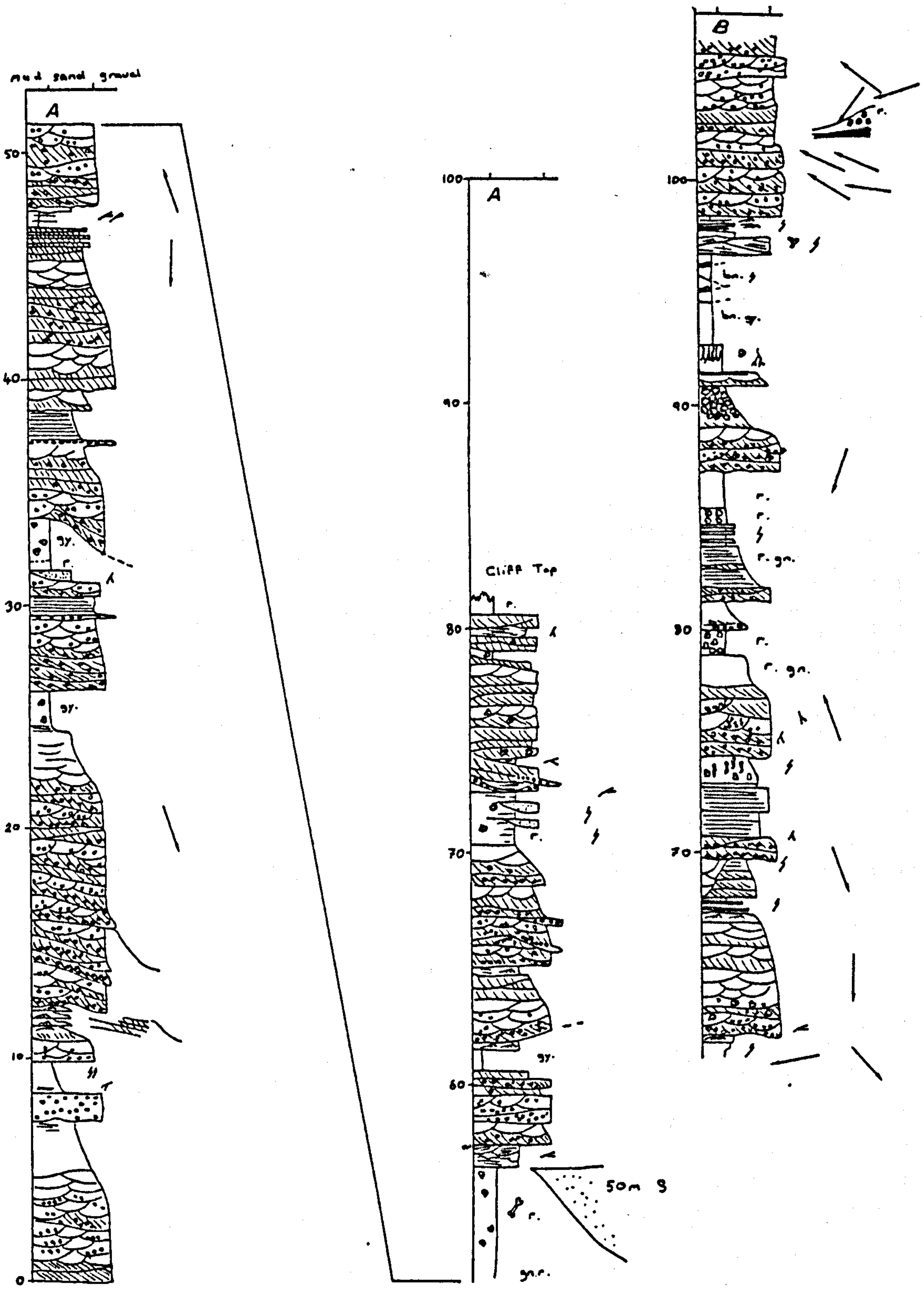


Figure 6.4 Graphic log for the 'type' section at Praia da Santa Rita redrawn from an original supplied by R. Hiscott. Note the dispersion of palaeocurrent vectors. Section 'A' was measured on the beach and up a cliff path immediately above the unconformity (see Figure 6.1). Section 'B' was measured several hundred metres to the south up a cliff path punctuated by red tile steps. Vertical scale in metres.

6.2.1 Sand and gravel facies

These facies occur in erosive based, sheet sand bodies, with a lateral extent usually in excess of that available at outcrop (i.e. in excess of hundreds of metres). This is particularly apparent in the Santa Rita section (Figure 6.1), where lateral extents in excess of a kilometer are not uncommon. The bases of sand bodies rarely exhibit significant relief, 0.5m is usually the maximum, and groove casts are a common feature. There are rare examples with highly irregular basal erosive surfaces.

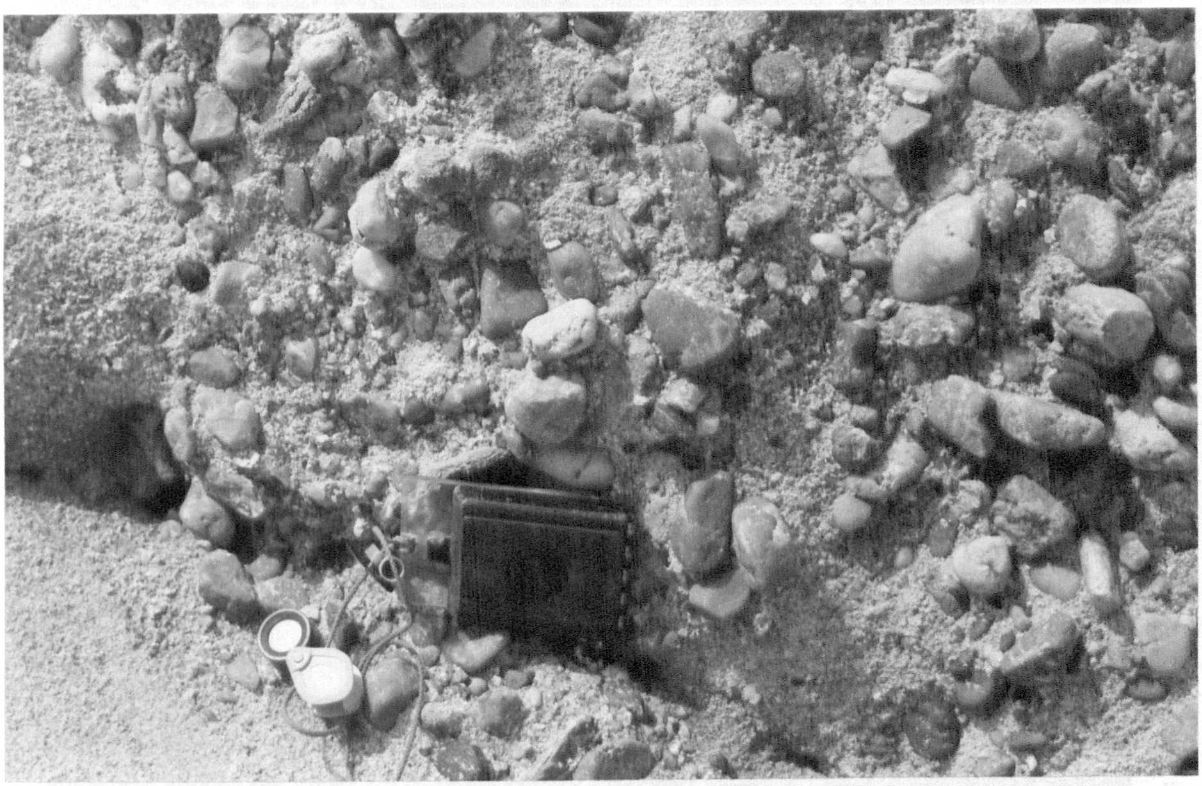


Figure 6.5 Typical lithology of the sand and gravel facies. Sub- to well-rounded quartz rich metasediment pebbles supported by a matrix of coarse sand.

Sand body thicknesses vary between 1 and 11 m, with most between 3 and 8m. A number are clearly multi-storey, with laterally persistent erosion surfaces separating units with differing palaeocurrent vectors. Pebble, cobble and occasionally, boulder grade mudclasts, are common in their lower parts and also above the erosion surfaces in multi-storey units

Lithologically, the sand and gravel facies consist of sub-to-well rounded pebble to cobble grade clasts, supported in a matrix of poorly sorted coarse angular sand (Figure 6.5). Clast types are more varied than those in other members, being dominated by quartz rich metasediments, plus granites and vein quartz. Sands are buff, bright yellow or red in colour. The latter may be due to infiltration of clays from the associated highly coloured mudrocks. This is considered a possibility because the sands are very poorly lithified and highly porous. Not even the concretionary cements found in the other, finer members of the Lourinhã formation are present. The buff coloured sands in particular are rather kaolinitic.

Figure 6.6 A number of excellently preserved dinosaur vertebrae on a bedding plane. Location approx. 300m N of Foz do Arelho.

The author has undertaken no further petrographic analysis. However, Hiscott (pers. comm.) reports the presence of berthierine cements in the topmost sand body in the Santa Rita section. This clay mineral, related to chlorites, is generally thought to be of marine origin (Odin and Matter 1981). The sand body from which these specimens are derived tends to have a different appearance lithologically. It is a dark yellow in colour compared to the buffs and greys of the sand bodies beneath. In addition there is an unusual shale unit beneath this sand body (described in section 6.2.2).

Plant debris is relatively rare, being less common in the Praia da Amoreira member only. Fossil logs are very rare indeed. Local concentrations of plant debris do appear, usually in finer lithologies. Other exotic clasts include dinosaur bones, in particular some exquisite vertebrae (Figure 6.6). These are most common in the Foz do Arelho section.

Mud drapes appear within the channel fills, though they are very thin, laterally impersistent, and rare. Clasts of pedogenically derived carbonate are locally common in the Santa Rita section (R. Hiscott pers comm.).

Planar bedded pebble conglomerates consist of thin sheets of clast to matrix supported conglomerate which are crudely horizontally stratified. Although most common in the upper parts of sand bodies, this facies does appear at lower levels, though it has never been observed at their bases. Planar bedded gravels are commonly interbedded with cross bedded facies. Beds are rarely more than 0.4m thick (Figure 6.7). This is not a common facies, representing only about 4% of the total thickness of sand/gravel facies logged.

Planar cross bedded gravels consist of planar tabular cross beds of matrix supported pebbles and cobbles, up to 2 metres thick. Mudclasts may be found on foresets. Set thicknesses are commonly between 0.3-0.4 metre.

Figure 6.7 Typical arrangement of structures in a sand body 20m above the base of the Santa Rita member north of Ferrel. A large tabular cross bed about 2m above the base, lies above an erosion surface marking the base of the upper tier of a multi-storey sand body. The cross bed is succeeded by trough cross beds of variable but upward decreasing scale. A veneer of horizontally bedded gravel lies at the top of the section by the metre pole.

Large scale cross beds are most often found at channel bases, with smaller scale sets and cosets appearing at higher levels. Toesets are usually angular, tangential geometries being relatively uncommon. A number of examples observed at Foz do Arelho do however have a strongly sigmoidal foreset geometry with locally, a few centimetres of parallel laminae beneath and in front of the structure. As discussed in Chapters 3 and 4, this probably reflects the presence of 'humpback bars' (Allen 1983). This is the second most common facies developed, representing about 27% of the coarse facies logged.

Trough cross bedded gravels represent the most common facies type developed. They represent some 55% of the sand/gravel facies logged. The size of trough sets varies considerably, from 0.1m thick up to 1m thick and 5-6m wide. The scale of cross beds often decreases upwards (vis. Figure 6.7). This facies is so common that it completely dominates a number of sand bodies and is the most frequent structure encountered in most others. There are local examples of trough shaped scour fills at sand body bases. Where these occur within multi-storey units, they may be up to a metre thick.

Planar bedded medium and fine sands is a restricted facies, largely confined to the topmost parts of sand bodies. Thicknesses vary from a thin veneer up to about 1 metre. Some examples are highly coloured (purple or red). This facies may grade into rather structureless sands.

Structureless sands are relatively uncommon, but do occur in the uppermost parts of sand bodies. Such sediments are often colour mottled and locally, exhibit a few faint invertebrate burrows. Caliche nodules are also present locally (R. Hiscott pers. comm.)

6.2.2 Mudrock and sand facies

This facies consists dominantly of brightly coloured mudrocks, plus limited volumes of sand in units too thin (< 1m) to be considered channel deposits.

Most commonly, scarlet or purple structureless silts and silty claystones lie abruptly over the coarse gravelly sediments. Often no more than a metre or two of silts occur between sand bodies, though thicknesses up to 5 metres appear locally.

Soil profiles are rarely well developed in the Ferrel section, though this is not the case in the type section (see below). At Ferrel, these consist of diffuse accumulations of small irregular carbonate nodules which may also be associated with mudcracks with sandy fills. The mudcracks penetrate up to half a metre into the silts, and usually appear below channel sand bodies: i.e. at the top of the mudrock profile.

In the type section at Praia da Santa Rita, better developed 'stage III' (Gile et al. 1966; Machette 1985) calcic soil profiles appear between most of the channels. A further feature of the mudrocks in this section, when not associated with calcrete, is that they may contain significant volumes of plant debris. Mudrocks in this association are usually grey.

There are rare examples of the silts exhibiting preserved structures. These usually take the form of parallel laminae. Sands associated with these silts may be packed with coarse plant debris. This is true of many of the thin, isolated beds of fine to medium sand which locally appear within the 'overbank' association. Structures are rarely preserved, but usually consist of faint parallel laminae or a series of cross-cutting scours with concentric drapes. One sand bed exhibits basal load casts with a relief of about 0.2m, on its base. Hiscott (pers. comm.) reports the presence of an unusual mudrock unit immediately below the last sand body in the Santa Rita section (92.5-98.5m Figure 6.4). Features of this unit include; broad low angle scours, climbing ripple laminae and possible *Cruziana* facies (Frey and Pemberton 1981) burrows. A palynological sample suggested stagnant anoxic conditions, possibly in a marsh.

There are a few examples of decimetre-scale sand beds interbedded with mudrocks directly above sand bodies. One of these heterolithic units consists of a metre thick packet of very fine to medium sands some of which fill small, concave-up scours a few metres in width. The entire unit fills a scour in the top of a sand body which has a width of tens of metres (only one margin is exposed). The sand beds lap out onto the margins of the scour and are cross cut by burrows, consisting of sinuous, sub-vertical tubes up to 12mm diameter. A further example is thicker, (2.5m) and lies above a relatively thin gravel unit. The lateral extent of this example is not apparent, but is thought to be several tens of metres. The spacing of sand beds increases upwards. None of the sand beds in either example, exhibit any primary sedimentary structures.

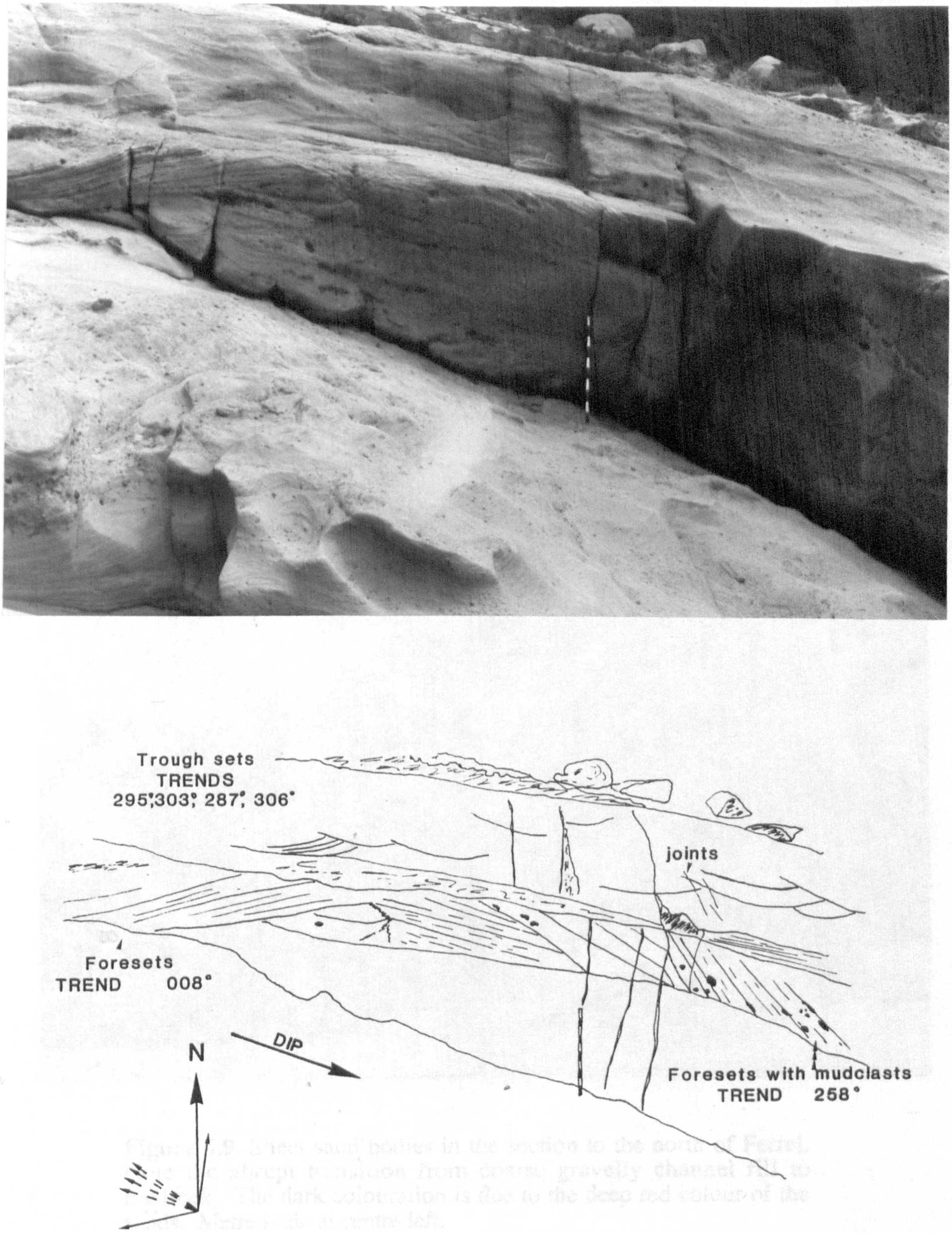


Figure 6.8 Photograph and sketch illustrating facies relationships observed to the north of Foz do Arelho. A planar cross bed with complex foresets which have diverging dip orientations is overlain by trough cross beds the foreset orientation of which is between those of the foresets below. The vector diagram indicates the relative orientation of the planar (solid lines) and trough (dashed lines) cross bed foresets. Such a relationship suggests that large tabular bars were mantled by sinuous dunes. Metre pole for scale.

6.3 FACIES RELATIONSHIPS

Because of the relatively smaller range of facies types present, facies relationships in the Santa Rita member appear rather simple.

A degree of fining-up is normally only observed in the upper parts of a number of sand bodies. This may be represented by a decrease in the proportion of pebble grade material or by the appearance of sandy lithologies. Fining-up with a progressive decrease in grade is best developed within the Santa Rita section.

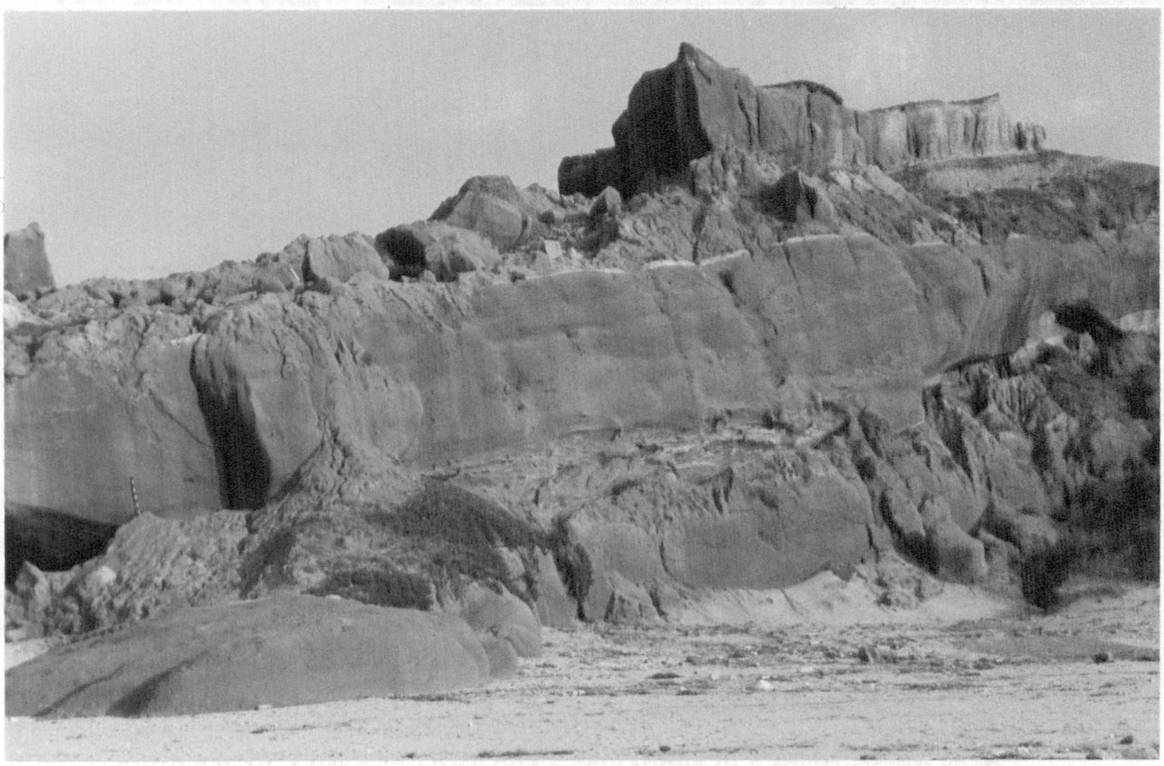


Figure 6.9 Sheet sand bodies in the section to the north of Ferrel. Note the abrupt transition from coarse gravelly channel fill to mudrock. The dark colouration is due to the deep red colour of the sands. Metre scale at centre left.

Large scale (>1m) tabular cross beds are most common at the bases of sand bodies (Figure 6.2). These most often pass up into dominantly trough cross bedded gravels in the middle regions of the sand bodies (Figure 6.7). A few well exposed examples of these facies, north of Foz do Arelho, exhibit an interesting association

(Figure 6.8). A large planar tabular cross bed has complex, composite foresets, the dip orientation of which vary by over 100° (see palaeocurrent vectors in Figure 6.8). Trough cross beds which overlie this structure have orientations within this range. Such a relationship suggests that sinuous dunes may have mantled large bars on the river bed. Similar successions have been recorded in a number of modern examples (Collinson 1970; Bluck 1971; Jackson 1976; Levey 1978).

Figure 6.10 Opposed cross bed orientations in a multi storey sand body. The cross beds are separated by an erosion surface which passes the metre pole about 35cm from its top. A few mudclasts lie on this surface. Evidence such as this leads to the suggestion that the fluvial system was sinuous as opposed to low sinuousity (braided).

The topmost parts of sand bodies are composed of a mixed, cross bedded and parallel laminated gravel and sand facies. Transitions into muddy facies are commonly abrupt (Figure 6.9) and may be associated with structureless and discoloured sands and gravels at the tops of the sand bodies.

Erosion surfaces between storeys in a multi-storey sand body are irregular and may have a relief of up to a metre. Their recognition is aided by an examination of palaeocurrent vectors which may indicate considerable variability above and below such surfaces (Figures 6.3 and 6.10). In a few examples, erosion surfaces have a persistent dip which has an angle of only a few degrees. Palaeocurrent vectors in these units tend to show little difference above and below the erosion surfaces.

6.4 INTERPRETATION

The association of locally unidirectional cross strata in the sand bodies, plus terrestrial indicators such as reddening and carbonate nodules in the mudrocks indicate that the deposits of the Santa Rita member are those of a fluvial system. Leinfelder (1986,1987) suggested that the sediments were the deposits of low-sinuosity 'braided' streams (see Miall 1977). This conclusion probably follows a preliminary interpretation by Wilson (1979) of the Santa Rita section (his facies vii) as the inland outcrop that Leinfelder studied was rather poor (pers. comm.). However, palaeocurrent patterns recorded north of Ferrel (Figure 6.1) suggest a high dispersion between channels, though locally within a single channel variance is relatively restricted. Such patterns are usually associated with sinuous fluvial systems (Collinson 1978,1986).

It has been clear for some time that gravelly, sheet-like sandbodies are not neccessarily the deposits of low-sinuosity 'braided' rivers and that fining-up is often not developed in sinuous/meandering systems (McGowen and Garner 1970; Jackson 1978; Collinson 1978). In addition, the absence of clear epsilon cross-strata cannot be cited as evidence for a fluvial system not to have been meandering (Allen 1970; Collinson 1978). The identification of lateral accretion bedding is usually aided by the presence of a heterolithic channel fill, mud drapes and cross cutting erosion surfaces. In the absence of such a variable fill in coarser grades of sediment which are characterised by very low angles of point bar dip (Bluck 1971; Gustavson 1978; Campbell and Hendry 1987), the identification of lateral accretion surfaces becomes difficult. However, lateral accretion surfaces have been observed in some modern gravelly systems where gravel sheets, rather than bars with steep slip faces, have developed (Campbell and Hendry 1987). Further complications may arise because of the development of terraces in coarse grained point bars (McGowen and Garner 1970). Examples of cut banks at Santa Rita and north of Ferrel do show terracing and, if these are inner banks, suggests that some analogy with the systems documented by McGowen and Garner (1970) may be possible..

Therefore, there is some limited sedimentological evidence for the Santa Rita member being the product of a moderately sinuous (meandering ?), gravelly fluvial system. There are a few low angle erosion surfaces which, coupled with the palaeocurrent data, strongly suggest sinuous laterally migrating channels. The palaeocurrent data, though limited, comprises a single or mean measurement from each channel sandbody, or storey, where measurement was possible. A similar pattern is produced by data (also sparse) collected from the Santa Rita section (R. Hiscott pers. comm.; Figure 6.4). It suggests highly dispersed palaeochannel orientations.

The lower parts of the palaeochannels were locally dominated by large bars, probably with a crest height of 1-2 metres. Bar morphologies probably resembled the linguoid or transverse bars commonly associated with braided systems, with obliquely facing foresets (Figure 6.8). These bedforms were probably mantled by smaller scale, sinuous crested dunes. The repetition of this facies association suggests that such bedforms were superimposed upon one another up the face of the point bar, thus resulting in terracing, or the production of 'platforms' (Figure 6.11). Mid-point-bar regions, appear to have been dominated by a mixture of smaller-scale straight and sinuous crested dunes. Where the major cross beds are absent, decimetre-scale cross bed sets suggest that the entire lower and middle point bar face consisted of sinuous crested bars.

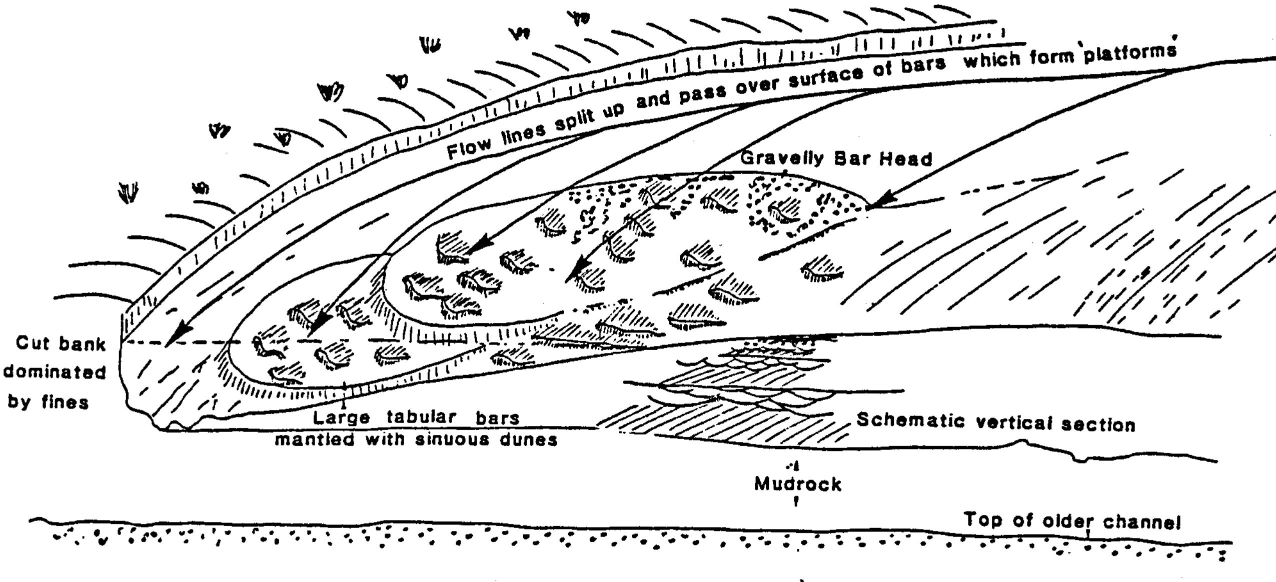


Figure 6.11 Environmental model for the deposits of the Santa Rita member and the resulting vertical profile.

The presence of poorly to horizontally stratified pebble conglomerates at higher levels suggests that these formed 'armoured' veneers on the upper parts of point bars, perhaps mantling other bedforms, or flooring chute channels. Parallel laminated sands may have been deposited in a similar position, perhaps forming the upper margins of point bars or downstream bar 'tails' which tend to be finer grained in such rivers eg.the Endrick in Scotland (Bluck 1971), Nueces in Texas (Gustavson 1978) and Babbage in western Canada (Nanson 1983).

Structureless and discoloured sediments, being generally confined to the topmost parts of the sandbodies, are probably the result of pedogenic modifications.

Overbank associations generally exhibit strong evidence for active pedogenesis, indicated by their frequently intense colouration, the absence of primary sedimentary structures and particularly at Santa Rita, in the presence of calcic soil profiles. These calcretes are very similar to those from older sediments in the Lourinhã formation, and indicate a similar, semi-arid/sub-tropical climate, with marked seasonality.

The presence of very little levee-type facies does not pose a problem for the meandering fluvial interpretation for the member (see Jackson 1978). This in fact is probably due to the coarse nature of much of the channel fill. Overbank sedimentation was therefore largely confined to suspended sediment. The presence of mudrocks testifies to the existence of significant suspended load.

The unusual heterolithic scour fill described above was probably an abandonment fill. The limitations of the outcrop prevented the determination of whether the unit thickened or terminated laterally. Arche (1983) described similar facies from recent systems in Spain.

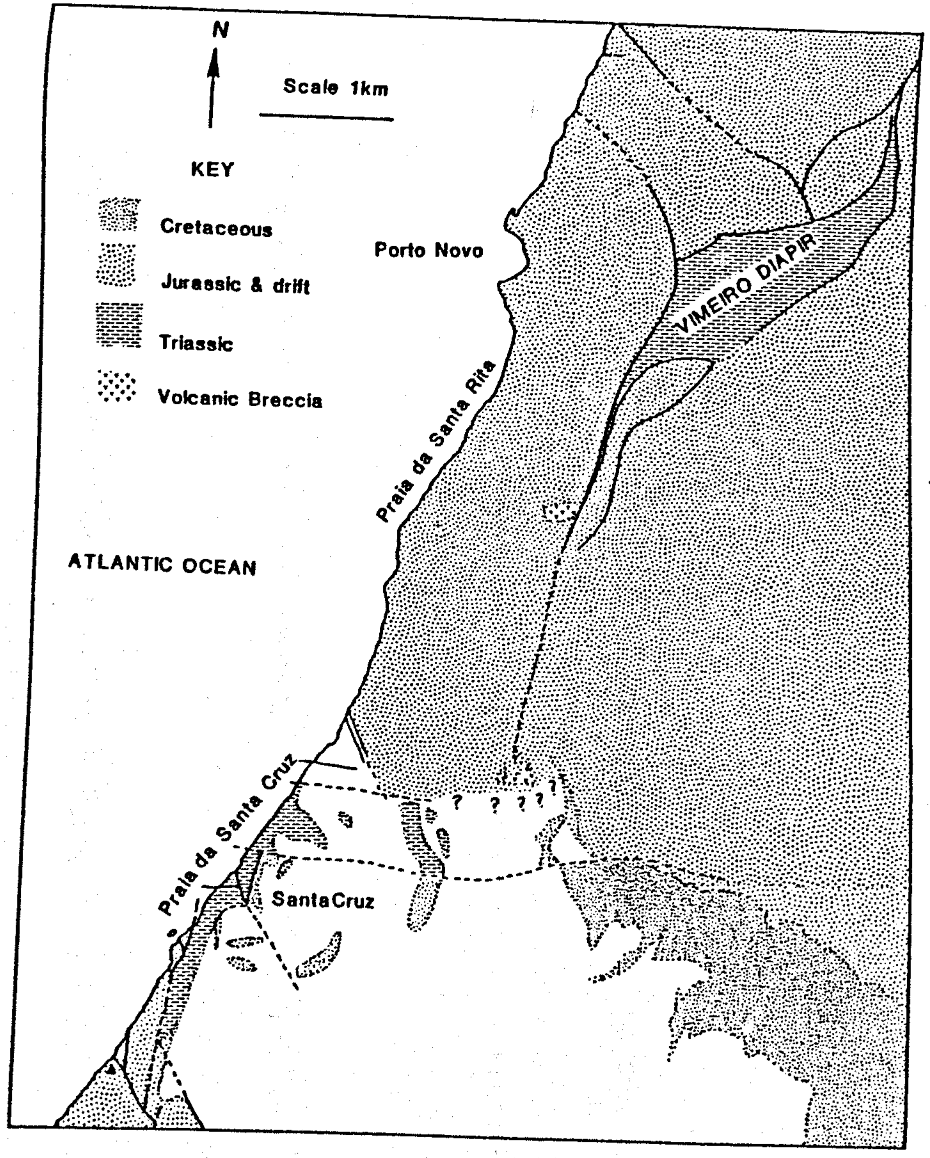
The rare occurrences of structures within the silts, and the association with laminated and deformed sands and plant debris, suggests subaqueous deposition, probably in small lakes on the floodplain. This seems the most likely origin for the facies documented by Hiscott at Praia da Santa Rita. The scours observed may be the result of floodwaters spilling into a shallow lake and subjecting its bottom to tractional currents.

A low incidence of mud drapes within channel sand bodies need not be an indicator of infrequent discharge fluctuations. Such conditions are suggested by the climatic indicators discussed above. Thin, fine sediment drapes would have a poor preservation potential within channels having flashy discharge and poorly cohesive bedload sediment. Thus virtually all fine sediment was preserved on the floodplain.

The presence of the berthierine cements within the topmost sand body at Praia da Santa Rita, if they are marine, are a little difficult to explain. It is possible that ingress of saline ground waters occurred. Theşe may have infiltrated during one of the marine transgressions recorded by the deposits of the Assenta member.

6.5 TECTONIC SETTING

The most interesting and unusual aspect of the Santa Rita member is its restricted distribution in the western part of the basin. Outcrops are confined to inland and coastal areas north of a fault which runs from Torres Vedras westwards, intersecting



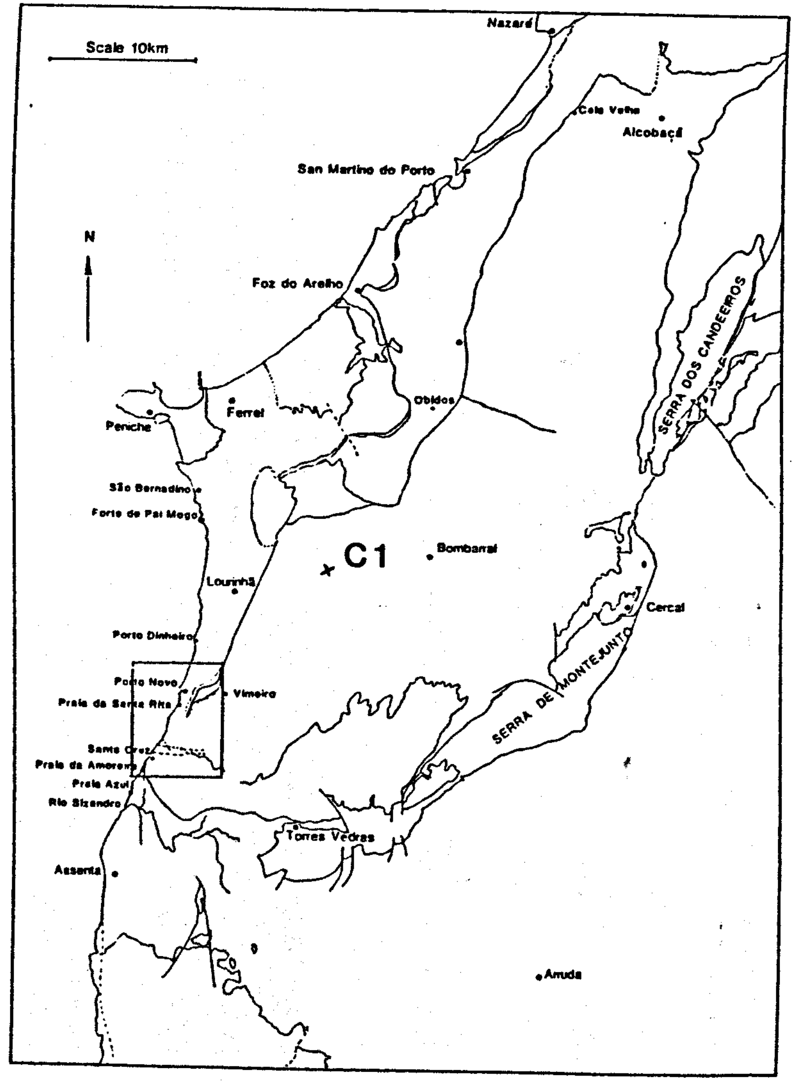


Figure 6.12 Maps illustrating the position of the faults which mark the southern limit of the Santa Rita member outcrop on the coast. The fault zone marks the intersection, on the coast, of a fault which extends westwards from Torres Vedras, with the Caldas diapiric trend. Much of the solid geology is obscured by recent colluvium making interpretation of the relationships difficult. The regional geology is illustrated on Figure 1.7 which uses the same base map as the right hand map above. 'C1' marks the location of the Campelos 1 borehole.

with a complex zone of faulting and diapirism on the coast around Santa Cruz (Figures 6.12, 1.10). Inland, this fault is very poorly exposed and does not express itself in any topographic manner. It is also partly covered by drift. The principle evidence for its existence is the juxtaposition of white, Cretaceous, continental clastics to the north, against Praia Azul member facies. The distribution of coarse facies mapped by Leinfelder (1986, 1987) indicate that coarse grained sediment did occur in the NE and E parts of the Arruda region (Figure 1.7).

The southern-most coastal outcrop of the Santa Rita member occurs at GR 6788 3363, to the north of Praia de Santa Cruz. Cretaceous clastics are downfaulted to the south. This locality sits at the northern margin of diapirically disturbed and faulted sediment which extends 1.75km southwards to Santa Cruz (Figure 6.12),(Camarate Franca et al. 1961). Here, Triassic continental sediments are faulted against marine slope facies of the lower Kimmeridgian Abadia formation (Ellwood 1987). The published map shows isolated blocks of Cretaceous and Triassic sediments much of which is covered with drift.

To the south of this line, only facies of the Assenta member occur above the Praia Azul member. As indicated in Chapters 1 and 5, about 120m of the Assenta member succeeds the Praia Azul member at Praia da Santa Rita. The Santa Rita member is situated unconformably above the Assenta member. The unconformity was probably due to syn-depositional uplift on the flanks of Vimeiro Diapir. This movement is most likely intra-Tithonian in age (see Chapter 1). It is not possible to demonstrate a similar unconformity at the base of the Santa Rita member in either the Foz do Arelho or the Ferrel sections.

The position and orientation of the zone separating the Santa Rita member from the finer facies developed to the south is the same as that which separated the Bombarral sub-basin from the Turcifal sub-basin during the deposition of the Abadia formation (Figure 1.3). The fundamental difference between the basins in the early Kimmeridgian was in their structural styles (Chapter 1). The Bombarral sub-basin was characterised by salt withdrawal and rapid subsidence. In contrast, the Turcifal and neighbouring Arruda basins were half graben structures (Ellwood 1987; Wilson et al. in press). The nature of the divide between the northern and southern sub-basins is poorly known for the Lower Kimmeridgian (Ellis et al. in press).

It seems that the differentiation into sub-basins may have occurred, at least partially in the west, once more during the Tithonian with the Santa Rita member restricted to the former Bombarral sub-basin, plus areas to the east of the Serra de Candieros, the northern and eastern parts of the Arruda sub-basin and west of the Caldas diapir. Over 1300m of Lourinhã formation sediments, all younger than the Santa Rita member, appear in the Campelos 1 borehole (RCL Wilson pers comm.)(Figure 6.12 map), compared to a probable maximum formation thickness of 1300m on the coast at Foz do Arelho. As some movement can be demonstrated on the flanks of the Vimeiro structure, it is probable that salt withdrawal was actively occurring in the axial zone of the sub-basin.

Inland, mudrocks are apparently more prevalent in the section than on the coast. There are no brick clay quarries near coastal outcrops of the Lourinhã formation which are near the flanks of the salt diapirs. This also suggests that mudrock may be less common in these coastal areas. Numerous models (e.g. Allen 1978; Leeder 1978; Bridge and Leeder 1979) have shown that as subsidence, and therefore alluviation rate, increases, the proportion of overbank mudrocks preserved is also likely to increase. It may be that field observations of the proporton of mudrock to channel sediment have been biased in this study, due to the location of the main coastal sections on or near the flanks of diapirs where the reduced rate of subsidence tended to *increase* the sandbody density at the expense of overbank sediments. A tendency to produce multistorey sandbodies would also be a consequence of the slower subsidence (Allen 1978).

Most of the pebbles and cobbles of quartz rich metamorphics in the Santa Rita member are sub- to well-rounded, suggesting considerable transport. Such material was not observed to be abundant in the westerly derived clastics of the Praia da Amoreira fan systems (Chapter 2). In addition, some of the coarsest fractions in the Santa Rita member were found to the east of the Serra de Candieros. Basement rocks to the east include Lower Palaeozoic quartzites (Ribeiro et al. 1979). In view of this, and where the coarser material was observed, it seems possible that much of the sediment was derived from the east. The distribution of facies mapped by Leinfelder (1986, 1987; see Chapter 7) supports an easterly source. This implies some rejuvenation of sourcelands, perhaps due to footwall uplift during continued extension of the basin. Unfortunately, outcrop is so poor that the significant number of palaeocurrent measurements required to give some strength to the arguments regarding sourcelands are unobtainable.

The final question that requires answering is, 'what was the nature of the structural divide between the Bombarral and Turcifal sub-basins ?'. Two factors may have been responsible. It has already been indicated that salt movement was occurring during the Tithonian and it is known that salt lies beneath the Serra de Montejunto (RCL Wilson pers comm.). It is possible that a westward extension of this salt distribution may have provided a 'high' which formed a partial barrier to the southward migration of the Santa Rita fluvial systems in the western part of the basin. An alternative solution might be that the fault, illustrated in Figure 6.12, for which there is evidence of more recent activity, was active during the Tithonian, with downthrow to the north (its present day sense) merging westwards with the southern limb of the Caldas diapiric trend. These are speculative solutions in the presence of the circumstantial evidence discussed above. The nature of the divide seems no more certain for these Tithonian successions than it does for the earlier basin sub-division in the Lower Kimmeridgian.

6.6 CONCLUSIONS

The coarse gravelly fluvial systems of the Santa Rita member were probably characterised by sinuous channel courses. It is likely that point bars existed, with deposition in their lower parts dominated by straight crested or transverse bars mantled with sinuous dunes. A mixture of smaller straight and sinuous crested dunes occupied the mid point bar regions. Chute channels with a gravelly base and parallel laminated sands occupied upper bar regions, and probably bar tails also.

Overbank sedimentation of coarse fractions was very restricted due to their large grain sizes. Clay rich mudrocks, often highly oxidised, were the primary floodplain deposits. There is some evidence for small lakes having existed in places on the floodplain. Elsewhere pedogenic processes operated and these resulted in the formation of caliche soil profiles. In the Santa Rita section, where these are particularly well developed, re-worked caliche pebbles are found within channel sandbodies.

Santa Rita member facies differ markedly from those of the Assenta member and are not found south of a line between Santa Cruz and Torres Vedras. This line also marked a basin sub division during the lower Kimmeridgian. At their closest approach, on the coast, Santa Rita and Assenta member facies of probably similar age are only some 6km apart at the present day. This seems an impossibly short distance to imagine a decrease in mean grainsize from gravel to medium sand. It is therefore thought that the two environments were separated by a structural divide, controlled by faulting and/or salt tectonics.

In the Bombarral sub-basin differential subsidence probably occurred between the basin axis and 'highs' over salt structures, thus producing differences in the thickness of the sediment pile and the preservation potential of overbank sequences between these two areas.

The nature and shape of most of the large clastic grains suggests considerable transport. Their apparently coarser nature to the east, overall distribution and the limited palaeocurrent data available suggests a source from the interior of the Iberian Meseta, rather than the basin's western margin.

CHAPTER 7 CONCLUSIONS

7.1 INTRODUCTION

The purposes of this chapter are five fold:

1) to summarise the palaeoenvironmental models presented in Chapters 2-6;

2) to discuss the palaeogeography of the Lusitanian Basin during Kimmeridgian /Tithonian times;

3) to place the events in a plate tectonic setting;

4) in the light of (3), to discuss the dominant controls on the distribution of sedimentary environments and the limitations of these speculations;

5) to conduct a comparison of the succession with those found in basins on the Western Atlantic margin and briefly consider their importance in the context of the North Atlantic.

7.2 PALAEOENVIRONMENTS

7.2.1 The Praia da Amoreira member

This is probably the most restricted of all the members of the Lourinhã formation. The muddy distal alluvial fan sediments are confined to the western margin of the basin, sections outcropping from Ferrel south to Praia da Amoreira. The fans were sourced from the western margin of the basin and the environment of deposition was characterised by flashy sheet-flood sedimentation which tended to precede the establishment of channelised flow. Switching of the active area of sedimentation on the surface of the fans produced a repetetive fining-upward cyclicity. This autocyclically controlled process overprinted a mega-scale fining-upward cyclicity which is attributed to periods of increased activity along the basin margin faults.

7.2.2 The Porto Novo member

The Praia da Amoreira member does not extend far inland, nor does it appear in the section south of San Martinho do Porto. It is replaced both laterally and vertically by the meandering fluvial deposits of the Porto Novo member.

The environments of deposition of the Porto Novo member were characterised by a major trunk river 5-9m-deep with a number of tributaries (usually 3-4m-deep). These were subject to significant discharge fluctuations resulting, during floods, in the deposition of re-worked pedogenic carbonate nodule and mudflake conglomerates and current lineated sandstone and, at low stages, thick mud drapes which were locally desiccated on sub-aerial exposure. Channel sand body dimensions suggest that the system may have had a channel length of about 500km and drained a basin area of 15-35 000 km^2. Locally, there is direct evidence for syn-sedimentary extension, with multistorey sand bodies present in hanging walls close to normal faults, and small-scale extensional faults present below channel sand bodies. A significant increase in the thickness of the Porto Novo member occurs northwards.

7.2.3 The Praia Azul member

In the southern part of the study area, continental facies are interspersed with the deltaic and low energy mud-dominated shoreline deposits of the Praia Azul member. The establishment of these environments was due to a (basal) Tithonian rise in relative sea level. The palaeoshoreline was characterised by delta distributaries and crevasse channels which discharged significant quantities of suspended sediment into shallow brackish and marine bays. There is no evidence in the marine strata for water depths greater than a few metres. As the transgression (probably) occurred over a broad, low-gradient alluvial plain, it was likely to have been rapid and resulted in a broad shallow marine shelf over which the delta system prograded rapidly.

As marine environments were shallow, marine circulation was restricted. This was enhanced by the presence of patch reefs and shell banks. There is very little evidence of re-working of sediment, save for the presence of a few erosive based storm beds and, locally, sandy facies interpreted as the deposits of shoreface and beach environments. The broad, shallow and restricted nature of the marine depocentres associated with the deltas allowed fluvial discharge variations to cause salinity fluctuations in these areas.

7.2.4 The Assenta member

The lower delta plain environments of the Praia Azul member passed landwards into the upper delta plain/fluvial environments of the Assenta member. The deposits of this unit are essentially those of trunk distributary and, locally, crevasse channels which may have been a few kilometres to tens of kilometres from the marine depocentres. Lacustrine sediments feature significantly in the sediment pile. Like the Praia da Amoreira and Porto Novo members, sand bodies in the Assenta member exhibit evidence for fluctuating discharges, which probably had a climatic origin.

The largely continental facies are punctuated by thin marine units which bear no evidence of a deltaic association. The strong marine signature and dearth of clastic sediment associated with nodular carbonates, and a coral bed in particular, suggest that potential sources of clastic input had been displaced a considerable distance landward by rapid relative rises in sea level.

7.2.5 The Santa Rita member

Both the Porto Novo and Assenta members passed rapidly into a gravelly continental fluviatile succession in the northern and eastern parts of the basin. Locally, e.g. on the flanks of the Vimeiro diapir (Figure 1.7), this transition occurs across a small angular unconformity. Lithologically the Santa Rita member is quite unlike any of the other members of the Lourinhã formation. It is much coarser grained and the sub-to well-rounded shape of the quartzose clasts suggests significant transport. Palaeocurrents suggest a moderately sinuous meandering system with a source to the north and/or east. In addition a degree of coarsening is detected to the east. This evidence plus the facies distributions mapped by Leinfelder (1986, 1987) suggests an eastern origin, from the interior of the Iberian Meseta, for this clastic sediment.

7.3 PALAEOGEOGRAPHIES

Figures 7.1A-C are a series of palaeogeographic maps representing facies distributions at three points during the deposition of the Lourinhã formation. In addition to field data gathered by the author, detail has ben provided by an examination of Fürsich and Schmidt-Kittler (1980), Felber et al. (1982), Ellis (1984), Leinfelder (1986, 1987) and Ellwood (1987).

The first of the maps covers the late Kimmeridgian. It shows the position of the basin's western (faulted) margin from which the Praia da Amoreira fan systems prograded. These pass laterally into the meandering fluvial systems of the Porto Novo member which existed in northern, central and eastern areas. Though data from the eastern outcrops inland is somewhat sparse, there is no evidence for a coarse clastic influx from the east. In this direction it is not known just how far to the east the basin margin extended, largely because Mesozoic sediments have either been eroded from the basement where this has been inverted, or are blanketed by thick Tertiary cover. However, in the south-eastern part of the outcrop area Leinfelder (1986, 1987) documented a belt of mixed carbonates and deltaic clastics separating high-and low energy carbonates from the clastics to the northwest.

It is possible that Kimmeridgian fluvial systems documented by Ellwood (1987) in the Cabo Mondego area equate with the early (Porto Novo) continental systems, though the age of the Cabo Mondego succession is uncertain.

North of the Berlengas Islands the facies distribution is uncertain. Offshore well data has shown that below the Cretaceous unconformity the Upper Jurassic section preserved is very thin (only a few hundred metres), and dominated by marls (RCL Wilson pers. comm). This suggests that these units equate with the Alcobaça beds which are stratigraphically beneath the Lourinhã formation in onshore areas. To the south, Ellis (1984) documented the deposits of an easterly dipping carbonate ramp system on the coast west of Lisbon. The westerly source for this material suggests

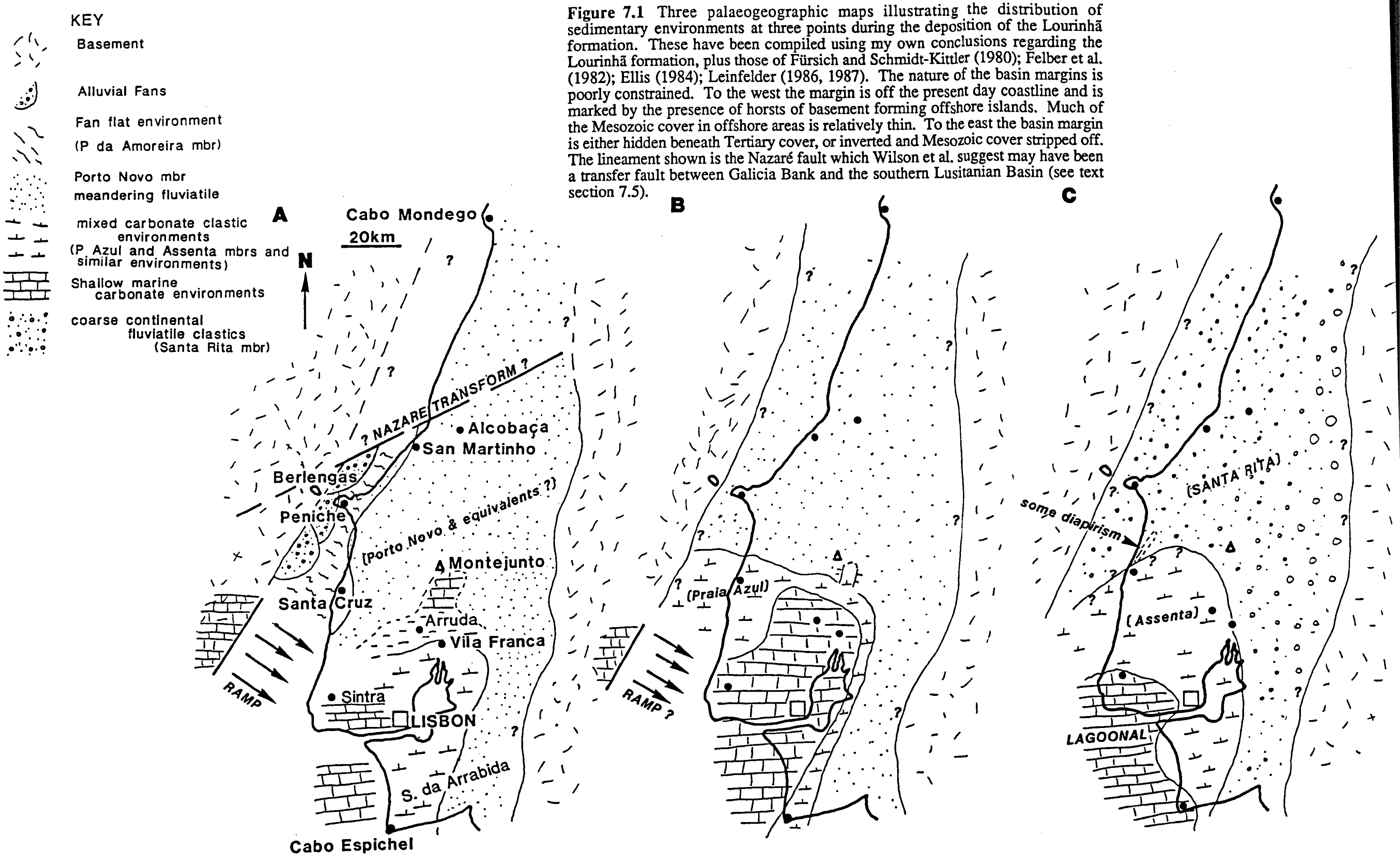


Figure 7.1 Three palaeogeographic maps illustrating the distribution of sedimentary environments at three points during the deposition of the Lourinhã formation. These have been compiled using my own conclusions regarding the Lourinhã formation, plus those of Fürsich and Schmidt-Kittler (1980); Felber et al. (1982); Ellis (1984); Leinfelder (1986, 1987). The nature of the basin margins is poorly constrained. To the west the margin is off the present day coastline and is marked by the presence of horsts of basement forming offshore islands. Much of the Mesozoic cover in offshore areas is relatively thin. To the east the basin margin is either hidden beneath Tertiary cover, or inverted and Mesozoic cover stripped off. The lineament shown is the Nazaré fault which Wilson et al. suggest may have been a transfer fault between Galicia Bank and the southern Lusitanian Basin (see text section 7.5).

A Late Kimmeridgian; Low-relief alluvial fans, sourced from the western margin of the basin distribute sediment eastwards to where their distal facies (Praia da Amoreira member) interfinger with meandering fluvial systems (Porto Novo member). The probable submerged counterpart of the Berlengas block, west of Sintra, provides a carbonate platform (ramp) which supplies carbonate sediment to a marine basin in the Sintra /Lisbon area. The shallow marine counterpart of the ramp lay to its northeast in the Arruda region. A narrow belt of mixed carbonate/clastic shelf/deltaic environments separated the continental clastic and carbonate environments.

C. Late Tithonian; A probable change in the main sediment sourcelands to the eastern margin of the basin. Coarse grained (Santa Rita member) sinuous fluvial systems have replaced the Porto Novo fluvial environments in northern and eastern parts of the basin. In the southwestern parts of the basin lagoonal carbonate environments existed. These were separated from the coarse clastic systems by a belt of sandy fluvio-deltaic environments (Assenta member). Periodic marine incursions were widespread and isolated much of the formerly fluvio-deltaic areas from clastic input.

B. Basal Tithonian; A marine transgression results in the establishment of shoal water deltas (Praia Azul member) in the Santa Cruz region. A transition to predominantly shallow marine carbonate environments occurs to the east. In the northern part of the basin meandering fluviatile environments persist (Porto Novo member). To the southwest, carbonate ramp sedimentation continued.

that a 'high', perhaps the submerged south-westward continuation of the Berlengas block, existed to the northwest of Sintra.

Figure 7.1B illustrates the probable palaeogeography during the maximum extent of the (basal) Tithonian transgression, which appears to coincide with a global eustatic highstand (Haq et al. 1987), and re-introduced marine environments to the southern part of the study area. Tongues of marine strata are found at least as far north as the Forte de Pai Mogo (Figure 1.8). In the approximately coeval fluvial strata there is no indication of the fan facies observed in the older successions. This may be due to relatively subdued relief in the clastic sourcelands following their earlier denudation. However, in associated parts of the Porto Novo member in sections as far south as Porto de Barcas/Porto Dinheiro, and especially in the northern Ferrel and Foz do Arelho sections (Figures 1.7, 1.8), coarser sediment (up to very coarse sand) does appear. This suggests the continued presence of lower relief fans or an alluvial plain supplying sediment from the north and west.

South of Torres Vedras and Montejunto, sedimentary environments were progressively more carbonate dominated eastwards. Leinfelder (1986) attributed the facies distribution to the influence of diapirism acting as a N-S trending clastic 'fence' with rapid subsidence to the north of Arruda forming a clastic 'trap'. The idea of diapirism in particular seems unlikely, as the original Triassic evaporitic depocentres did not extend as far south as the proposed 'fence'. Simpler explanations are that either the environnments of deposition were progressively further removed from the western margin of the basin, which may still have acted as the main clastic sourceland, or that tilted fault blocks acted as clastic fences. Lagoonal carbonate environments dominated the southwestern part of the basin around Lisbon and Sintra.

Figure 7.1C illustrates the disposition of sedimentary environments in the mid.-late Tithonian. The sandy rivers of the Porto Novo and Assenta members became gravelly with time to form the Santa Rita member. Locally, this is observed as a dramatic lithological transition across an unconformity.

The general palaeogeographic position was however, relatively unchanged. Shallow marine/lagoonal carbonate environments now dominated the southwestern part of the basin around Lisbon and Sintra (Ellis 1984; Ellis et al., in prep). The west-east clastic-carbonate transition between Santa Cruz and Arruda no longer existed, and a belt of fine grained fluvial and delta plain facies separated the carbonate environments in the southwest from the coarse clastic environments to the north and east.

There appears to have been some form of structural divide in the western part of the basin, resulting in a marked contrast between the Santa Rita and Assenta members (thought to be coeval) in outcrops only 6km apart. Palaeocurrent data from the Ferrel section suggests that the Santa Rita clastics were sourced from the east which would be consistent with the known facies distributions.

Occasionally the fine grained fluvial/delta plain environments (Assenta member) were affected by rapid marine transgressions which resulted in the establishment of marine/lagoonal conditions over much of the southern part of the basin. The development of ostensibly marine cements in one of the fluvial sand bodies at Praia da Santa Rita and the presence of unusual ichnofauna in associated mudrocks (R. Hiscott pers. comm.) suggests that a marine transgression may have extended north of Santa Cruz.

The palaeogeographies presented are all rather generalised due to the poor environmental data available from inland outcrop and also the poor biostratigraphic control. The major feature apparent in the maps is that the western margin of the basin was more important as a source of clastic sediment during the late Kimmeridgian and early Tithonian, though the clastics in the successions to the south of Lisbon were sourced from the east. By the late Tithonian, the more important coarse clastic influx was probably from the east. In the western part of the basin around Torres Vedras-Santa Cruz the southward distribution of the coarse clastic Santa Rita member appears to have been prevented. To the south of Santa Cruz, fine grained continental clastic plains were periodically inundated by widespread, short-lived marine incursions from the south.

7.4 PLATE TECTONIC SETTING

The Lusitanian Basin is one of a family of formerly adjacent Atlantic marginal basins, many of which have a history of rifting and subsidence as far back as the Triassic (Wilson 1975; Masson and Miles 1986). Until the main emplacement of oceanic crust began in the late Aptian (ca 110-115my, Sibuet, Ryan et al. 1979; Montadert et al. 1979; Groupe Galice 1979; Masson and Miles 1984), the tectonic and sedimentary history of the Lusitanian Basin was closely linked to that of the other basins. A number of similar events can be identified in the Lusitanian, Jeanne d'Arc, Cantabrian, Asturian, Aquitaine and NW European basins, all of which occur north of the Gibraltar-Azores fracture zone (Figure 1.1) (Hiscott et al. 1988). There is recent evidence, from seismic, magnetic and gravity data, for an earlier (Tithonian/Berriasian) sea floor spreading event in the Tagus Abyssal Plain (Mauffret et al. in press) 150-450 km SE of the study area. A late Jurassic age for initial oceanic crust emplacement was earlier favoured by Ribeiro et al. (1979). This age suggests that deposition of the Lourinhã formation may have heralded, and been coincident with, an earlier phase of oceanic crust emplacement. Evidence for late Kimmeridgian/Tithonian rifting events in offshore areas was noted following DSDP and ODP cruises (Mougenot et al. 1979; Mauffret and Montadert 1987).

7.5 PALAEOENVIRONMENTAL CHANGES AND THEIR CONTROLS

The palaeoenvironments and palaeogeographies outlined in sections 7.2 and 7.3 can be considered in their plate tectonic context and the relative importance of tectonic versus other (e.g. eustatic) controls assessed. The most important changes that took place in palaeoenvironments and palaeogeographies during the late Kimmeridgian to late Tithonian were:

1) the marine regression and progradation of continental environments at the base of the Lourinhã formation during the mid- to late Kimmeridgian;

2) the minor oscillations between marine and terrestrial facies in both the Praia Azul member and lower and topmost parts of the Assenta member;

3) the basal Tithonian transgression resulting in the deposition of the Praia Azul member and the widespread, but short-lived transgressions which resulted in the deposition of the thin carbonate units within the Assenta member;

7.5.1 The establishment of continental sedimentary environments

The continental facies of the Praia da Amoreira and Porto Novo members are preceded by marine shelf sediments of the Alcobaça beds in the northern part of the study area (see Chapter 1) and marine shelf and slope facies of the Amaral and Abadia formations (studied by Ellwood 1987) to the south. As the Kimmeridgian marine/continental transition marked the upper boundary of the sequences studied by Ellwood (op cit.) he discussed the nature of the facies transition and the possible controls on this. He concluded that it was a diachronous transition resulting from the progradation of a mixed carbonate/clastic shelf into the marine Abadia marl basin. In his model (Figure 7.2) fluviodeltaic systems transported siliciclastic sediments from NW to SE , across the shelf, into basinal areas. I broadly agree with this conclusion because:

1) there is a transition over much of the northern part of my study area from shallow marine facies of the Alcobaça beds and Amaral formation to the continental facies of the Lourinhã formation;

2) this transition is exposed at Quebrada da Amejoada (see Chapter 1); as far south as Concolaçâo on the coast (Werner 1986; Ellwood 1987); on both flanks of the Montejunto anticline (Leinfelder 1986; Ellwood 1987); and in the Arruda area (Leinfelder 1986);

3) no evidence for any angular unconformity has been observed;

4) the Lourinhã formation thickens substantially northwards; which may be indicative of more prolongued terrestrial sedimentation in this area.

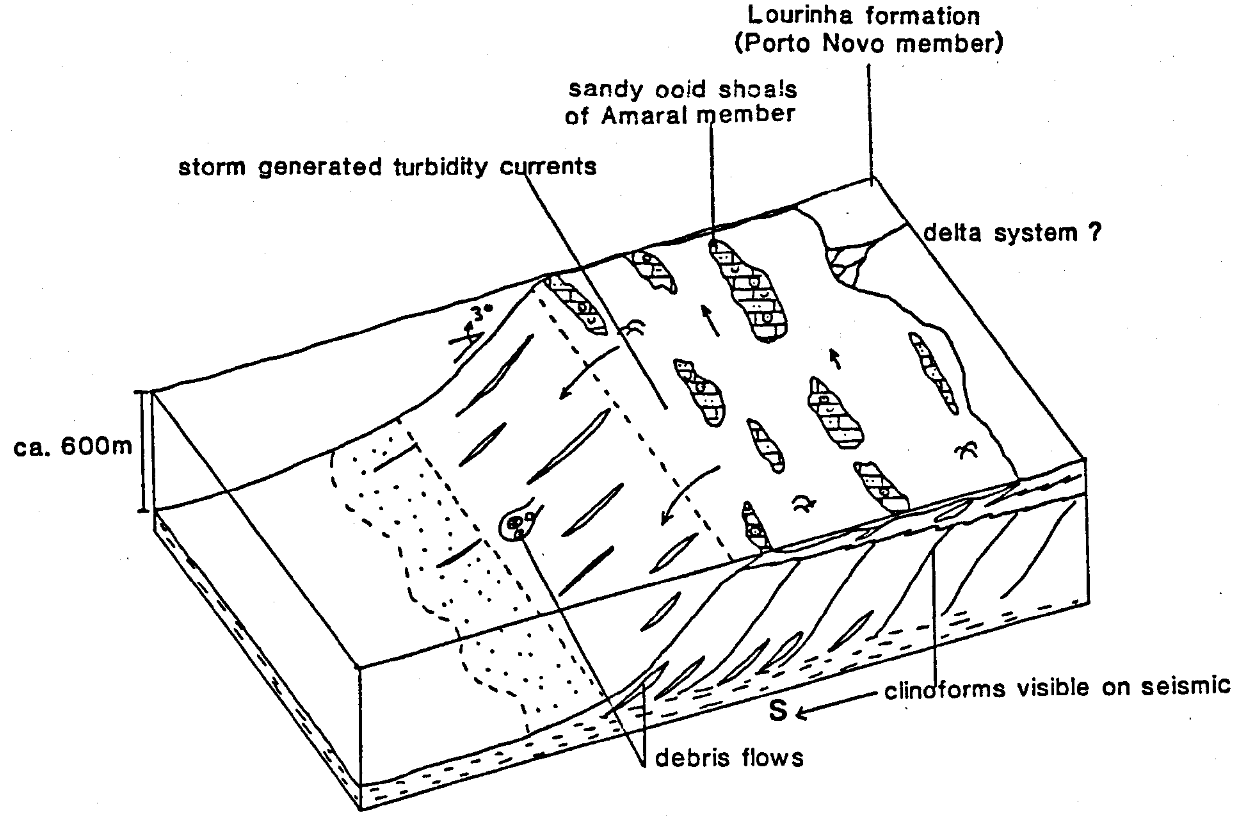


Figure 7.2 Environmental model of Ellwood (1987) for the upper part of the Abadia formation and associated Amaral and Lourinhã formations

A rather more abrupt transition to terrestrial facies is observed on the coast south of Santa Cruz. The Abadia formation preceding the Lourinhã formation suggests that relatively rapid shallowing followed by a period of relatively constant depth and low sediment input occurred, the latter indicated by a thick bioturbated marine sandstone (see 1.6.3) (Ellwood 1987). North of São Bernadino there is an abrupt increase in grain size and change in lithological appearance at the transition to continental facies from buff, well sorted medium sands to grey/brown, poorly sorted kaolinitic arkoses full of basement derived lithic clasts. The transition to distal alluvial fan facies at these localities and others as far north as Ferrel indicate that a period of tectonic activity occurred which may have been coincident with marine regression. In the light of this rather abrupt facies transition, it seems possible that uplift preceded renewed rift activity and so may have accelerated the advance of the shoreline in the southern part of the basin. Eustacy seems to have played little part in the changes observed. Haq et al. (1987) suggest an overall global eustatic sea level rise during the late Kimmeridgian.

The succession on the flanks of the Vimeiro diapir in the southern part of the study area (see section 1.6.3) also shows a rapid transition from marine to terrestrial environments. Kimmeridgian carbonates, which may have occupied a linear 'high' starved of clastic input on the crest of the Vimeiro diapir, appear to have been rapidly replaced by the distal fan facies. This conclusion is supported by palaeocurrent data from a coastal section to the north west (see Figure 1.14) indicating flows directly towards the diapir. This implies that the diapir was no longer emergent when the fans prograded.

Variations in the thickness of the Praia da Amoreira member between the type section and that north of Ferrel (see Figure 1.10), plus the intertonguing of Praia da Amoreira and Porto Novo member facies in the latter area, suggest that differential subsidence may have occurred, perhaps between different fault blocks (the sections are 23km apart). The section at Ferrel is closer (ca. 12km compared to 17km) to the postulated basin margin and some increase in thickness of a fan succession would be expected as the basin margin is approached.

7.5.2 Minor fluctuations in relative sea level

In any modern deltaic system, minor marine transgressions and regressions occur as a consequence of the switching of a delta distributary network and consequent abandonment of a delta lobe, or progradation of a newly initiated distributary network (Coleman and Gagliano 1964). The minor oscillations between marine and continental facies observed in the Praia Azul member and upper and lower parts of the Assenta member are attributed to such autocyclic processes. This is because of the relatively small scale and high frequency of the transgressive events and also the fact that it is not posssible to correlate the events observed (in the Praia Azul member) between sections. The rapid, short-term changes responsible for the repetitive appearance of marine strata every few metres are unlikely to be caused by large scale external controls which might be expected to operate over longer timescales.

7.5.3 Large scale fluctuations in sea level

The remaining two groups of transgressive 'events' are likely to be due to more large scale processes. In the case of the Praia Azul transgression, this was the first and most widespread transgressive event recorded within the Lourinhã formation and changed the palaeogeography significantly. The thin carbonate units within the Assenta member must also have been the product of widespread and relatively rapid transgressions/regressions because although they are thin, their presence indicates that the environment was starved of clastic input and any potential source of siliciclastic sediment must have retreated a significant distance. If the basin gradient were between 4-13cm/km (values from a range of meandering fluvial systems presented by Schumm et al. (1972)) then a sea level rise of 4-10m over several thousand years, would cause a transgression to penetrate up to 80km inland. The three carbonate units in the Assenta member have a spacing of 20-30m. Following the discussion on alluviation rates in Chapter (3), the associated alluvial sediments could take 2×10^4 to 3×10^5 years to accumulate. Thus transgressions are likely to have a spacing of tens to hundreds of thousands of years.

The sea level curves of Haq et al. (1987) do show a transgression and highstand in the late Kimmeridgian so it is concievable that this may have been responsible for the transgression at the base of the Praia Azul member. In addition there are a number of minor fluctuations on the curves which might be invoked to explain the later transgressive/regressive events.

There are only two known mechanisms for sea level changes taking place at rates of up to 1m per 10^3 years. These are glacially controlled eustatic sea level changes (Donovan and Jones 1979; Mörner 1987) and 'intraplate stress' mechanisms (Cloetingh et al. 1985; Cloetingh 1986 a,b, 1988).

For some time it has been thought that ice caps were relatively small or absent during most of the Mesozoic era (Frakes 1979). This was due partly to the apparent extension of hot semi-arid or tropical environments to high palaeolattitudes. In addition there is no sedimentological evidence for glaciation in the Triassic anywhere in the world. These concepts have recently been questioned by Frakes and Francis (1988), because of the discovery of Cretaceous glacial sediments in Australia and other regions at lower palaeolattitudes (ca. 65° S) than previously known. However, highstands on the charts of Haq et al. (1987) during the late Jurassic and the 'broad zone of relatively warm and arid climates extending to about lattitudes of 45°' (Frakes 1979), developed at that time, suggests that very large volumes of water were not tied up in ice sheets. Therefore, it is possible that glacial processes were not responsible for the sea level fluctuations observed.

If glacial processes are not thought a probable control on relative sea-level fluctuations at this time then we are left with variations in intraplate stress (Cloetingh et al. 1985; Cloetingh 1986a,b, 1988). The imposition of axial compressional stress on a basin previously subject to axial tensional stresses would, according to Cloeting et al. (1985), cause more rapid differential subsidence in the basin centre coupled with uplift at its margins. An examination of the regional correlation chart (Figure 1.10) shows that, assuming the correlations made are correct, the later marine horizons in the Assenta member appear at the same level as the coarse fluviatile clastics of the Santa Rita member (Chapter 6). Deposition of coarse, basement-derived sediments following relatively fine-grained ones (Chapters 3&4) shows a 'rejuvenation' of sourcelands at the margins of the basin. Possible further evidence of tectonic activity is provided by the local unconformities on the flanks of diapirs and the probable existence of a structural divide between the depositional environments of the Santa Rita and Assenta members in the western part of the basin. Uplift at the basin margins coupled with subsidence in central regions during the deposition of the Assenta and Santa Rita members fits with a scenario predicted by the intraplate stress model. Therefore it is possible that fluctuations in intraplate stress regimes provide an explanation for the nature of the marine incursions observed to occur in the Praia Azul and Assenta members.

A potential cause of changes in western Iberian regional stress fields has recently come to light. Mauffret et al. (in press) believe that there may have been an earlier (Tithonian/Berriasian) phase of sea floor spreading (later abandoned) in the region of the present-day Tagus Abyssal Plain which lies only 150-450km to the southwest of the outcrops in question. If Mauffret et al. (in press) are correct, then the early initiation of sea floor spreading could be responsible for a change in the stress regime during the deposition of the Praia Azul and Assenta members.

7.6 RELATIONSHIP OF THE LUSITANIAN BASIN TO OTHER NORTH ATLANTIC MARGIN BASINS, NOTABLY THE JEANNE D'ARC BASIN

Recent reconsructions of the pre-Atlantic continental relationships by Masson and Miles (1984) place the Flemish Cap and Galicia Bank in adjacent positions (Figure 1.1). This implies the juxtaposition of the Lusitanian and Jeanne d'Arc basins. These basins have broadly similar stratigraphic histories from the Triassic through to their 'breakup' unconformities in the Cretaceous. Other features the two basins have in common is the strong influence that inherited tectonic fabrics have played in their development and the existence of significant deformation associated with diapirs cored by Triassic marine evaporites (Tankard and Welsink 1987; Willis 1988; Wilson et al. in press). An important difference between the basins is that the fill of the Jeanne d'Arc Basin, at up to 18km, is significantly thicker than that of the Lusitanian Basin which is at most only about 4km thick. Events the two basins have in common are:

1) an initial rifting event in the Triassic, resulting in the deposition of continental sediments, succeeded by the deposition of thick evaporites;

2) a reduction of fault controlled subsidence and the establishment of broad saucer shaped carbonate basins due to thermal subsidence;

3) renewed rifting in the Late Jurassic to Early Cretaceous, resulting in rapid subsidence and the deposition of thick siliciclastic successions below the 'breakup' unconformity;

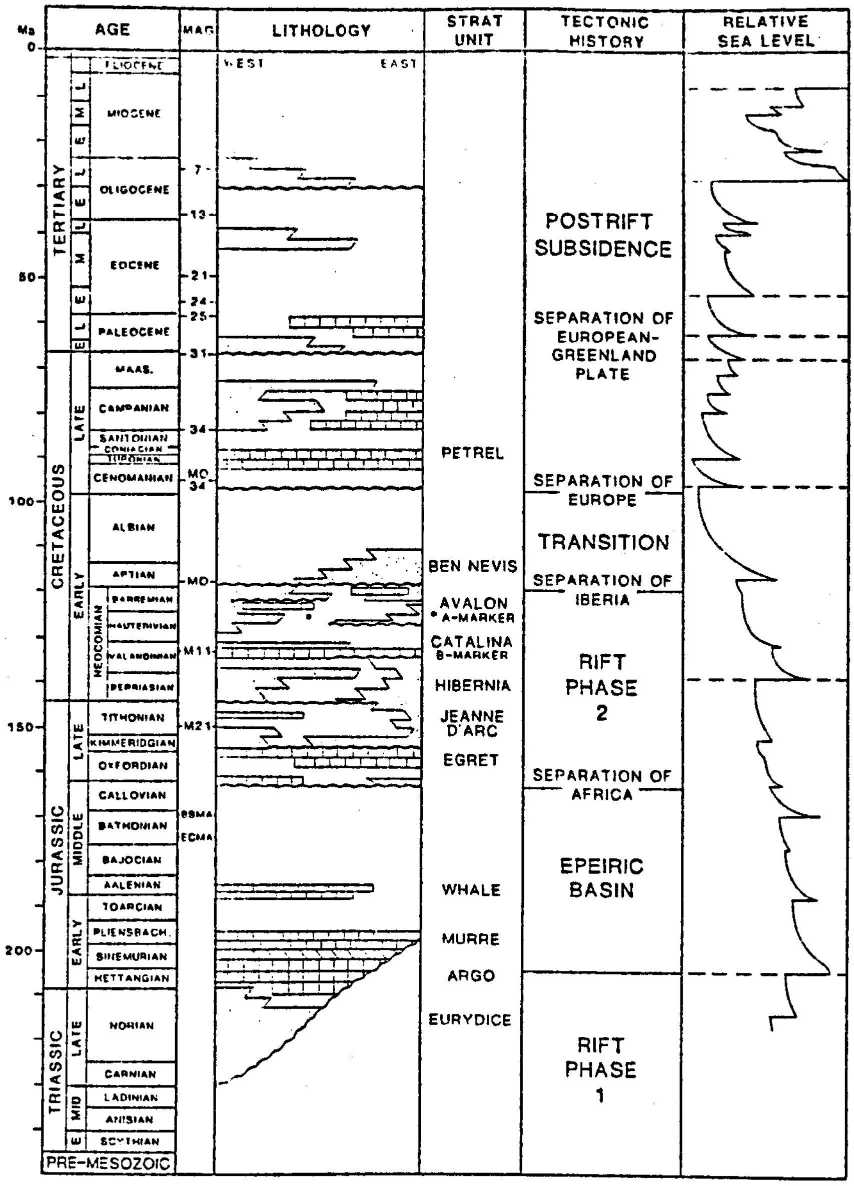


Figure 7.3 Lithostratigraphic column for the Jeanne d'Arc basin after Tankard and Welsink (1987). The Jeanne d'Arc formation is the equivalent of the Lourinhã formation.

Figure 7.3 is the stratigraphic column of the Jeanne d'Arc Basin. The Jeanne d'Arc Formation is the equivalent unit to the Lourinhã formation of the Lusitanian Basin. The Jeanne d'Arc Formation is significantly thicker than the Lourinhã formation, up to 2000m being preserved in the central parts of the basin. It is probably Middle-Kimmeridgian to Tithonian/Berriasian in age and is bounded by unconformities with Middle Jurassic carbonates and the fan-deltaic clastics of the younger Hibernia Formation and is the main reservoir interval in the basin. Sandstones and conglomerates deposited by braided rivers formed an apron round the basin margin. The rivers terminated in a fan delta complex. Accelerated subsidence resulted in a

Tithonian marine transgression and the widespread deposition of shallow marine sandstones and mudstones. Subsequently a coarsening-up unit, with cycles up to 100m thick, resulted from the progradation of the fan delta system (Tankard and Welsink 1987).

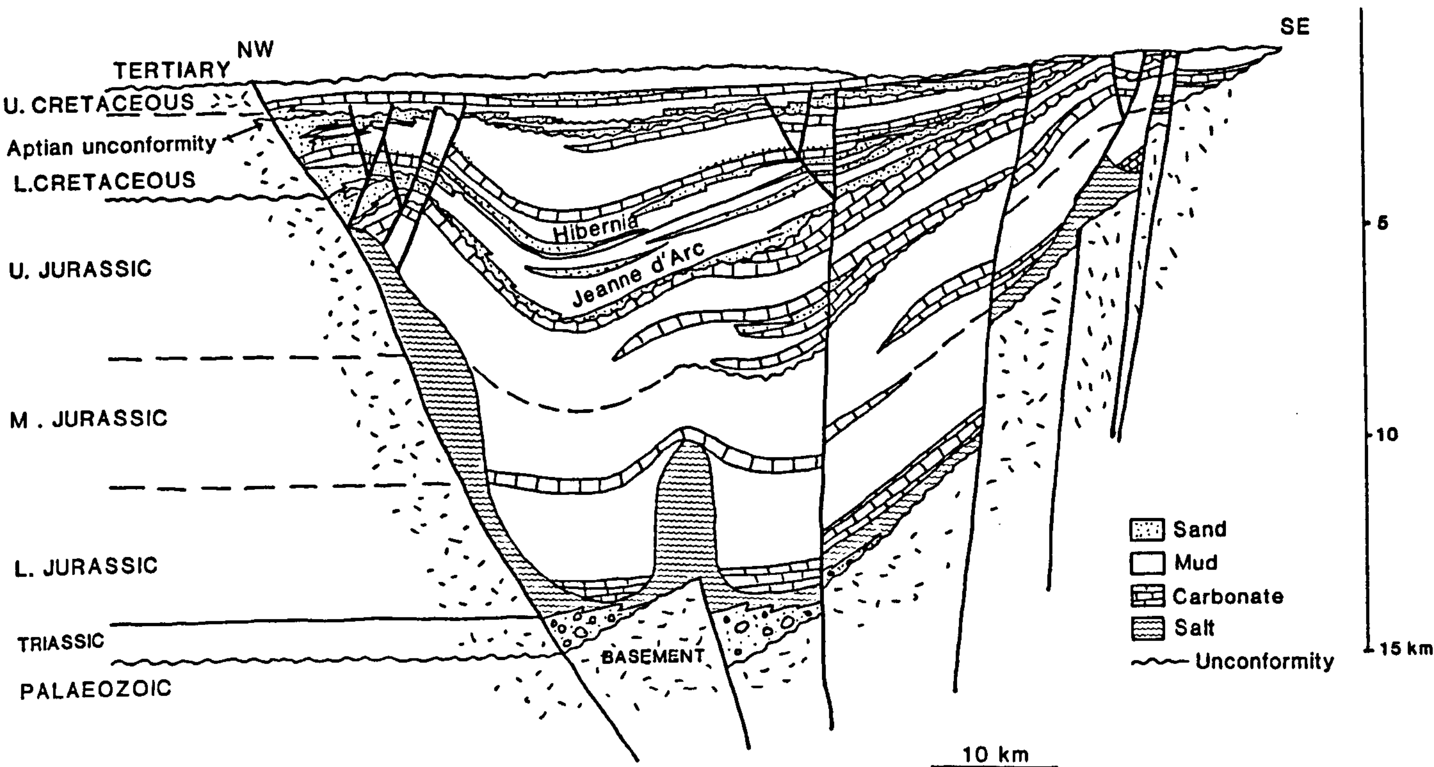


Figure 7.4 Geological cross section across the Jeanne d'Arc basin, after Tankard and Welsink (1987).

Clearly there are broad similarities between the Lourinhã and Jeanne d'Arc formations, both being the deposits of fluvial and shallow marine environments. The substantially greater thickness and coarser grained nature of the Jeanne d'Arc fluvial deposits were caused by significantly greater rates of subsidence which produced steeper river gradients and consequently low sinuosity gravelly fluvial systems. The only comparable coarse fluvial unit at outcrop in the Lusitanian Basin is the Santa Rita member, though coarser systems may have existed near the basin margins particularly to the west, off the present day coastline. A further point is that the timing of the period of maximum subsidence is earlier in the Lusitanian Basin where the Middle-Oxfordian to Upper Kimmeridgian Abadia formation records the period of maximum subsidence (Guery et al 1986; Ellwood 1987; Wilson et al. in press). It is possible that the upper boundaries of the Lourinhã and Jeanne d'Arc formations are synchronous.

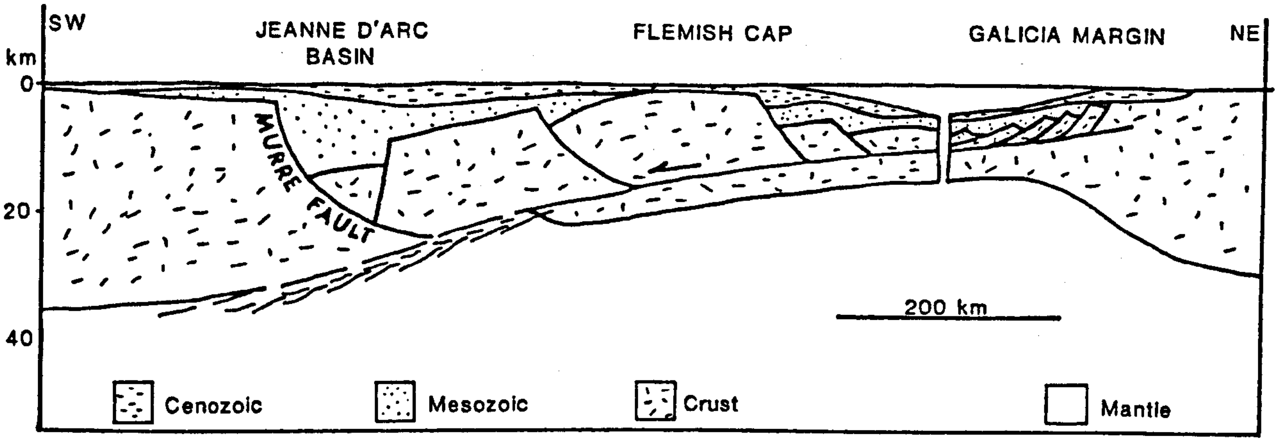


Figure 7.5 Extensional model for the Jeanne d'Arc Basin and Galicia Bank after Tankard and Welsink (1987); suggests failure along a low-angle crustal detatchment. The model sets out to account for the assymetry observed, in amounts of extension, style and scale of faulting and depth to detatchment.

The Jeanne d'Arc Basin is essentially a half graben basin which has extended over a major crustal detatchment; the Murre fault, which soles out at about 26km depth (Tankard and Welsink 1987; Enachescu 1987)(Figure 7.4). Three major transfer faults (Gibbs 1984) separated adjacent areas which were extended by different amounts (Tankard and Welsink 1987). For the Jeanne d'Arc basin, Tankard and Welsink (1987) compute a ß factor (Mackenzie 1978) of 1.2 compared to 1.45 for the Galicia Bank. In contrast to the deep detatchment into which the Murre fault soles out, a shallower detatchment (the S-reflector of de Charpal et al 1978) is thought to exist below the Galician margin at about 9.5km (Tankard and Welsink 1987). In addition fault block spacing is considerably less (10-30km) on the Galicia margin compared to the Jeanne d'Arc Basin (40-100km)(Figure 7.5). Tankard and Welsink (1987) explain the differences in the amount of extension, fault block spacing and depth to detatchment by a large-scale detatchment model (Figure 7.5) which includes a large westward dipping decollment. This is based upon the simple shear model of Wernicke (1985) developed in the Basin and Range province in the US.

Compared to both the Galicia Bank and the Jeanne d'Arc Basin, the Lusitanian Basin underwent remarkably little post-'breakup' subsidence. Willis (1988) computed a ß factor of only about 1.1 for the onshore part of the basin, based on the limited amount of seismic refraction data available (Mendes Victor et al. 1980). This suggests that there had in fact been relatively little crustal extension. Willis also noted the high angle of many of the normal faults in the basin. High angle faults cannot accomodate significant amounts of extension.

Relatively little post Aptian extension is postulated for the Lusitanian Basin (Willis 1988) compared to the Galicia margin, which forms a conventional passive margin. In order to explain this, Wilson et al. (in press) suggested that the Lusitanian Basin was separated from the Galicia Bank by a transfer fault, the Nazaré fault (Figures 1.4 & 7.1A), the offshore extension of which gives rise to the Nazaré submarine canyon and forms the northern bounding fault of the Berlengas block (Figure 1.2). It is possible that this structure explains the absence of alluvial fan facies in the northern coastal section examined. If the basement to the north of the Berlengas block had a fundamentally different structural style, then it may not have been elevated at the time the Berlengas blocks were supplying the Praia da Amoreira fan systems with coarse clastic sediment.

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APPENDIX 1

ASSUMPTIONS MADE IN THE CONSTRUCTION OF FIGURE 1.10

AND COMMENTS UPON THE RESULT

1) The two main horizons used to correlate the sections are the base of the Praia Azul member and in the absence of this i.e. in the Ferrel section, the base of the Santa Rita member.

2) No assumption made re the fault at the top of São Bernadino section.

3) Areia Branca fault assumed to be restored by correlation of marine horizons. If correct, throw is 80m.

4) No displacement assumed across break in section at Porto Dinheiro.

5) About 15m of section assumed to be covered at the base of the Praia Azul member at Praia da Santa Rita.

6) Approximately 50m of section covered at the top of the Praia Azul section. Assumes no displacement across the break in section occupied by the Rio Sizandro.

7) Fault at Ponta da Vela, which breaks the Assenta member section, shown with a displacement of 30m. It is likely to be greater, possibly substantially so.

8) No cognisance taken of fault within the upper part of the Ferrel section when computing the thickness of the Santa Rita member, 200m of which is assumed. Loss of outcrop could be a consequence of severe faulting and may have introduced significant error to the estimate.

MAIN POINTS

1) Thickness of the Lourinhã formation in the southern part of the basin 600-860m.

2) Continental facies of the Praia da Amoreira member at Ferrel are thicker than those to the south though units 75 & 140m thick are separated by a channel sand body pertaining to the Porto Novo member. Type section 140m thick.

Porto Novo member:	at Praia da Amoreira	47m-thick
	Porto Novo	67m
	Ferrel	330m
	Foz do Arelho	800m?(not on section)

Praia Azul member type section	<120m
Assenta member type section	>300m?
Santa Rita member	310m? (only110m ever seen at outcrop)

APPENDIX 2 LIST OF ENCLOSURES

ENCLOSURE 1	Correlation diagram for graphic logs São Bernadino south - Areia Branca. Plus Master Key for all graphic logs.
2	Correlation diagram for graphic logs Porto Dinheiro north-Areia Branca.
3	Graphic logs São Bernadino south-Areia Branca.
4	Graphic Logs Porto Dinheiro north-Areia Branca.
5	Graphic Logs Foz da Arelho section.
6	Graphic Logs Ferrel section.
7	Graphic Logs Porto Novo-Porto Dinheiro (Praia da Amoreira and Porto Novo members).
8	Graphic Logs Praia da Santa Rita Section (Praia Azul and Assenta members).
9	Graphic Logs Praia da Amoreira section (Praia da Amoreira and Porto Novo members).
10	Graphic Logs Praia Azul section.
11	Graphic Logs Rio Sizandro section.
12	R. Hiscott's Graphic Logs of the upper part of the Assenta member section, above the fault at Ponta da Vela.
13	Orientation Map; illustrating the location of sections studied, relevant graphic logs and relative stratigraphic positions.

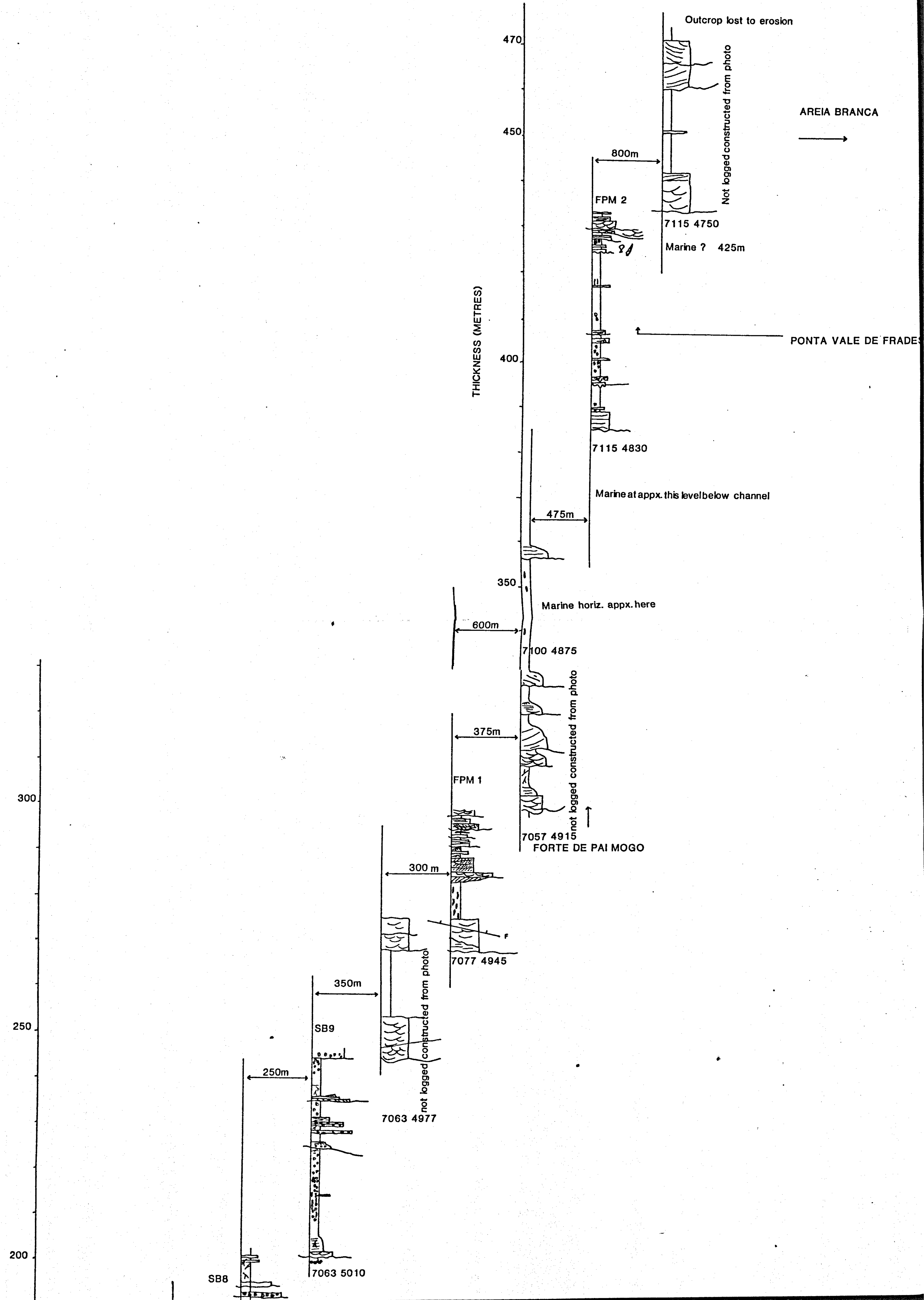
Enclosures

ENCLOSURE 1

CORRELATION DIAGRAM FOR LOGS SÃO BERNADINO SOUTH-AREIA BRANCA (SB 1-9, FPM 1/2)

(TOP)

PLUS MASTER KEY FOR ALL LOGS


Outcrop lost to erosion
Not logged constructed from photo
AREIA BRANCA
800m
FPM 2
7115 4750
Marine ? 425m
PONTA VALE DE FRADES
THICKNESS (METRES)
470
450
400
350
300
250
200
7115 4830
Marine at appx. this level below channel
475m
Marine horiz. appx. here
600m
7100 4875
not logged constructed from photo
375m
FPM 1
7057 4915
FORTE DE PAI MOGO
300 m
F
7077 4945
350m
not logged constructed from photo
SB9
250m
7063 4977
SB8
7063 5010

KEY TO LOGS

PLANAR CROSS-BEDS	PEBBLY SANDSTONE
TROUGH CROSS-BEDS	SANDY SILTSTONE
PLANAR BEDS	CALICHE
OR RIPPLE LAMINAE	CALICHE CONCRETIONS
CLIMBING RIPPLE LAMINAE	RHIZOCRETIONS
DISTURBED LAMINAE	ROOTLETS
FINE SEDIMENT DRAPES	MUDCRACKS
FINE SEDIMENT INTRACLASTS	INVERTEBRATE BURROWS (INTENSITY)
PALAEOCURRENT MEASUREMENTS	---- GRADATIONAL BOUNDARY

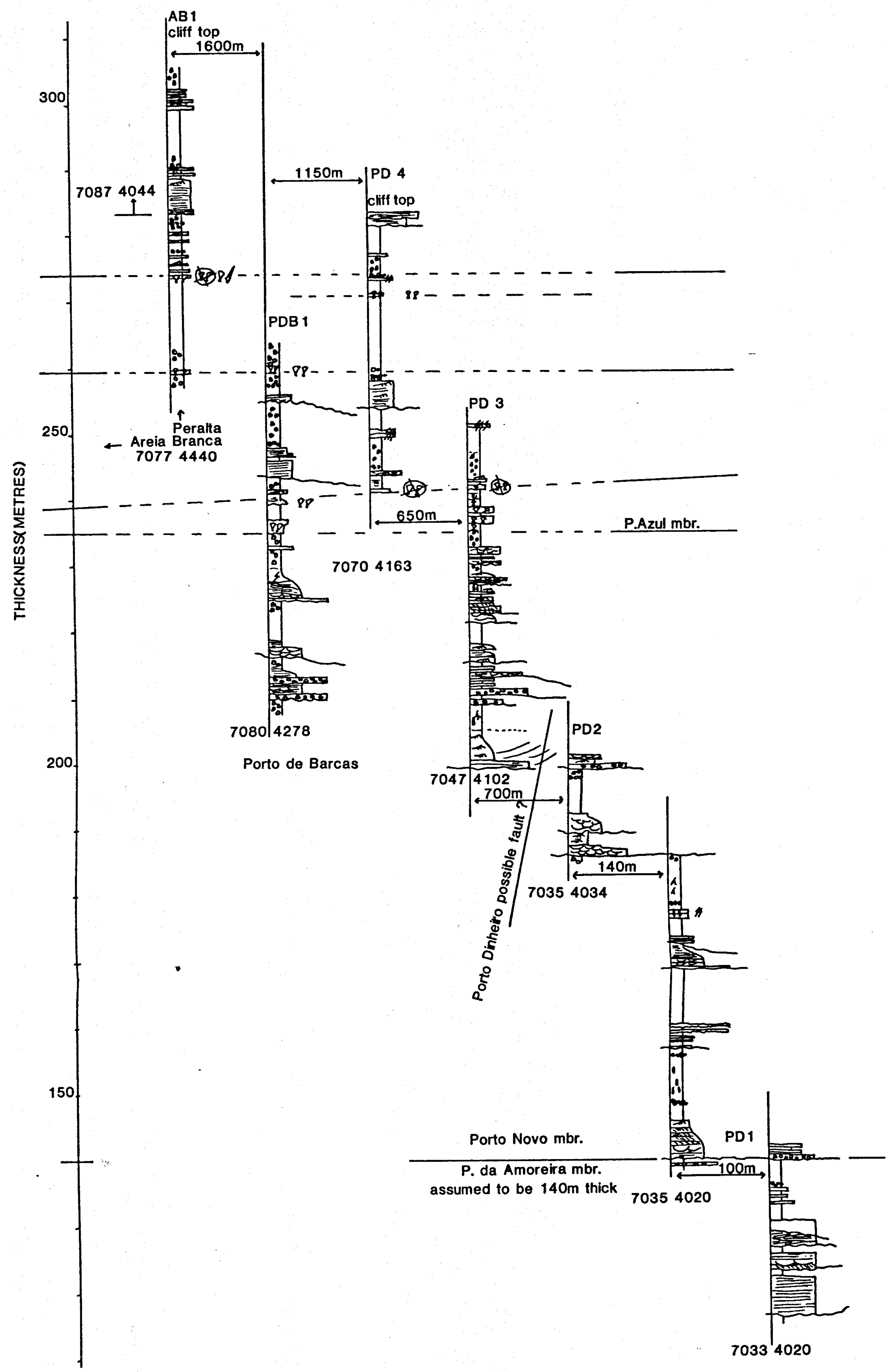
COLOURS

gn.	green	w.	white	dk.	dark
r.	red	bn.	brown	l.	light
p.	purple	mott.	mottled	weath.	weathering

BIVALVES	I. ISOGNOMON
GASTROPODS	E. EOMIODON
ECHINODERMS	My. MYOPHORELLA
CORALS	O. OYSTERS
BONES TEETH	SHELL DEBRIS
CARBONATE	DIPLOCRATERION
LIGNITIC DEBRIS	THALLASSINOIDES
LOGS	PLANOLITES

ENCLOSURE 2

CORRELATION DIAGRAM FOR LOGS PORTO DINHEIRO NORTH-AREIA BRANCA (PD 1-4, PDB 1,AB1)



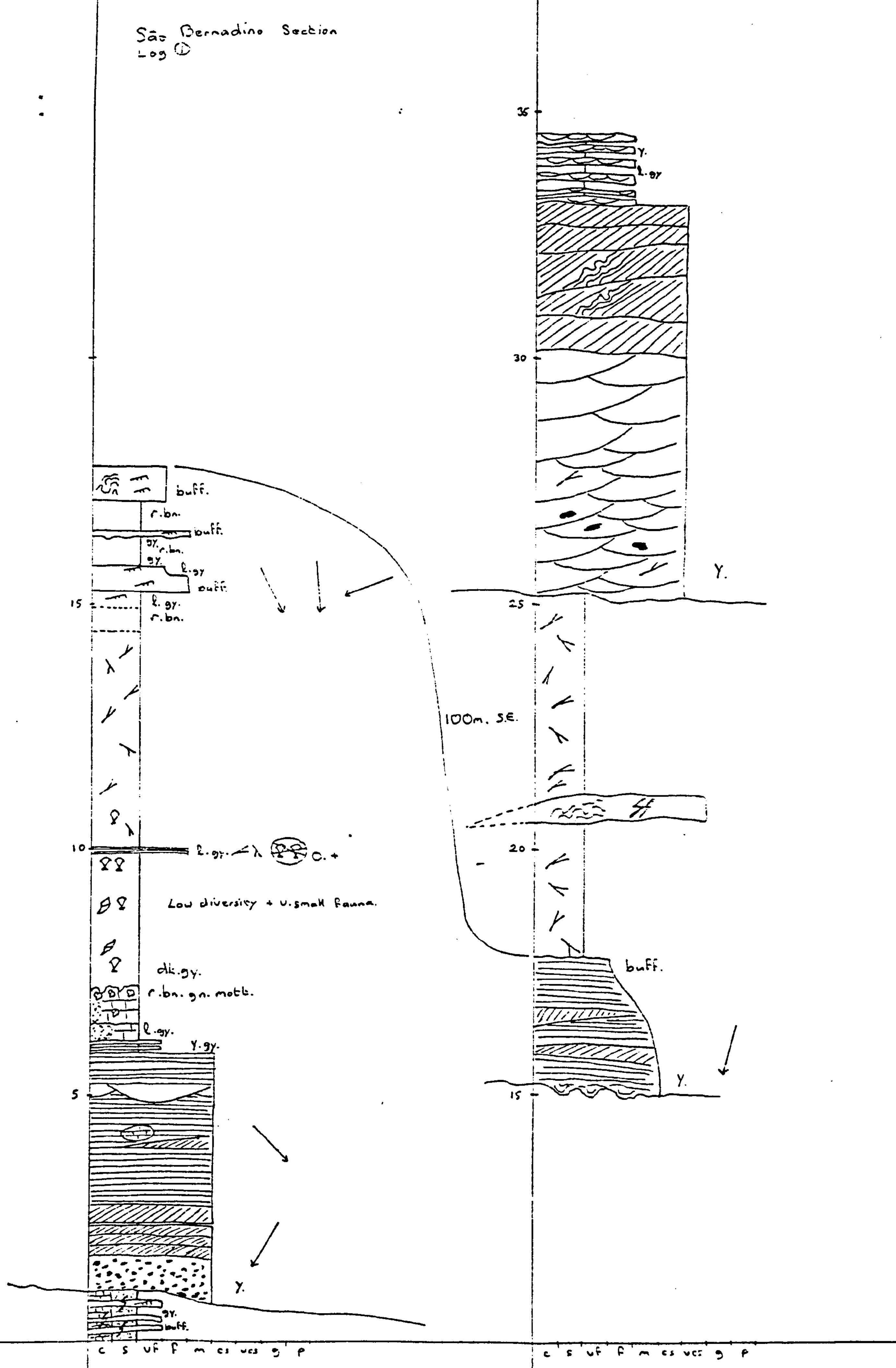
THICKNESS(METRES)
300
250
200
150
AB1
cliff top
1600m
7087 4044
Peralta
Areia Branca
7077 4440
PDB1
7080 4278
Porto de Barcas
1150m
PD 4
cliff top
7070 4163
650m
PD 3
P.Azul mbr.
7047 4102
700m
PD2
Porto Dinheiro possible fault ?
140m
7035 4034
Porto Novo mbr.
P. da Amoreira mbr.
assumed to be 140m thick
PD1
100m
7035 4020
7033 4020

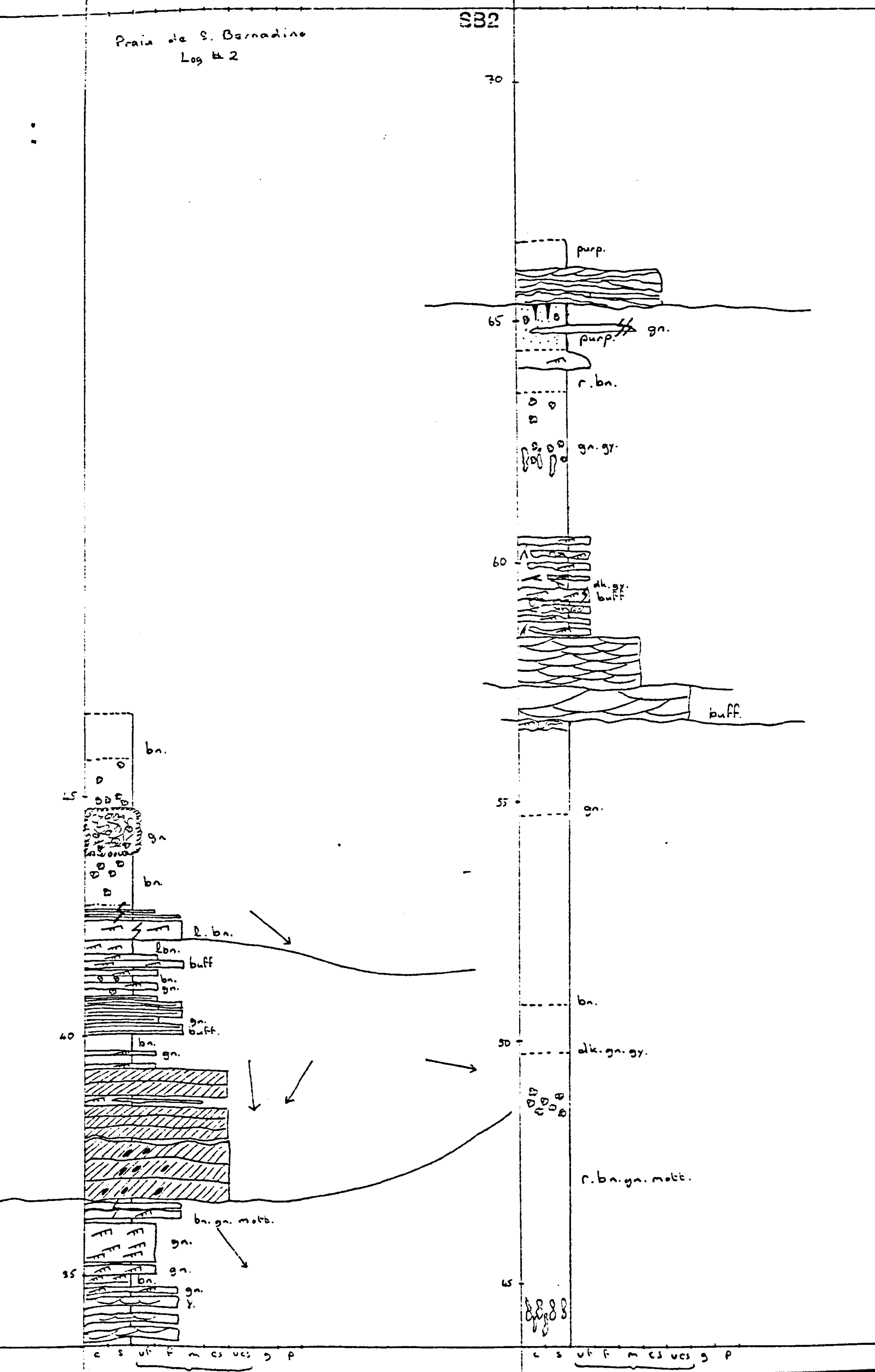
ENCLOSURE 3

LOGS: SÃO BERNADINO SOUTH-AREIA BRANCA

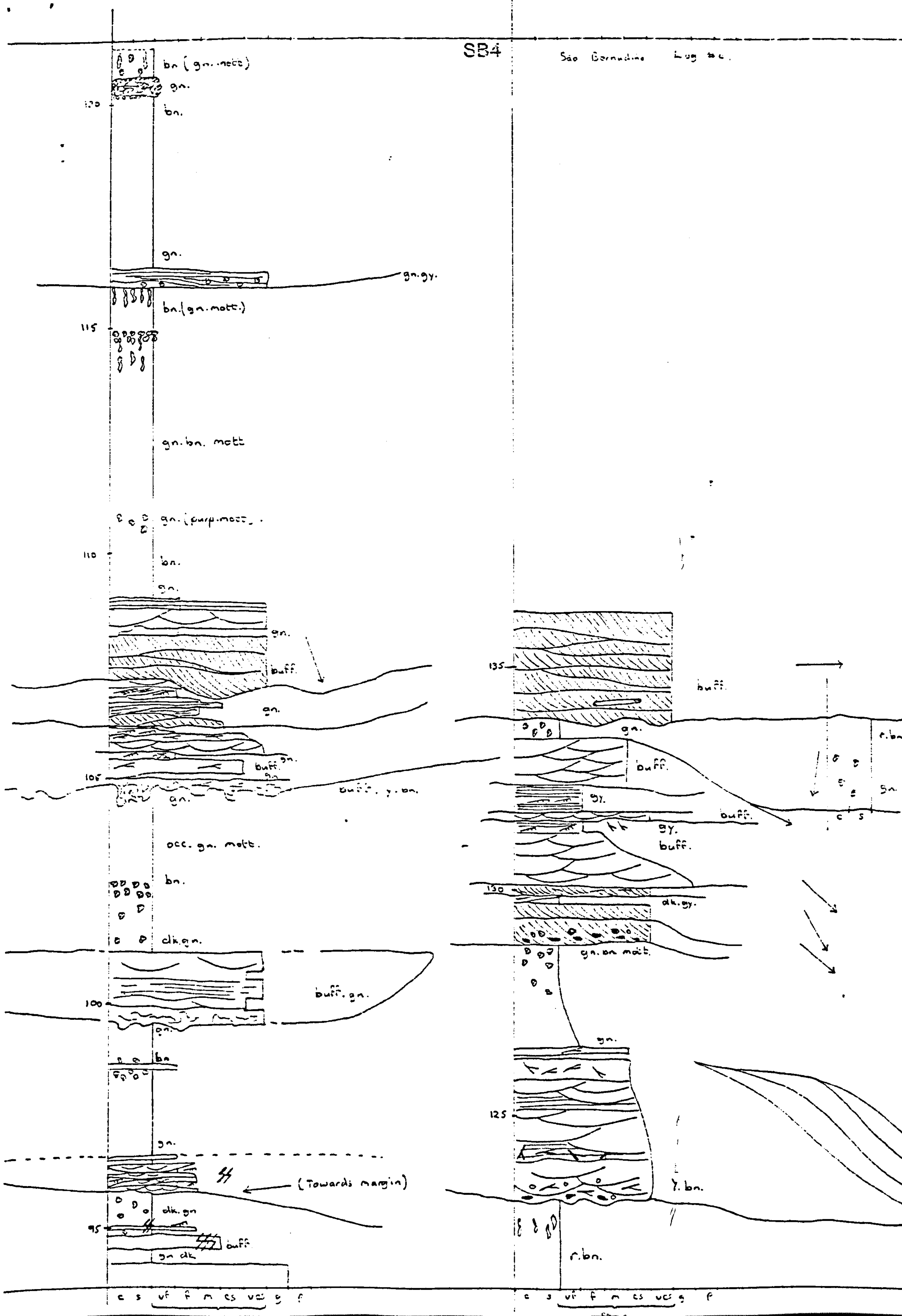
(SB 1-9, FPM 1/2)

SB1

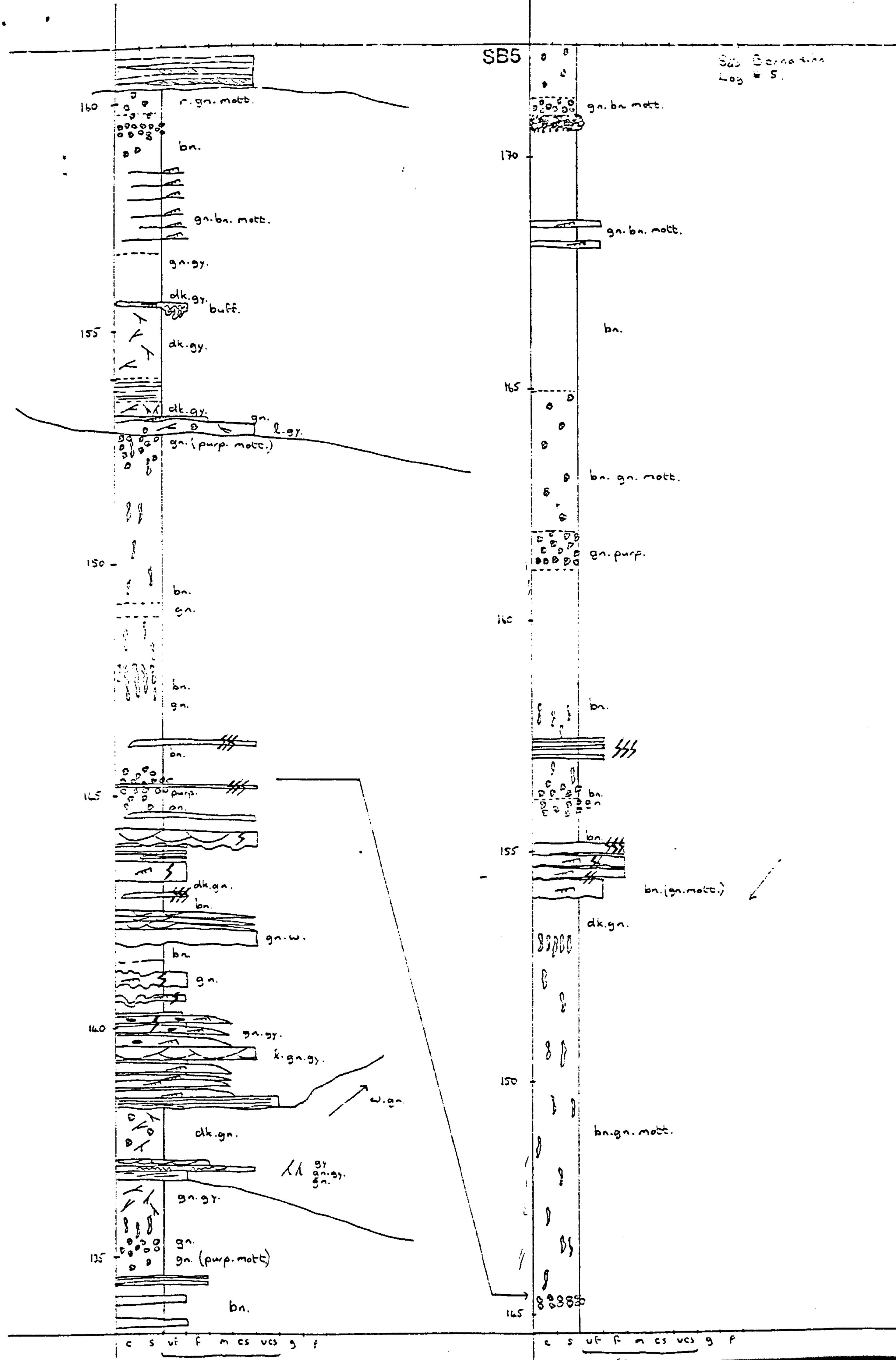

São Bernadino Section
Log ①
buff.
r.bn.
buff.
gy. r.bn.
gy.
l.gy
buff.
l. gy.
r.bn.
l.gy.
O. +
Low diversity + v.small fauna.
dk.gy.
r.bn. gn. mott.
l.gy.
y.gy.
y.
gy.
buff.
c s vf f m cs vcs g p
35
y.
l.gy
30
y.
25
100m. S.E.
20
buff.
y.
15
10
5
c s vf f m cs vcs g p


SB2
Praia de S. Bernadino
Log # 2
70
purp.
65
gn.
purp.
r. bn.
gn. gy.
60
dk. gy.
buff
buff.
55
gn.
bn.
50
dk. gn. gy.
r. bn. gn. mott.
45
bn.
gn
bn.
l. bn.
l bn.
buff
bn.
gn.
gn.
buff.
40
bn.
gn.
bn. gn. mott.
gn.
35
gn.
bn.
gn.
y.
c s vf f m cs vcs g p
c s vf f m cs vcs g p

SB3
bn.
bn. gn. purp. mott.
bn.
gn.
80
bn.gn. mott
gn.
gn. w.
gn.
bn.
gn. bn. mott.
75
gn.
gn.
Y.
70
Y.
dk. gy
Y.
buff.
gn.
purp.
purp. gn. mott.
purp.
65
purp.
bn
gn.
bn.
gn. w.
60
buff
Y. bn.
c s vf f m cs vcs g p
buff.
purp. gn. mott
buff
dk. gn.
90
gn.
gn.
bn.
85
c s vf f m cs vcs g p

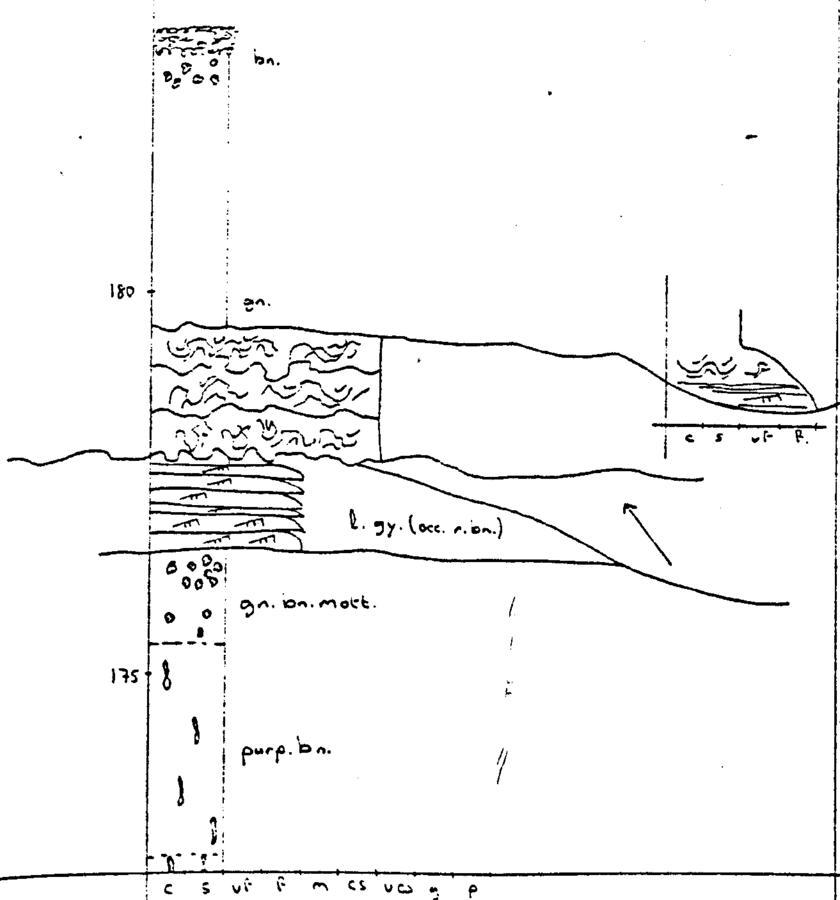


SB4
São Bernardino Log
120
bn (gn. mott)
gn.
bn.
gn.
gn. gy.
bn. (gn. mott.)
115
gn. bn. mott
gn. (purp. mott.)
110
bn.
gn.
gn.
buff.
gn.
gn.
buff.
gn.
105
buff. y. bn.
gn.
occ. gn. mott.
bn.
dk. gn.
buff. gn.
100
gn.
bn
gn.
(Towards margin)
dk. gn
95
buff
gn dk
c s vf f m cs vc g p
135
buff.
gn.
r. bn.
buff.
gy.
gn.
buff.
c s
gy.
buff.
130
dk. gy.
gn. bn mott.
gn.
125
Y. bn.
r. bn.
c s vf f m cs vc g p

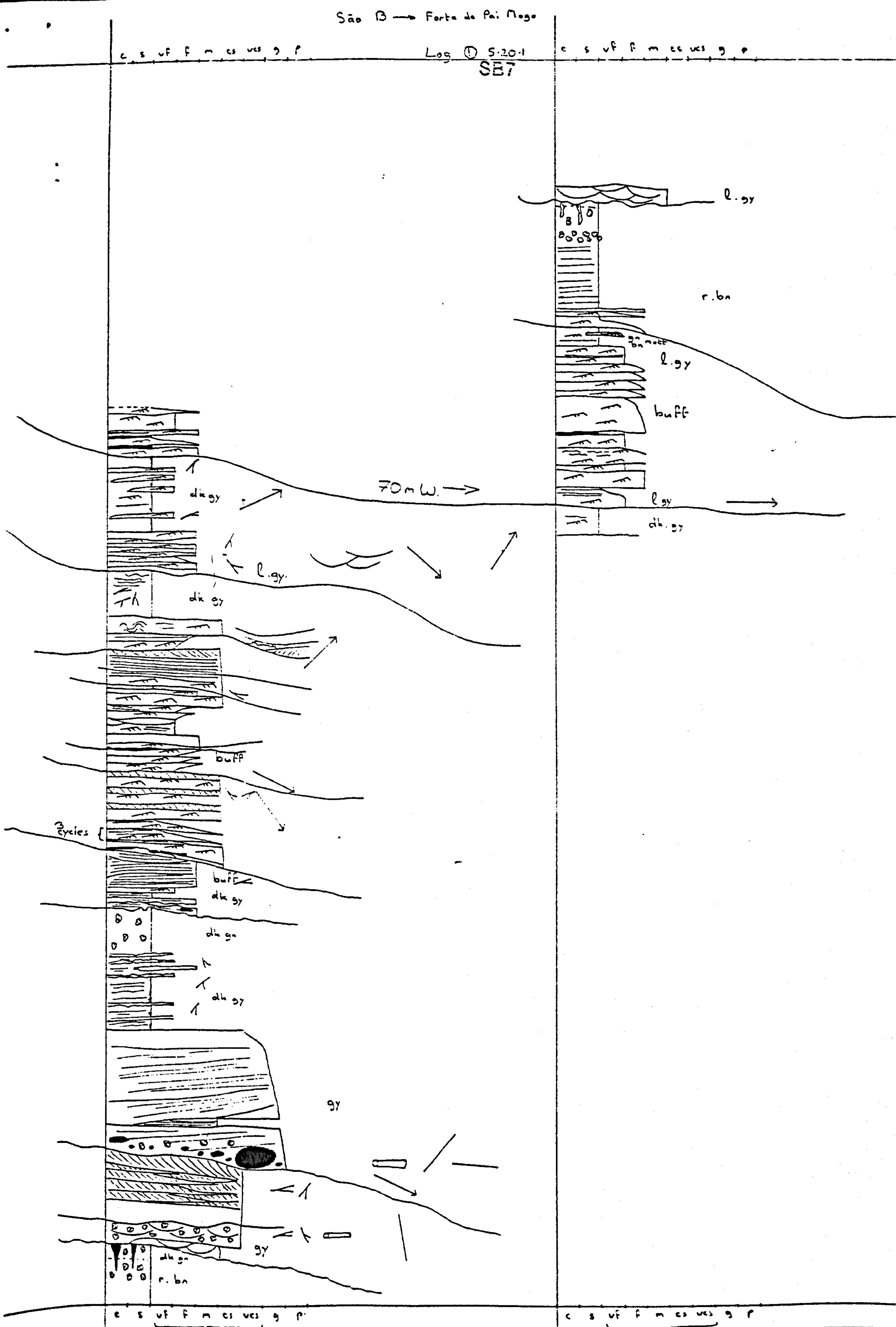

SB5
San Bernardino
Log # 5
r. gn. mott.
bn.
gn. bn. mott.
gn. gy.
dk. gy.
buff.
dk. gy.
dk. gy.
gn.
l. gy.
gn. (purp. mott.)
bn.
gn.
bn.
gn.
bn.
purp.
bn.
dk. gn.
bn.
gn. w.
bn.
gn.
gn. gy.
l. gn. gy.
w. gn.
dk. gn.
gy
gn. gy.
gn.
gn. gy.
gn.
gn. (purp. mott)
bn.
gn. bn. mott.
gn. bn. mott.
bn.
bn. gn. mott.
gn. purp.
bn.
bn.
bn.
bn. (gn. mott.)
dk. gn.
bn. gn. mott.
c s vf f m cs vcs g p
c s vf f m cs vcs g p

SB6

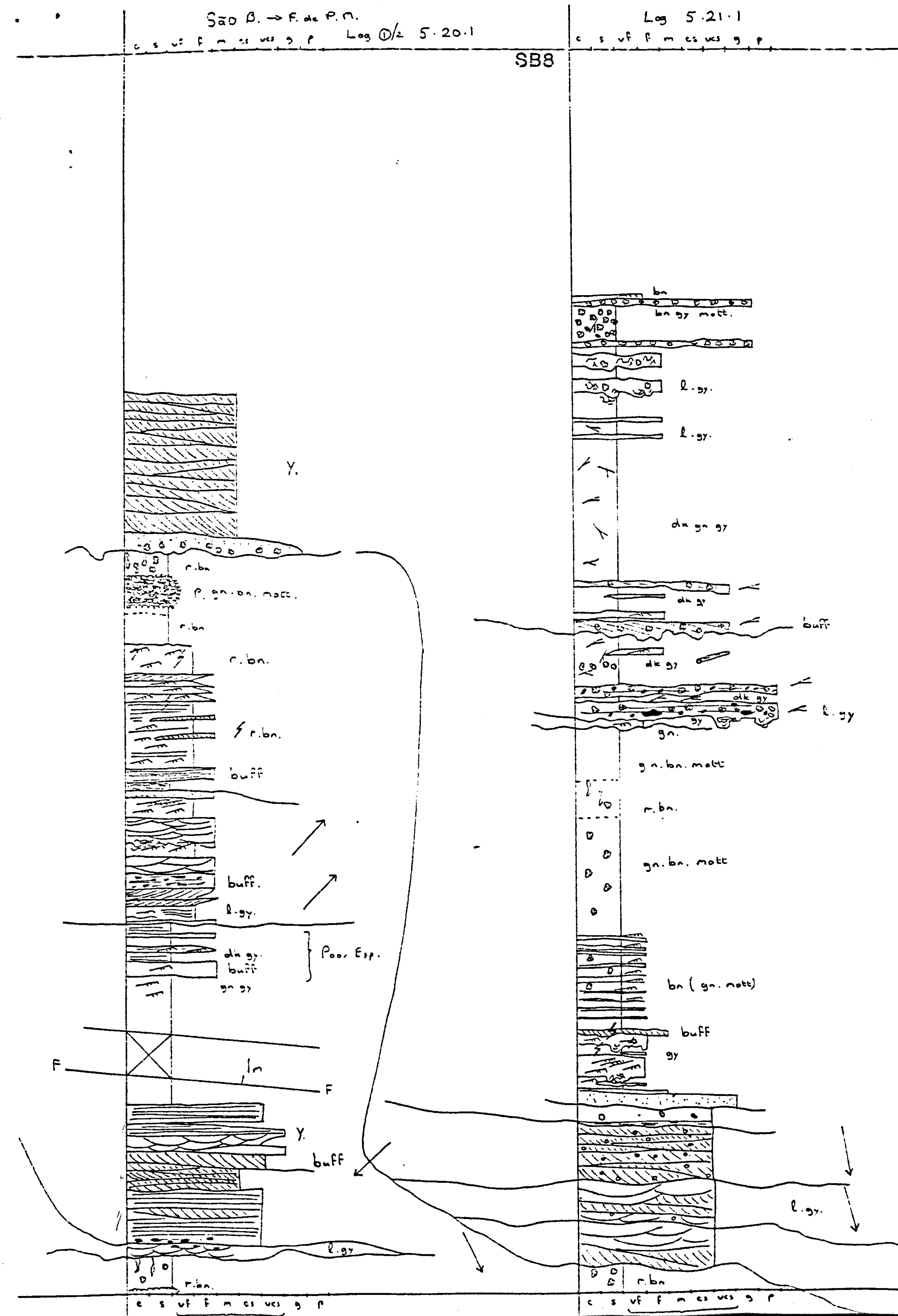
São Bernardino Log #6

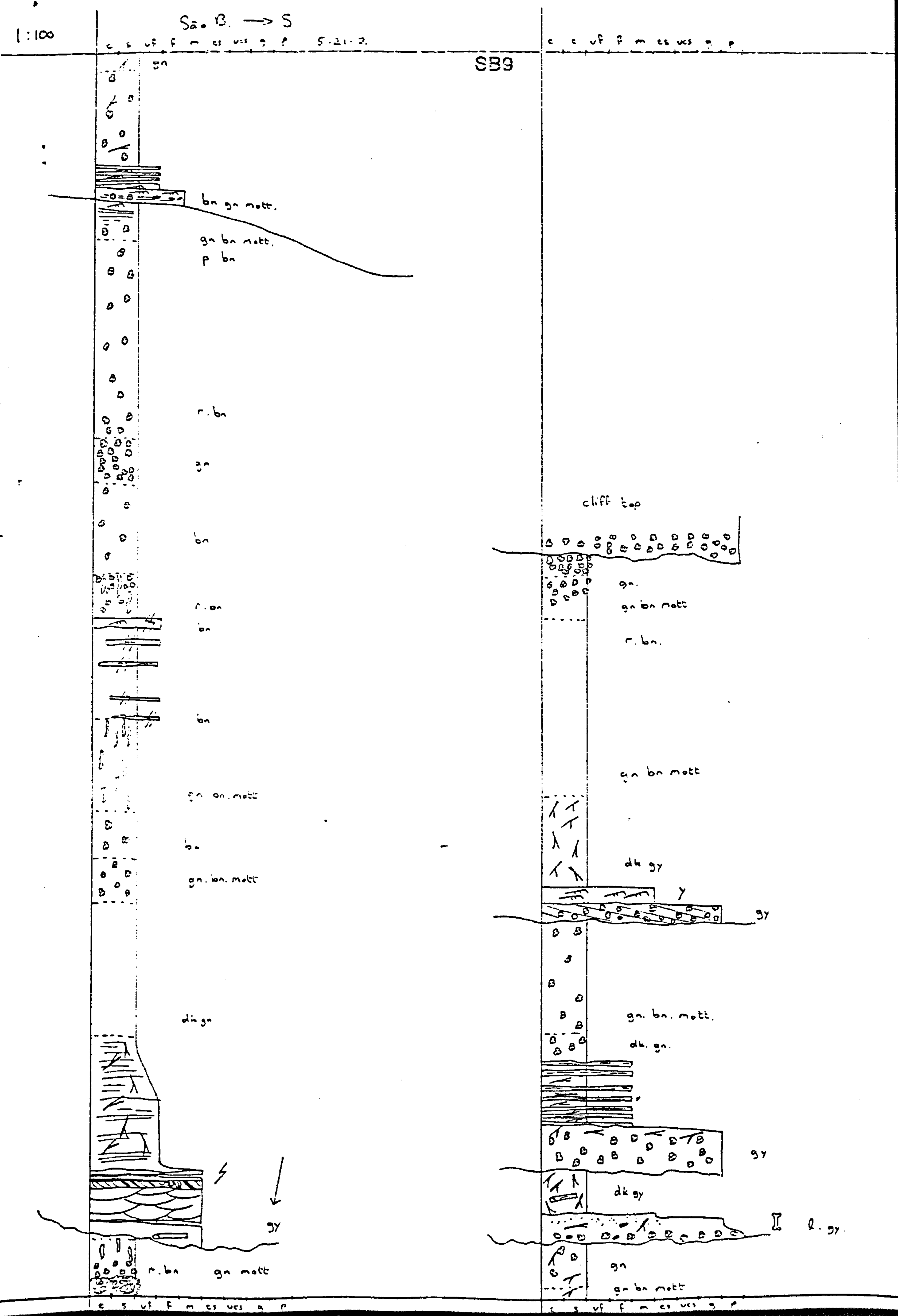


c s vf f m cs vcs g p



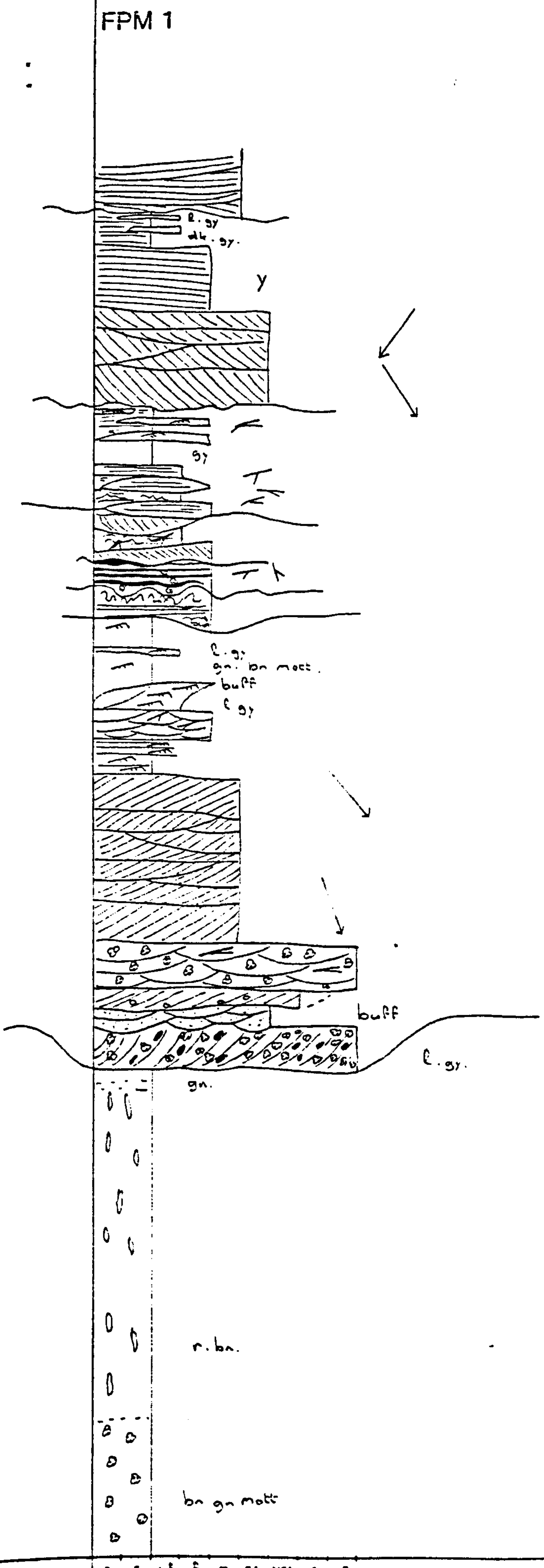
São B → Forte de Pai Mogo
Log ① 5·20·1
SB7
c s vf f m cs vcs g p
l. gy
r. bn
l. gy
buff
l gy
dk. gy
70 m W.
dk gy
l. gy.
dk gy
buff
3 cycles
buff
dk gy
dk gn
dk gy
gy
gy
dk gn
r. bn


São B. → F. de P.M.
Log ①/2 5·20·1
c s vf f m cs vcs g p
SB8
Y.
r.bn
P. gn.-bn. mott.
r.bn
r.bn.
r.bn.
buff
buff.
l.gy.
dk gy.
buff
gn gy
Poor Exp.
F
In
F
Y.
buff
l.gy
r.bn.
c s vf f m cs vcs g p
Log 5·21·1
c s vf f m cs vcs g p
bn
bn gy mott.
l.gy.
l.gy.
dk gn gy
dk gy
buff
dk gy
dk gy
l.gy
gy
gn.
gn.bn.mott
r.bn.
gn.bn.mott
bn (gn. mott)
buff
gy
l.gy.
r.bn
c s vf f m cs vcs g p

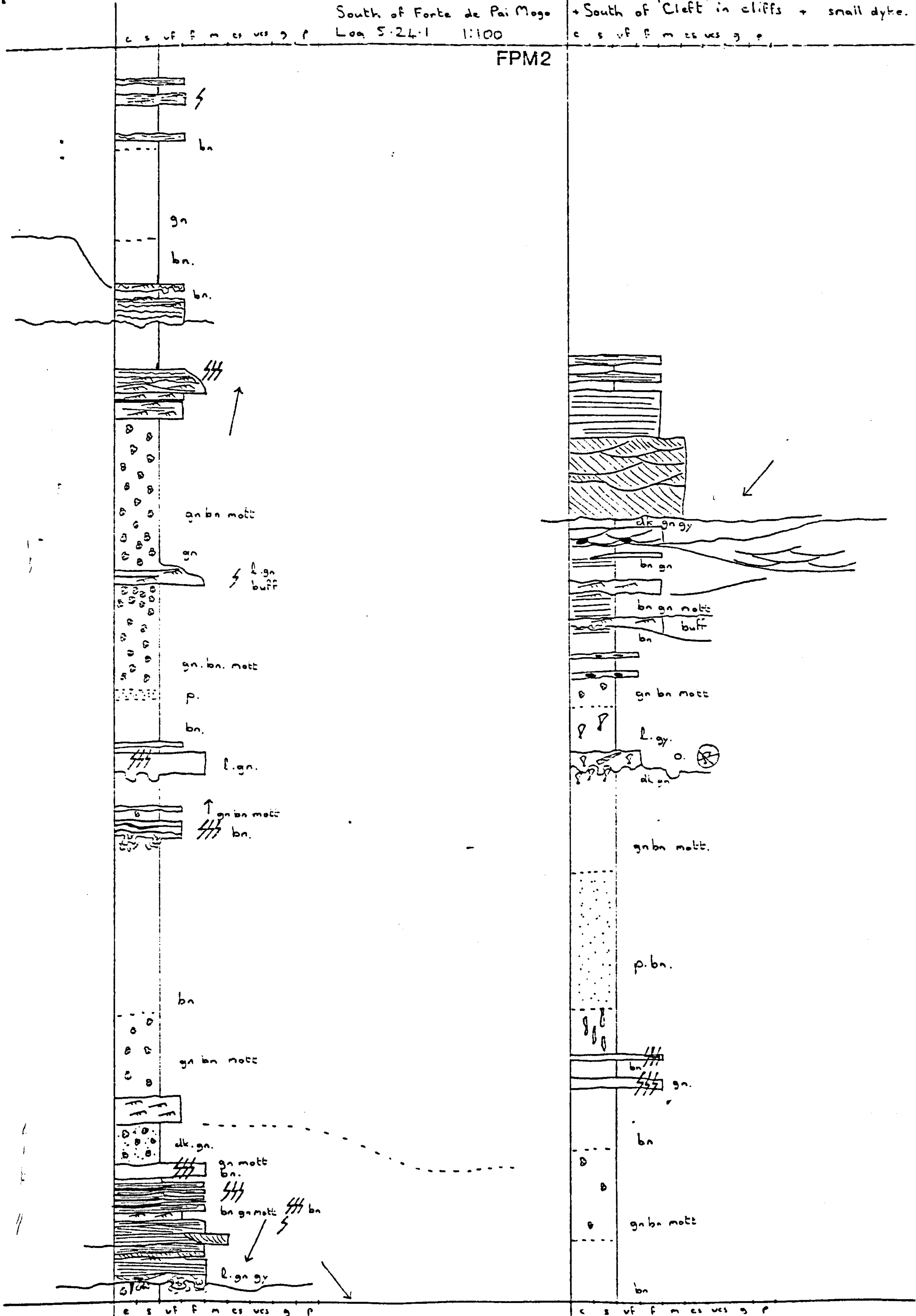

1:100
São B. → S
5.21.2.
SB9
c s vf f m cs vcs g p
c s vf f m cs vcs g p
gn
bn gn mott.
gn bn mott.
p bn
r. bn
gn
bn
r. bn
bn
bn
gn bn. mott
bn
gn. bn. mott
dk gn
gy
r. bn
gn mott
cliff top
gn.
gn bn mott
r. bn.
gn bn mott
dk gy
gy
gn. bn. mott.
dk. gn.
gy
dk gy
l. gy.
gn
gn bn mott
c s vf f m cs vcs g p
c s vf f m cs vcs g p

N. of. F. de Pai Mogo

c s vf f m cs vcs g p   Lug 5·22·1   c s vf f m cs vcs g p



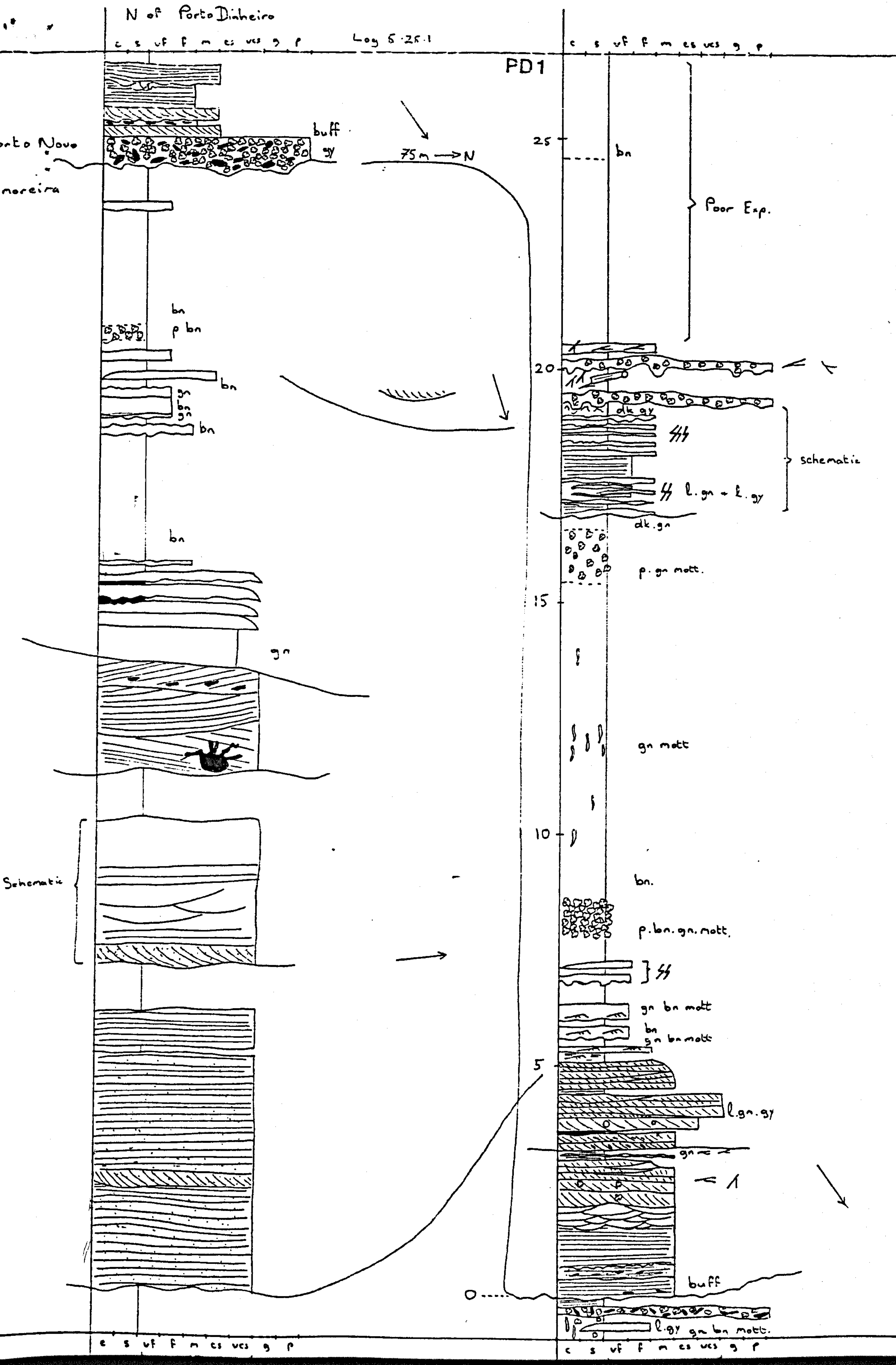
c s vf f m cs vcs g p   c s vf f m cs vcs g p


South of Forte de Pai Mogo
Log 5.24.1 1:100
FPM2
+ South of 'Cleft' in cliffs + small dyke.
c s vf f m cs vcs g p
bn
gn
bn.
bn.
gn bn mott
gn
l. gn buff
gn. bn. mott
p.
bn.
l. gn.
gn bn mott
bn.
bn
gn bn mott
dk. gn.
gn mott bn.
bn gn mott
bn
l. gn gy
dk gn gy
bn gn
bn gn mott
buff
bn
gn bn mott
l. gy.
dk gn
gn bn mott.
p. bn.
bn
gn.
bn
gn bn mott
bn

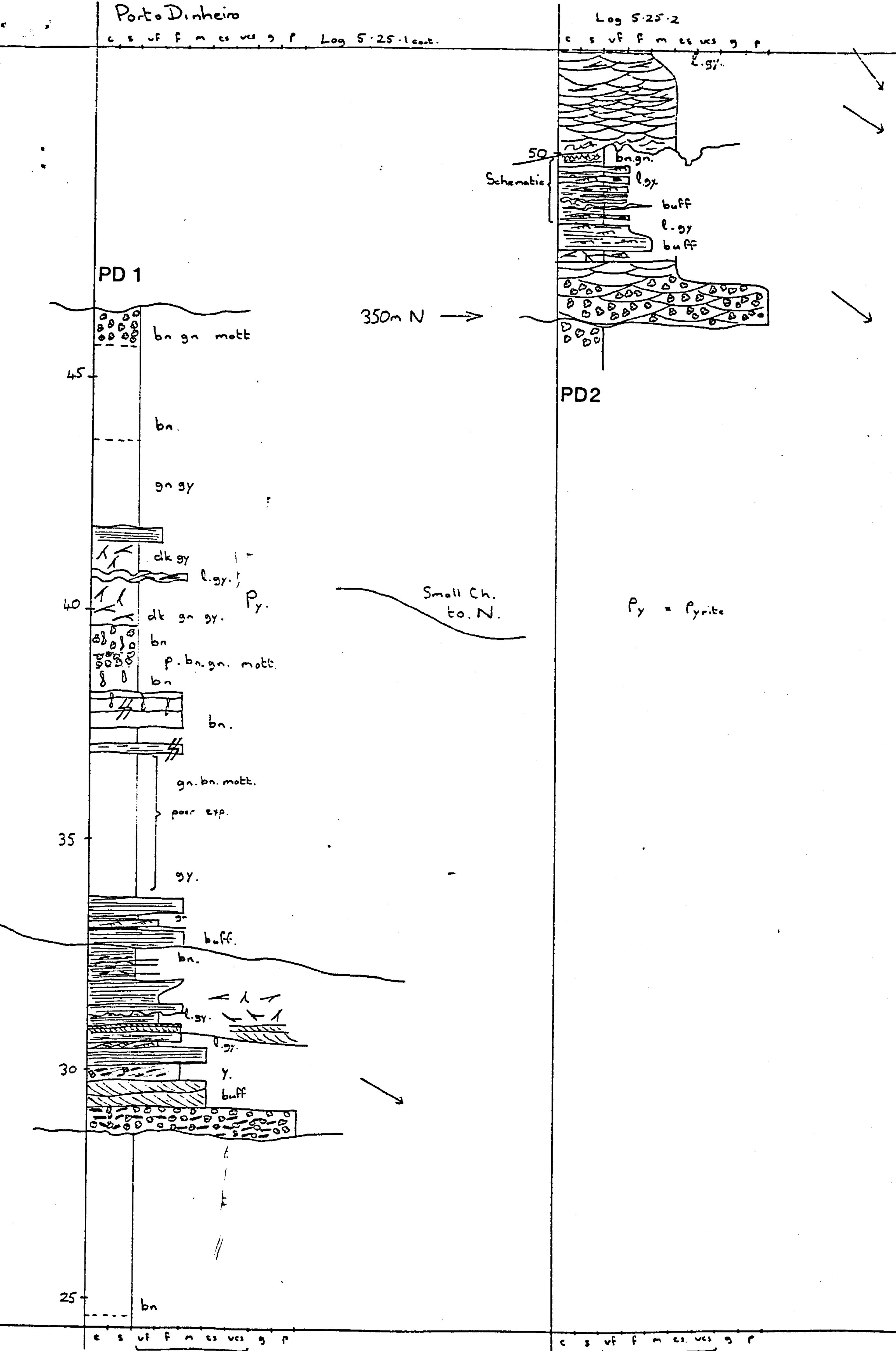
ENCLOSURE 4

LOGS: PORTO DINHEIRO NORTH-AREIA BRANCA

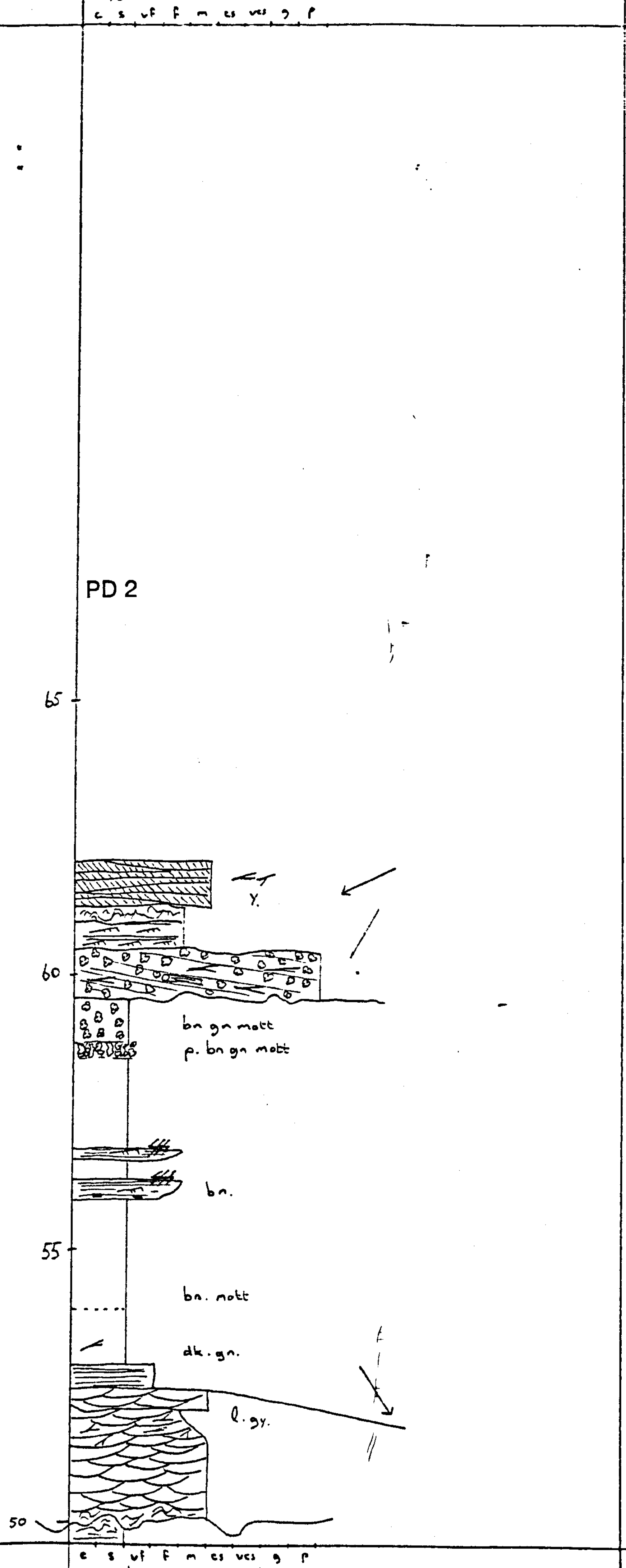
(PD 1-4, PDB 1,AB1)

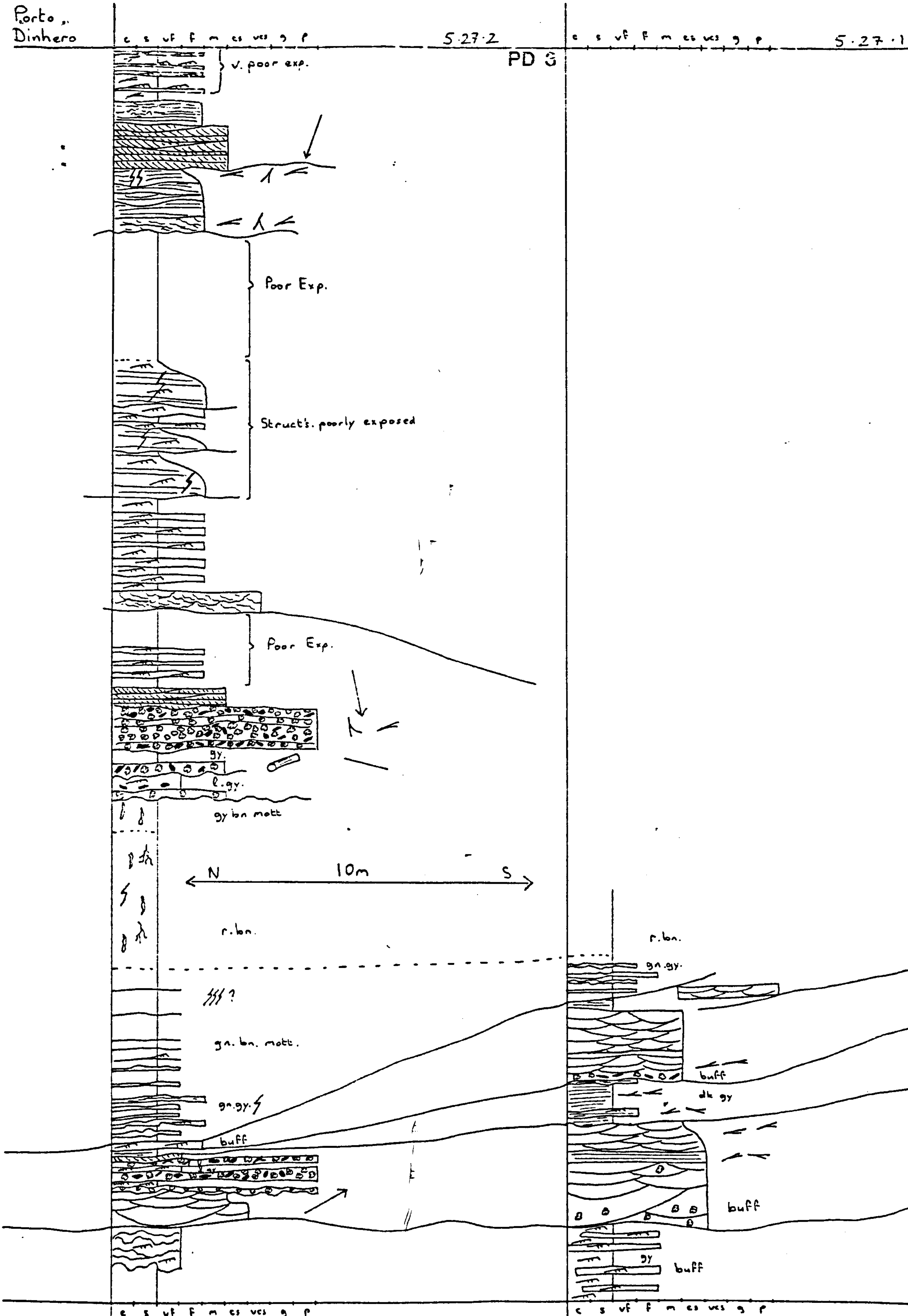


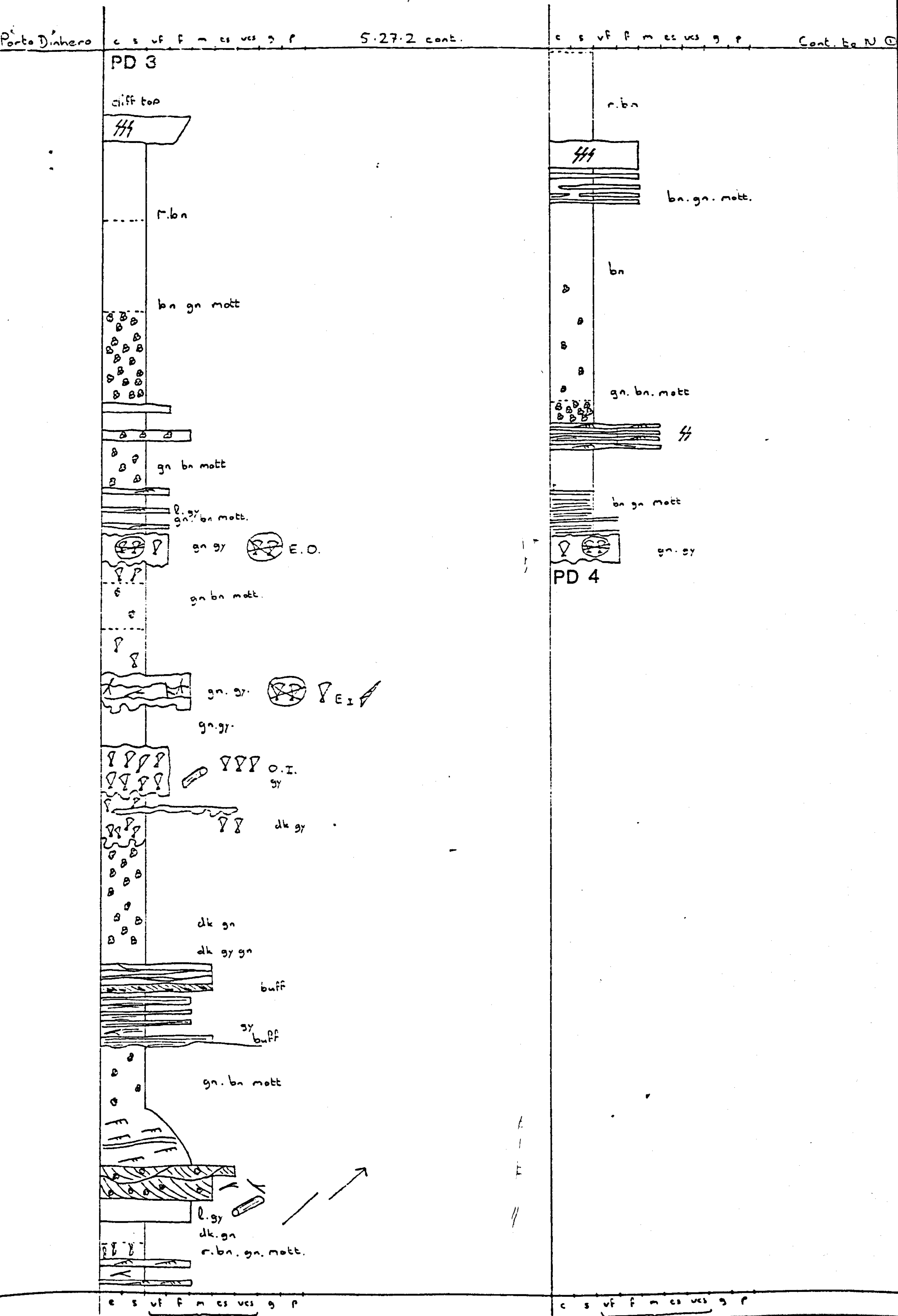
N of Porto Dinheiro
Log 5.25.1
PD1
c s vf f m cs vcs g p
buff
gy
Porto Novo
moreira
75m → N
25
bn
Poor Exp.
bn
p bn
bn
gn
bn
gn
bn
20
dk gy
schematic
l. gn + l. gy
dk. gn
bn
p. gn mott.
15
gn
gn mott
10
Schematic
bn.
p. bn. gn. mott.
gn bn mott
bn
gn bn mott
5
l. gn. gy
gn
buff
0
l. gy gn bn mott.
c s vf f m cs vcs g p


Porto Dinheiro
c s vf f m cs vcs g p
Log 5·25·1 cont.
Log 5·25·2
c s vf f m cs vcs g p
l.gy.
50
bn.gn.
Schematic
l.gy
buff
l.gy
buff
PD 1
bn gn mott
45
350m N
PD2
bn.
gn gy
dk gy
l.gy.
Py.
40
dk gn gy.
Small Ch. to. N.
Py = Pyrite
bn
p. bn. gn. mott.
bn
bn.
gn. bn. mott.
poor exp.
35
gy.
gn
buff.
bn.
l.gy.
l.gy.
30
y.
buff
25
bn
c s vf f m cs vcs g p
c s vf f m cs vcs g p

Porto Dinhero 5·25·2 Cont.

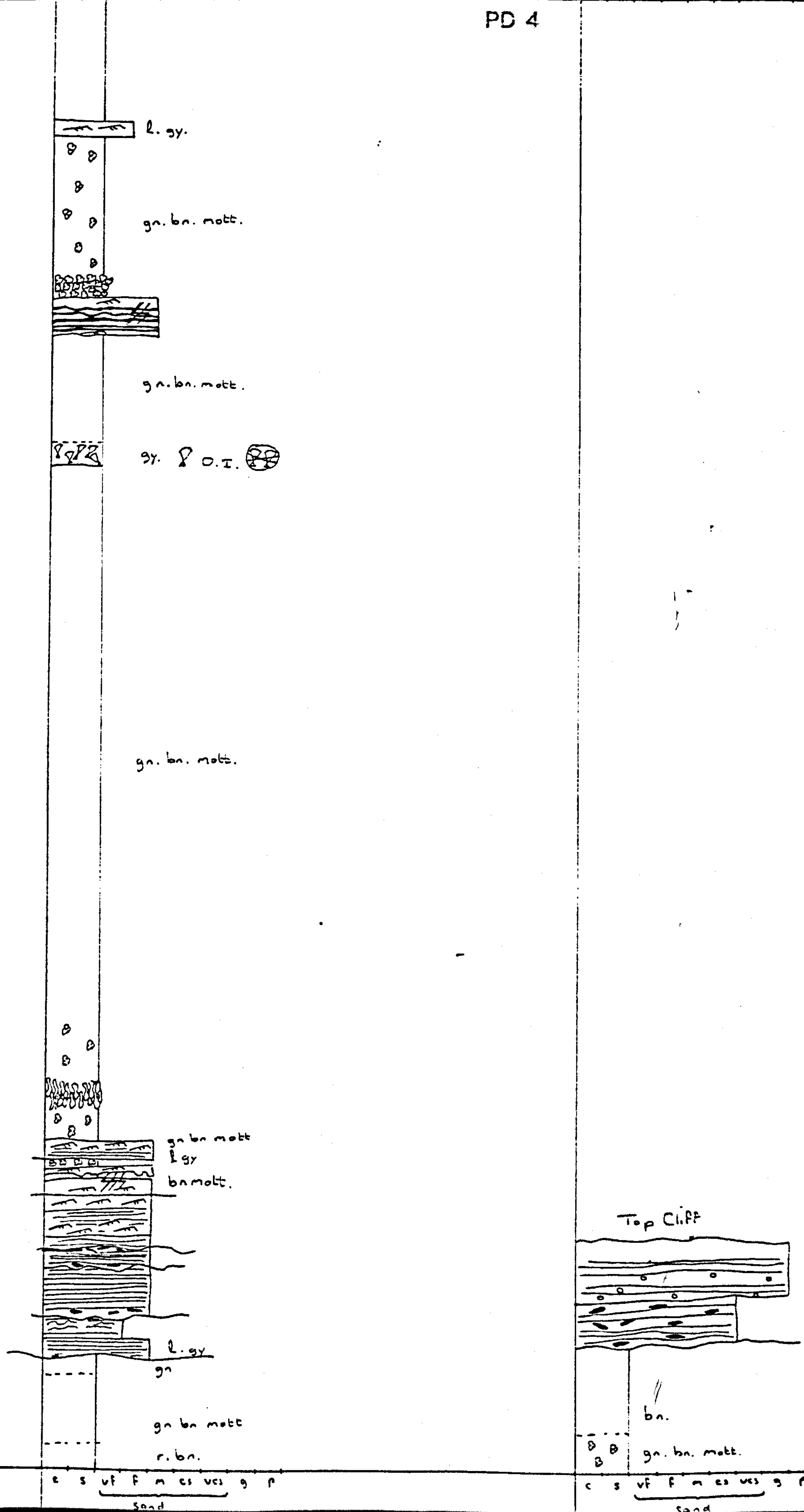



Porto Dinhero
c s vf f m cs vcs g p
5.27.2
PD 3
c s vf f m cs vcs g p
5.27.1
v. poor exp.
Poor Exp.
Struct's. poorly exposed
Poor Exp.
gy.
l. gy.
gy bn mott
N
10m
S
r. bn.
r. bn.
gn. gy.
gn. bn. mott.
buff
dk gy
gn. gy.
buff
buff
gy
buff
c s vf f m cs vcs g p
c s vf f m cs vcs g p

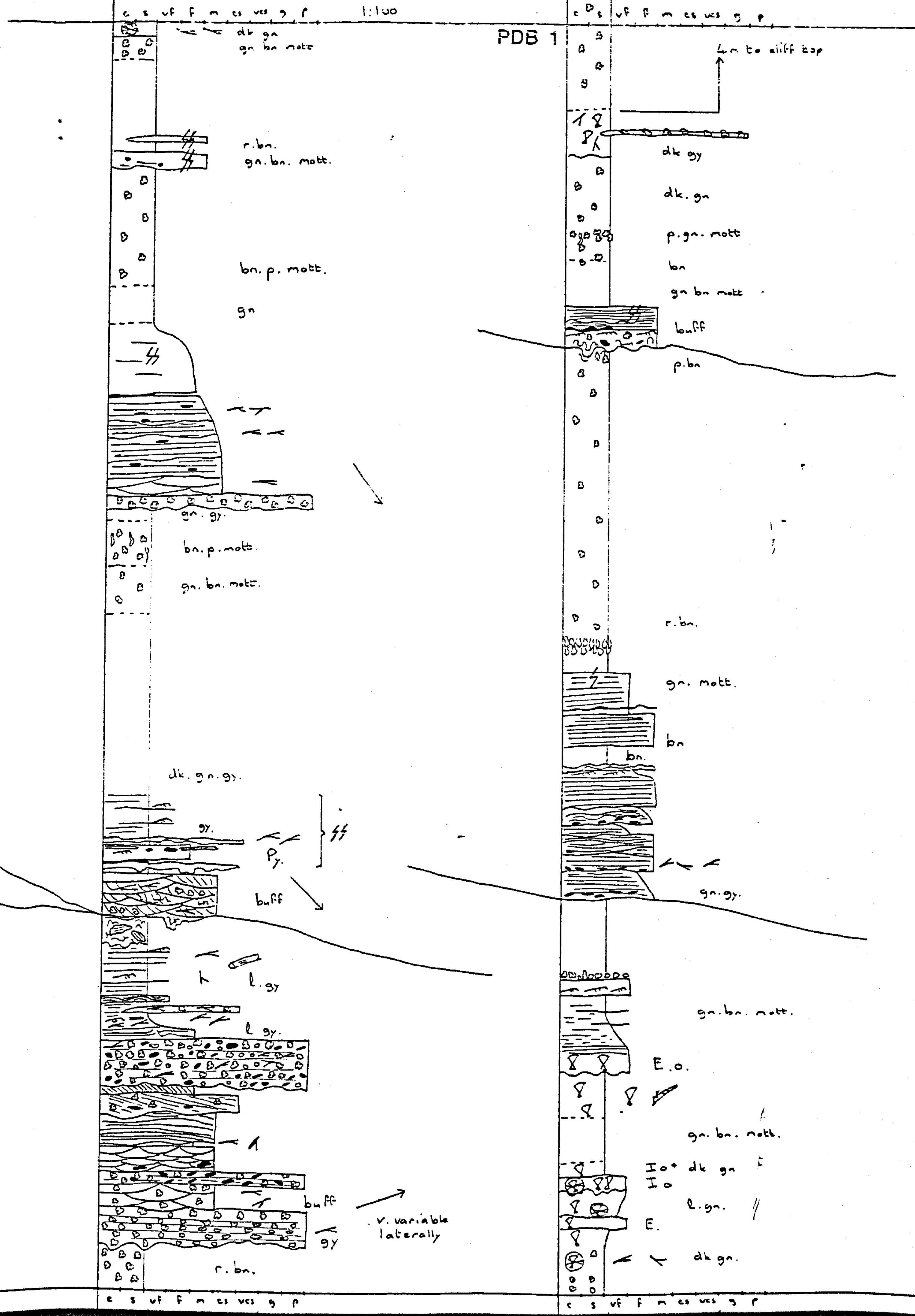

Porto Dinhero
c s vf f m cs vcs g p
5.27.2 cont.
PD 3
cliff top
r.bn
bn gn mott
gn bn mott
l.gy
gn bn mott.
gn gy
E.O.
gn bn mott.
gn. gy.
E I
gn.gy.
O.I.
gy
dk gy
dk gn
dk gy gn
buff
gy
buff
gn. bn mott
l.gy
dk.gn
r.bn. gn. mott.
c s vf f m cs vcs g p
Cont. to N ①
r.bn
bn. gn. mott.
bn
gn. bn. mott
bn gn mott
gn. gy
PD 4

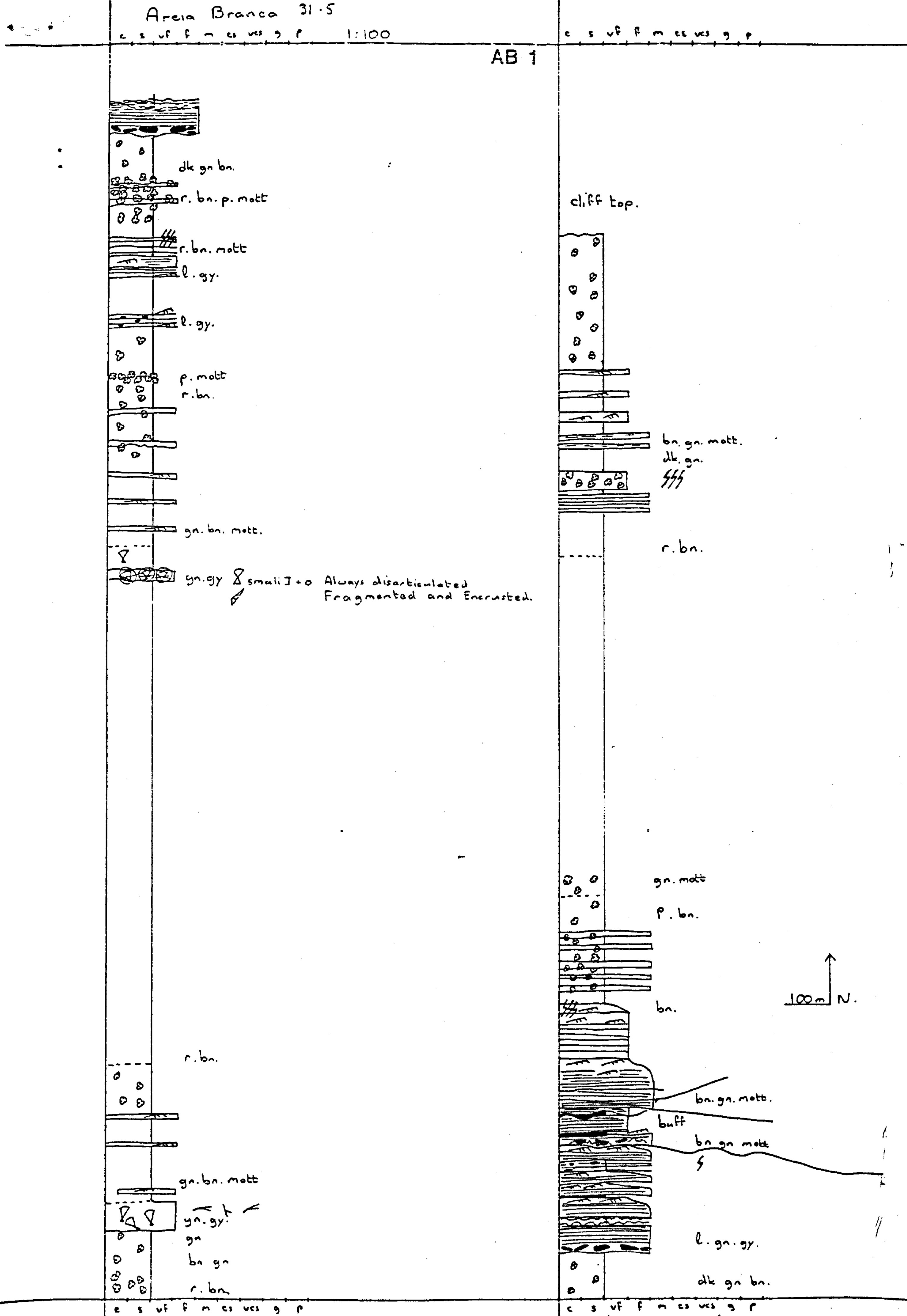
Porto Dinheiro | c s vF f m cs vcs g p | 5.27.2 Cont. to N (2) | c s vF f m cs vcs g p

PD 4



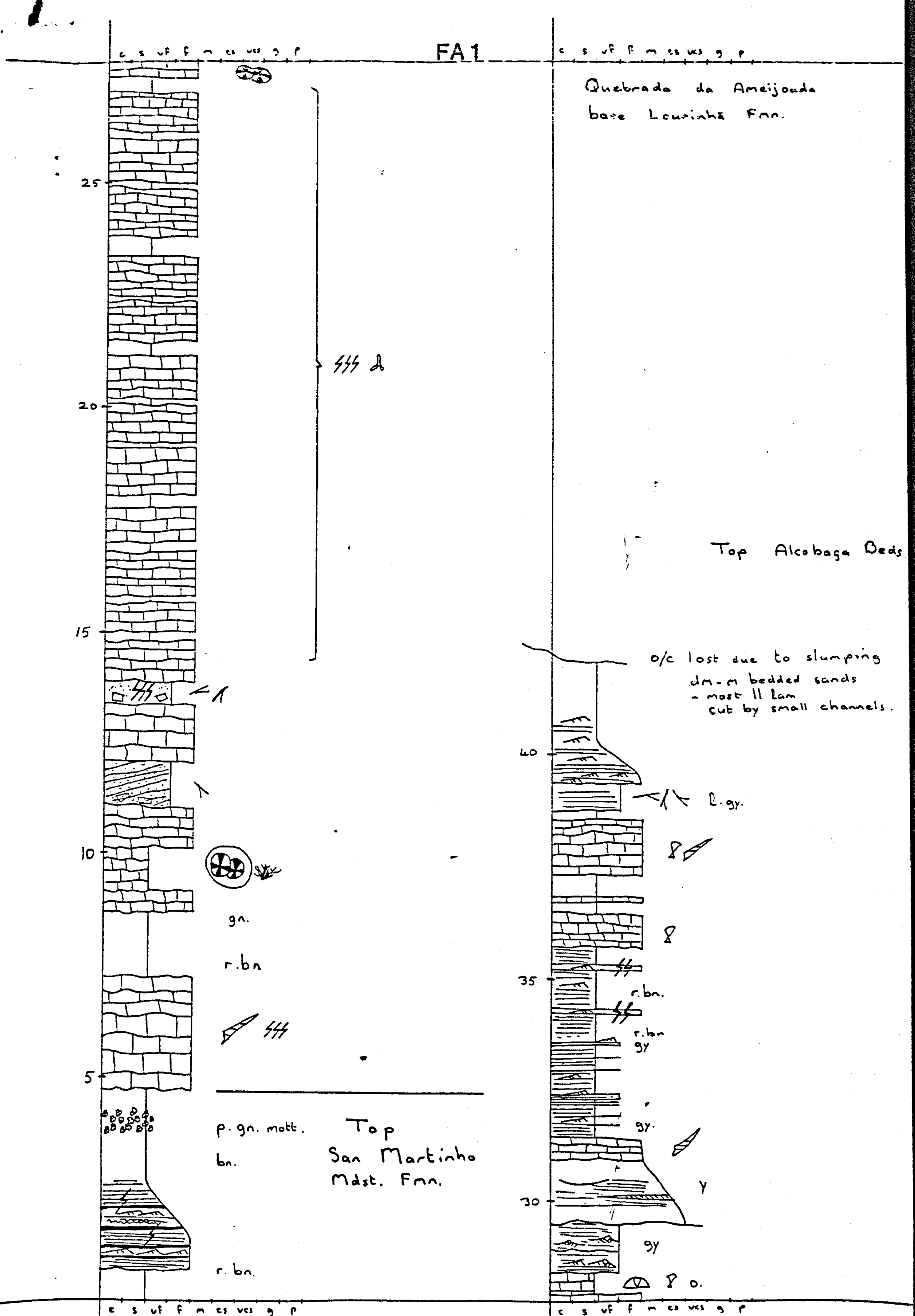
Porto de Barcas - "Fishermans Ladder" Section




Areia Branca 31·5
c s vf f m cs vcs g p
1:100
c s vf f m cs vcs g p
AB 1
dk gn bn.
r. bn. p. mott
r. bn. mott
l. gy.
l. gy.
p. mott
r. bn.
gn. bn. mott.
gn. gy
small J + o Always disarticulated
Fragmented and Encrusted.
r. bn.
gn. bn. mott
gn. gy.
gn
bn gn
r. bn
c s vf f m cs vcs g p
cliff top.
bn. gn. mott.
dk. gn.
r. bn.
gn. mott
p. bn.
bn.
100m
N.
bn. gn. mott.
buff
bn gn mott
l. gn. gy.
dk gn bn.
c s vf f m cs vcs g p

ENCLOSURE 5

LOGS: FOZ DO ARELHO SECTION (FA 1/2)

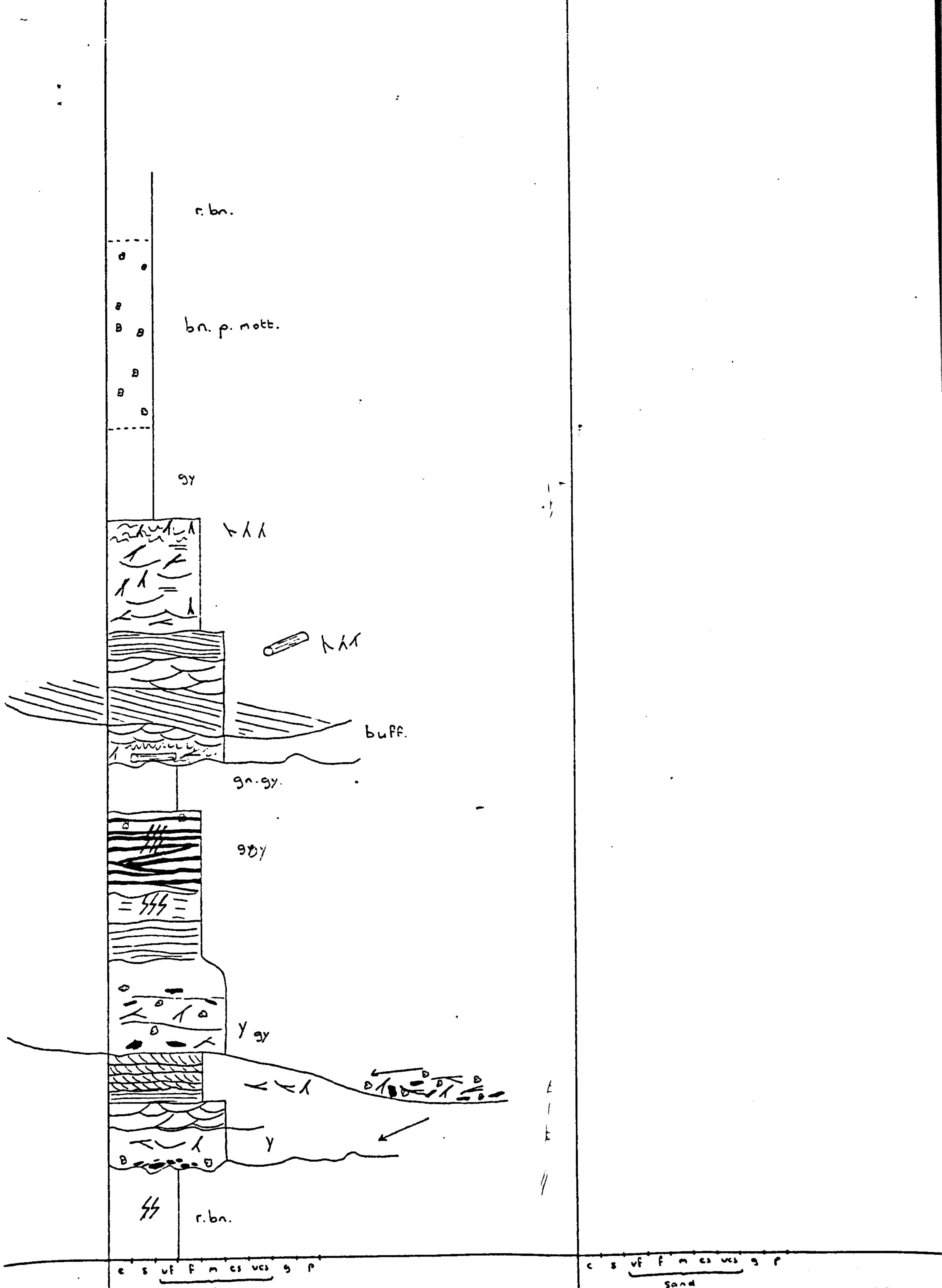


FA1
c s vf f m cs vcs g p
Quebrada da Ameijoada
base Lourinhã Fmn.
25
20
15
10
5
gn.
r.bn
p. gn. mott.
bn.
r. bn.
Top
San Martinho
Mdst. Fmn.
Top Alcobaça Beds
o/c lost due to slumping
dm-m bedded sands
- most ll lam
cut by small channels.
40
l. gy.
35
r.bn.
r.bn
gy
gy.
30
y
gy
o.
c s vf f m cs vcs g p

FA 2 Scale 1:50

10-20m above base of Lourinhã Fm.

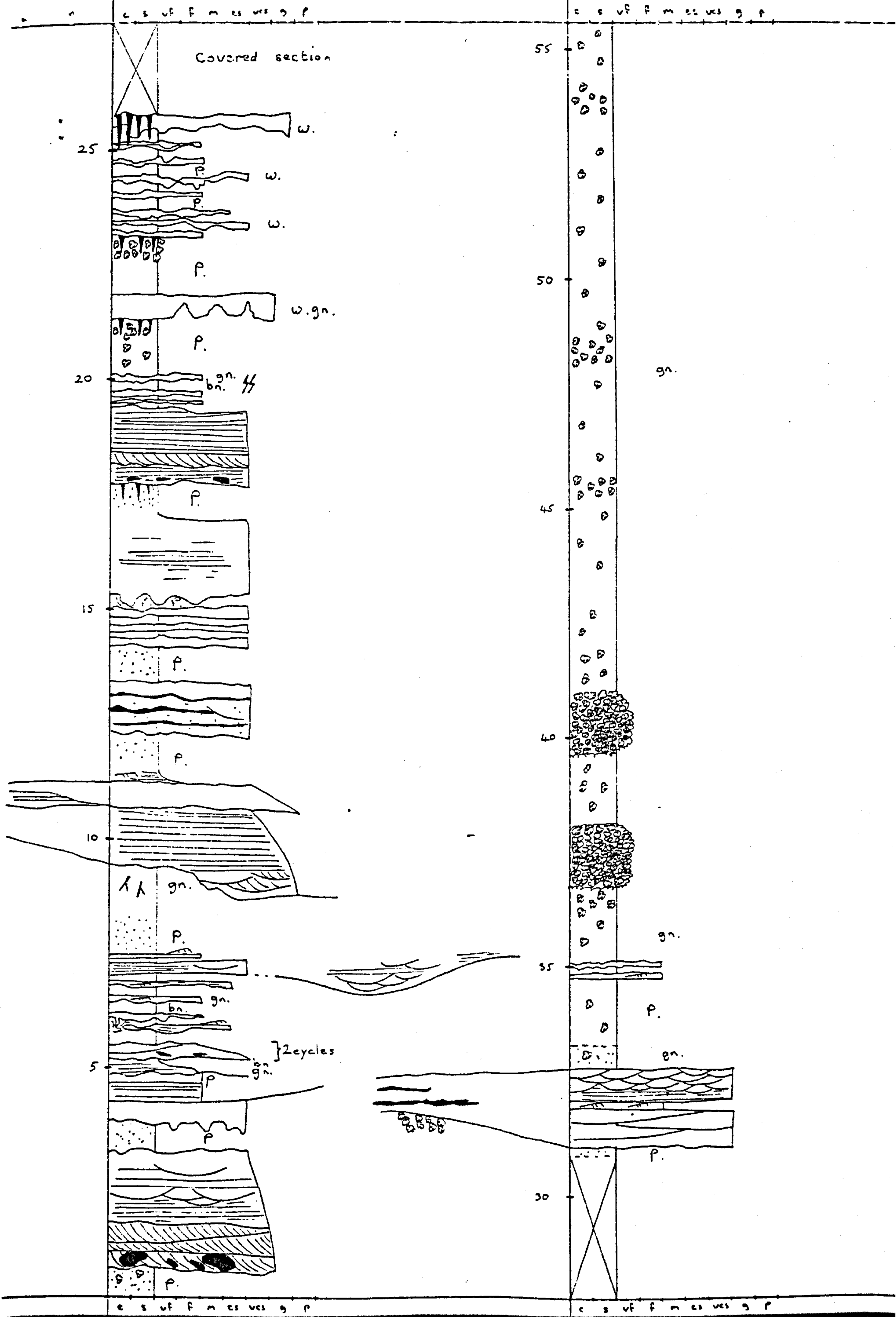
Quebrada da Ameijoada

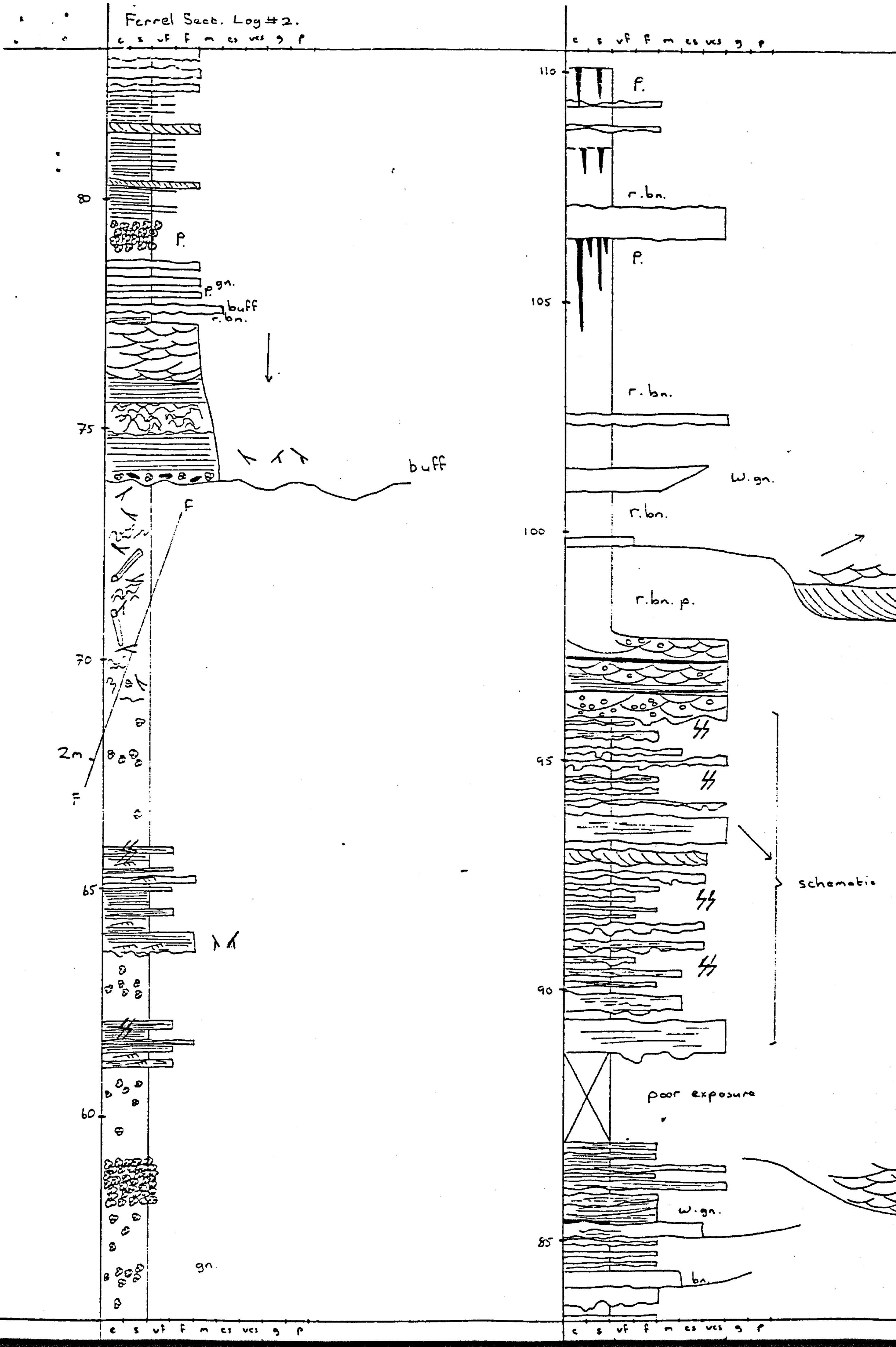


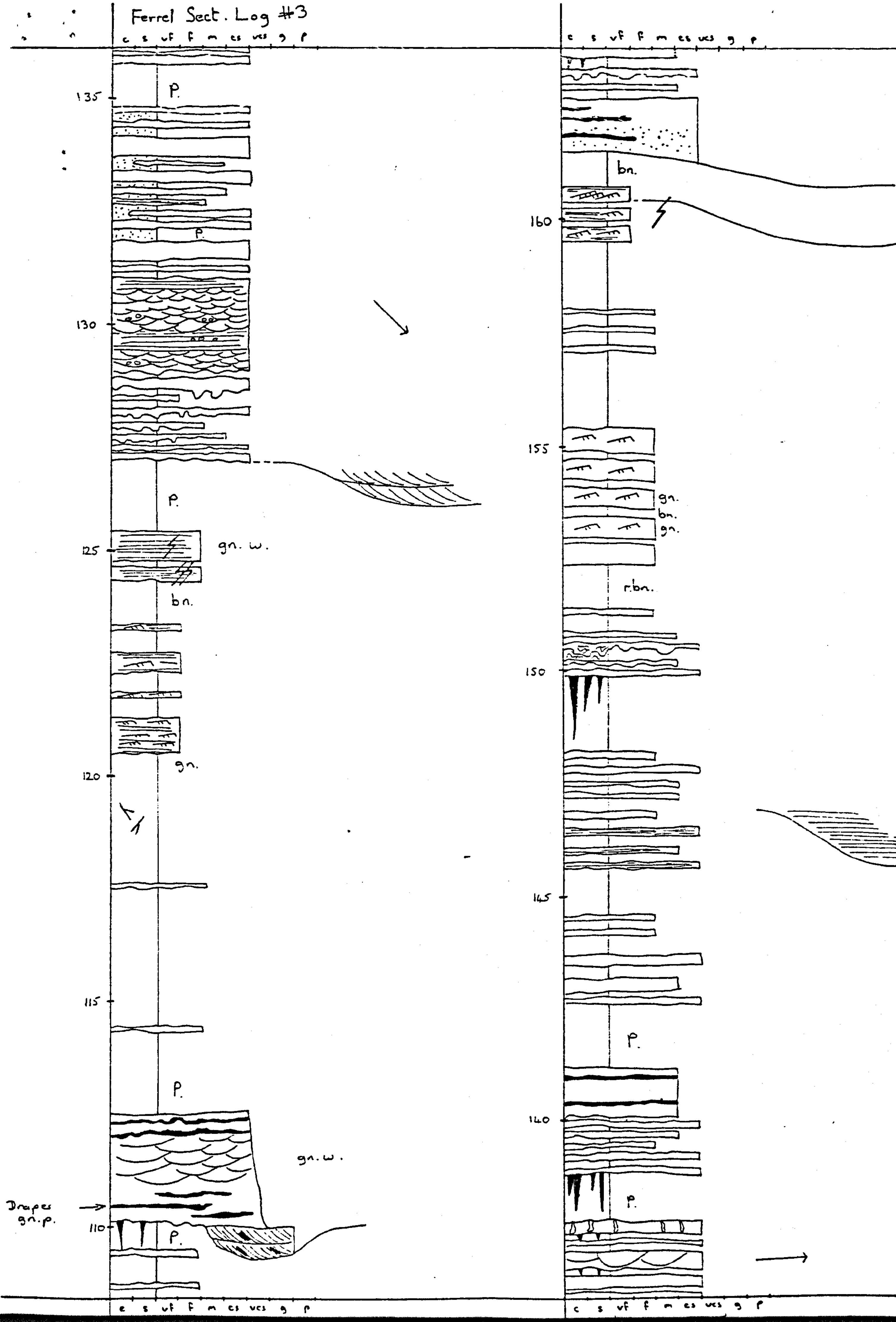
ENCLOSURE 6

LOGS: FERREL SECTION (1-13)

Ferrel Sect. Log #1

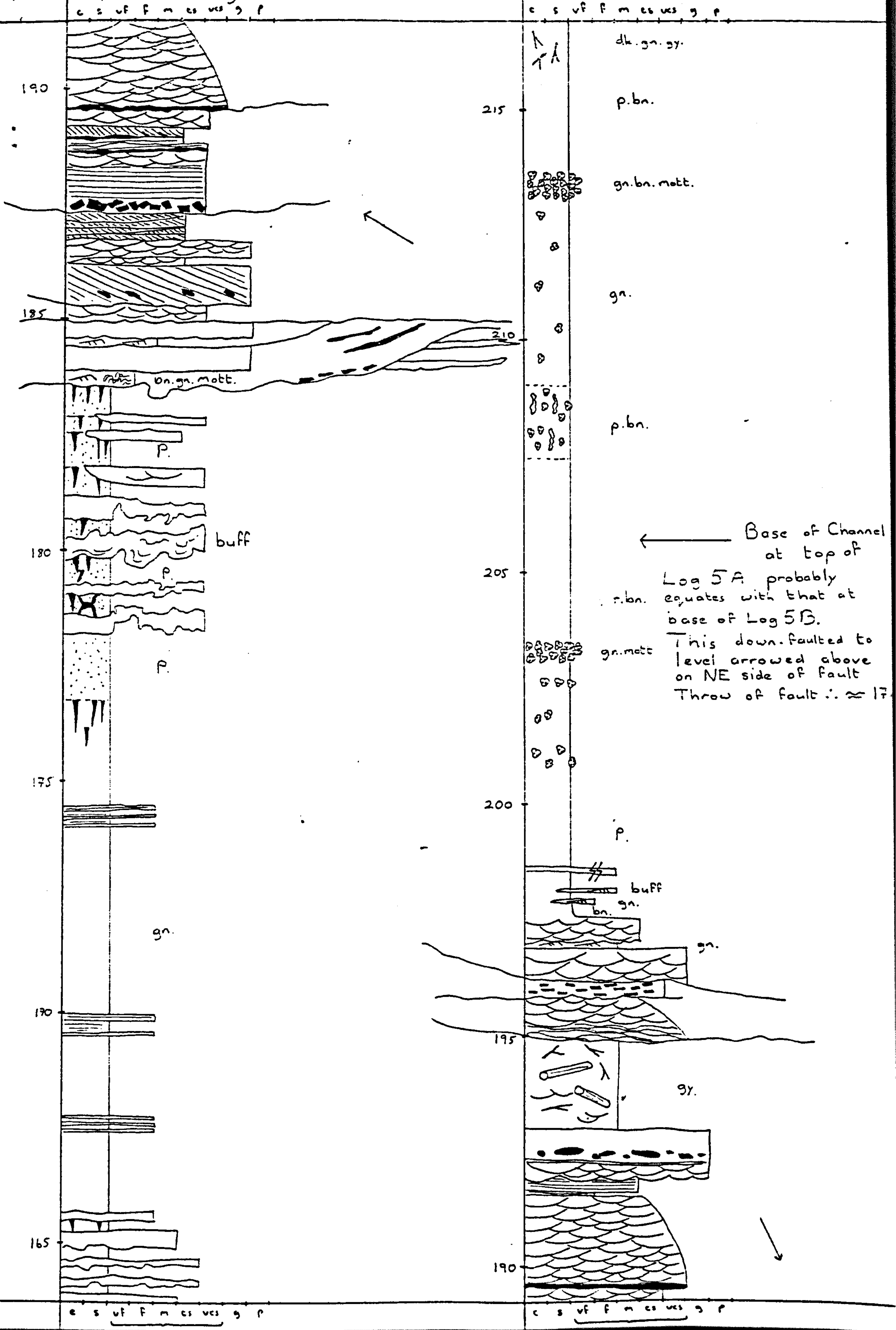



Ferrel Sect. Log #2.
c s vf f m cs vcs g p
80
75
70
65
60
2m
F
P.
p. gn.
buff
r.bn.
buff
gn.
110
105
100
95
90
85
P.
r.bn.
P.
r.bn.
w.gn.
r.bn.
r.bn.p.
schematic
poor exposure
w.gn.
bn.

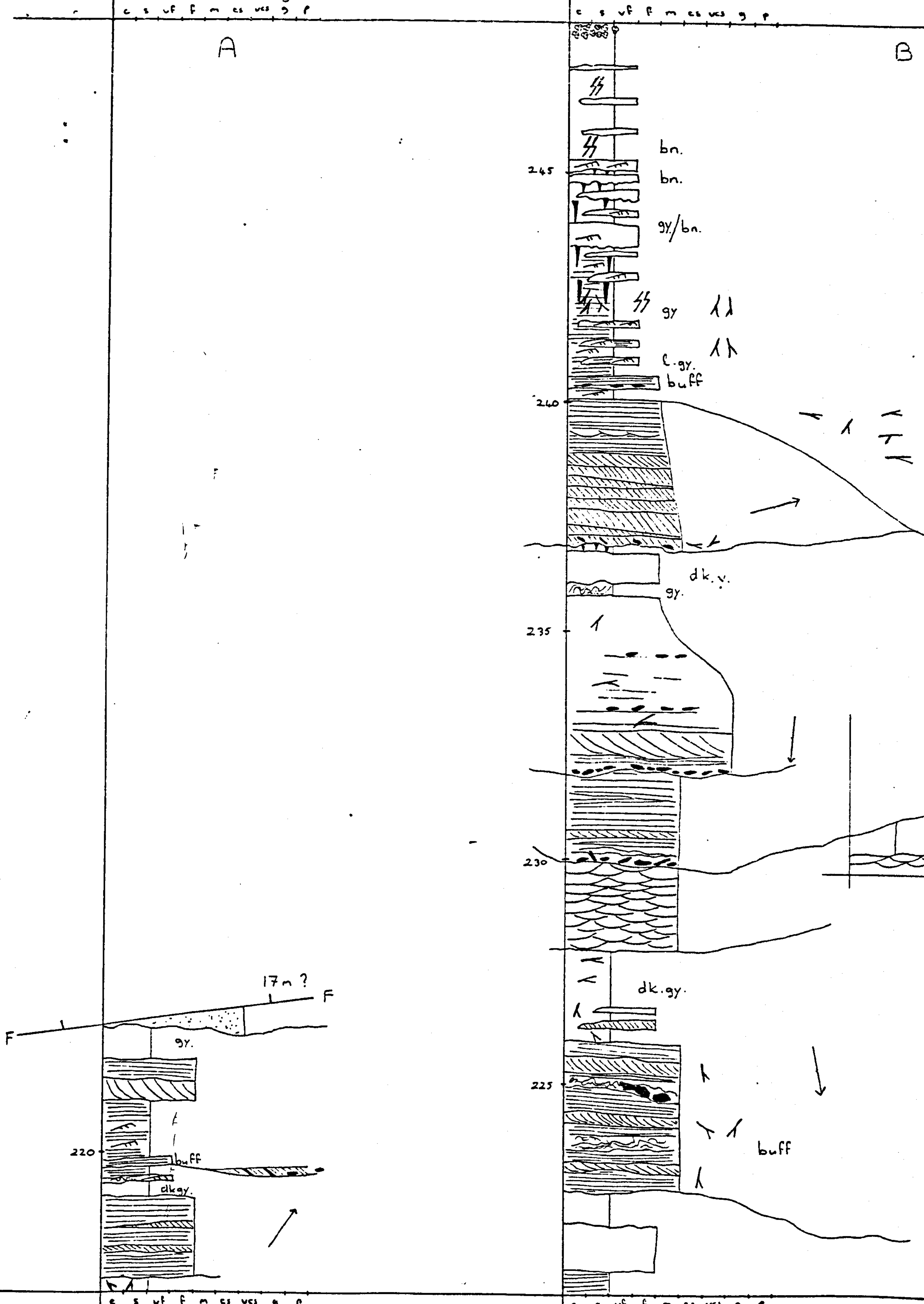


Ferrel Sect. Log #3
c s vf f m cs vcs g p
135
P.
P.
130
P.
125
gn. w.
bn.
gn.
120
115
P.
gn. w.
Drapes gn. p.
110
P.
c s vf f m cs vcs g p
bn.
160
155
gn.
bn.
gn.
r.bn.
150
145
P.
140
P.

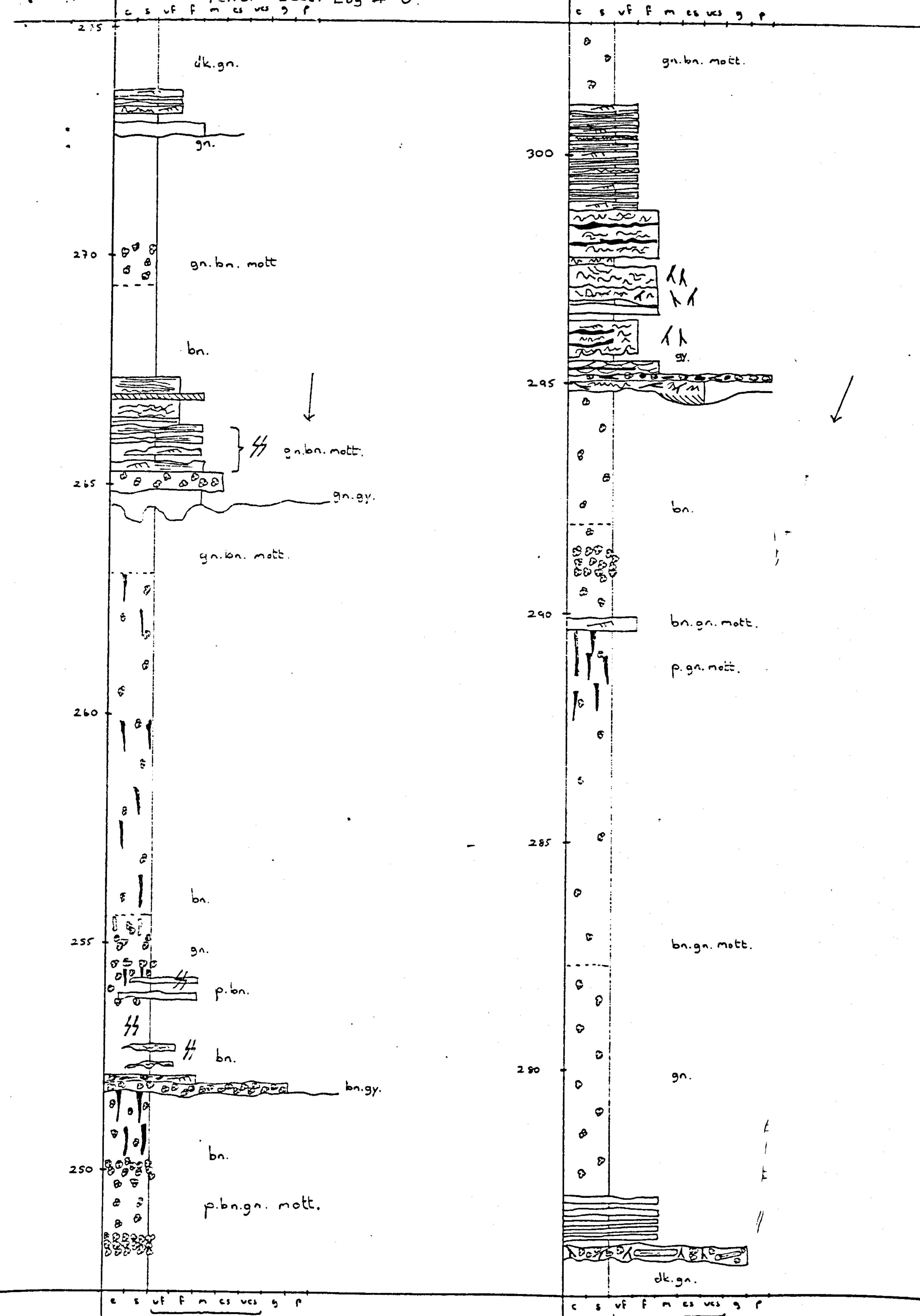
Ferrel Sect. Log # 4

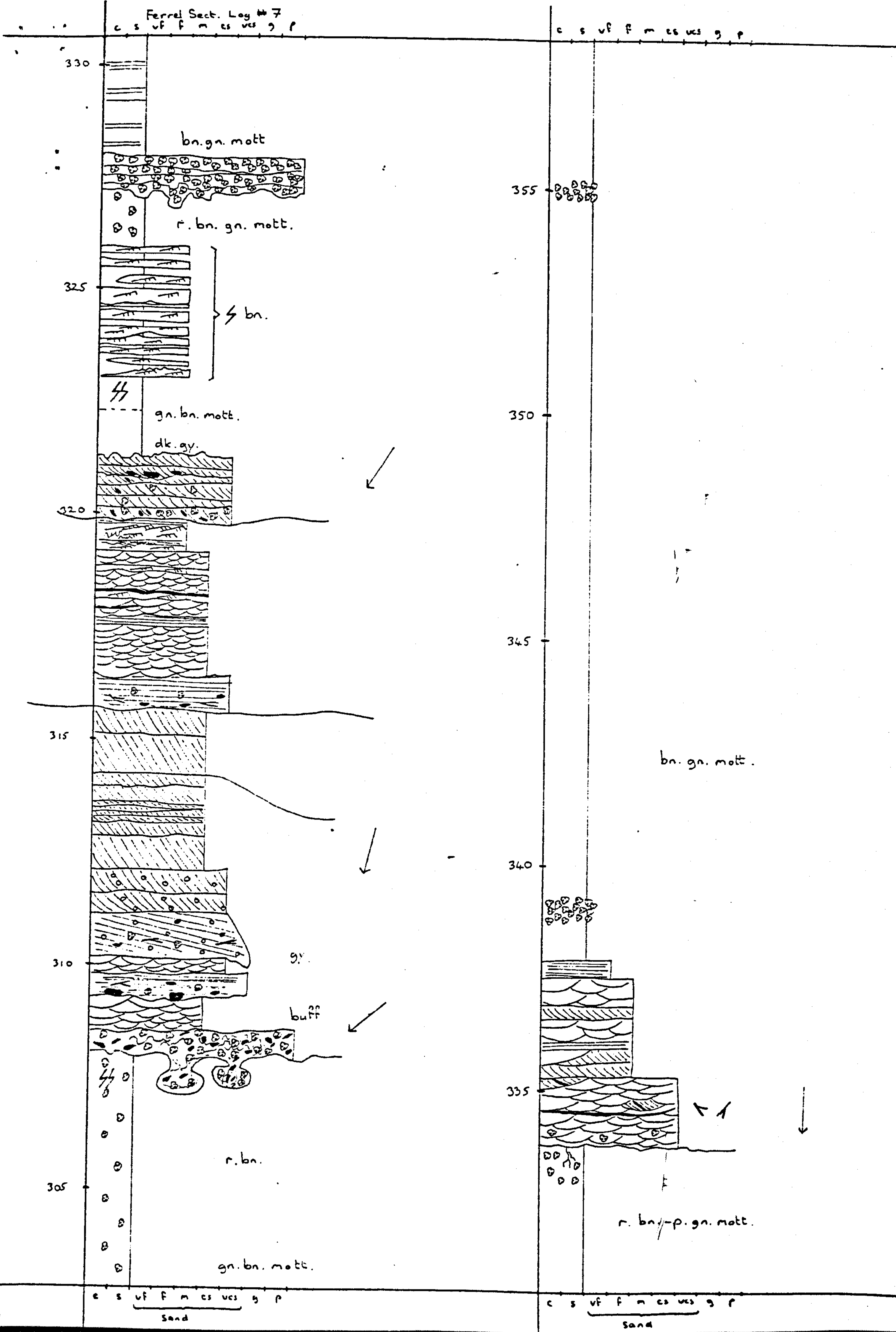


Ferrel Sect. Log # 5

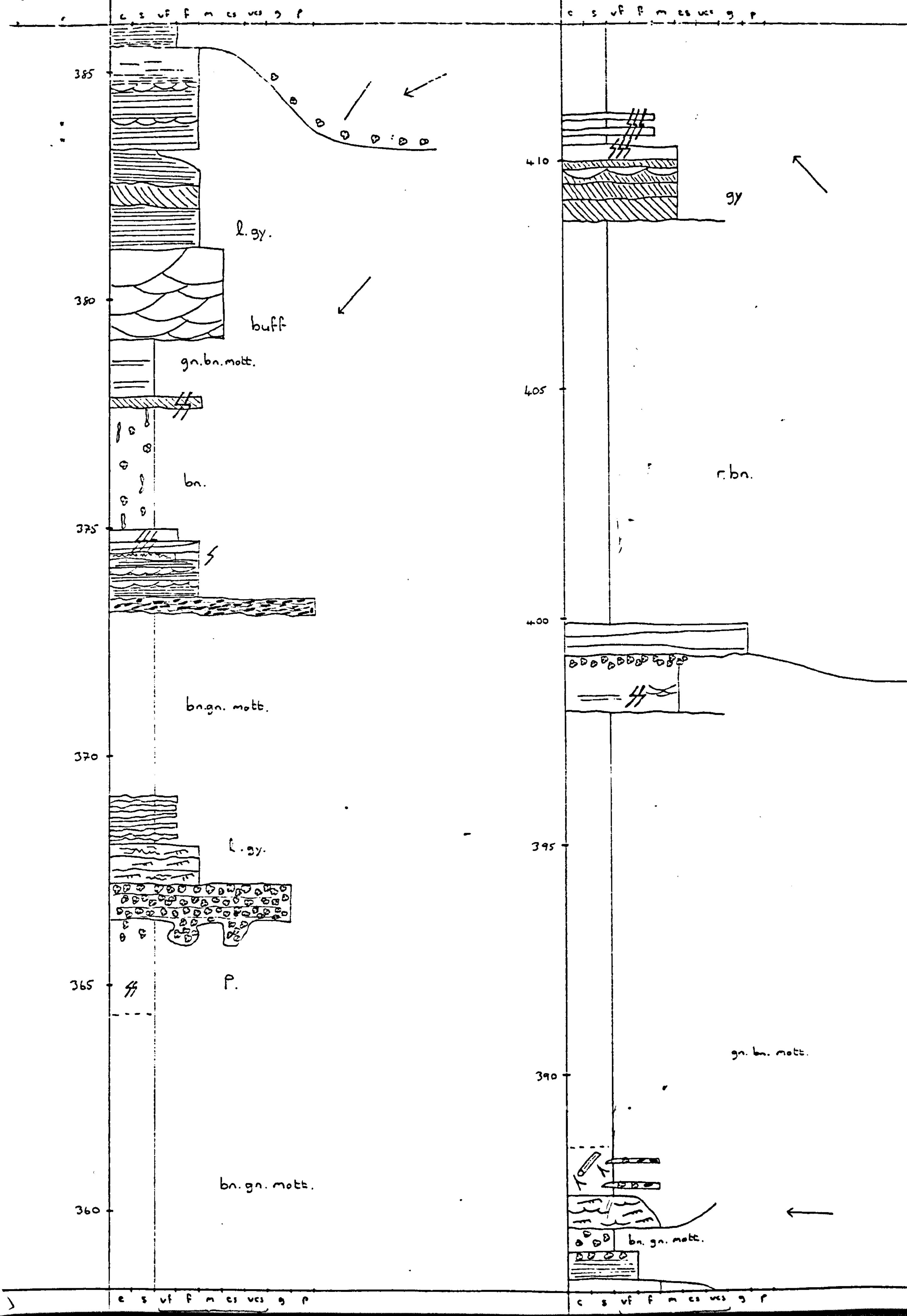


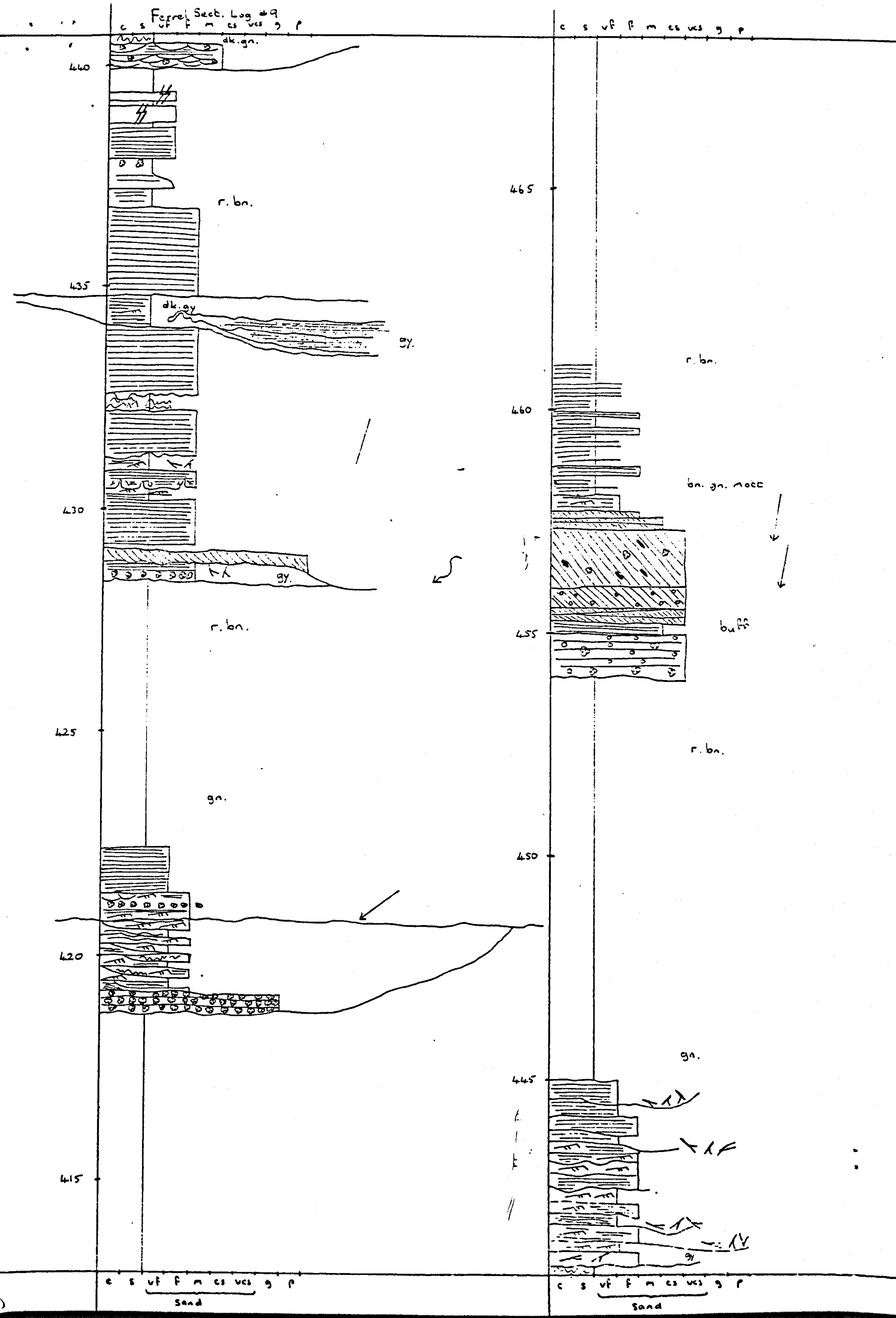
Ferrel Sect. Log # 6.

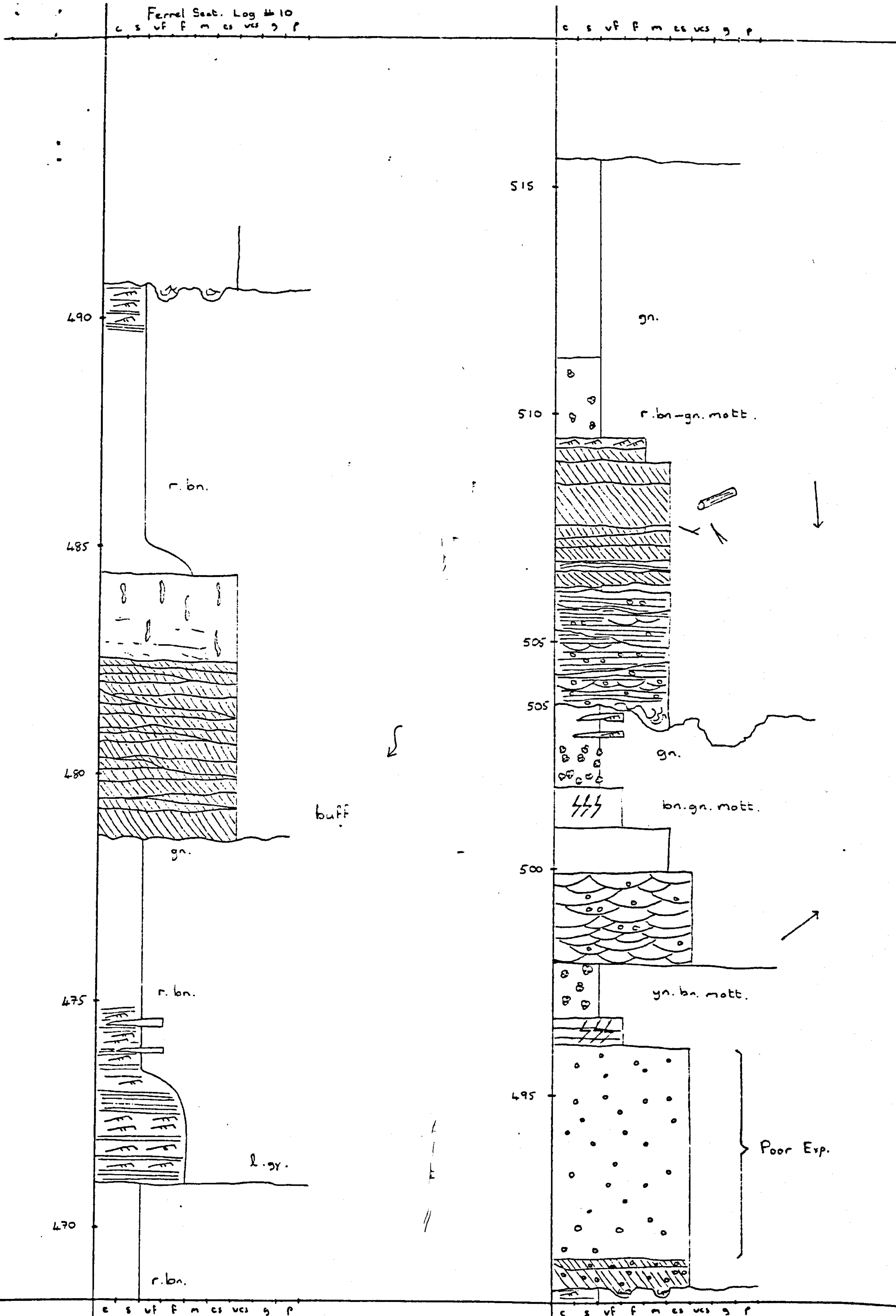


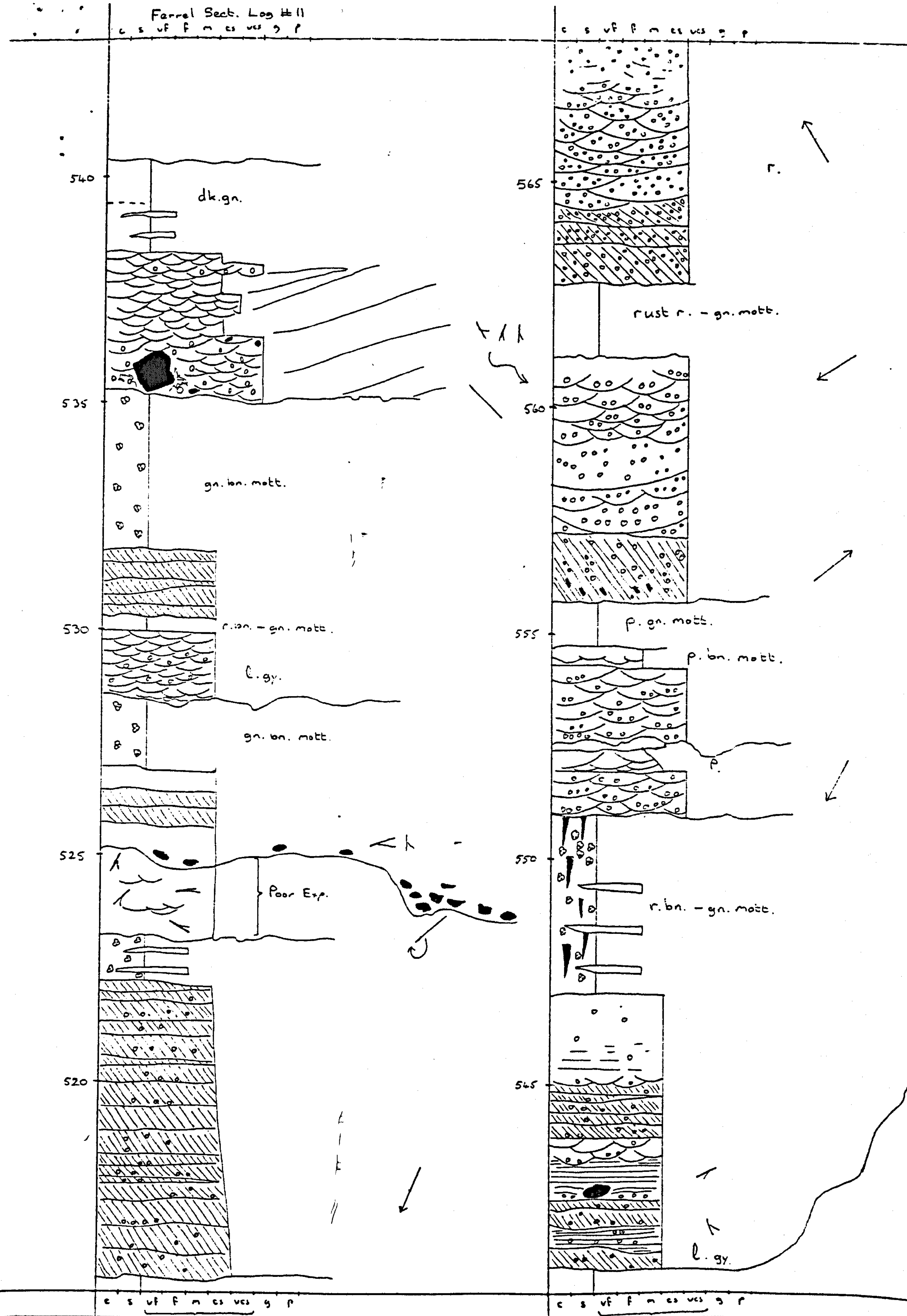

Ferrel Sect. Log # 7
c s vf f m cs vcs g p
330
325
320
315
310
305
bn. gn. mott
r. bn. gn. mott.
bn.
gn. bn. mott.
dk. gy.
gy.
buff
r. bn.
gn. bn. mott.
Sand
355
350
345
340
335
bn. gn. mott.
r. bn/-p. gn. mott.
Sand

Ferrel Sect. Log # 8

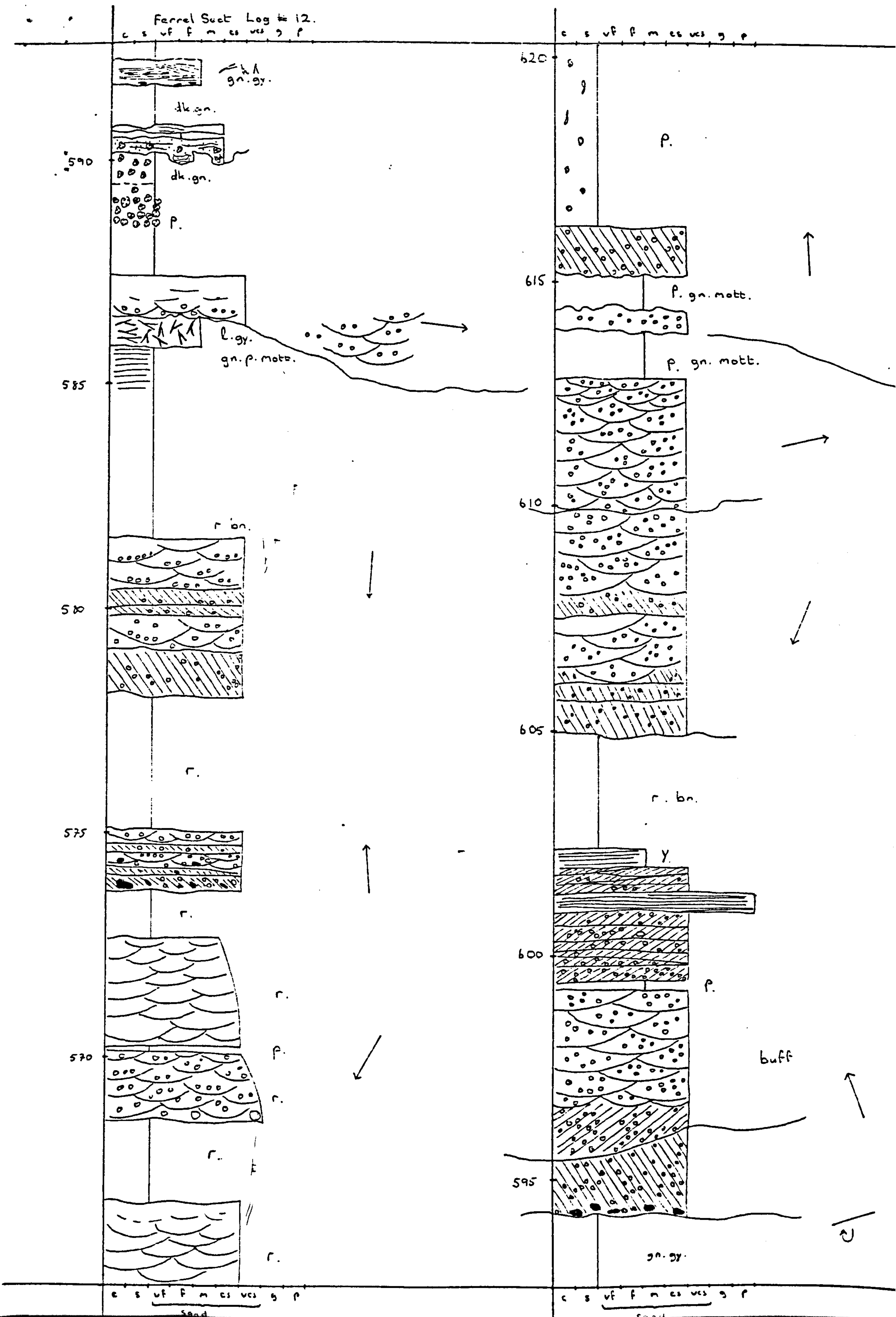



Ferrel Sect. Log #9
c s vf f m cs vcs g p
dk. gn.
440
r. bn.
435
dk. gy
gy.
430
gy.
r. bn.
425
gn.
420
415
c s vf f m cs vcs g p
Sand
c s vf f m cs vcs g p
465
r. bn.
460
bn. gn. mott
455
buff
r. bn.
450
gn.
445
gy
c s vf f m cs vcs g p
Sand

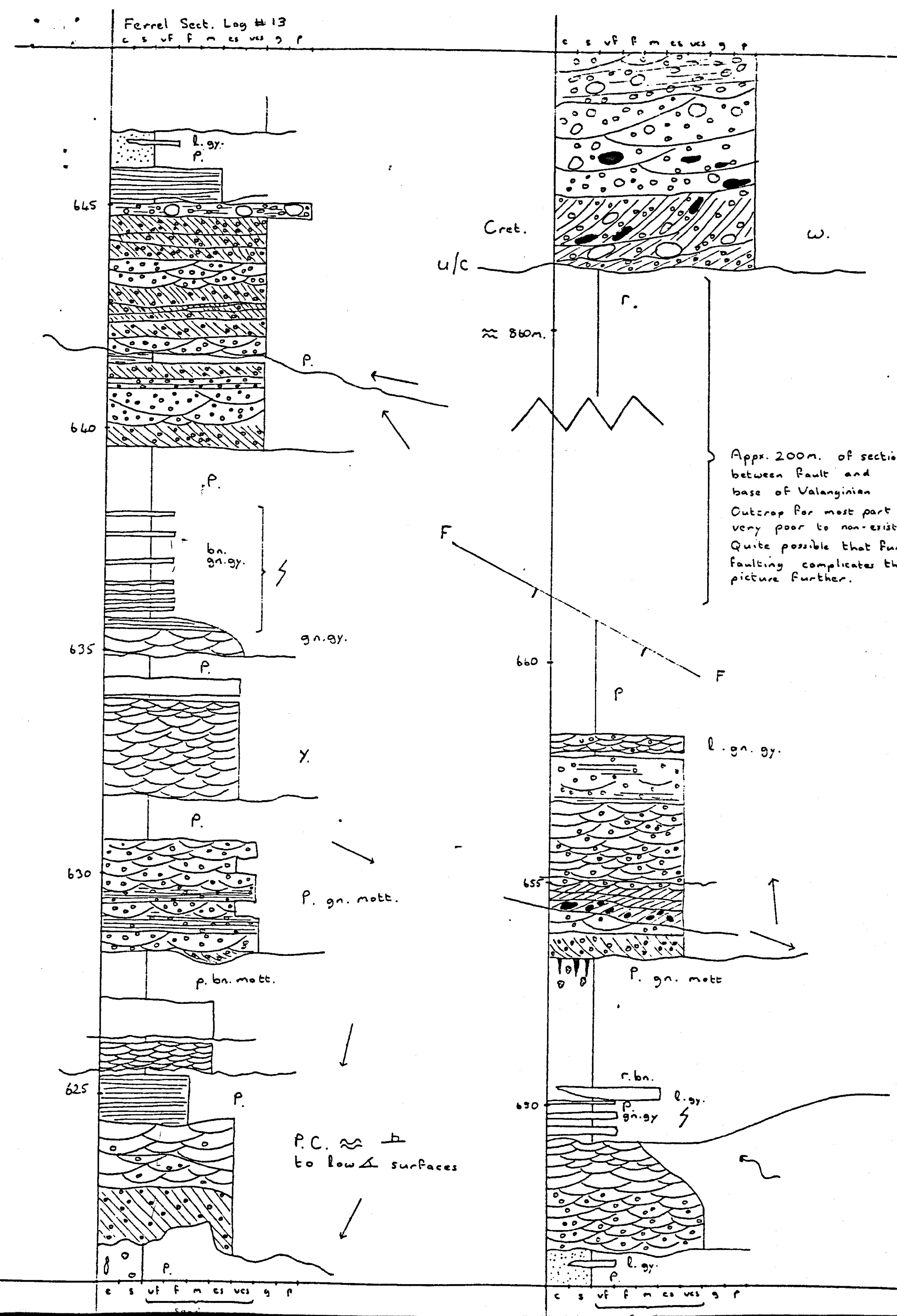

Ferrel Sect. Log # 10
c s vf f m cs vcs g p
490
485
480
475
470
r. bn.
buff
gn.
r. bn.
l. gy.
r. bn.
c s vf f m cs vcs g p
515
510
505
505
500
495
gn.
r. bn-gn. mott.
gn.
bn. gn. mott.
gn. bn. mott.
Poor Exp.
c s vf f m cs vcs g p
c s vf f m cs vcs g p



Farrel Sect. Log #11
c s vf f m cs vcs g p
540
dk.gn.
535
gn. bn. mott.
530
r. bn. - gn. mott.
l. gy.
gn. bn. mott.
525
Poor Exp.
520
565
r.
rust r. - gn. mott.
560
555
p. gn. mott.
p. bn. mott.
p.
550
r. bn. - gn. mott.
545
l. gy.



Ferrel Sect Log # 12.
c s vf f m cs vcs g p
gn. gy.
dk. gn.
590
dk. gn.
p.
l. gy.
gn. p. mott.
585
r bn.
580
r.
575
r.
r.
p.
570
r.
r.
r.
sand
620
p.
615
p. gn. mott.
p. gn. mott.
610
605
r. bn.
y.
600
p.
buff
595
gn. gy.



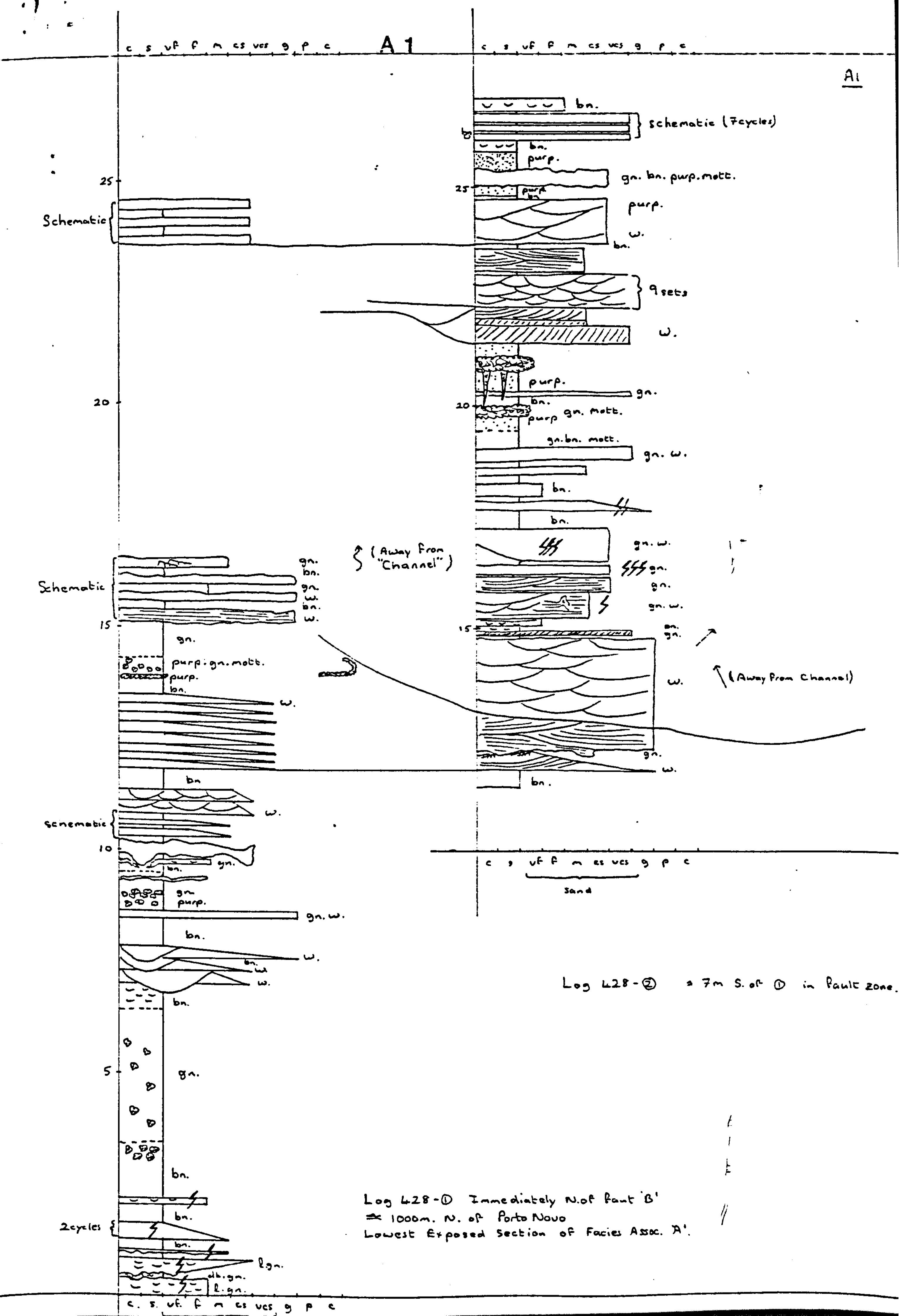
Ferrel Sect. Log # 13
c s vf f m cs vcs g p
l. gy.
P.
645
P.
640
P.
bn.
gn. gy.
gn. gy.
635
P.
Y.
P.
630
P. gn. mott.
p. bn. mott.
625
P.
P. C. ≈ ⊥
to low ∠ surfaces
P.
c s vf f m cs vcs g p
c s vf f m cs vcs g p
Cret.
W.
u/c
r.
≈ 860m.
Appx. 200m. of section between fault and base of Valanginian
Outcrop for most part very poor to non-exist
Quite possible that fu faulting complicates th picture further.
F
F
660
P
l. gn. gy.
655
P. gn. mott
r. bn.
l. gy.
650
P.
gn. gy
l. gy.
P.
c s vf f m cs vcs g p

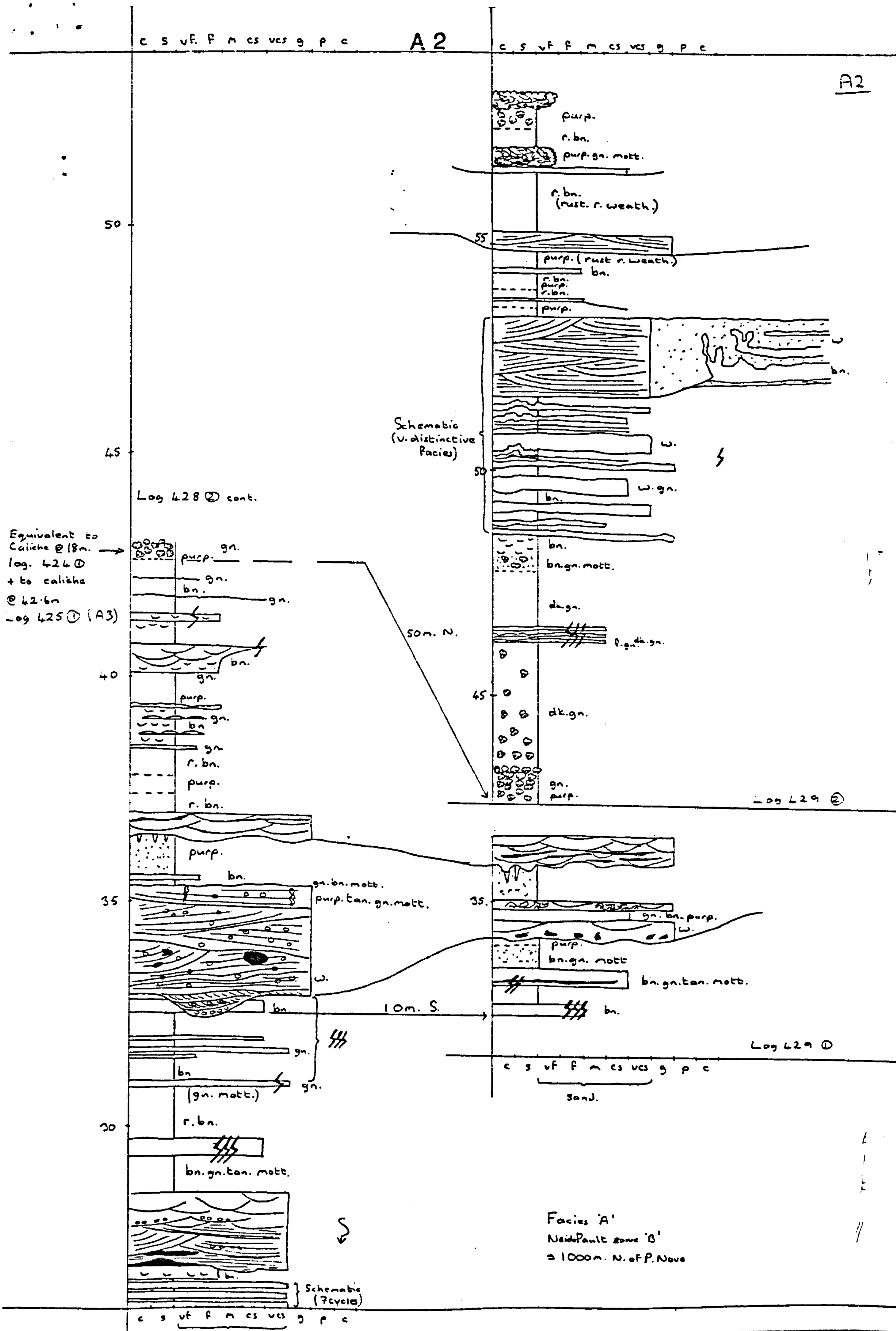
ENCLOSURE 7

LOGS: PORTO NOVO-PORTO DINHEIRO

(P. DA AMOREIRA MEMBER A 1-14,

PORTO NOVO MEMBER B 1-3)

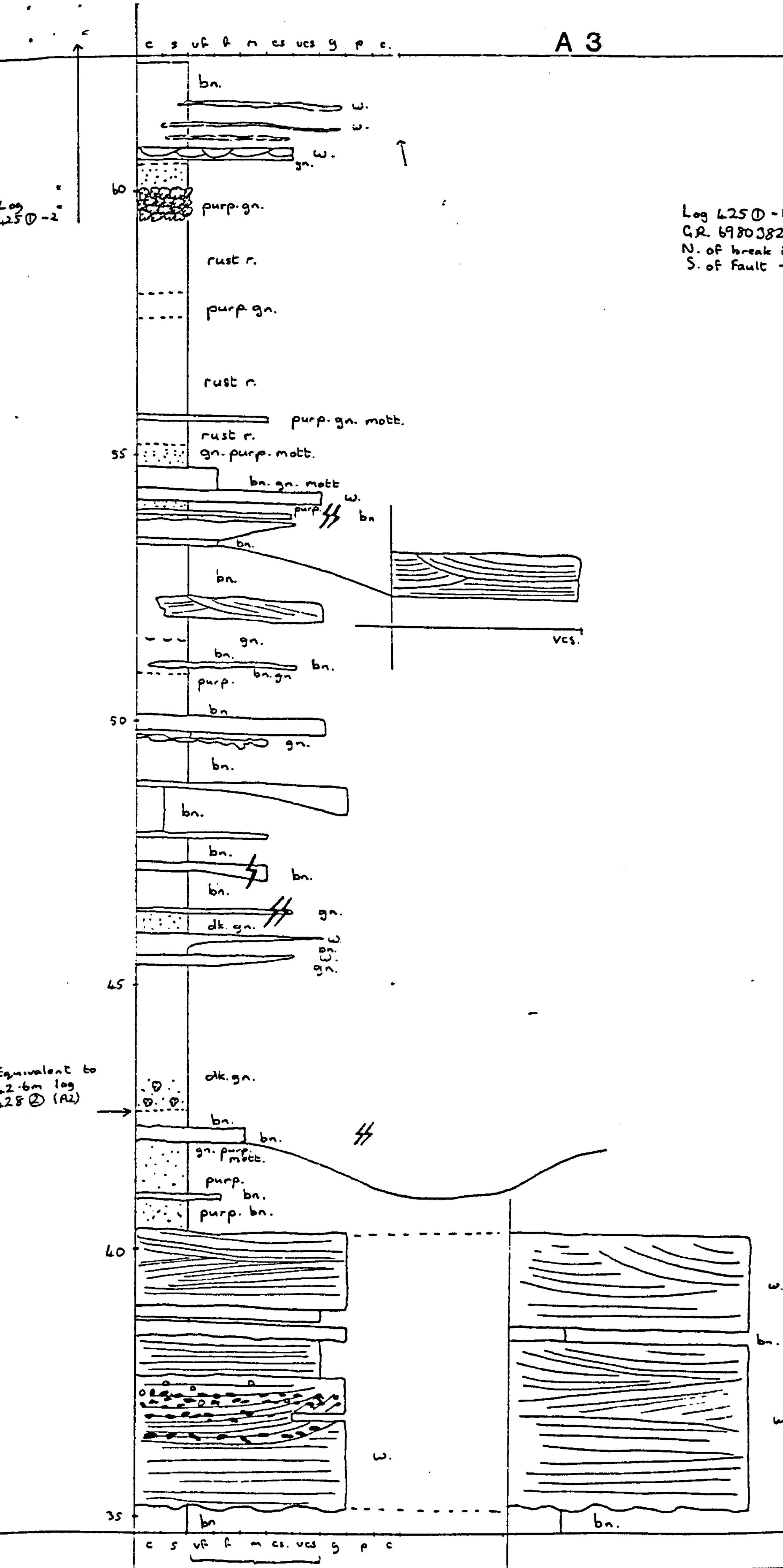

A 1
A1
c s vf f m cs vcs g p c
bn.
schematic (7cycles)
purp.
gn. bn. purp. mott.
purp.
w.
9 sets
w.
purp.
gn.
bn.
purp gn. mott.
gn. bn. mott.
gn. w.
bn.
bn.
gn. w.
gn.
gn.
gn. w.
bn.
gn.
w.
(Away from Channel)
gn.
w.
bn.
c s vf f m cs vcs g p c
Sand
Schematic
(Away from "Channel")
Schematic
gn.
bn.
gn.
w.
bn.
w.
gn.
purp. gn. mott.
purp.
bn.
w.
bn
w.
Schematic
gn.
gn
purp.
gn. w.
bn.
w.
bn.
w.
w.
bn.
gn.
bn.
bn.
2cycles
bn.
lgn.
db. gn.
l. gn.
c. s. vf. f m cs vcs g p c
25
20
15
10
5
Log L28-② ≈ 7m S. of ① in fault zone.
Log L28-① Immediately N. of fault 'B'
≈ 1000m. N. of Porto Novo
Lowest Exposed Section of Facies Assoc. 'A'.


A 2
A2
c s vf f m cs vcs g p c
purp.
r. bn.
purp. gn. mott.
r. bn.
(rust. r. weath.)
55
purp. (rust r. weath.)
bn.
r. bn.
purp.
r. bn.
purp.
w.
bn.
Schematic
(v. distinctive
facies)
w.
50
w. gn.
bn.
bn.
bn. gn. mott.
dk. gn.
f. gn. dk. gn.
45
dk. gn.
gn.
purp.
Log 429 ②
50m. N.
50
45
Log 428 ② cont.
Equivalent to
Caliche @ 18m.
log. 424 ①
+ to caliche
@ 42.6m
Log 425 ① (A3)
purp.
gn.
gn.
bn.
gn.
bn.
gn.
40
purp.
gn.
bn
gn
r. bn.
purp.
r. bn.
purp.
bn.
gn. bn. mott.
purp. tan. gn. mott.
35
w.
35
gn. bn. purp.
w.
purp.
bn. gn. mott
bn. gn. tan. mott.
bn
10m. S.
bn.
gn.
Log 429 ①
bn
gn.
(gn. mott.)
sand.
30
r. bn.
bn. gn. tan. mott.
bn.
Schematic
(7 cycles)
Facies 'A'
NeidPault zone 'B'
≥ 1000m. N. of P. Nova

A 3

A3

Log 425 ① -1 Facies Assoc. 'A'
G.R. 69803825
N. of break in cliffs W. of Outeiro do Seixo
S. of Fault - throw ≈ 10m.



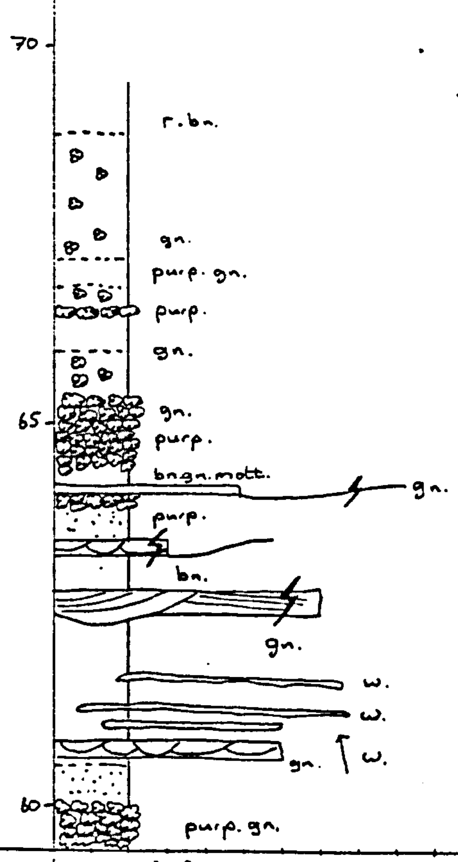
A4

A4

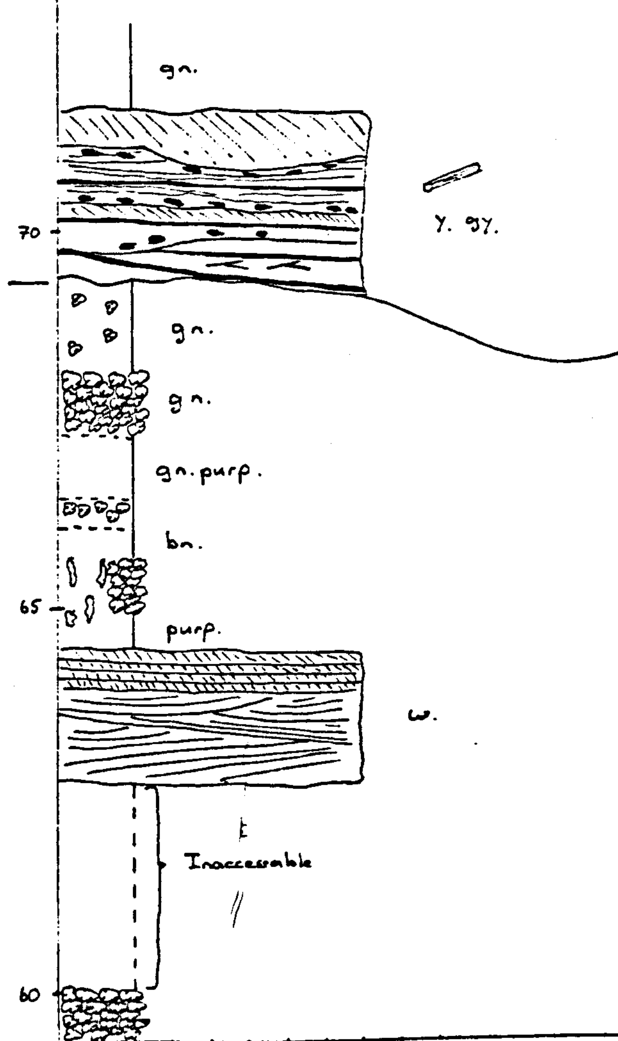
c s vf f m cs vcs g p c

Log 425 ①-2
GR 6980 3825 Near Cliff Top.
Continuation of Log 425①-1

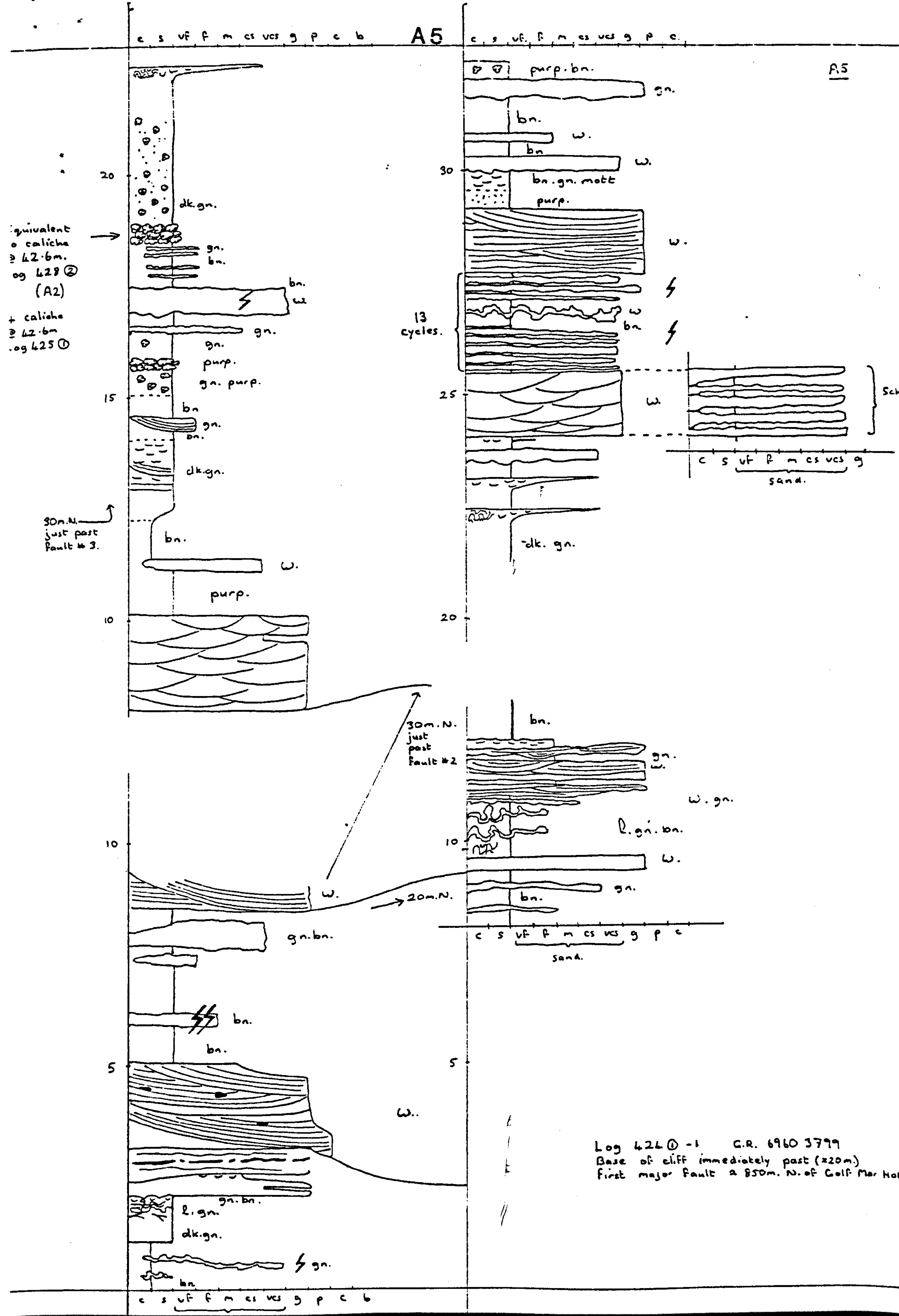
Log 425 ②
G.R. 6995 3870 Top of cliff
Overlap with Log 425 ①-2

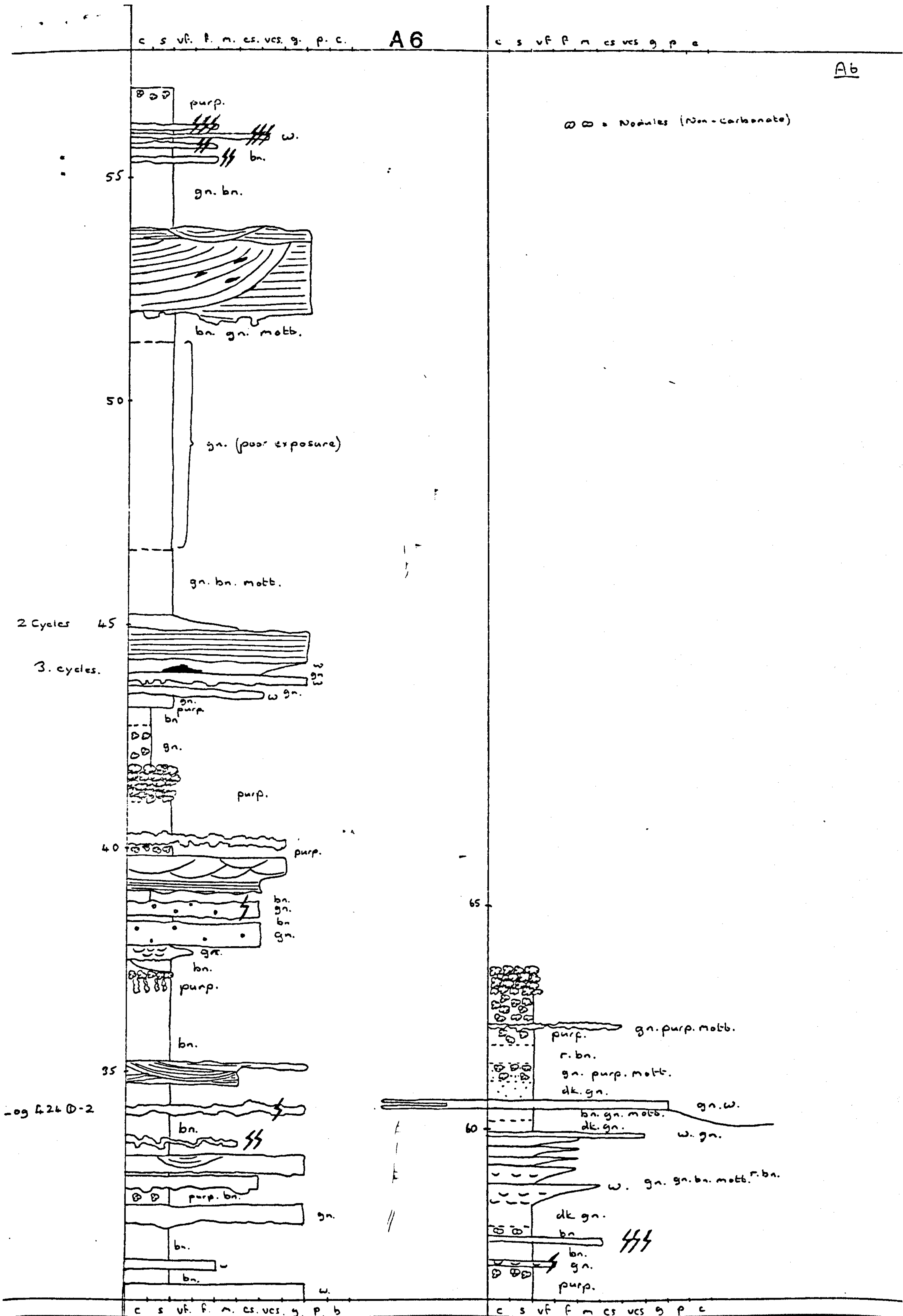


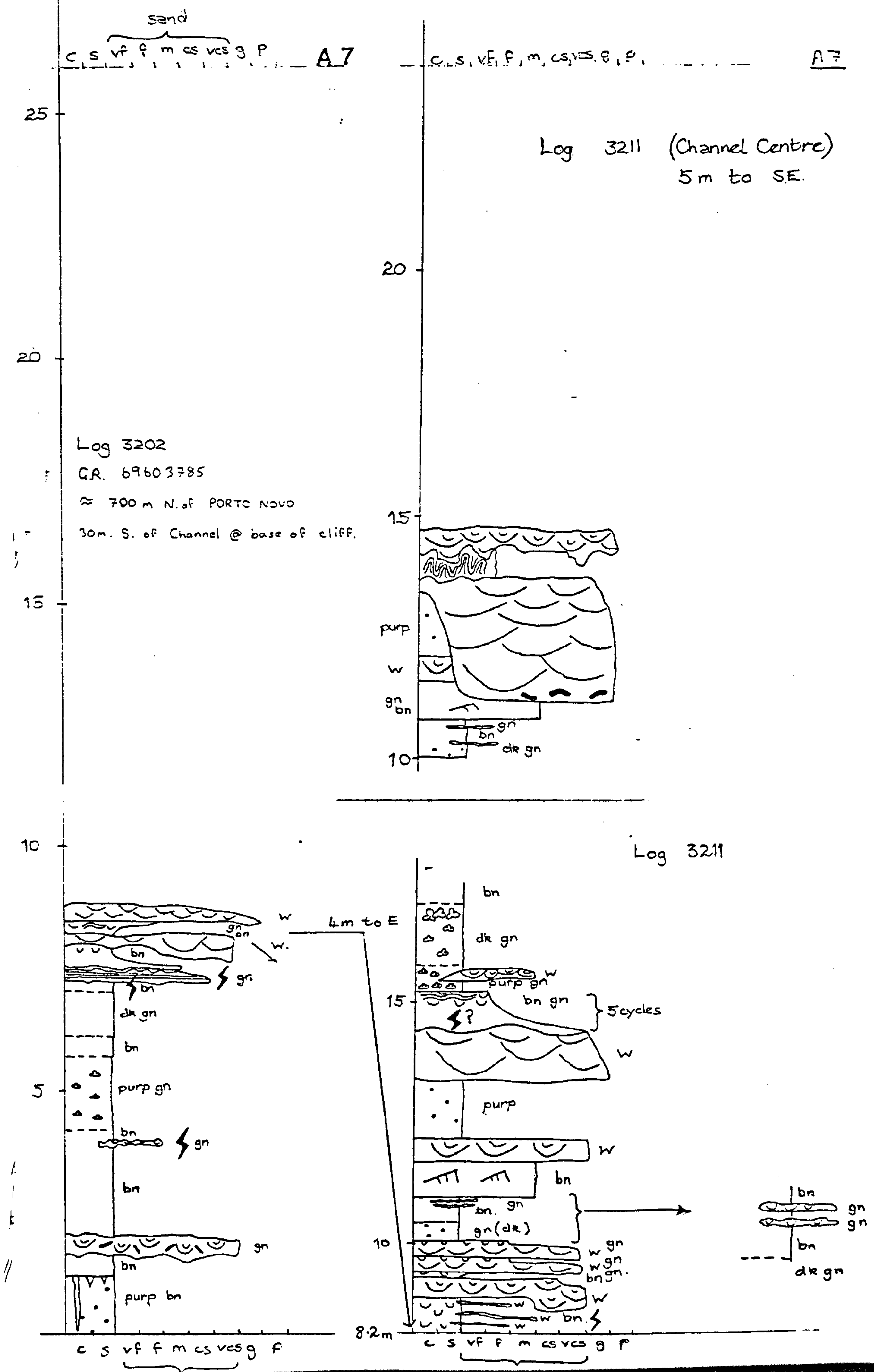
'B'
'A'



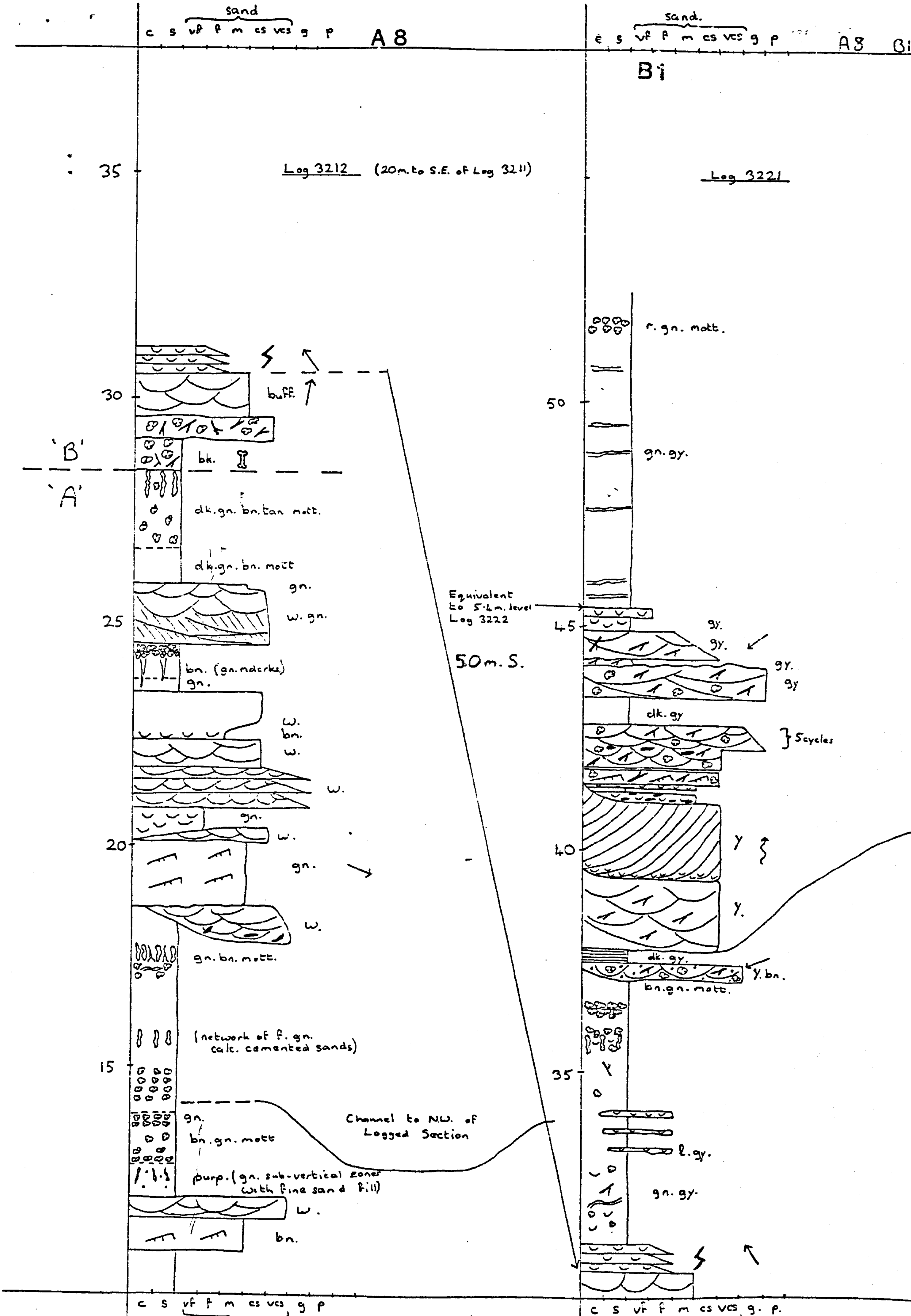
c s vf f m cs vcs g p c


A5
c s vf f m cs vcs g p c b
c s vf f m cs vcs g p c
A5
purp. bn.
gn.
bn.
w.
bn
w.
30
bn. gn. mott
purp.
w.
13 cycles.
w
bn.
25
w.
Sch
c s vf f m cs vcs g
sand.
dk. gn.
20
20
dk. gn.
Equivalent to caliche @ 42.6m. Log 428 ② (A2)
gn.
bn.
bn.
w.
+ caliche @ 42.6m Log 425 ①
gn.
gn.
purp.
gn. purp.
15
bn
gn.
bn.
dk. gn.
30m. N. just past fault # 3.
bn.
w.
purp.
10
30m. N. just past fault # 2
bn.
gn.
w.
w. gn.
l. gn. bn.
10
10
w.
w.
gn.
20m. N.
bn.
c s vf f m cs vcs g p c
sand.
gn. bn.
bn.
bn.
5
5
w..
Log 424 ① - 1 G.R. 6960 3799
Base of cliff immediately past (≈20m)
First major fault a 850m. N. of Golf Mar Hol
gn. bn.
l. gn.
dk. gn.
gn.
bn.
c s vf f m cs vcs g p c b


A6
Ab
c s vf. f. m. cs. vcs. g. p. c.
ꝏ = Nodules (Non-carbonate)
purp.
w.
bn.
55
gn. bn.
bn. gn. mott.
50
gn. (poor exposure)
gn. bn. mott.
2 Cycles
45
3. cycles.
w gn w
w gn.
gn. purp
bn
gn.
purp.
40
purp.
bn. gn. bn gn.
gn.
bn.
purp.
bn.
35
-og 424 D-2
bn.
purp. bn.
gn.
bn.
bn.
w.
65
gn. purp. mott.
purp.
r. bn.
gn. purp. mott.
dk. gn.
gn. w.
bn. gn. mott.
60
dk. gn.
w. gn.
w. gn. gn. bn. mott. r. bn.
dk gn.
bn
bn.
gn.
purp.

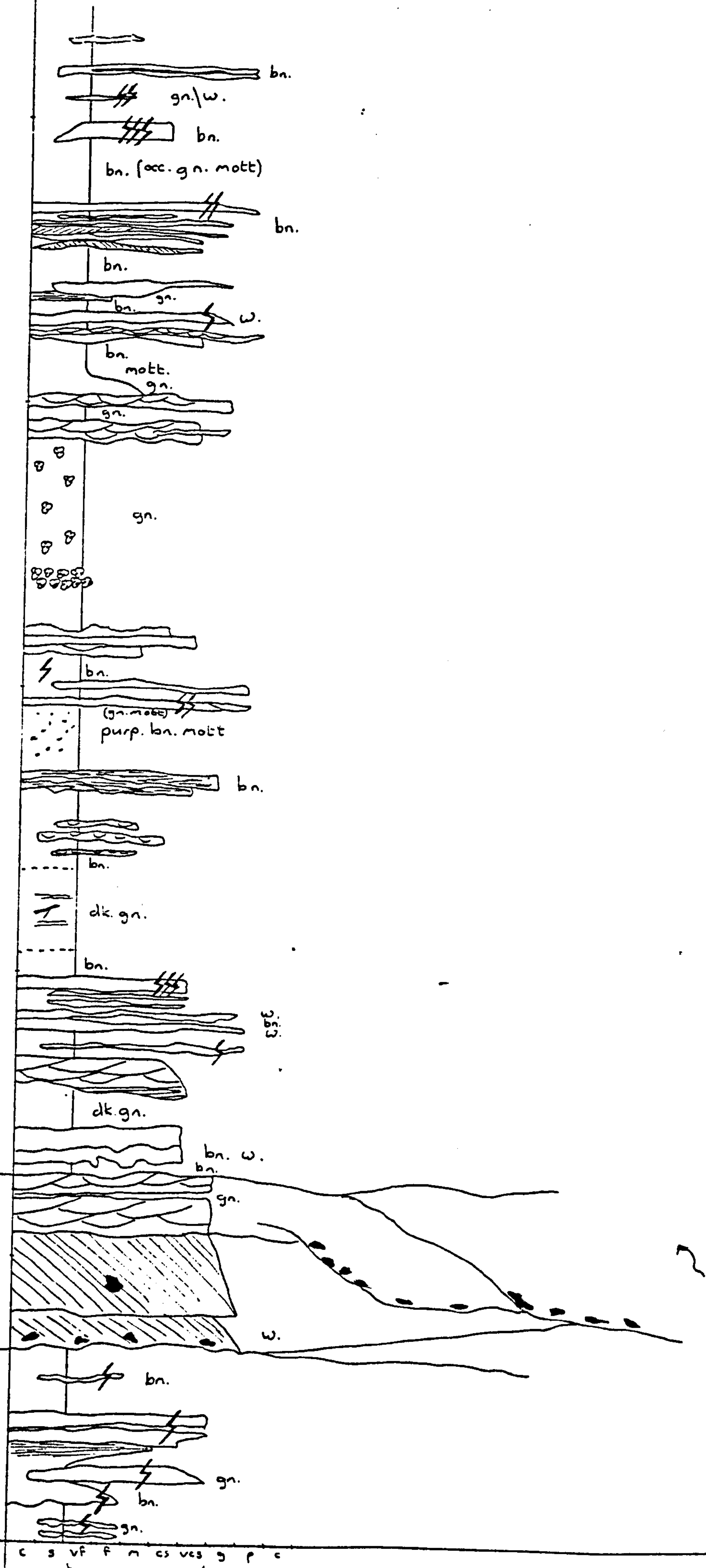


sand
c s vf f m cs vcs g p
A7
25
20
Log 3202
G.R. 69603785
≈ 700 m N. of PORTO NOVO
30m. S. of Channel @ base of cliff.
15
c s vf f m cs vcs g p
A7
Log 3211 (Channel Centre)
5 m to S.E.
20
15
purp
w
gn bn
gn
bn
dk gn
10
10
w
gn bn
w.
bn
gn
bn
dk gn
bn
5
purp gn
bn
gn
bn
gn
bn
purp bn
c s vf f m cs vcs g p
4m to E
Log 3211
bn
dk gn
w
purp gn
15
bn gn
5 cycles
w
purp
w
bn
gn
bn.
gn(dk)
10
w gn
w gn.
bn gn.
w
w
w bn.
w
8·2m
c s vf f m cs vcs g p
bn
gn
gn
bn
dk gn



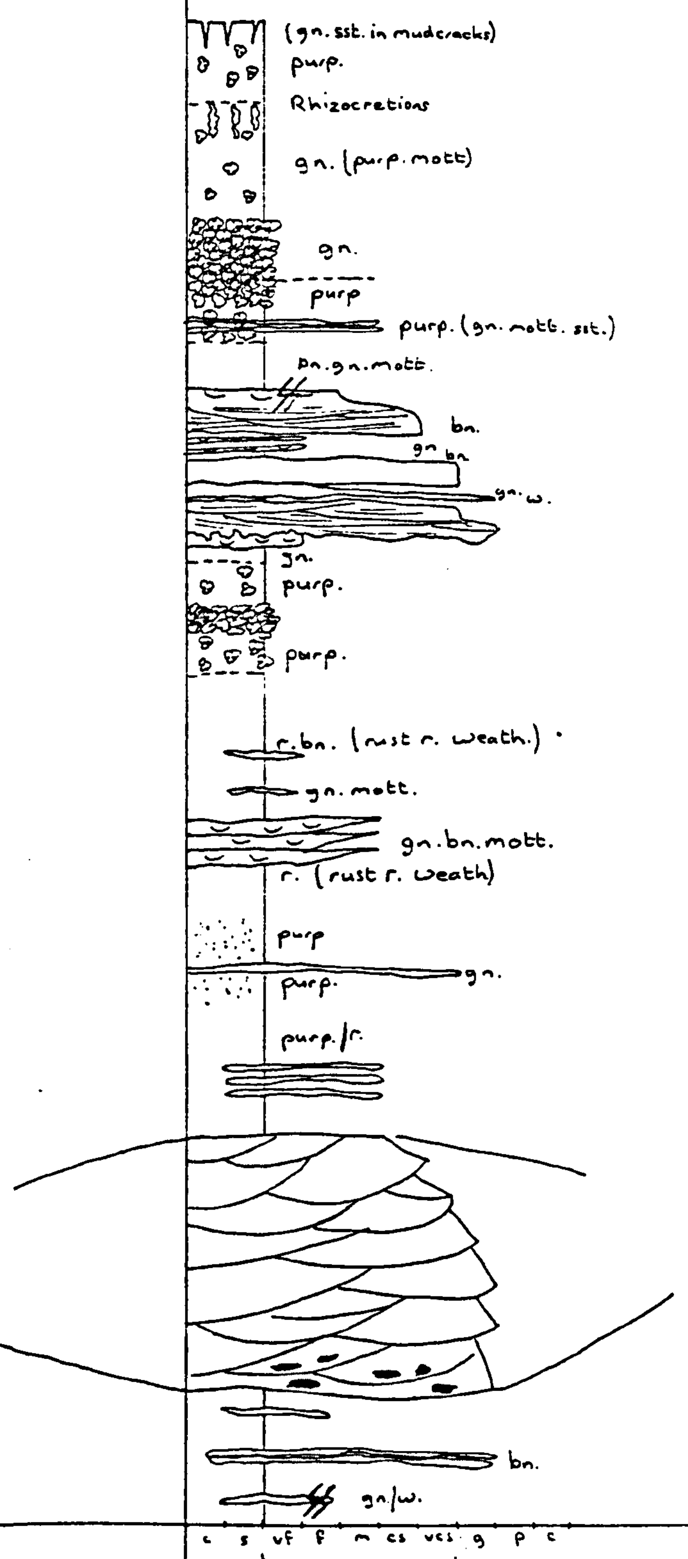
sand
c s vf f m cs vcs g p
A8
A8 Bi
Bi
Log 3212 (20m. to S.E. of Log 3211)
Log 3221
35
30
25
20
15
50
45
40
35
'B'
'A'
buff.
bk.
dk.gn. bn. tan mott.
dk.gn. bn. mott
gn.
w. gn.
bn. (gn. nodules)
gn.
w.
bn.
w.
w.
gn.
w.
gn.
w.
gn. bn. mott.
(network of f. gn. calc. cemented sands)
gn.
bn. gn. mott
purp. (gn. sub-vertical zones with fine sand fill)
w.
bn.
Channel to N.W. of Logged Section
Equivalent to 5.4m. level Log 3222
50m. S.
r. gn. mott.
gn. gy.
gy.
gy.
gy.
gy
dk. gy
5 cycles
y
y.
dk. gy.
y. bn.
bn. gn. mott.
l. gy.
gn. gy.

A9



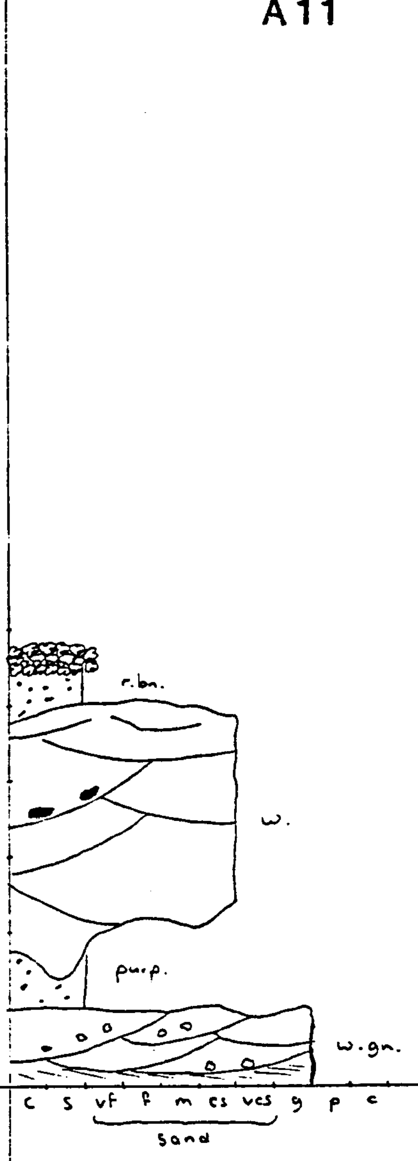
Log 1119 ① - 1
N. of fault 'A'
≈ 850m. N. of Porto Novo

Log 11 19 ① - 2


(gn. sst. in mudcracks)
purp.
Rhizocretions
gn. (purp. mott)
gn.
purp
purp. (gn. mott. sst.)
bn. gn. mott.
bn.
gn bn
gn. w.
gn.
purp.
purp.
r. bn. (rust r. weath.)
gn. mott.
gn. bn. mott.
r. (rust r. weath)
purp
gn.
purp.
purp./r.
bn.
gn./w.
c s vf f m cs vcs g p c

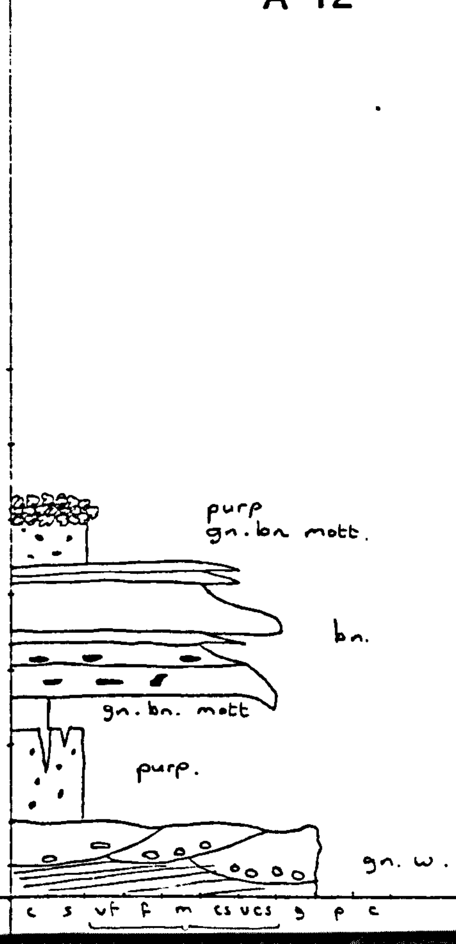
A11

Log 11182
700m. N. of Porto Novo

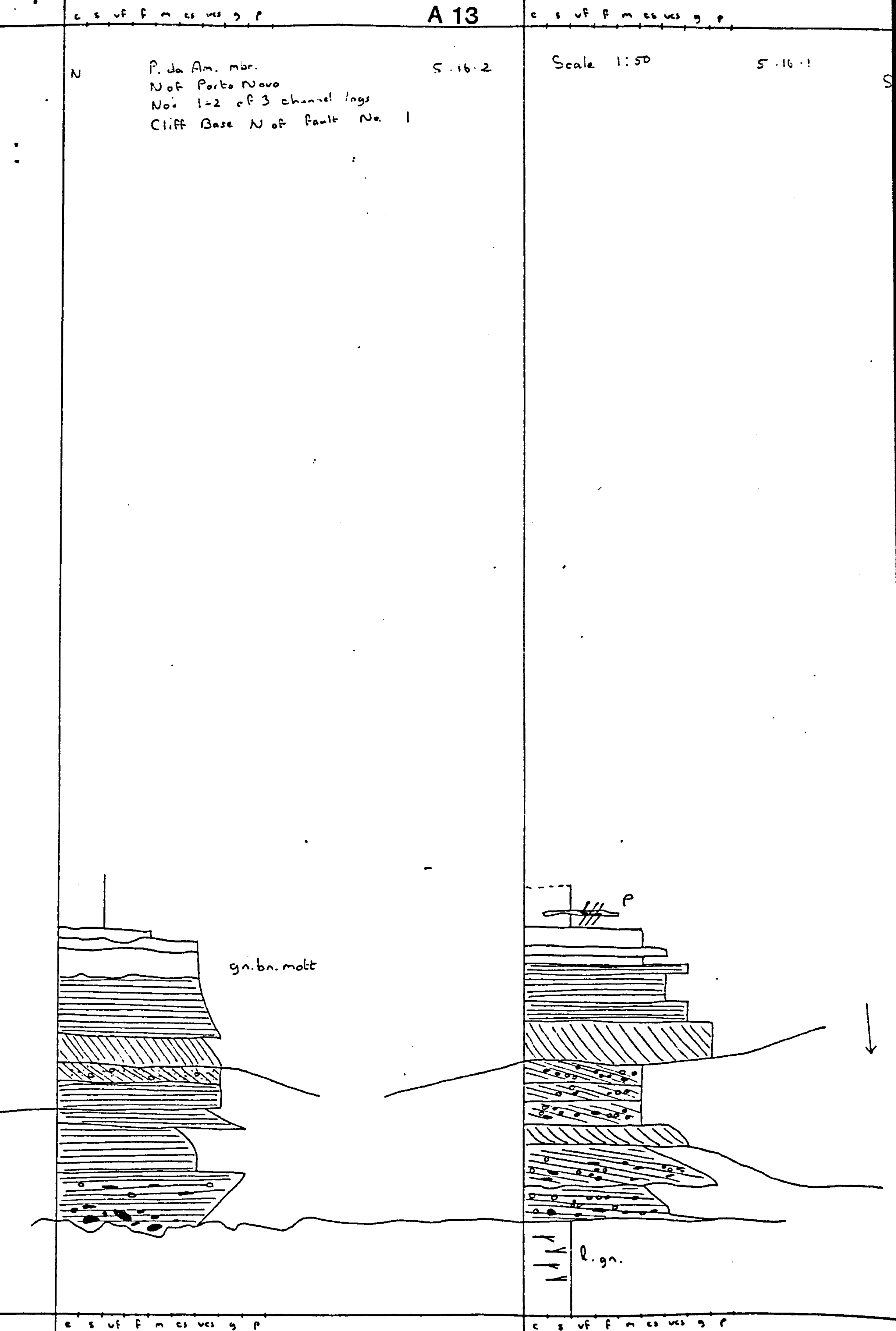


Log 11181
700m. N. of Porto Novo

A 12



A 13



c s vf f m cs vcs g p
N
P. da Am. mbr.
N of Porto Novo
No's 1+2 of 3 channel logs
Cliff Base N of fault No. 1
5.16.2
gn. bn. mott
c s vf f m cs vcs g p
c s vf f m cs vcs g p
Scale 1:50
5.16.1
S
P
l. gn.
c s vf f m cs vcs g p

c s vf f m cs vcs g p

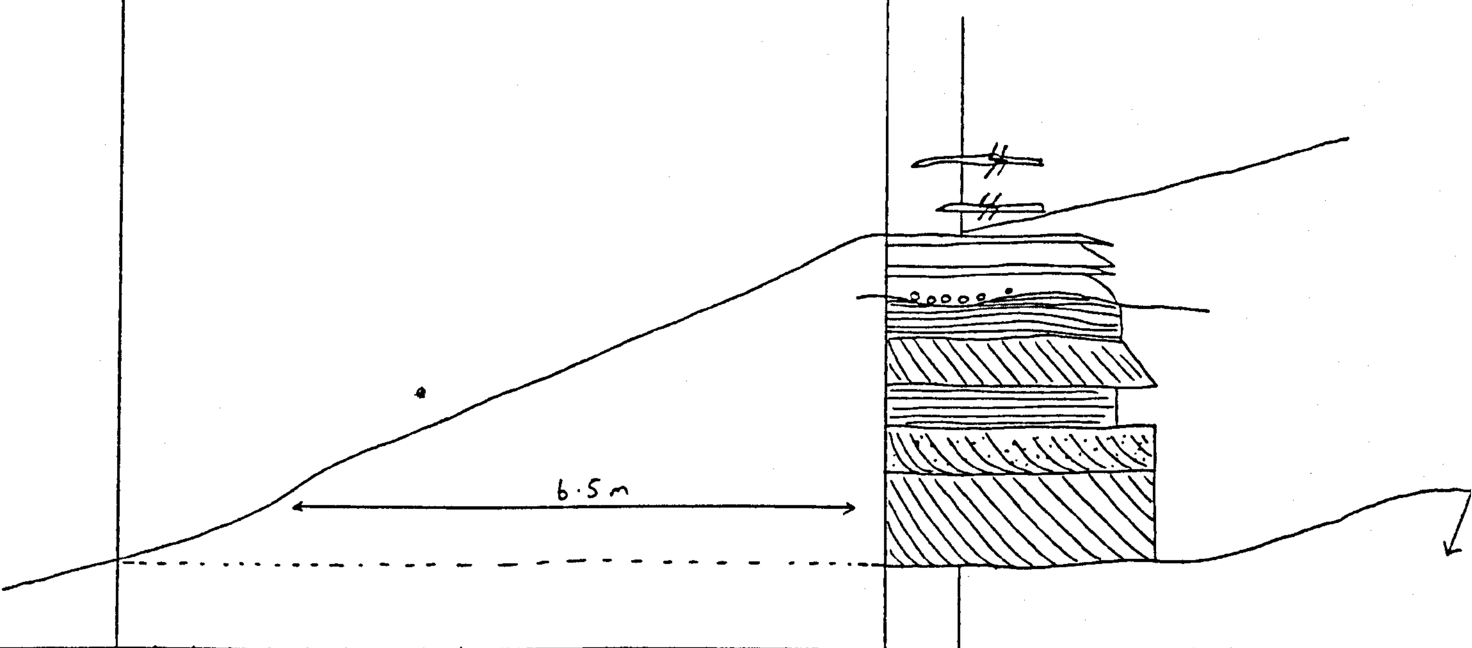
A 14

c s vf f m cs vcs g p

P. da Amoreira member
N. of Porto Novo
No. 3 of 3 channel logs

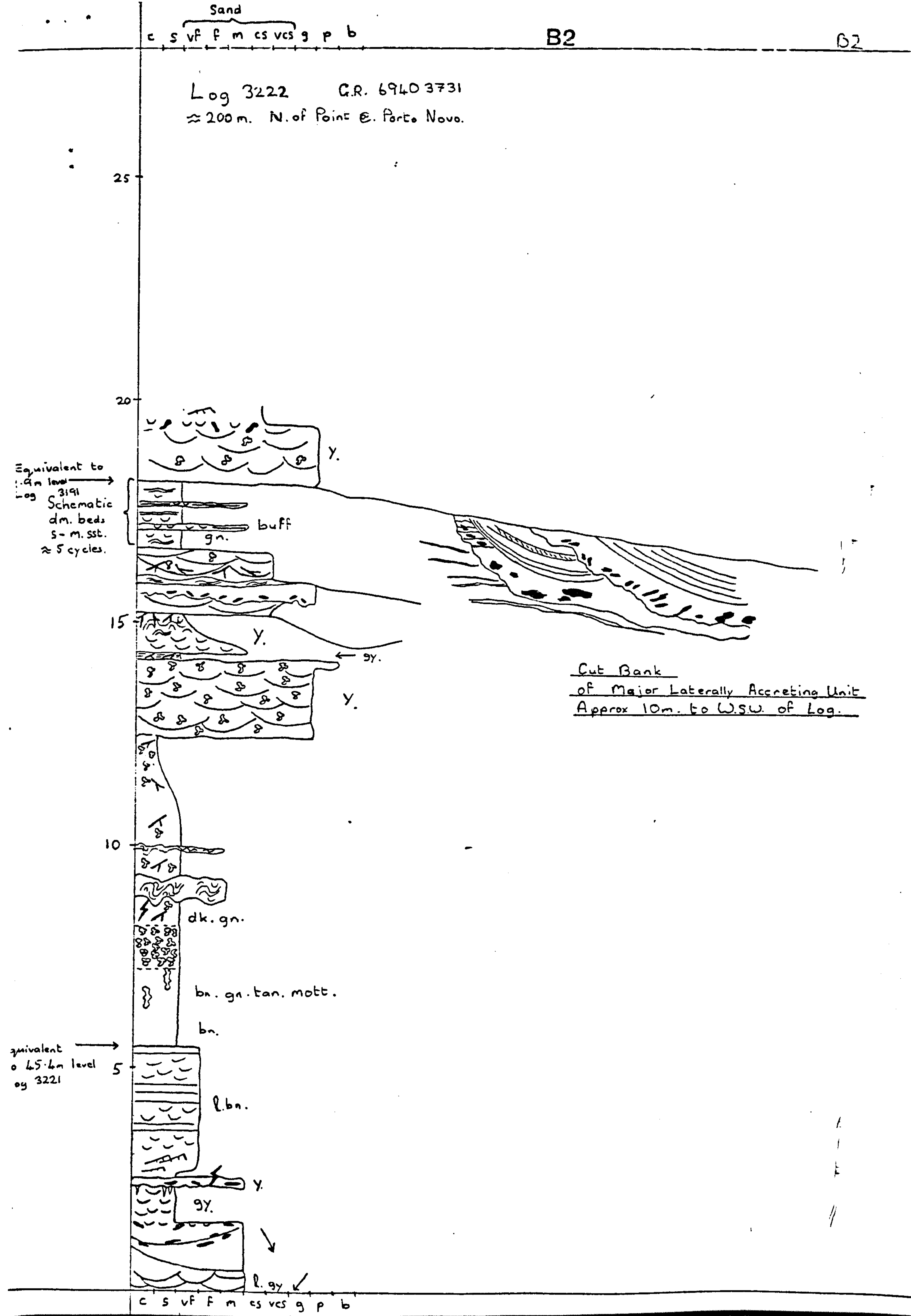
Scale 1:50

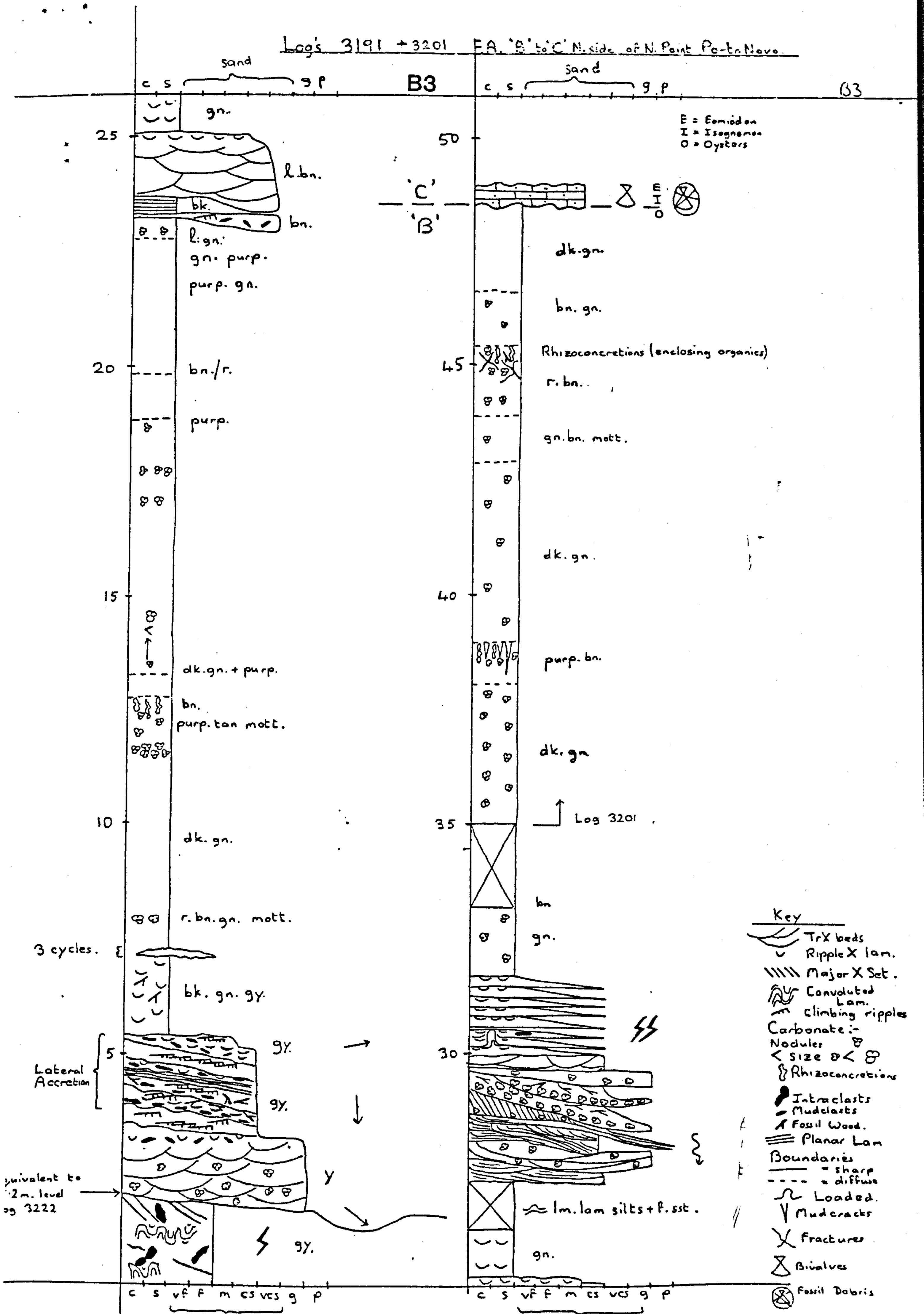
5·16·3



c s vf f m cs vcs g p

c s vf f m cs vcs g p


Sand
c s vf f m cs vcs g p b
Log 3222 G.R. 6940 3731
≈ 200 m. N. of Point E. Porto Novo.
25
20
15
10
5
Equivalent to
1.9m level
Log 3191
Schematic
dm. beds
s-m. sst.
≈ 5 cycles.
y.
buff
gn.
y.
gy.
y.
dk.gn.
bn. gn. tan. mott.
bn.
quivalent
o 45.4m level
og 3221
l.bn.
y.
gy.
l. gy
Cut Bank
of Major Laterally Accreting Unit
Approx 10m. to W.S.W. of Log.
c s vf f m cs vcs g p b

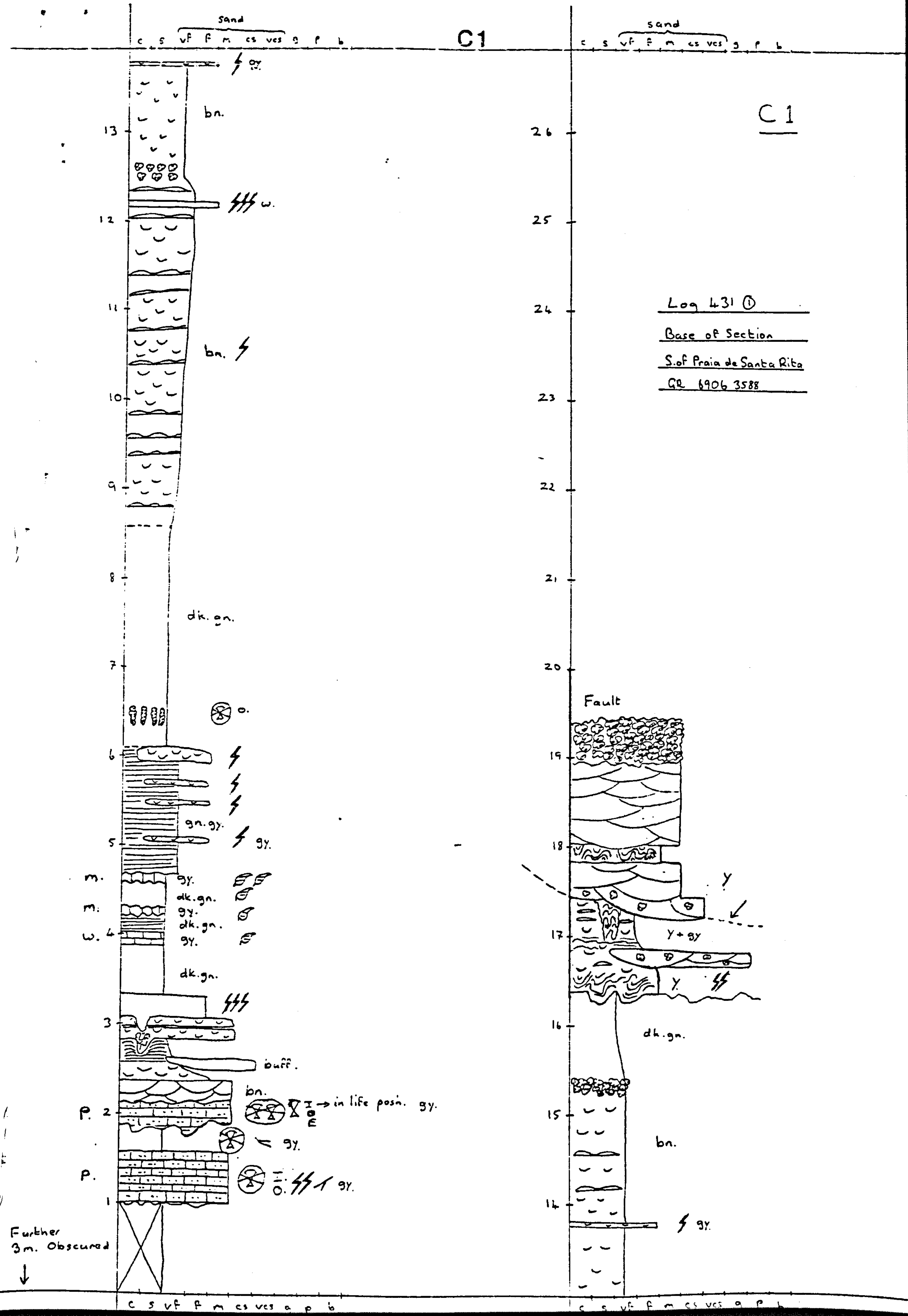


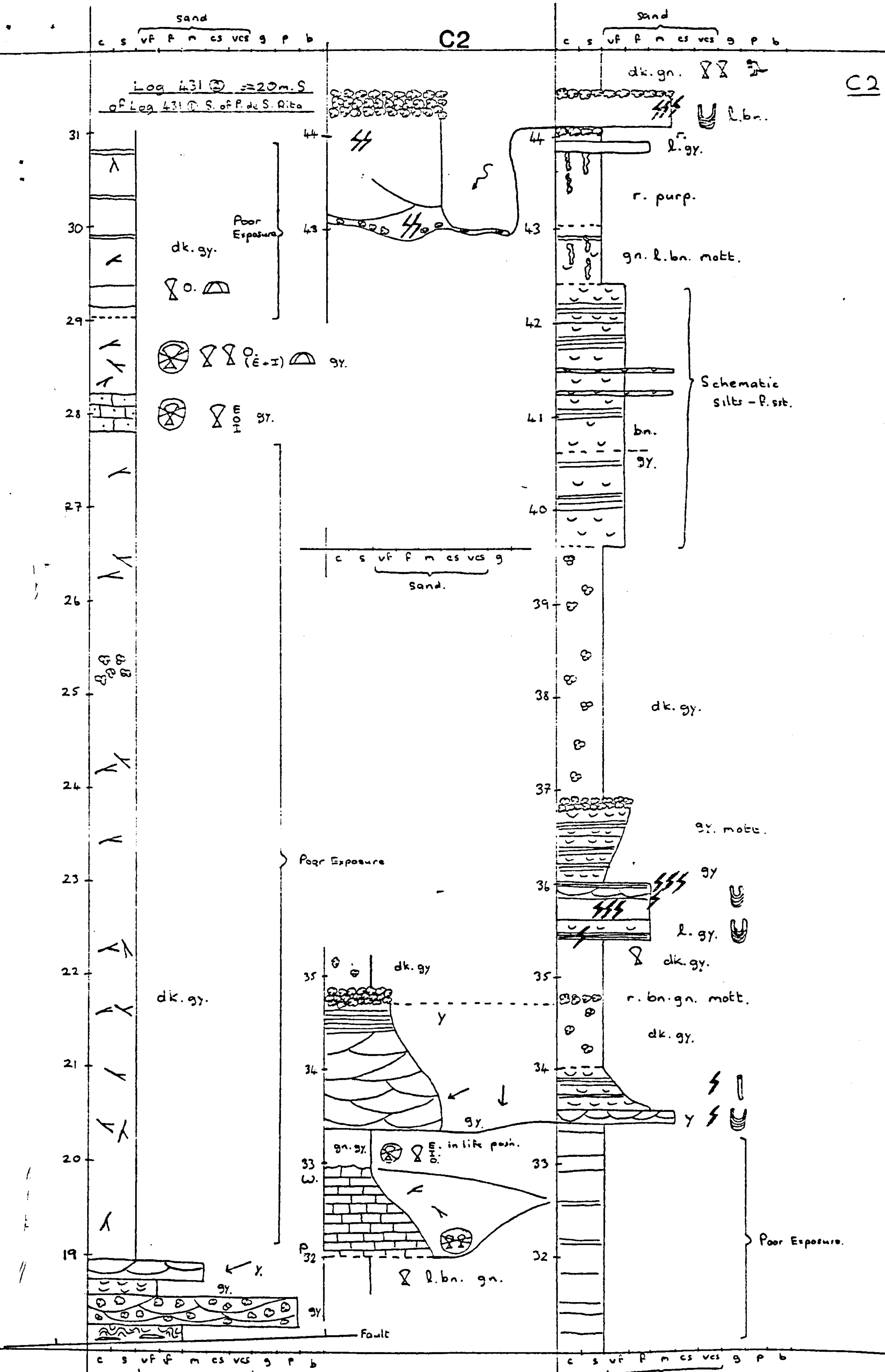
Log's 3191 + 3201 F.A. 'B' to 'C' N. side of N. Point Po-to-Novo
B3
sand
c s g p
gn.
25
l.bn.
bk.
bn.
l.gn.
gn. purp.
purp. gn.
20
bn./r.
purp.
15
dk.gn. + purp.
bn.
purp. tan mott.
10
dk. gn.
r. bn. gn. mott.
3 cycles.
bk. gn. gy.
5
Lateral Accretion
gy.
gy.
y
gy.
quivalent to ·2m. level og 3222
c s vf f m cs vcs g p
'C'
'B'
B3
E = Eomiodon
I = Isognomon
O = Oysters
50
E
I
O
dk.gn.
bn. gn.
45
Rhizoconcretions (enclosing organics)
r. bn.
gn. bn. mott.
dk. gn.
40
purp. bn.
dk. gn
35
Log 3201
bn
gn.
30
lm. lam silts + f. sst.
gn.
c s vf f m cs vcs g p
Key
TrX beds
Ripple X lam.
Major X Set.
Convoluted Lam.
Climbing ripples
Carbonate :-
Nodules
< size
Rhizoconcretions
Intraclasts
Mudclasts
Fossil Wood.
Planar Lam
Boundaries
- sharp
- diffuse
Loaded.
Mudcracks
Fractures
Bivalves
Fossil Debris

ENCLOSURE 8

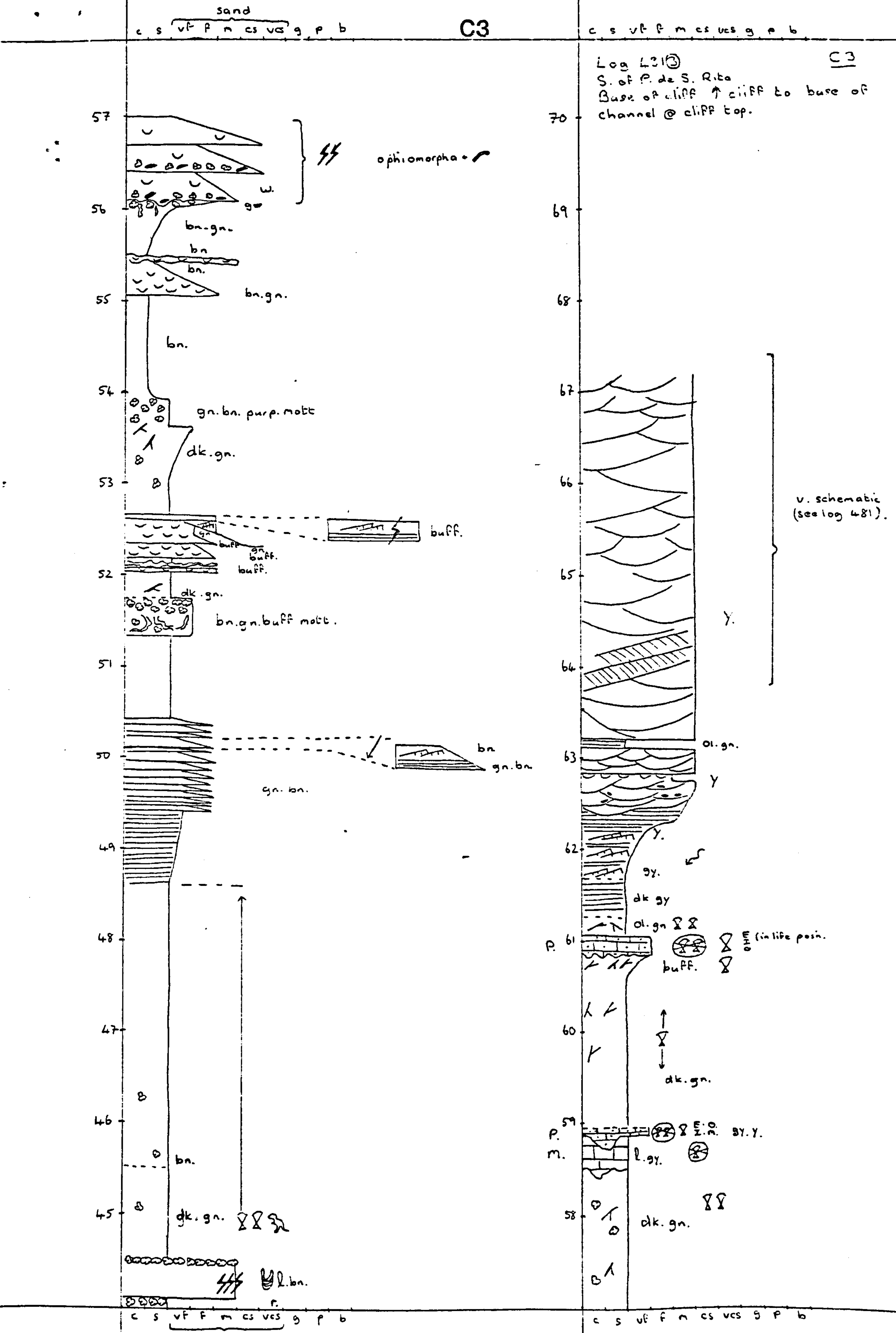
LOGS: PRAIA DA SANTA RITA

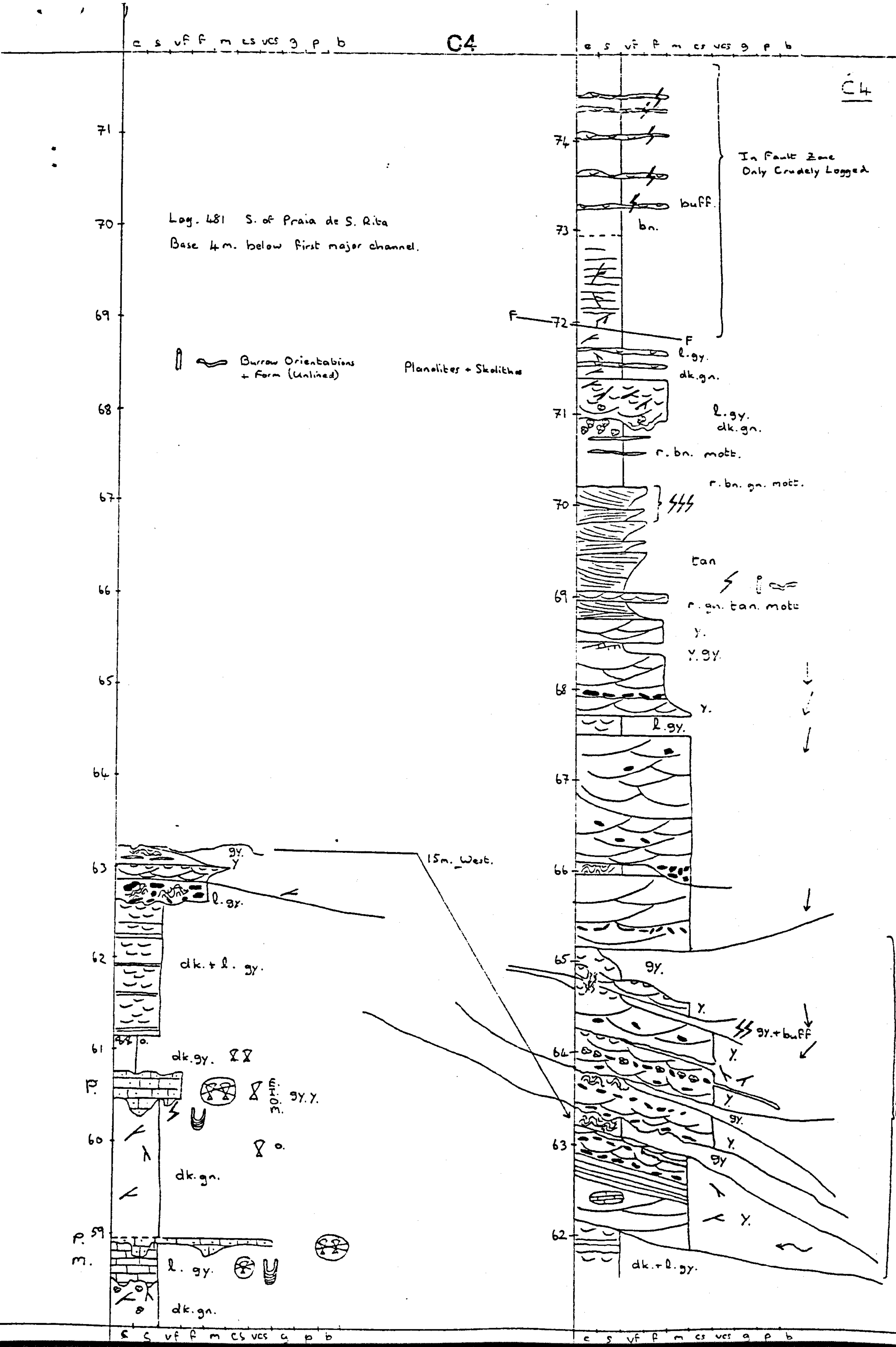
(C 1-7)

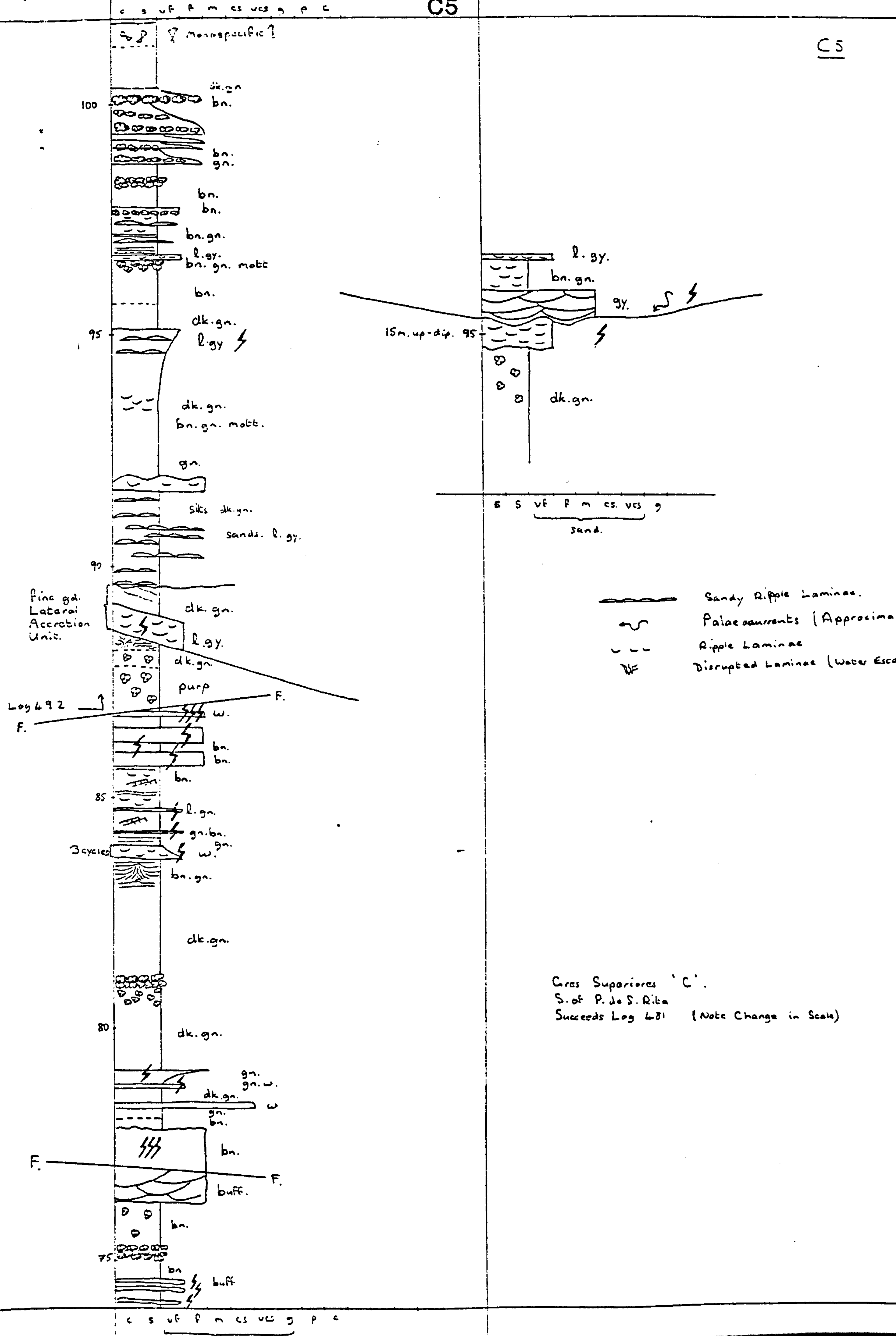

C1
sand
c s vf f m cs vcs g p b
C 1
Log 431 ①
Base of Section
S. of Praia de Santa Rita
GR 6906 3588
bn.
w.
dk. gn.
o.
gn. gy.
gy.
dk. gn.
buff.
bn.
in life posn.
Further 3m. Obscured
Fault
Y
Y + gy

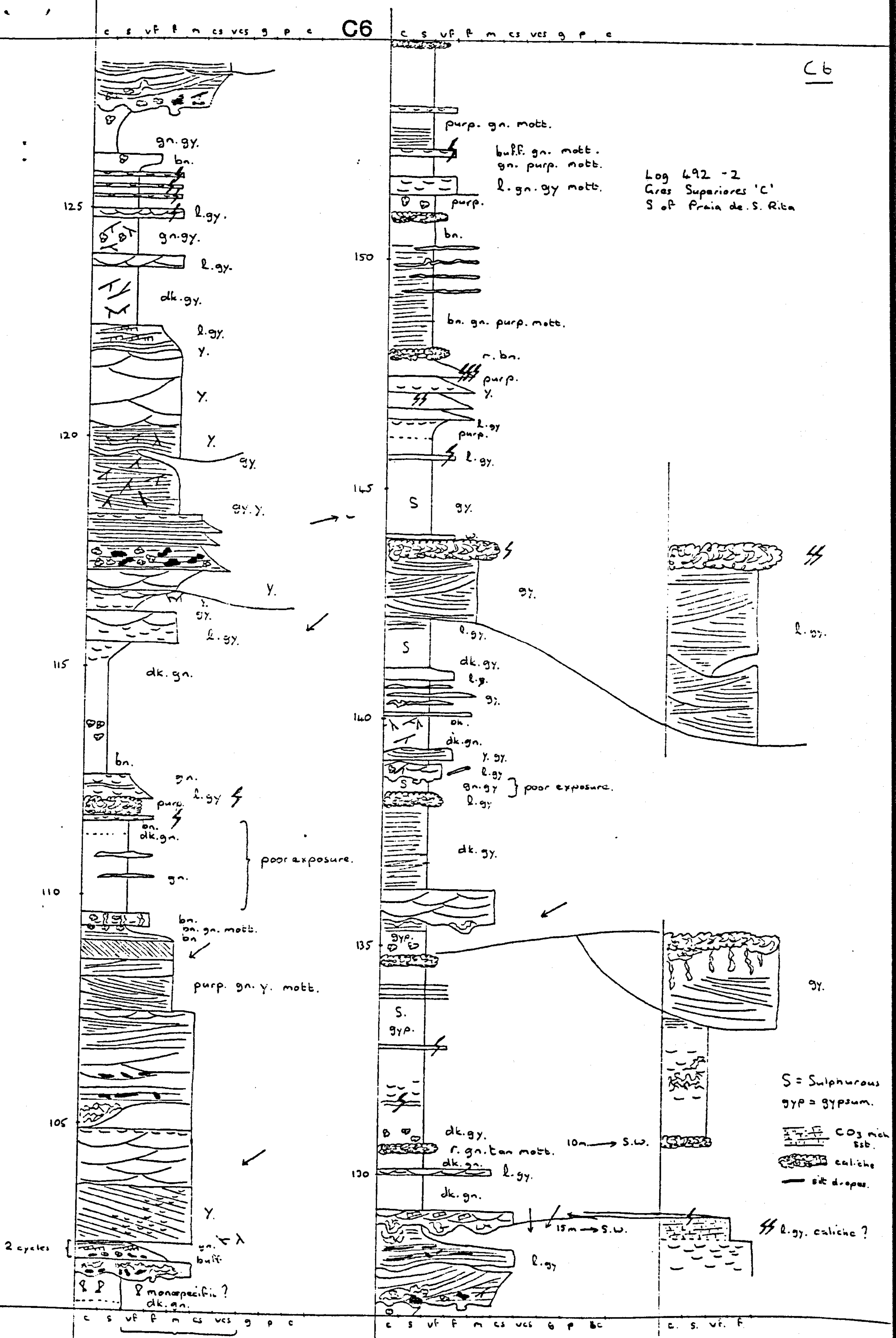

C2
sand
c s vf f m cs vcs g p b
Log 431 ② ≈20m. S of Log 431 ① S. of P. de S. Rita
Poor Exposure
dk. gy.
gy.
Poor Exposure
dk. gy.
Fault
dk. gy
y
gy.
gn. gy.
in life posn.
l. bn. gn.
dk. gn.
l. bn.
l. gy.
r. purp.
gn. l. bn. mott.
Schematic silts – f. sst.
bn.
gy.
dk. gy.
gy. mott.
gy
l. gy.
dk. gy.
r. bn. gn. mott.
dk. gy.
Poor Exposure.

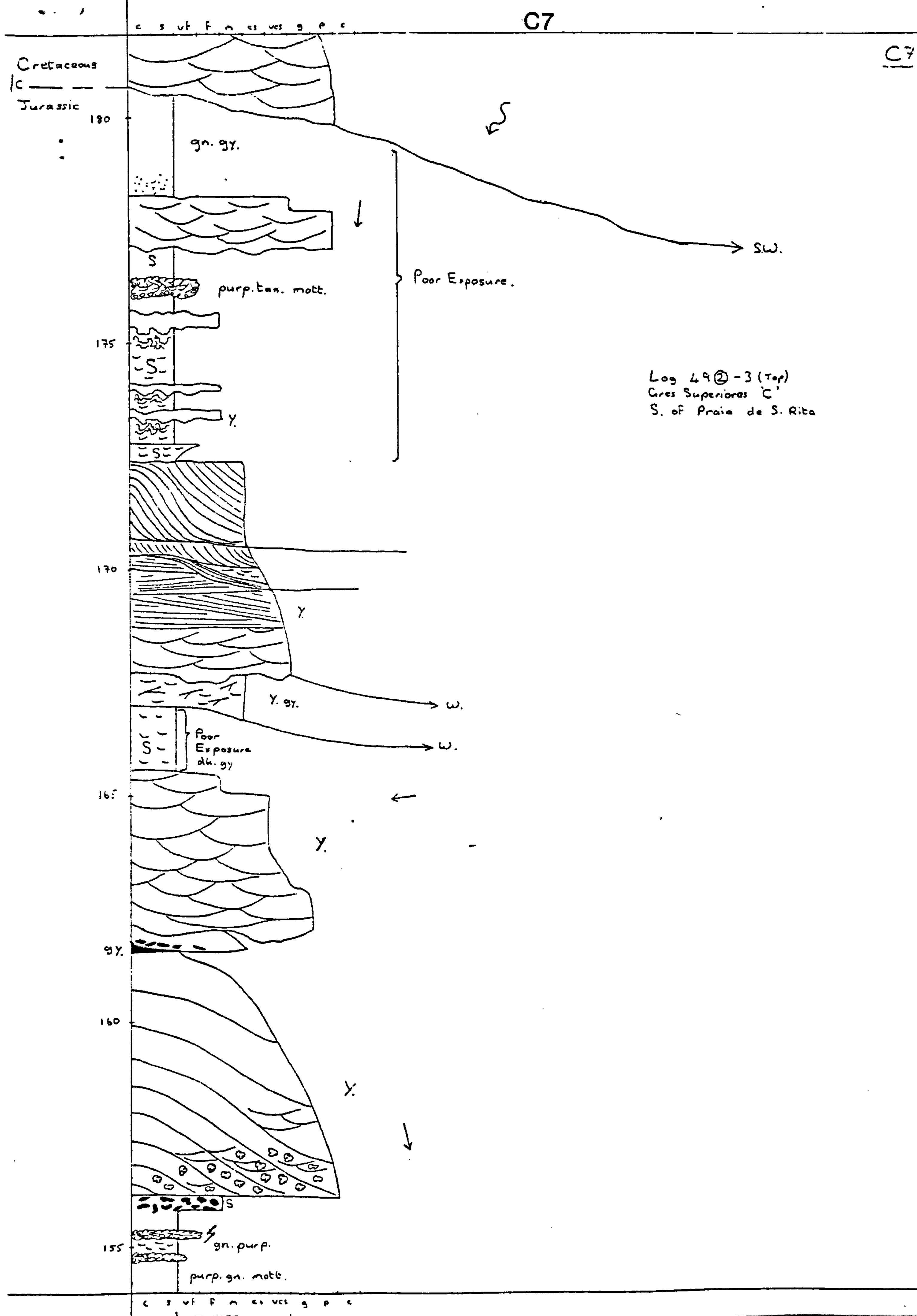
C3


sand
c s vf f m cs vcs g p b
Log L31③
C3
S. of P. de S. Rita
Base of cliff ↑ cliff to base of channel @ cliff top.
57
56
55
54
53
52
51
50
49
48
47
46
45
ophiomorpha
w.
g
bn-gn.
bn
bn.
bn.gn.
bn.
gn. bn. purp. mott
dk. gn.
buff.
gn
buff.
buff.
dk. gn.
bn. gn. buff mott.
bn
gn. bn
gn. bn.
bn.
dk. gn.
l. bn.
r.
70
69
68
67
66
65
64
63
62
61
60
59
58
v. schematic (see log 481).
y.
ol. gn.
y
y.
gy.
dk gy
ol. gn
P.
(in life posn.
buff.
dk. gn.
P.
m.
gy. y.
l. gy.
dk. gn.
c s vf f m cs vcs g p b


C4
c s vf f m cs vcs g p b
Log. 481 S. of Praia de S. Rita
Base 4m. below first major channel.
Burrow Orientations + Form (Unlined)
Planolites + Skolithos
gy.
y
l. gy.
15m. West.
dk. + l. gy.
o.
dk. gy.
P.
E. I. O. M.
gy. y.
o.
dk. gn.
P.
m.
l. gy.
dk. gn.
In Fault Zone Only Crudely Logged
buff.
bn.
F
F
l. gy.
dk. gn.
l. gy.
dk. gn.
r. bn. mott.
r. bn. gn. mott.
tan
r. gn. tan. mott
y.
y. gy.
y.
l. gy.
gy.
y.
gy. + buff
y.
y.
gy.
y.
gy
y.
dk. + l. gy.
71
70
69
68
67
66
65
64
63
62
61
60
59
74
73
72
71
70
69
68
67
66
65
64
63
62


c s vf f m cs vcs g p c
monospecific?
C5
dk.gn
100
bn.
bn.
gn.
bn.
bn.
bn.gn.
l.gy.
bn. gn. mott
bn.
dk.gn.
95
l.gy
dk.gn.
bn.gn. mott.
gn.
Silts dk.gn.
sands. l. gy.
90
Fine gd.
Lateral
Accretion
Unit.
dk. gn.
l.gy.
dk.gn.
purp
Log 492
F.
F.
w.
bn.
bn.
bn.
85
l.gn.
gn.bn.
3cycles
w. gn.
bn.gn.
dk.gn.
80
dk.gn.
gn.
gn.w.
dk.gn.
w
gn.
bn.
bn.
F.
F.
buff.
bn.
75
bn
buff.
c s vf f m cs vcs g p c
l. gy.
bn. gn.
gy.
15m. up-dip. 95
dk.gn.
c s vf f m cs. vcs g
sand.
Sandy Ripple Laminae.
Palaeocurrents (Approxima
Ripple Laminae
Disrupted Laminae (Water Esca
Gres Superiores 'C'.
S. of P. de S. Rita
Succeeds Log L81 (Note Change in Scale)

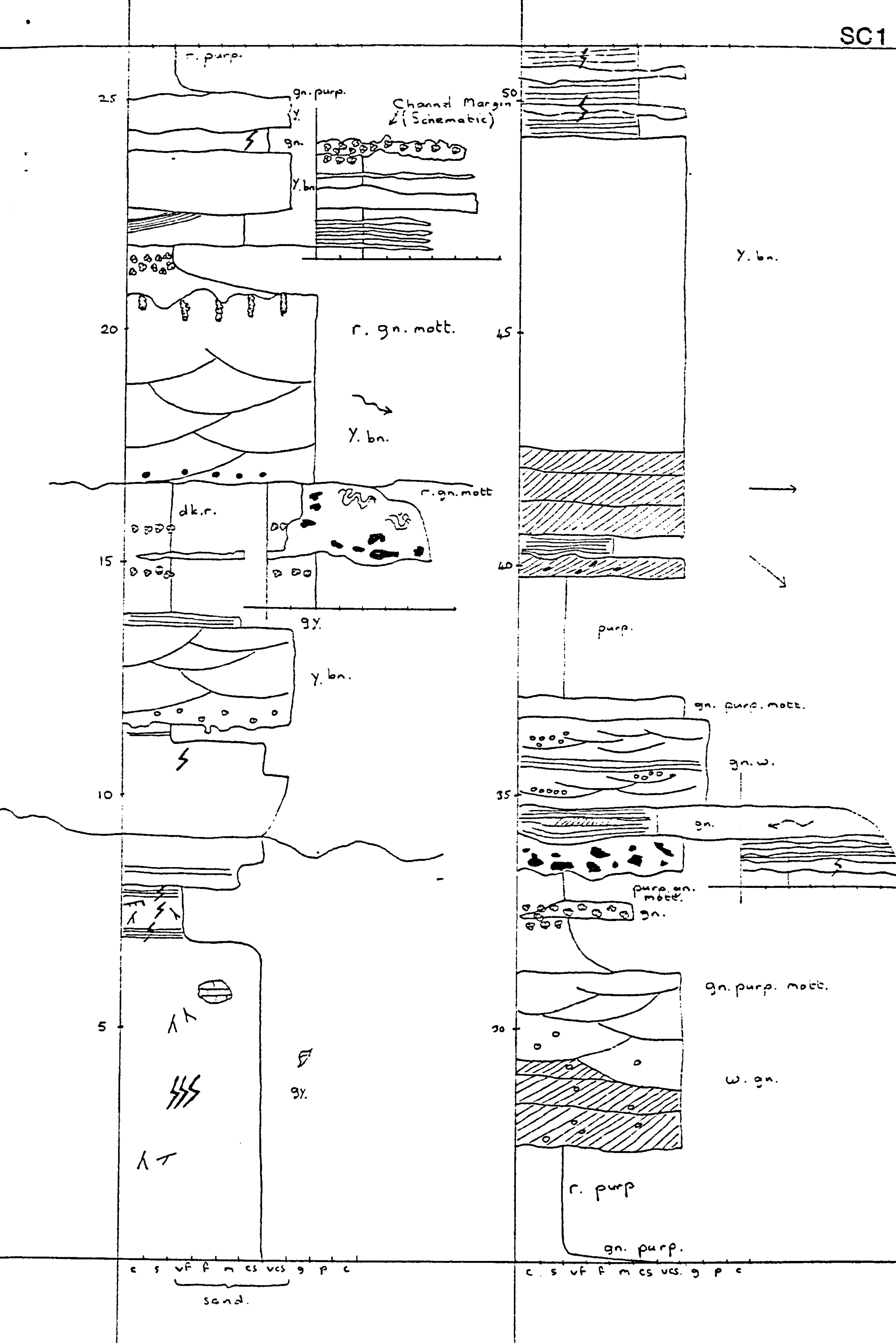

C6
c s vf f m cs vcs g p c
Log 492 -2
Gres Superiores 'C'
S of Praia de S. Rita
purp. gn. mott.
buff. gn. mott.
gn. purp. mott.
l. gn. gy mott.
purp.
bn.
150
bn. gn. purp. mott.
r. bn.
purp.
y.
l. gy
purp.
l. gy.
145
S
gy.
gy.
l. gy.
S
dk. gy.
l. g.
gy.
140
bn.
dk. gn.
y. gy.
l. gy
gn. gy
l. gy
poor exposure.
dk. gy.
135
gyp.
S.
gyp.
dk. gy.
r. gn. tan mott.
dk. gn.
l. gy.
130
dk. gn.
10m → S.W.
15m → S.W.
l. gy
l. gy. caliche ?
gy.
S = Sulphurous
gyp = gypsum.
CO3 rich sst.
caliche
silt drapes.
gn. gy.
bn.
125
l. gy.
gn. gy.
l. gy.
dk. gy.
l. gy.
y.
y.
120
y.
gy.
gy. y.
y.
y.
gy.
l. gy.
115
dk. gn.
bn.
gn.
l. gy
purp.
bn.
dk. gn.
gn.
poor exposure.
110
bn.
bn. gn. mott.
bn
purp. gn. y. mott.
105
y.
2 cycles
gn.
buff
monospecific ?
dk. gn.
c s vf f m cs vcs g p c
c s vf f m cs vcs G p Bc
c. s. vf. f.


c s vf f m cs vcs g p c
C7
Cretaceous
C
Jurassic
180
gn. gy.
S
purp. tan. mott.
Poor Exposure.
S.W.
175
S
Y.
S
Log 49② -3 (Top)
Gres Superiores 'C'
S. of Praia de S. Rita
170
Y.
Y. gy.
W.
Poor Exposure dk. gy
S
W.
165
Y.
gy.
160
Y.
S
155
gn. purp.
purp. gn. mott.
c s vf f m cs vcs g p c

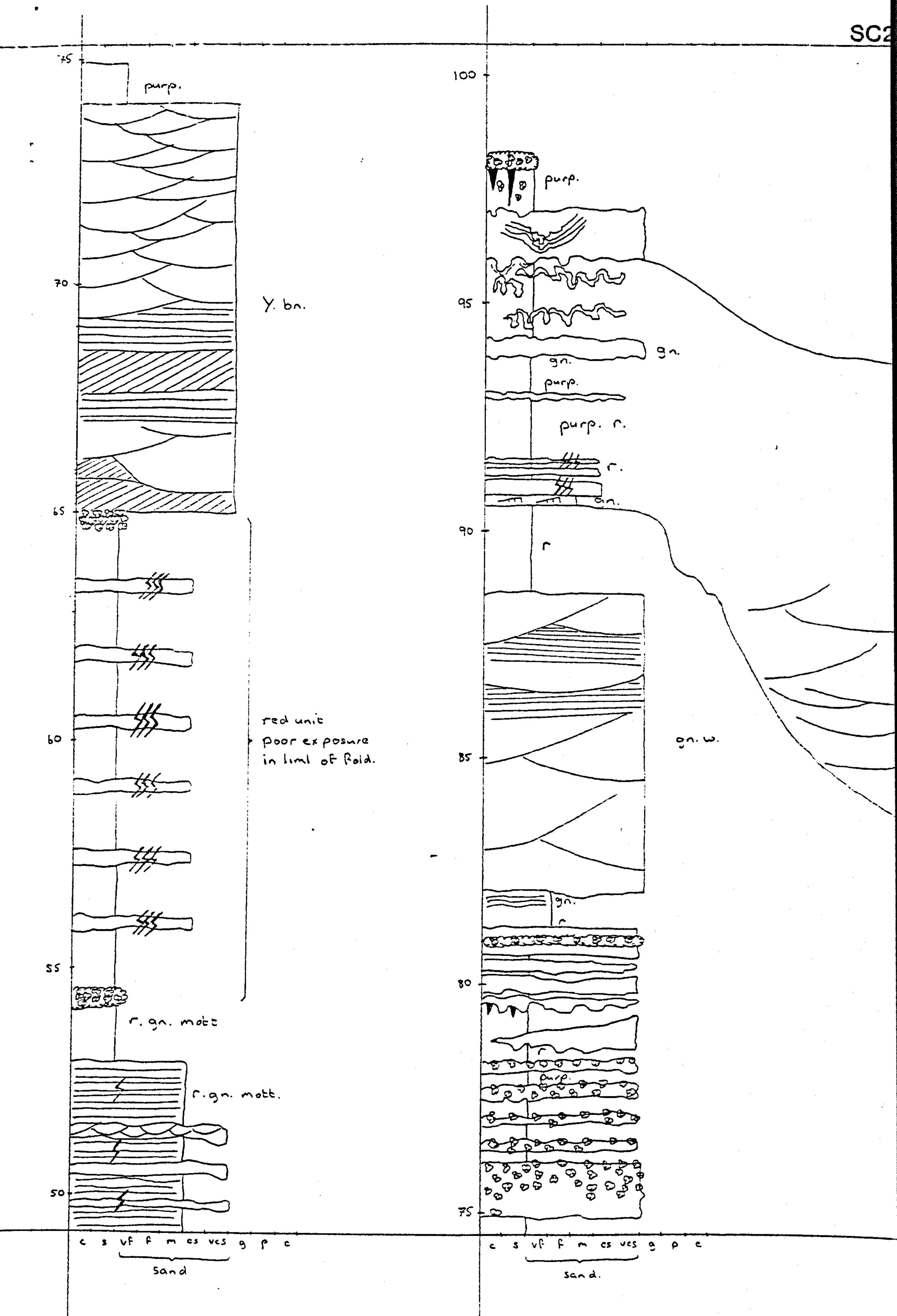
ENCLOSURE 9

LOGS: PRAIA DA AMOREIRA SECTION

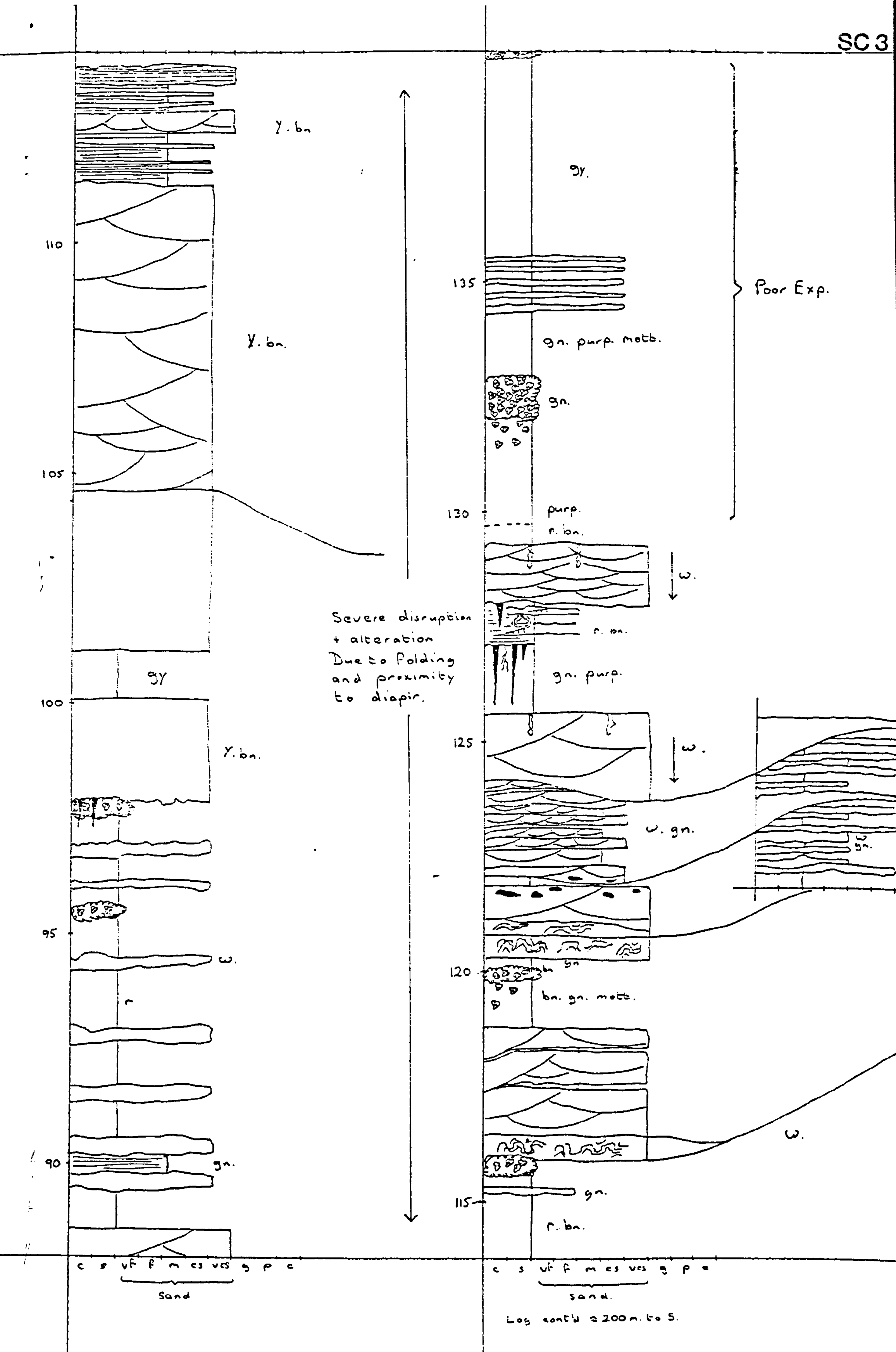
(SC1-7)

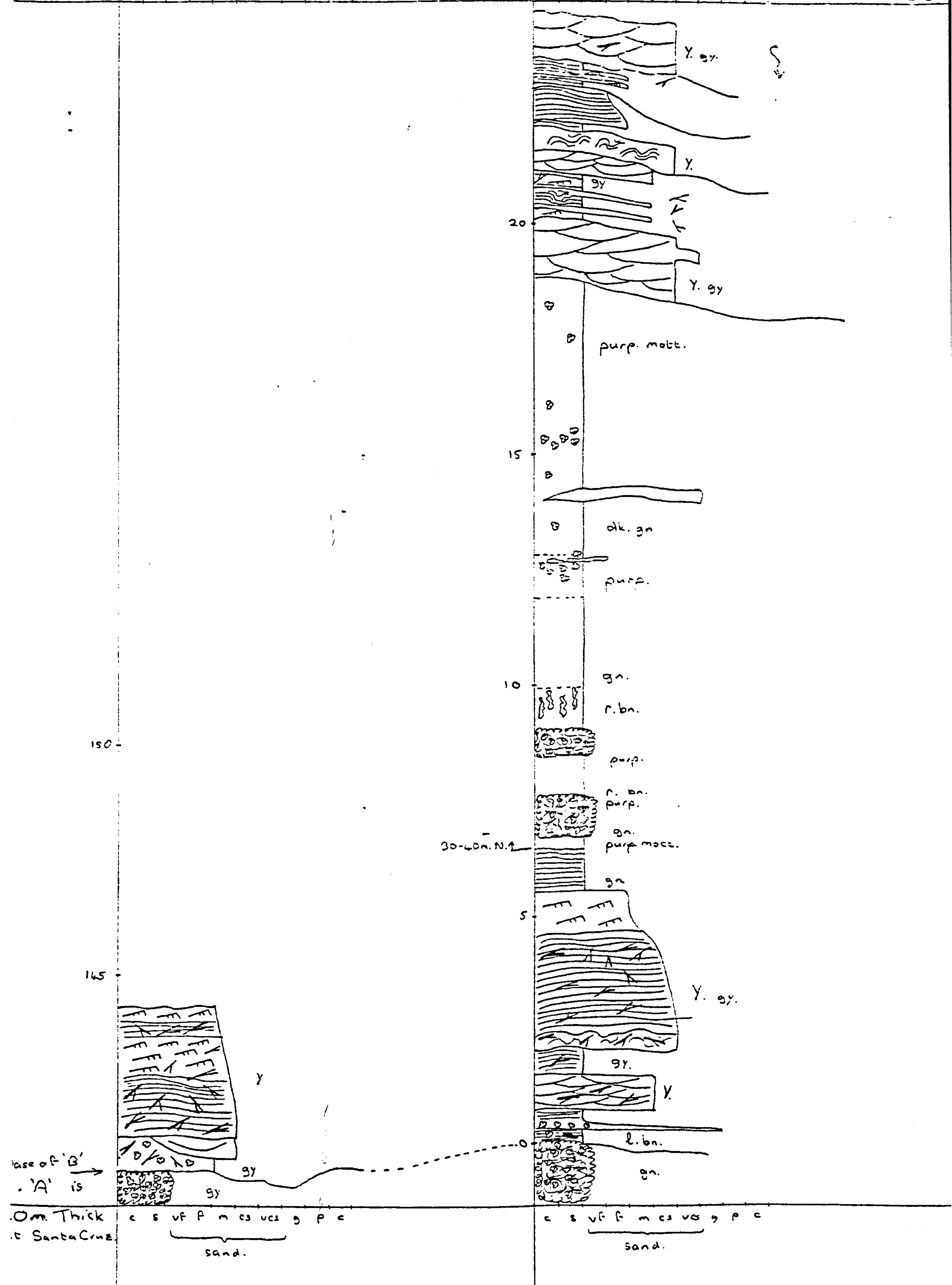


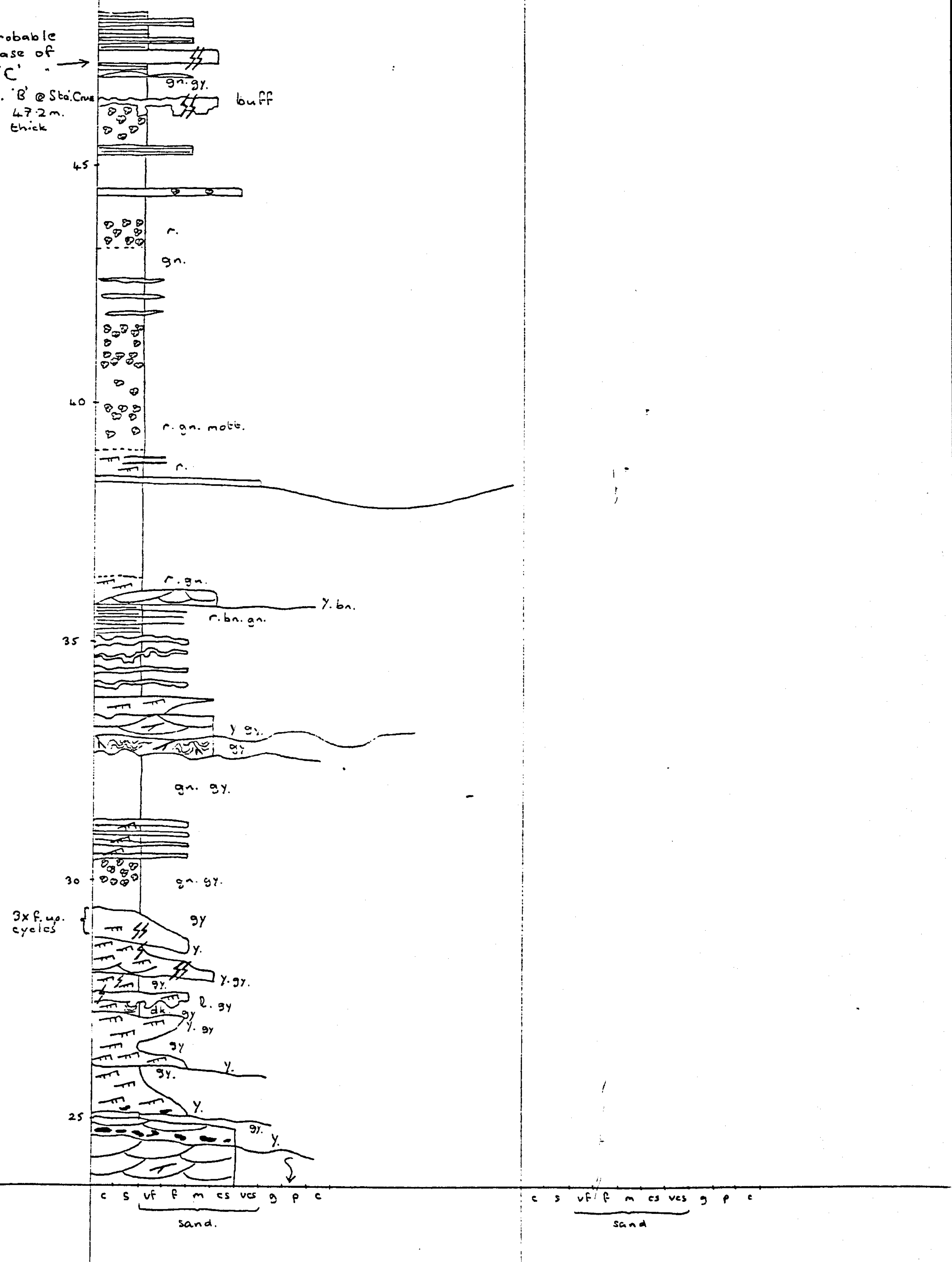
Santa Cruz Section Basal Log



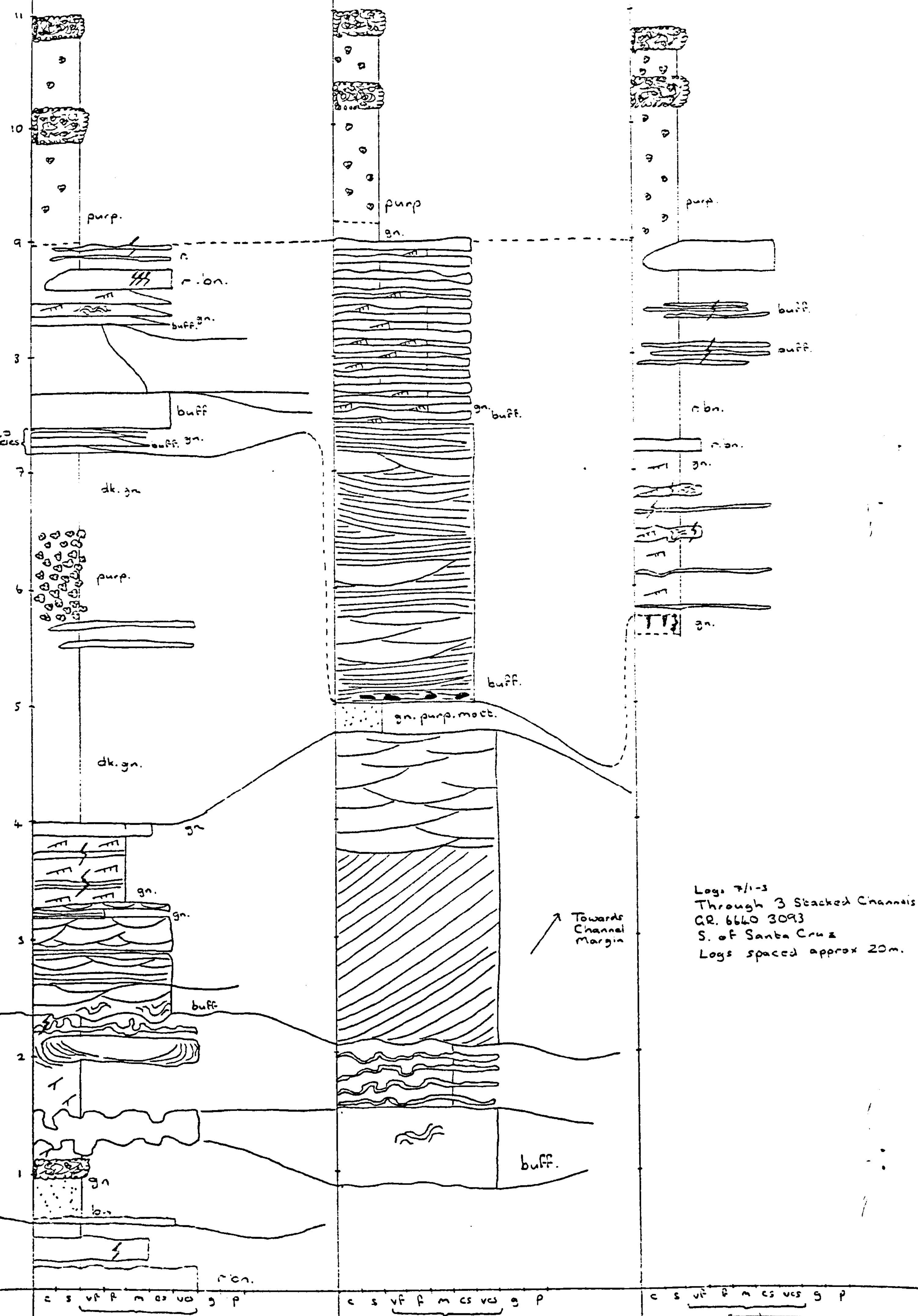
purp.
Y. bn.
red unit
poor exposure
in liml of fold.
r. gn. mott
r. gn. mott.
c s vf f m cs vcs g p c
sand
purp.
gn.
gn.
purp.
purp. r.
r.
gn.
r
gn. w.
gn.
r
r
purp.
c s vf f m cs vcs g p c
sand.


SC3
Y. bn
Y. bn.
gy
Y. bn.
w.
r
gn.
Severe disruption + alteration Due to Folding and proximity to diapir.
gy.
Poor Exp.
gn. purp. mott.
gn.
purp.
r. bn.
w.
r. bn.
gn. purp.
w.
w. gn.
gn.
bn gn
bn. gn. mott.
w.
gn.
r. bn.
110
105
100
95
90
135
130
125
120
115
c s vf f m cs vcs g p c
Sand
c s vf f m cs vcs g p c
sand.
Log cont'd ≈ 200m. to S.


Y. gy.
Y.
gy
20
Y. gy
purp. mott.
15
dk. gn
purp.
10
gn.
r. bn.
150
purp.
r. bn.
purp.
gn.
30-40m. N.
purp mott.
gn
5
145
Y. gy.
gy.
Y.
y
ase of 'B'
. 'A' is
gy
0
l. bn.
gn.
gy
.Om. Thick
t Santa Cruz
c s vf f m cs vcs g p c
sand.
c s vf f m cs vcs g p c
sand.

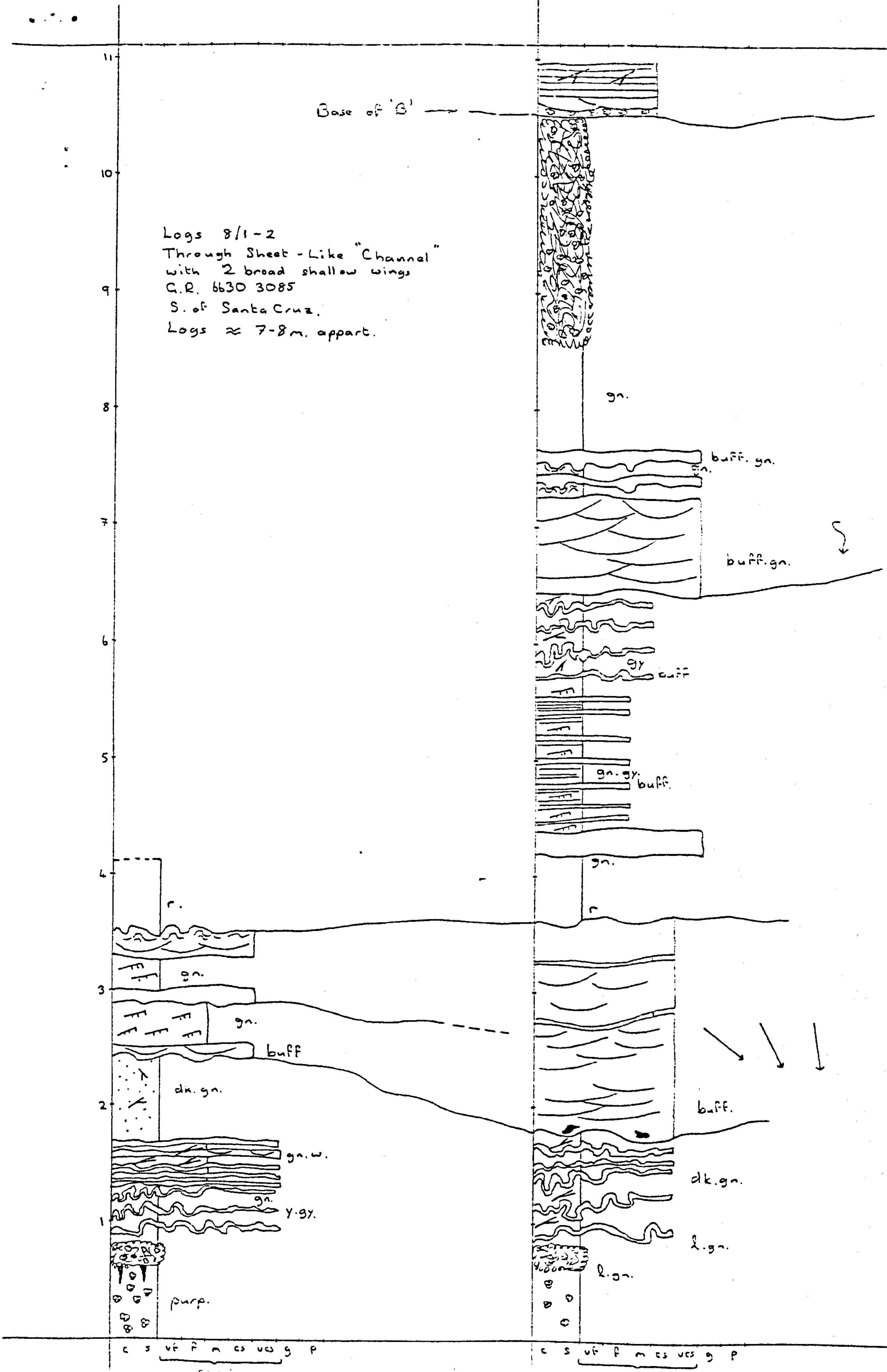


probable
base of
'C'
'B' @ Sta. Cruz
47.2 m.
thick
gn. gy.
buff
45
r.
gn.
40
r. gn. mott.
r.
r. gn.
y. bn.
r. bn. gn.
35
y gy.
gy
gn. gy.
30
gn. gy.
3x f. up. cycles
gy
y.
y. gy.
gy.
l. gy
dk.
gy
y. gy
gy
y.
gy.
y.
25
gy.
y.
c s vf f m cs vcs g p c
sand.
c s vf f m cs vcs g p c
sand



Logs 7/1-3
Through 3 Stacked Channels
GR. 6640 3093
S. of Santa Cruz
Logs spaced approx 20m.

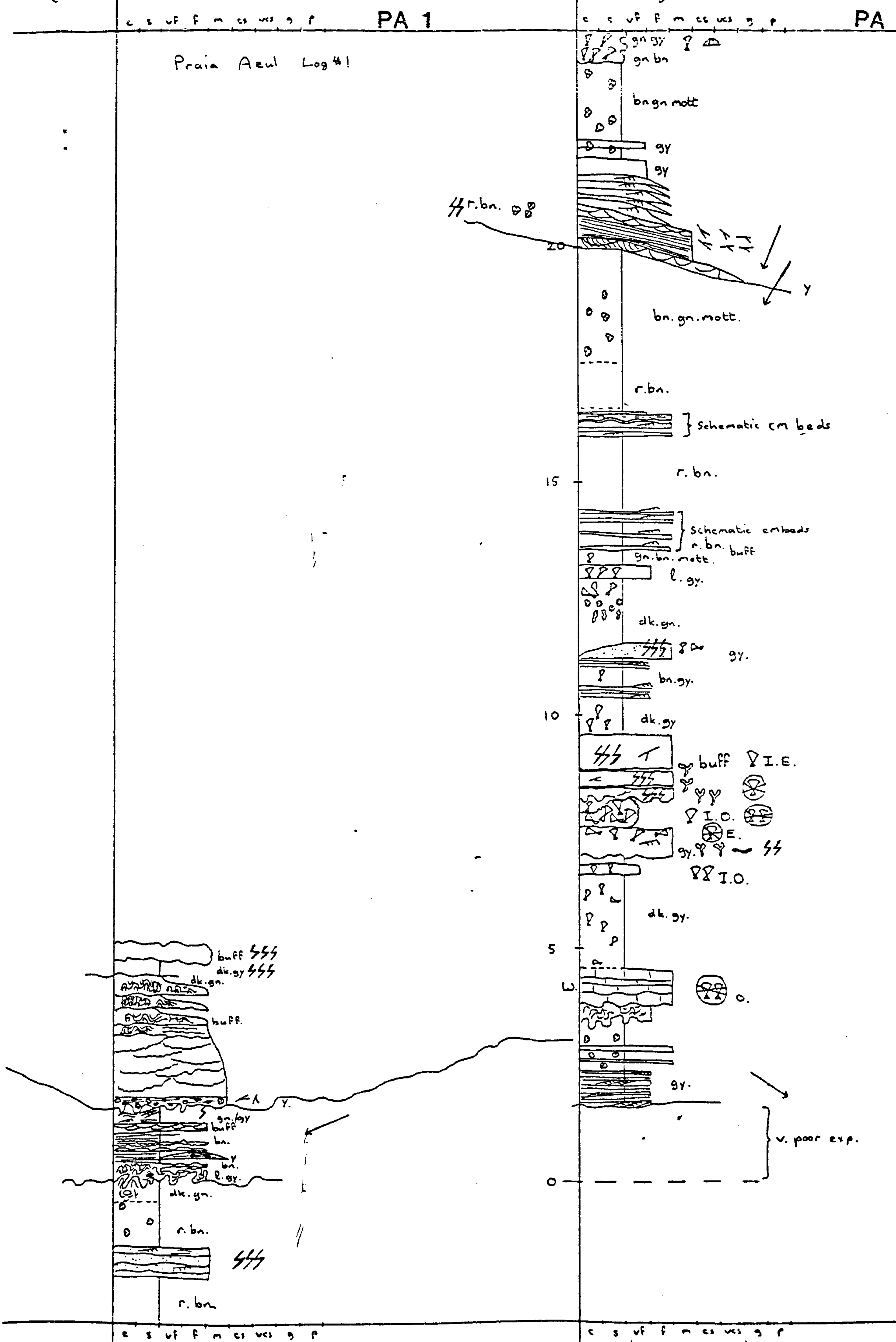
SC7


Base of 'B'
Logs 8/1-2
Through Sheet-Like "Channel"
with 2 broad shallow wings
G.R. 6630 3085
S. of Santa Cruz.
Logs ≈ 7-8m. appart.
11
10
9
8
7
6
5
4
3
2
1
gn.
buff. gn.
gn.
buff. gn.
gy
buff
gn. gy.
buff.
gn.
r.
r
gn.
gn.
buff
dk. gn.
buff.
gn. w.
dk. gn.
gn.
y. gy.
l. gn.
l. gn.
purp.
c s vf f m cs vcs g p
c s vf f m cs vcs g p

ENCLOSURE 10

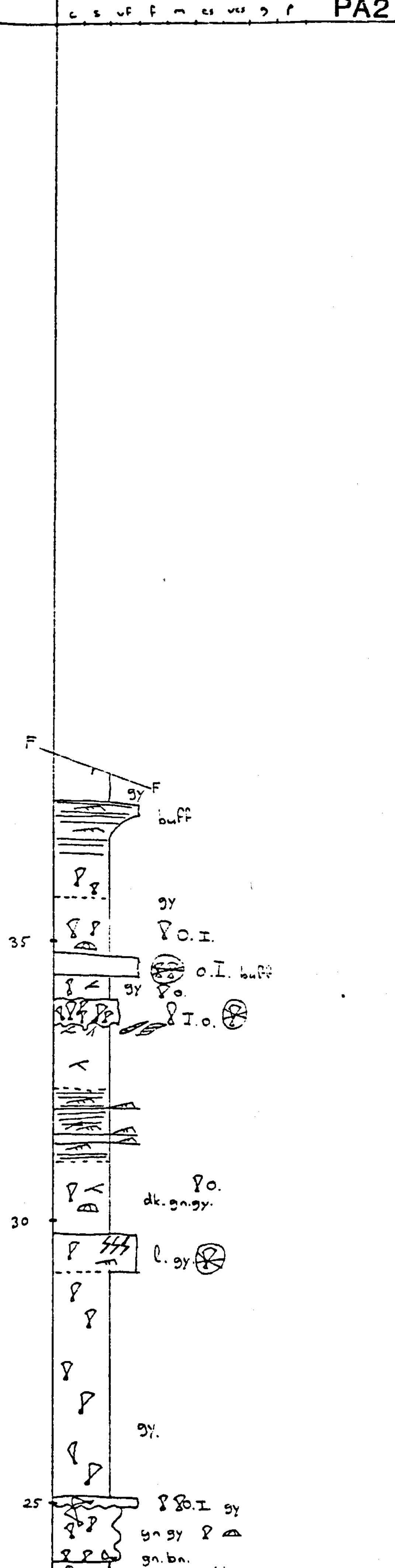
LOGS: PRAIA AZUL SECTION

(PA 1-4)


PA 1
PA
c s vf f m cs vcs g p
Praia Azul Log #1
gn gy
gn bn
bn gn mott
gy
gy
r.bn.
20
Y
bn. gn. mott.
r.bn.
Schematic cm beds
15
r. bn.
Schematic cm beds
r. bn.
buff
gn. bn. mott.
l. gy.
dk. gn.
gy.
gy.
bn. gy.
10
dk. gy
buff
I.E.
I.O.
E.
gy.
I.O.
dk. gy.
5
W.
O.
gy.
v. poor exp.
0
buff
dk. gy
dk. gn.
buff.
Y.
gn./gy
buff
bn.
Y
bn.
l. gy.
dk. gn.
r. bn.
r. bn
c s vf f m cs vcs g p
c s vf f m cs vcs g p

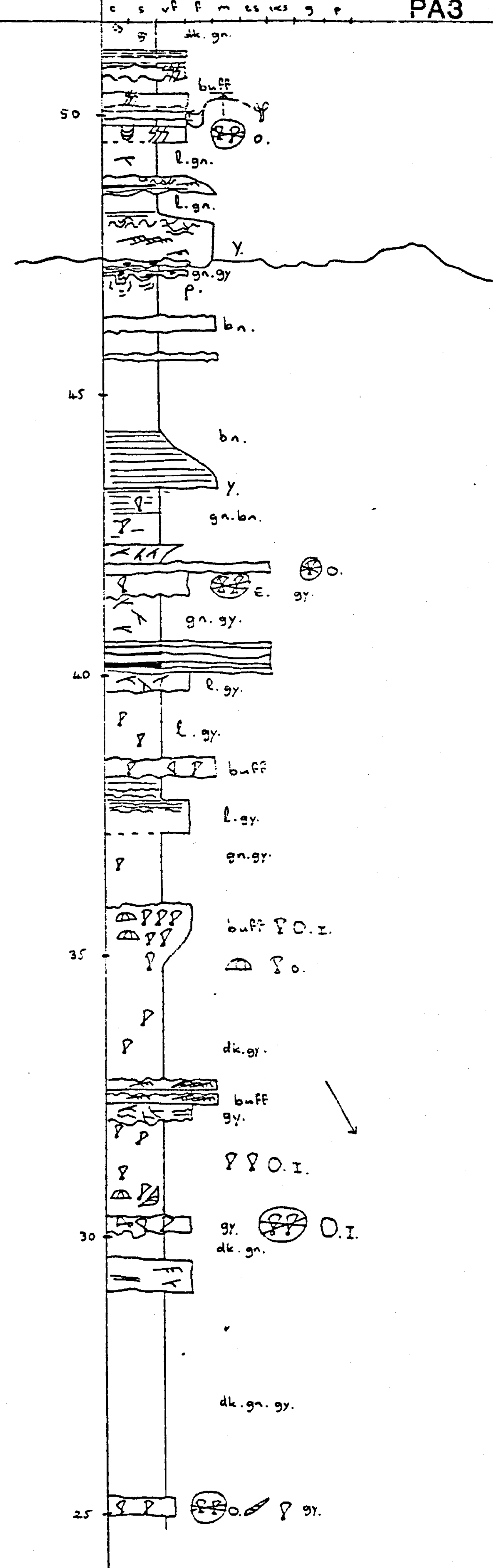
Praia Azul Log #2 Cont.

PA2

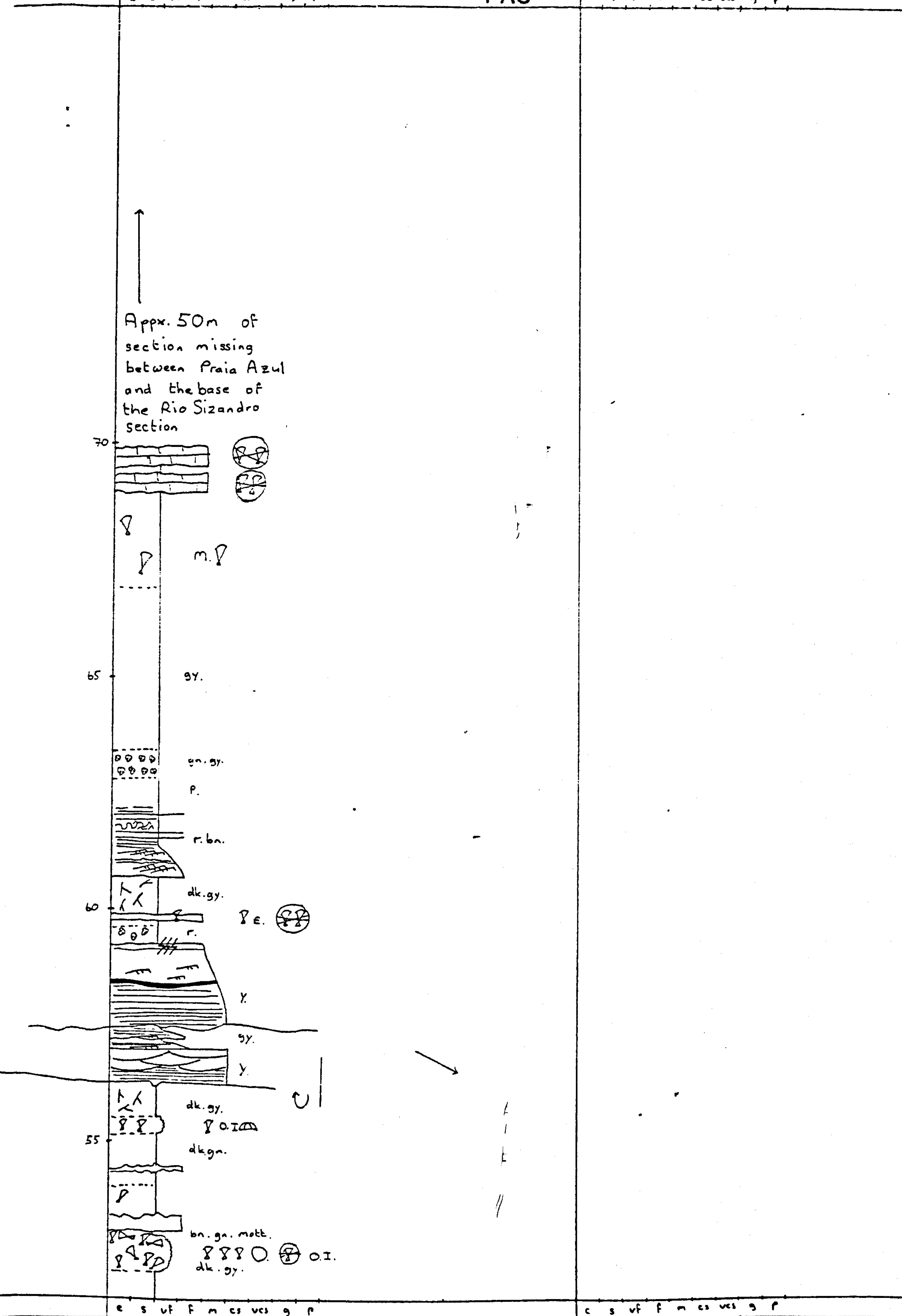


Praia Azul Log #3

PA3

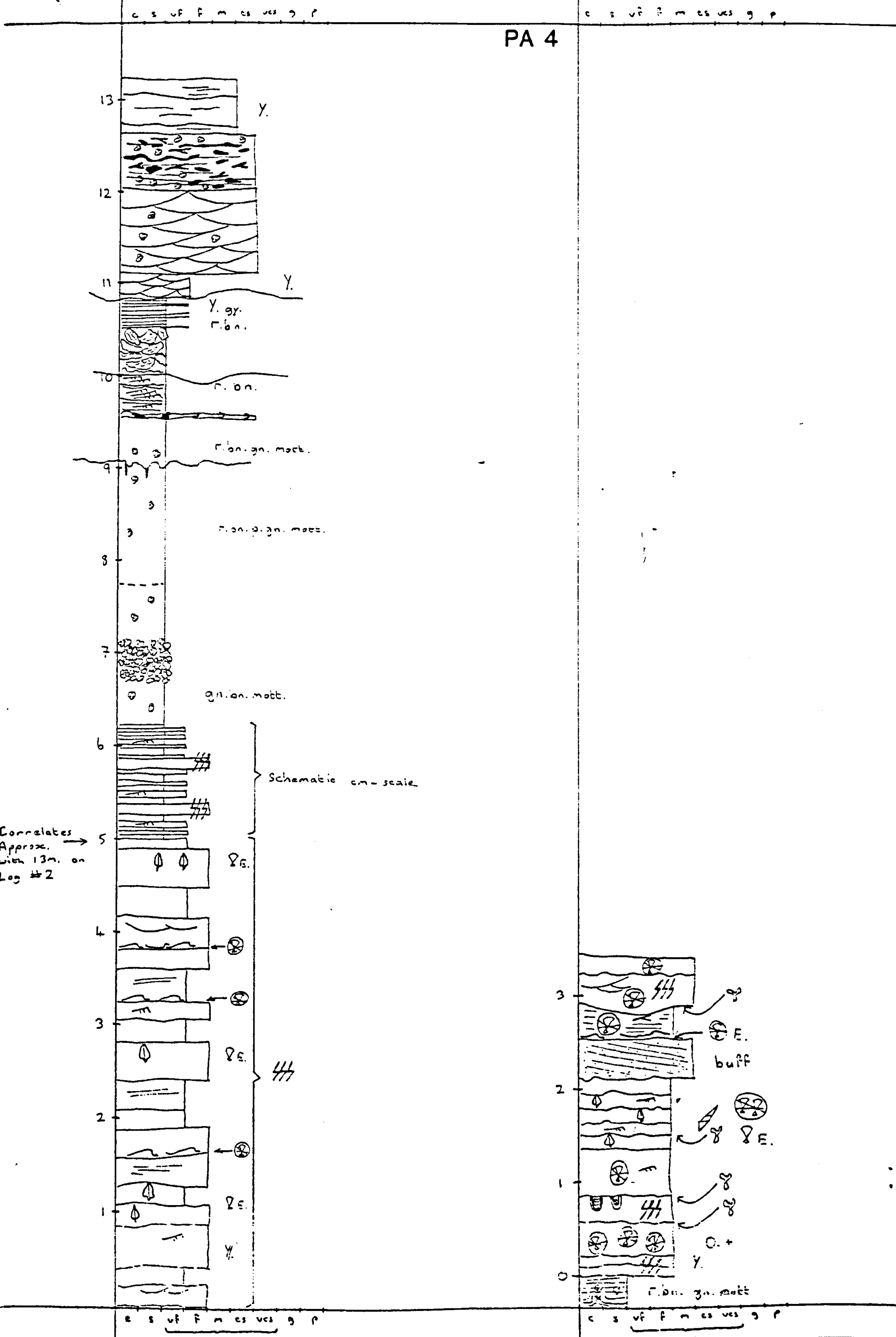


PA3



c s vf f m cs vcs g p
Appx. 50m of section missing between Praia Azul and the base of the Rio Sizandro section
70
m.
65
gy.
gn. gy.
p.
r. bn.
dk. gy.
60
E.
r.
y.
gy.
y.
dk. gy.
O.I.
55
dk. gn.
bn. gn. mott.
O.I.
dk. gy.
c s vf f m cs vcs g p

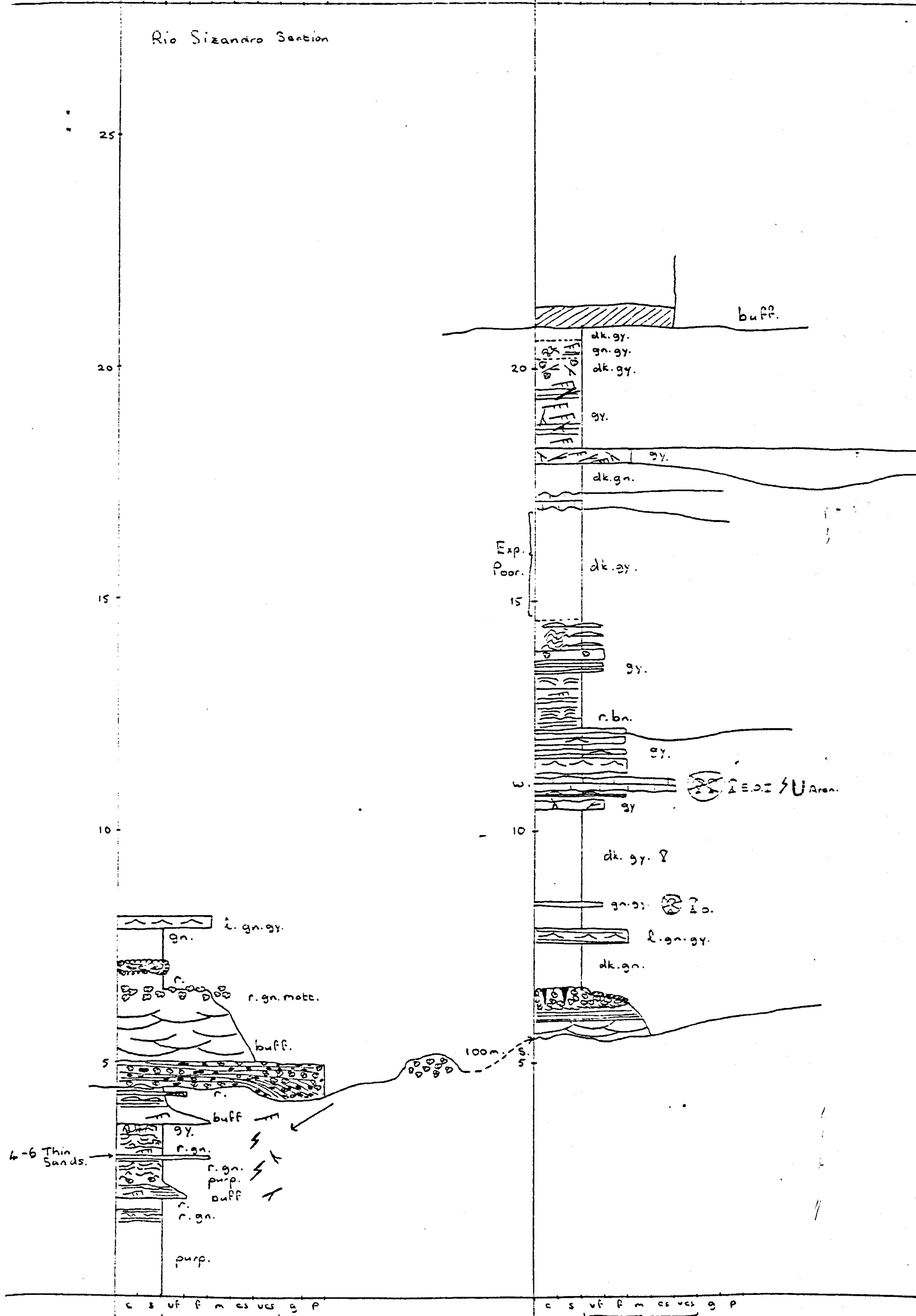
P. Azul Base Sect. 100m S of Alto da Veia.



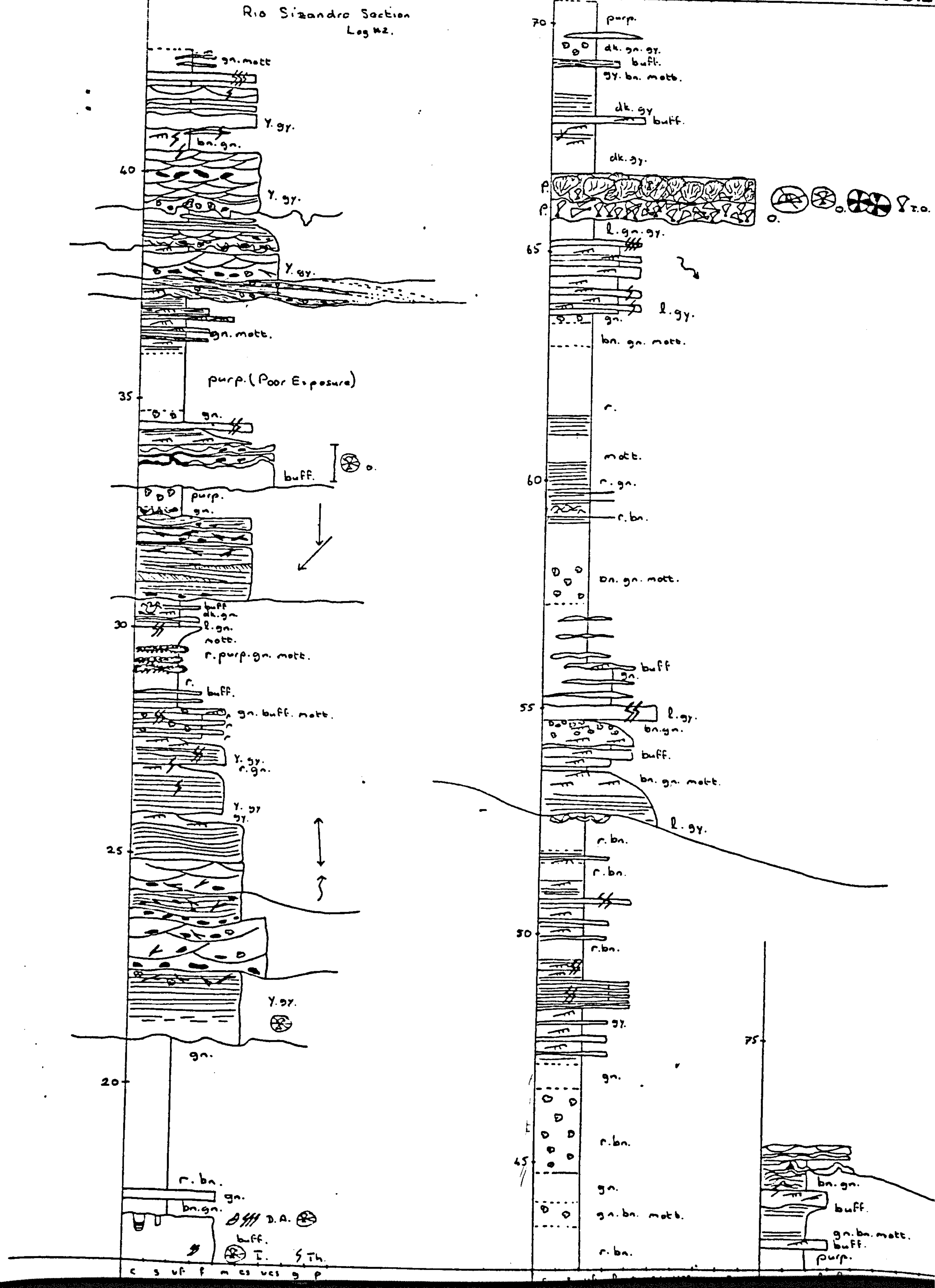
ENCLOSURE 11

LOGS: RIO SIZANDRO SECTION

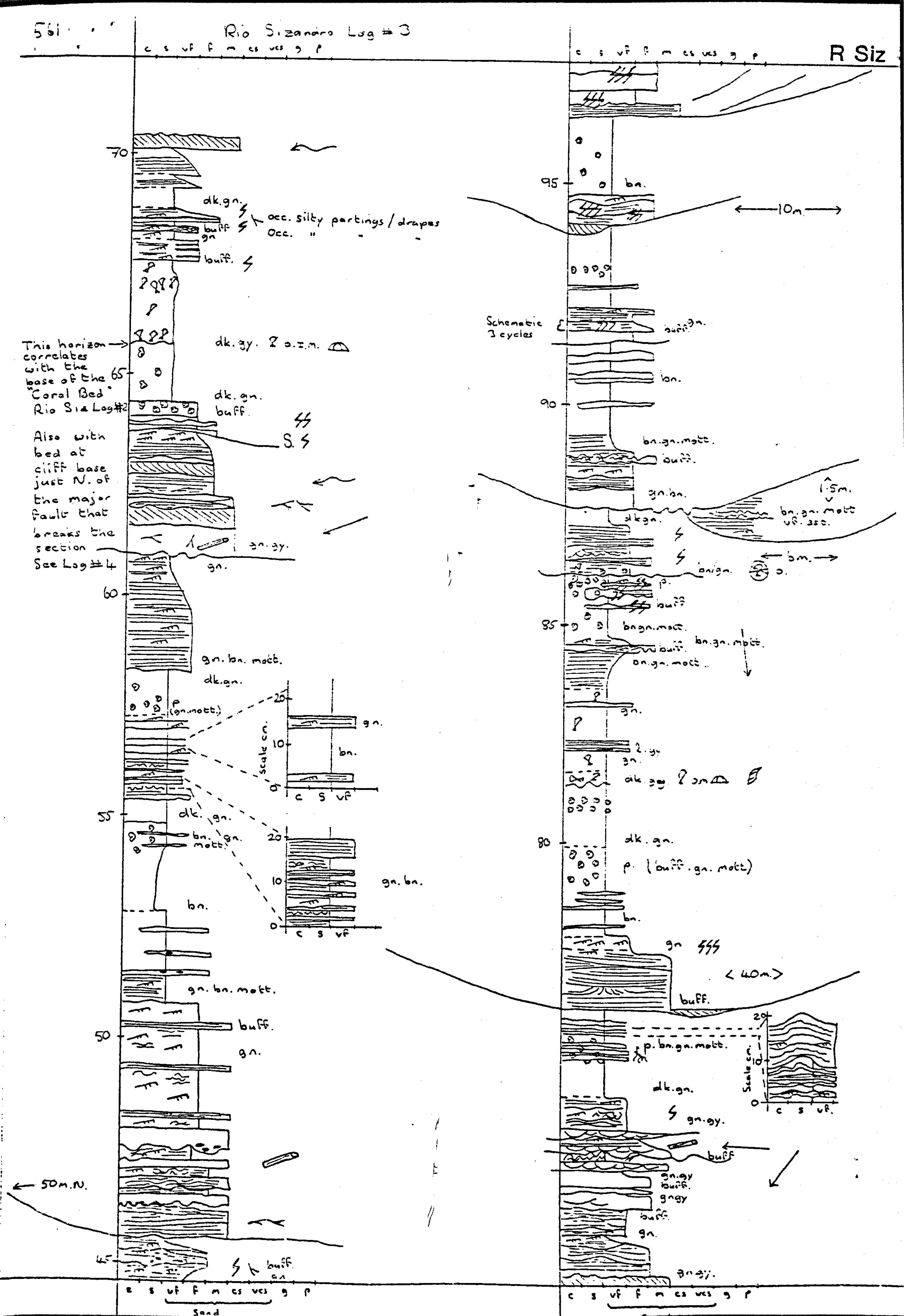
(R SIZ 1-7)


Rio Sizandro Section
25
20
15
10
5
ĺ. gn. gy.
gn.
r.
r. gn. mott.
buff.
r.
buff
gy.
r. gn.
4-6 Thin Sands.
r. gn.
purp.
buff
r.
r. gn.
purp.
c s vf f m cs vcs g p
100m.
S.
buff.
dk. gy.
gn. gy.
dk. gy.
gy.
gy.
dk. gn.
Exp. Poor.
dk. gy.
gy.
r. bn.
gy.
w.
Aren.
gy
dk. gy.
gn. gy.
l. gn. gy.
dk. gn.
c s vf f m cs vcs g p

R Siz

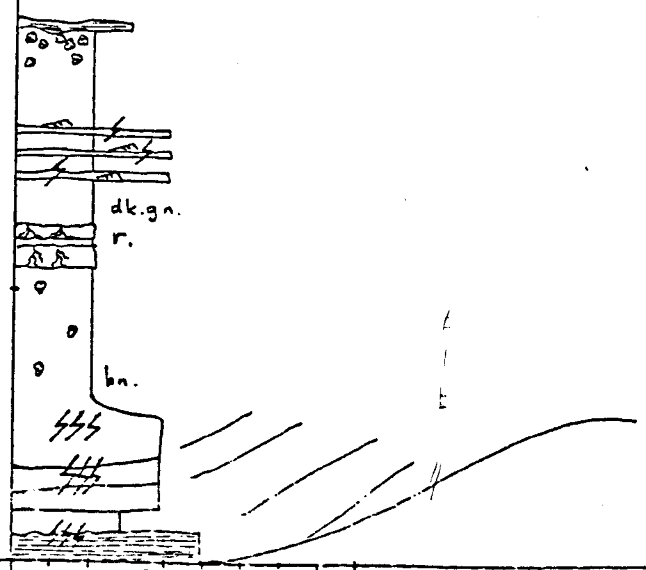


Rio Sizandro Section
Log №2.
gn. mott
Y. gy.
bn. gn.
40
Y. gy.
Y. gy.
gn. mott.
purp. (Poor Exposure)
35
gn.
buff.
o.
purp.
gn.
buff
dk. gn.
30
l. gn.
mott.
r. purp. gn. mott.
r.
buff.
gn. buff. mott.
Y. gy.
r. gn.
Y. gy
gy.
25
Y. gy.
gn.
20
r. bn.
gn.
bn. gn.
D.A.
buff.
I.
Th.
c s vf f m cs vcs g p
70
purp.
dk. gn. gy.
buff.
gy. bn. mott.
dk. gy
buff.
dk. gy.
o.
T.O.
l. gn. gy.
65
l. gy.
gn.
bn. gn. mott.
r.
mott.
60
r. gn.
r. bn.
bn. gn. mott.
buff
gn.
55
l. gy.
bn. gn.
buff.
bn. gn. mott.
l. gy.
r. bn.
r. bn.
50
r. bn.
gy.
gn.
r. bn.
45
gn.
gn. bn. mott.
r. bn.
75
bn. gn.
buff.
gn. bn. mott.
buff.
purp.


561
Rio Sizandro Log # 3
R Siz
c s vf f m cs vcs g p
70
dk.gn.
occ. silty partings / drapes
Occ. " " "
buff
gn
buff.
dk.gy. o.z.m.
This horizon correlates with the base of the 65 "Coral Bed" Rio Siz Log #2
dk.gn.
buff
Also with bed at cliff base just N. of the major fault that breaks the section
See Log # 4
gn.gy.
gn.
60
gn.bn.mott.
dk.gn.
p (gn.mott.)
Scale cm.
gn.
bn.
gn.bn.
55
dk.gn.
bn. mott.
gn.
bn.
gn.bn.mott.
buff.
50
gn.
50m.N.
45
buff
gn
Sand
95
bn.
10m
Schematic 3 cycles
buff gn.
bn.
90
bn.gn.mott.
buff.
gn.bn.
1.5m.
bn.gn.mott
vf. sst.
dk gn.
5m.
bn/gn.
p.
buff
85
bn.gn.mott.
buff.
bn.gn.mott.
bn.gn.mott.
gn.
gn.
dk.gy. gn
80
dk.gn.
p. (buff.gn.mott)
bn.
gn
40m.
buff.
p.bn.gn.mott.
dk.gn.
gn.gy.
buff
gn.gy
buff.
gngy
buff.
gn.
gn.gy.

R Siz 3

105



100

c s vf f m cs vcs g p

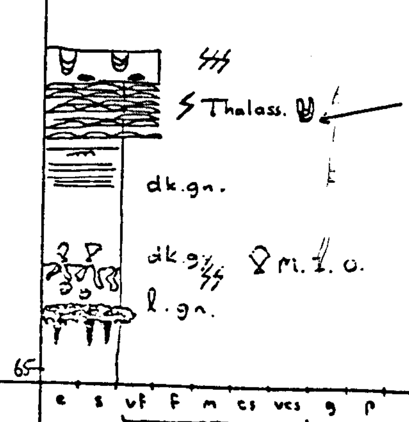
c s vf f m cs vcs g p

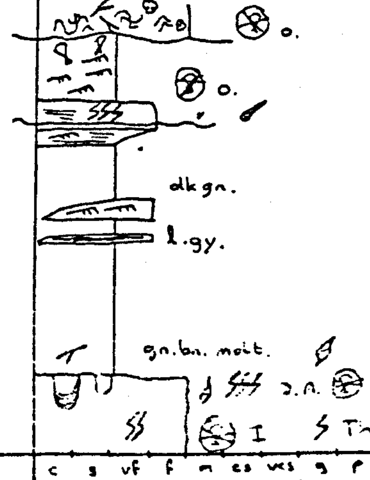
c s vf f m cs vcs g p R Siz 4

c s vf f m cs vcs g p R Siz5

Log . Rio Sizandro # 4
Terrestrial / Marine transition at beach level.
@ G.R.
Just N. of fault which breaks
the section -

Log # 5
Better exposed section
to N of Log # 2
- bioturbated bed ↑ base
of major channel.





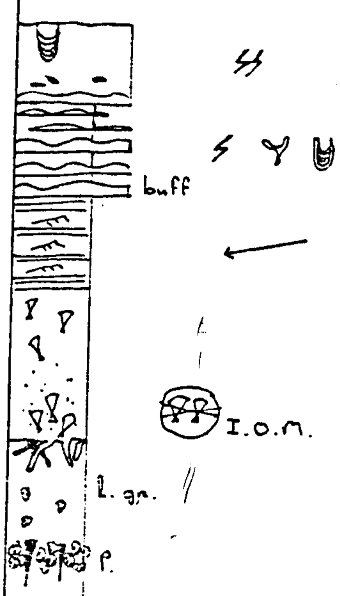
1:50

c s vf f m cs vcs g p

c s vf f m cs vcs g p

R Siz 6

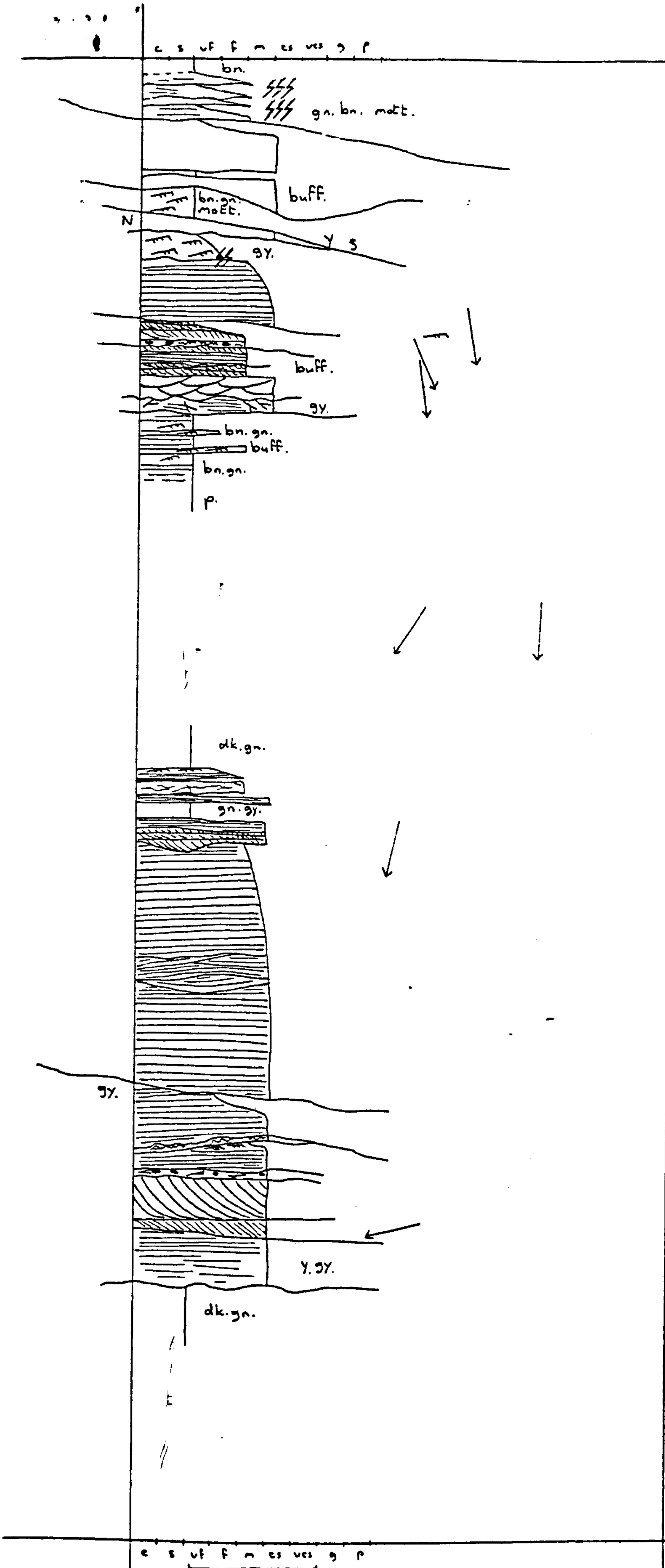
Cliff base 50m north of Fault which breaks section



c s vf f m cs vcs g p

c s vf f m cs vcs g p

R Siz 7



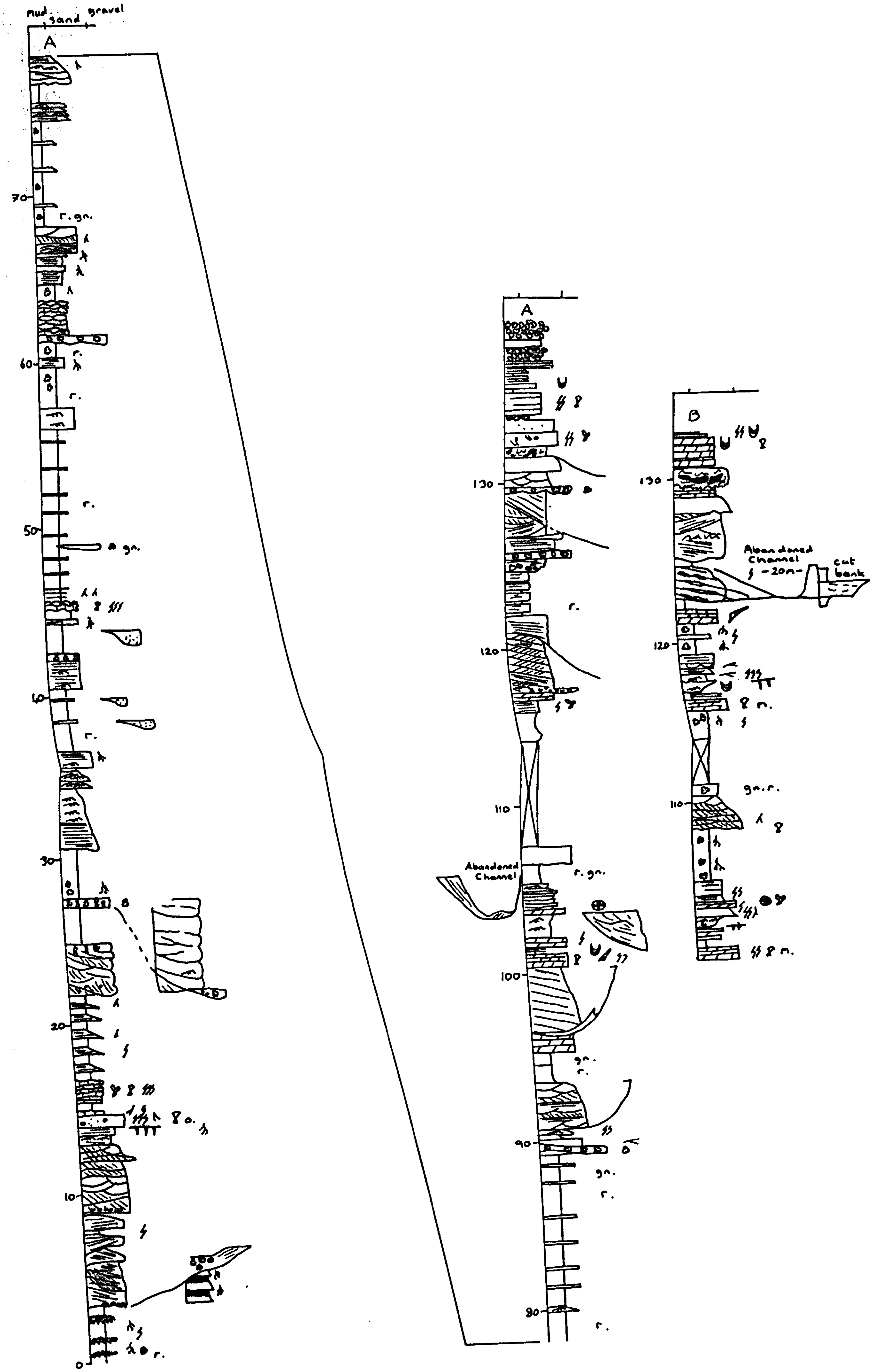
c s vf f m cs vcs g p

Rio Sizandro Logs
of 2 maj. channels in Log # 2
Logged ≈ 90m. S. of previous log.
80m.

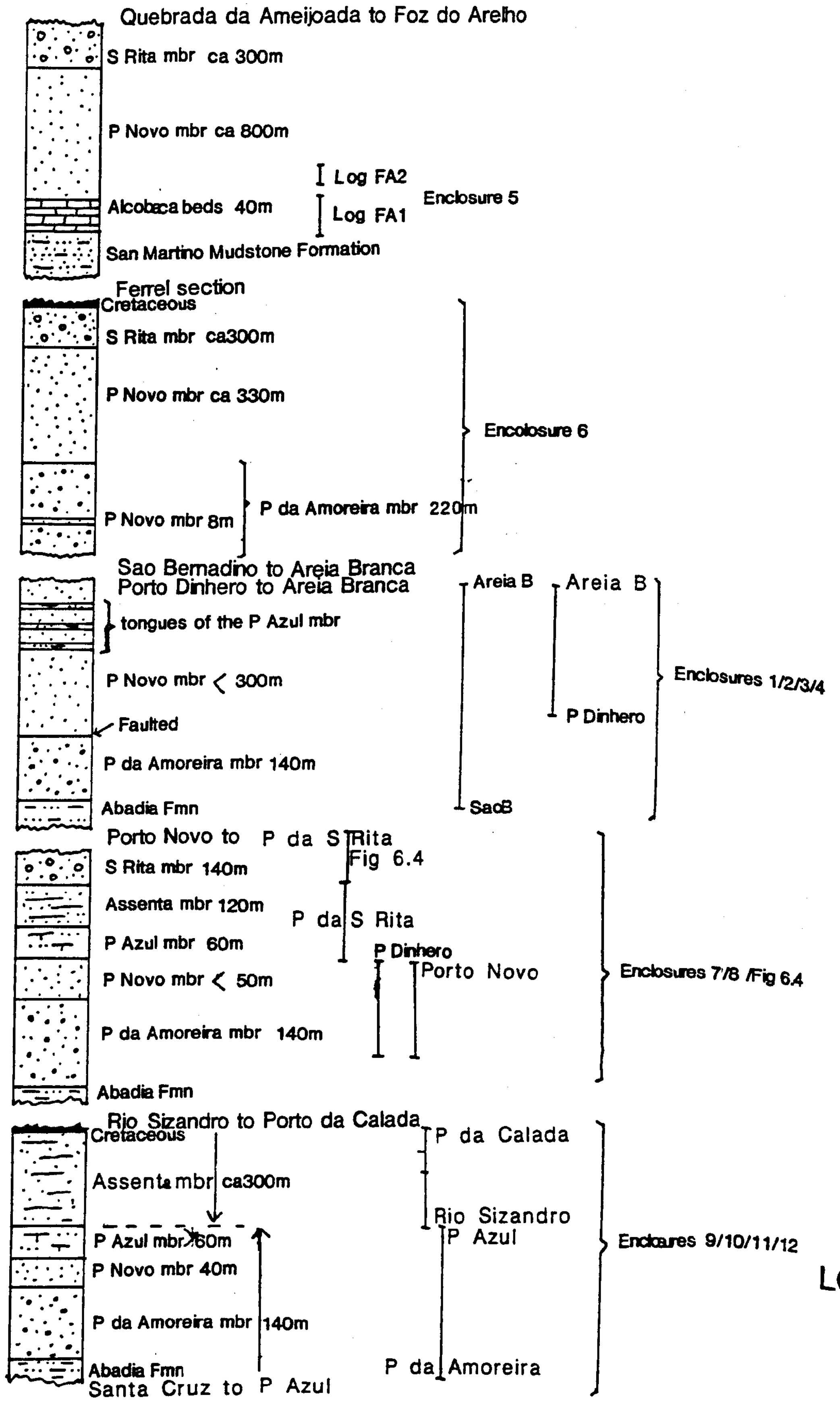
c s vf f m cs vcs g p

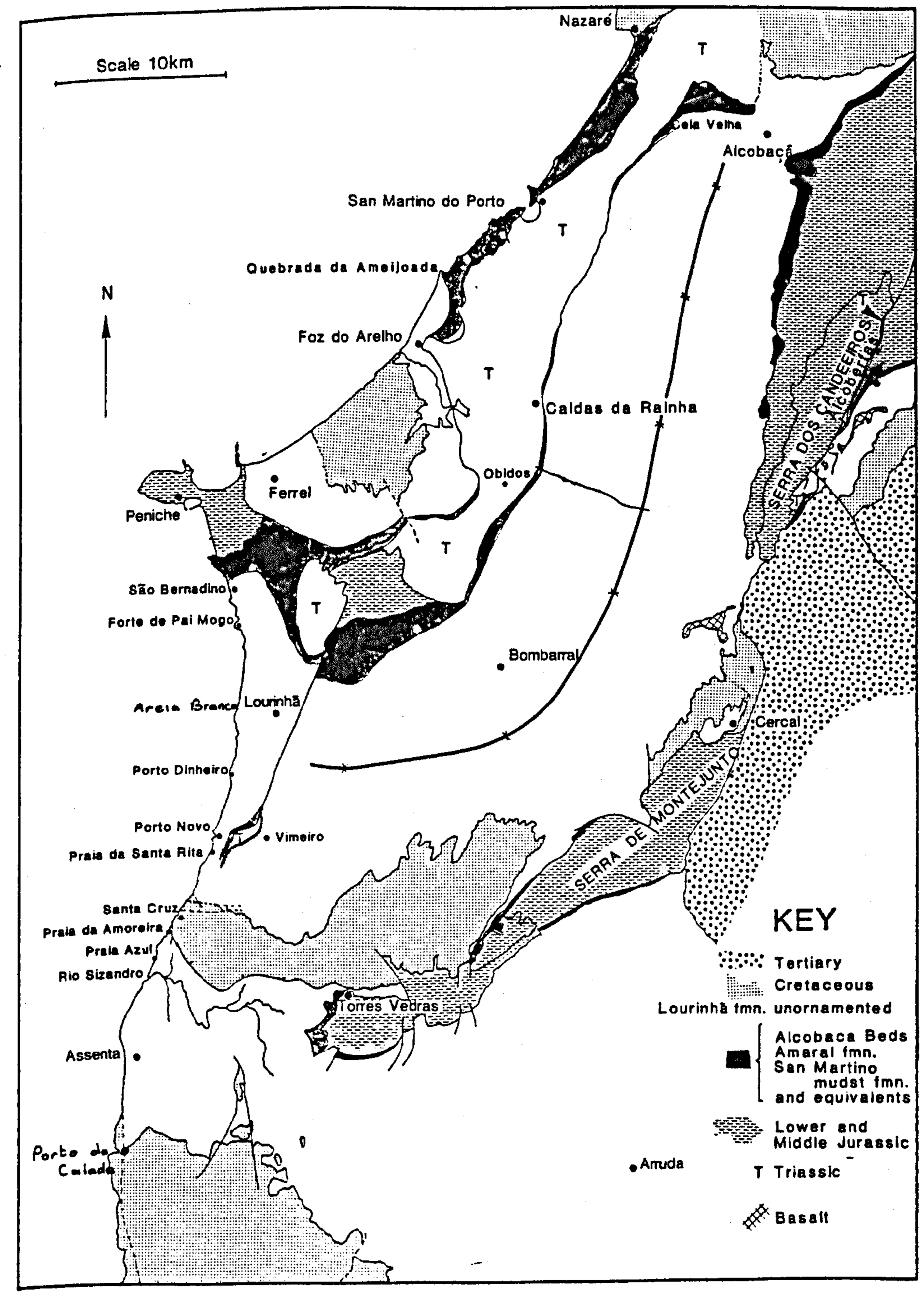
ENCLOSURE 12

R. HISCOTT'S LOGS OF THE ASSENTA MEMBER SECTION ABOVE FAULT


mud
sand
gravel
A
B
Abandoned Channel
Abandoned Channel
~20m
cut bank
r. gn.
gn. r.
0
10
20
30
40
50
60
70
80
90
100
110
120
130

SUMMARY LOGS NOT TO SCALE





ORIENTATION DIAGRAM

LOCATIONS LOGS AND RELATIVE STRATIGRAPHIC POSITION