**Master Thesis in Geosciences**

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UNIVERSITY OF OSLO
FACULTY OF MATHEMATICS AND NATURAL SCIENCES

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Oslo, June 2007

Samuel Ojong-Nsi Etta

# Abstract

This study comprises a geological reservoir characterization of paralic sandstone deposits with a case study of the Upper Jurassic Louriñha Formation in the Lusitanian Basin, which is an Atlantic rift basin located in central west Portugal.

The Louriñha Formation is thought to represent a low relief alluvial system directed from the north-western margin towards the south-east and distributing progradational clastic sediments in the Lusitanian Basin. The terrestrial clastic sediment distribution is suggested to have been triggered by local uplifts in the north and north-west from the granitic Hercynian Basement Horsts that today are exposed as the Berlengas and Ferilhões Islands. Petrographic analysis and palaeocurrent measurements go further to confirm this.

The architectural style, connectivity and heterogeneity of the sandstone bodies deposited in this area can be connected to base level fluctuations, though other minor factors may be present. Petrographic analysis also shows that porosity in the samples collected in the study area are generally poor and not of reservoir quality due to high calcite cementation and mud in the intergranular spaces.

The study is directed towards out-crop analogue approach, where data obtained from outcrops are used to enhance interpretation of comparable subsurface successions. In this study, description of facies, facies association, architectural style and depositional environments are integrated with other factors like heterogeneity in order to carry out computer based modeling of corresponding subsurface sandstone bodies. The above procedures are used as part of a multidisciplinary approach in order to characterize the reservoir.

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# 1 Introduction

Paralic sandstone bodies comprise sandstones from several environments and sub-environments and include fluvial sandstone deposits, estuarine sandstone deposits, delta sandstone deposits, shoreface and strand-plain sandstone deposits and associated mudstone deposits of their respective floodplain. The dimension, geometry, orientation, stacking pattern, sand/mud ratio and heterogeneity as a result of depositional environment and changes in relative sea level may affect sandstone interconnectedness. However, a general overview of paralic sedimentary environments can be read from Appendix A.

Predicting the occurrence, 3D distribution and interconnectedness of these sandstone bodies in the subsurface is usually a difficult task. Another problem is to understand the types and distribution of the facies in the sandstone bodies. In geological reservoir characterisation, the approach of outcrop analogue studies has recently been used as a method for characterisation of subsurface reservoirs, in order to better understand the architecture and internal heterogeneity of sandstone bodies (e.g. Flint and Bryant 1993). Studying the types and distribution of facies, and sandstone geometry, stratigraphic development, distribution and architectural style of outcrops believed to be analogous to the subsurface system might help in solving this problem.

The Upper Jurassic Louriñha Formation in the Lusitanian Basin is regarded as an analogue to the subsurface Statfjord Formation in the North Sea area. Outcrop data could be used to generate reservoir models of sandstone body architecture in the subsurface especially where data are restricted (e.g. Hornung and Aigner, 2002). Hence, collecting field data from the Louriñha Formation may assist in understanding the Statfjord Formation.

There is always a problem with analogue studies in that the data collected from outcrop is two dimensional, but numerical modelling usually requires 3D data sets (Hornung and Aigner, 2002). Hence, the restricted data obtained from the analogue Louriñha Formation to use as input data, makes it necessary to employ modelling techniques.

By using the soft-ware modelling program Petrel ™ different numerical values can stochastically be chosen on modelling parameters like thickness, width, sinuosity and orientation resulting in different geometrical output data of the modelled sinuous fluvial system. The models may be applied in further statistical studies on interconnectedness, fluid connectivity and uncertainties in reservoir management. Such application of modelling results is however, not been defined as part of the present Master project.

# 2 Methods and Data description

## 2.1 Study area

The study area is in the Lourinha formation in the Lusitanian basin, about 70 km NW of Lisboa, Portugal. The field-work was carried out between the periods of 26/09/06-16/10/06 in collaboration with Mona Nyrud. The work was supervised by Johan Petter Nystuen, Michael Heereman, Ivar Midtkandal, and Liv Hege Birkeland. The field area extended from Paimogo to Areia Branca with a total distance of ~ 2180 meters.

## 2.2 Methods and data description

Data collected during the field work which were recorded in as much detail as possible included grain size estimation, sedimentary structures present on bedding surfaces and within beds, the colour of sediments, the width, thickness, geometry and lateral extent of beds or sandstone units, and the nature, relationship and distribution of these sandstone bodies. Furthermore, each of the beds, bedsets, and sandstone body successions were described for thickness, boundary conditions, texture, structure, root and bioturbation traces, and presence of carbonate concretions.

Our task was to determine and describe the facies, facies association, geometry and spatial distribution of the sandstone bodies. Photographs were taken of the entire length of outcrop from Paimogo to Areia Branca and of individual features of interest. The photos were taken from a few distances on adjacent cliffs or at a closer distance if something of particular interest. Such features of interest include: sedimentary structures, distinct channel sandstones and any lithological feature that may assist in the various description phases. The scales of the photo mosaics obtained were determined from a pencil, a metre stick or a 1.6 m measured stick and direct measurements of sedimentary units on the lithology. The thickness variation of each sandstone body was measured using the 1.6 metre stick or the metre stick. The lateral extent was determined using the various GPS measurements.

Detailed sedimentological logs (consisting of mean grain size, colour, sedimentary structures, palaeocurrents and stratal geometry) of the entire succession of the exposed strata in the outcrops were measured at locations which were accessible. Also examined were lateral and vertical continuity and extent of beds and bedsets. Though at some instances assumptions were made where there was no accessibility (e.g. fig. 2.1). Also sedimentological logs were constructed at certain positions along some sandstone bodies that were thought to be described in detail. Our interest was also based on how the sandstone and mudstone bedding units change in depositional architecture across the study area and how their architecture is related to variations in sandstone thickness and spatial distribution.

Representative samples were collected at certain logs and along the major sandstone bodies that is believed to cover the entire lithofacies of the whole section. The samples collected were used for microscopic analysis of texture, composition, porosity, and possible permeability. Sandstone petrology of three representative samples was determined using the point counting method.

However, we encountered some problems while collecting the data. The vertical thickness of the entire stratigraphic column was however, unclear since the deposits are tilted toward south. So constructing a new log, the stratigraphic level at top of the previous log is traced to the bottom of the new log. This process is not always perfect since errors may arise during the tracing process. Quantifying the thickness and lateral extent of the various architectural elements was not always perfect. Difficulties arise in situations where a sandstone body is partially exposed due to their tilting nature. The lateral extent in such a case is based on assumptions of which might be incorrect.

**Fig. 2.1:** North-south orientation of outcrop in study area. Assumptions were made for places that were not accessible e.g. top of the cliff.

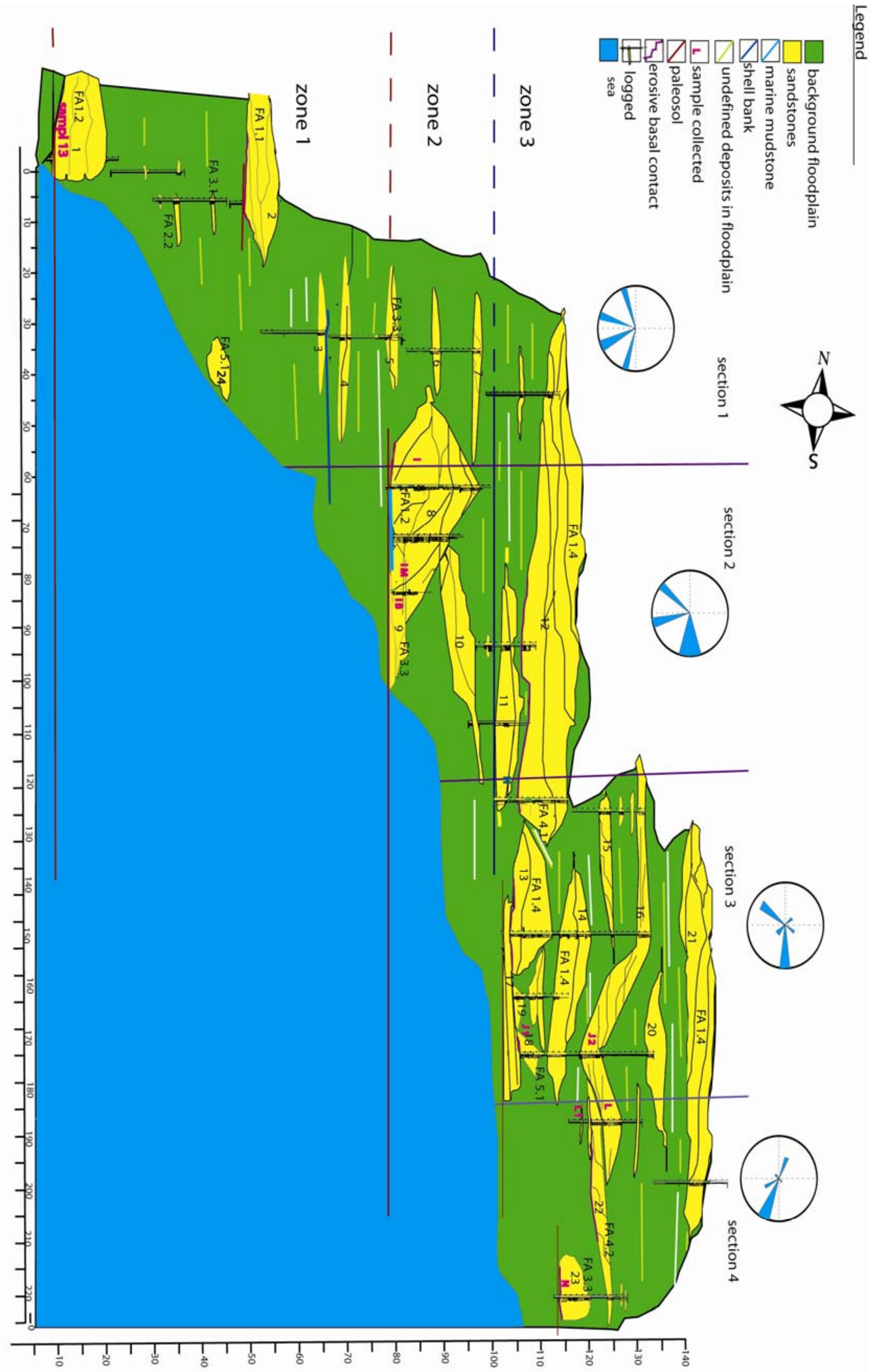



Figure 2.2 : NW - SE oriented panel diagram through the study area. The rose diagrams above represents the mean palaeocurrent direction for each of the section

# 3 Geological framework

### 3.1 Regional setting and structural development

The Lusitanian Basin is a Mesozoic rift basin located along the western Iberian margin, which formed as a response to extension and subsequent opening of the North Atlantic (Guery et al., 1986). The basin extends in length ~350 km north of Lisbon and in width ~150 km, including its offshore extension (Azeredo, 2002). The onshore area of the basin is approximately 23,000 $km^2$ and has a width of ~100 km and a length of 250 km. The Lusitanian Basin is bounded to the east by the Hercynian massif, and to the west the basin is delimited by several basement horsts, which are today exposed on the Berlangas and Ferilhoes islands (Wilson et al., 1989)

The development of the basin is related to three late Triassic – early Cretaceous rift phases that was responsible for the opening of the North Atlantic Ocean, i.e. break-up of Pangea (Rasmussen et al., 1998). Before the rifting commenced the Iberian plate was positioned between the Tethys Ocean and the initial seaway of the Atlantic Ocean and was affected by the rift events that occurred during the opening of the Central and north Atlantic Ocean (Wilson et al., 1989) (figs. 3.1 and 3.2) . The rift phases include (i) late Triassic (Triassic-Hettangian), (ii) Sinemurian- Pliensbachian, and (iii) the Late Oxfordian rift phase(Rasmussen et al., 1998).The depositional model that characterizes the Lusitanian Basin is related to syn-rift to fault-controlled graben/half-graben structures affected by the influence of halokinesis on the depositional environment (Alves et al., 2002)

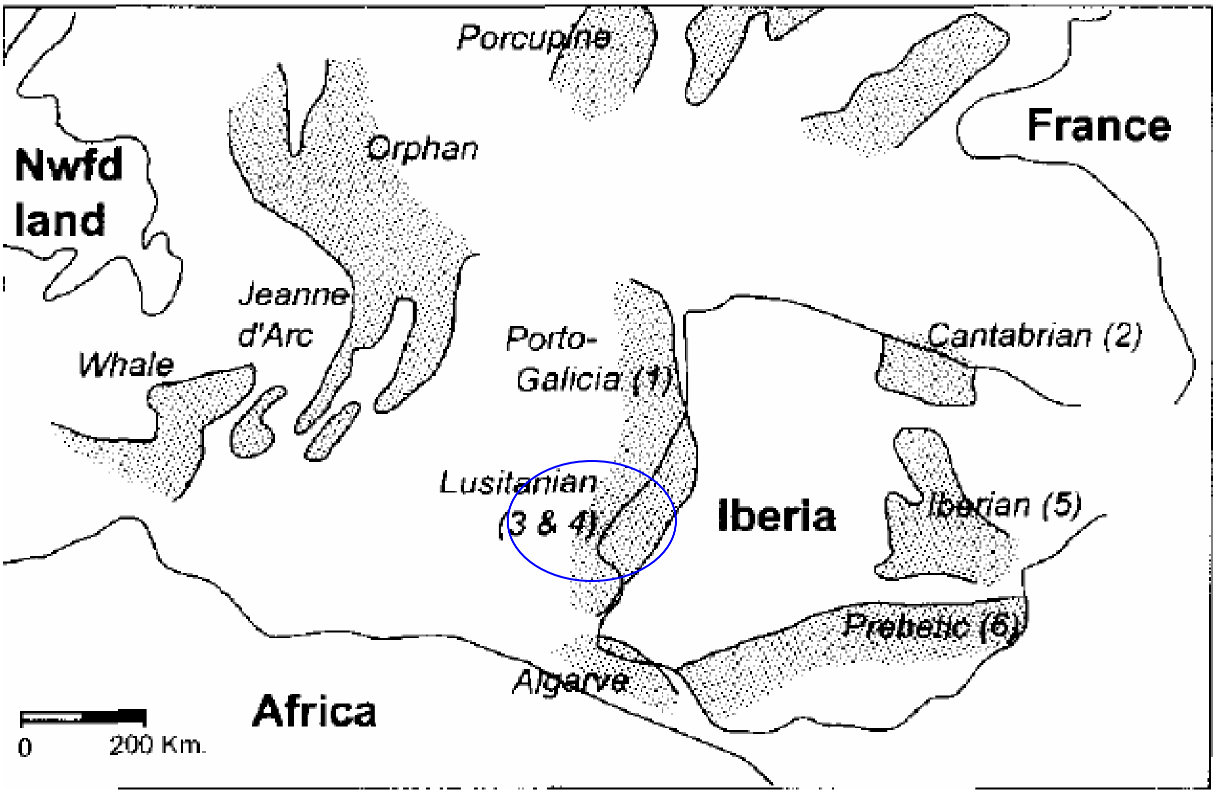



Fig. 3.1: Schematic overview of the rift basins of the Iberian continent and its conjugate margin in pre-drift situation (after Stapel, G et al. 1996) 7

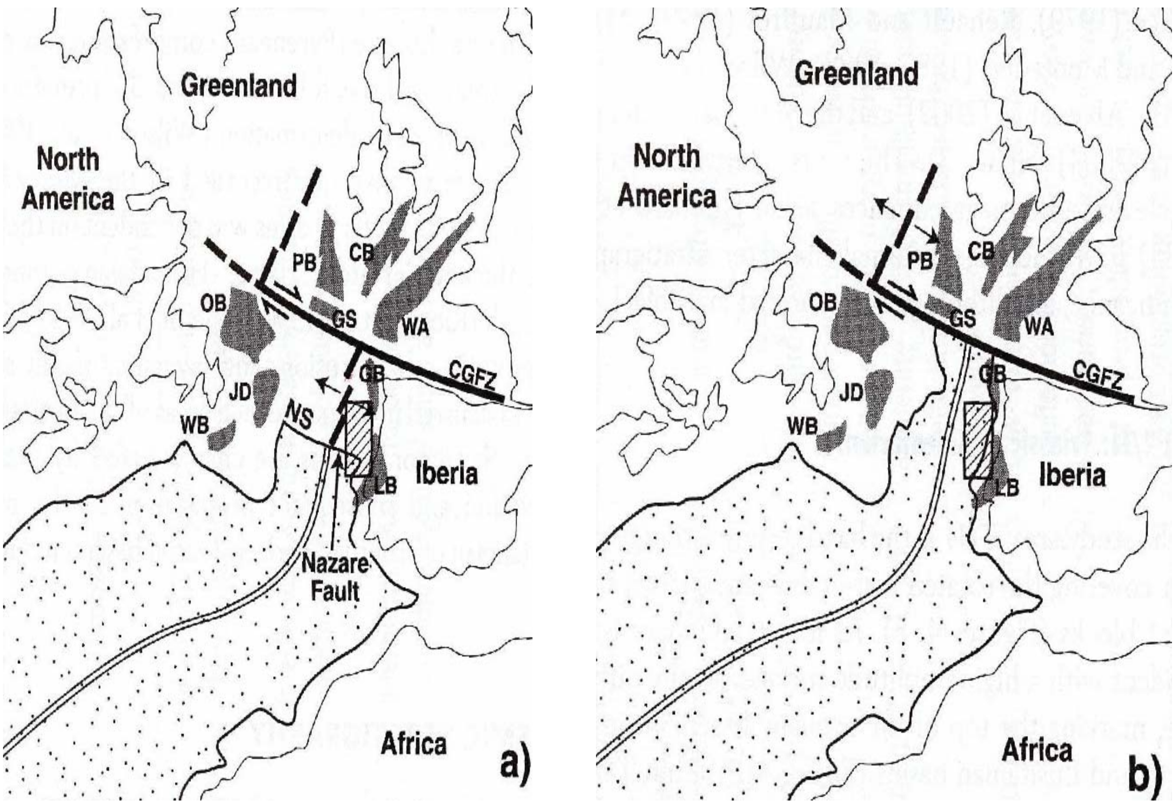



**Fig. 3.2:** Main stages of development of the North Atlantic Ocean during the Late Jurassic and Early Cretaceous period. After Driscoll et al (1995).

The basin can be divided into three sectors, each bounded by major faults. These faults were active during the formation of the basin and are thought of linking to structures formed during the Variscan Orogeny (Wilson et al., 1989). The Nazare Fault limits the northern and central sectors of the Lusitanian Basin and separates the northern, Beira Littoral Trough from the southern Estramadura Trough. Hence, the northern sector is bounded by the Aveiro Fault to the north, by the Nazare Fault to the south. The central sector of the basin is bordered to the south by the Arrife and Tagus-Gargalo Faults (fig. 3.3) The southern sector is bounded by these same faults to the north and to the south by the Arrabida Fault (Carvalho et al., 2005)

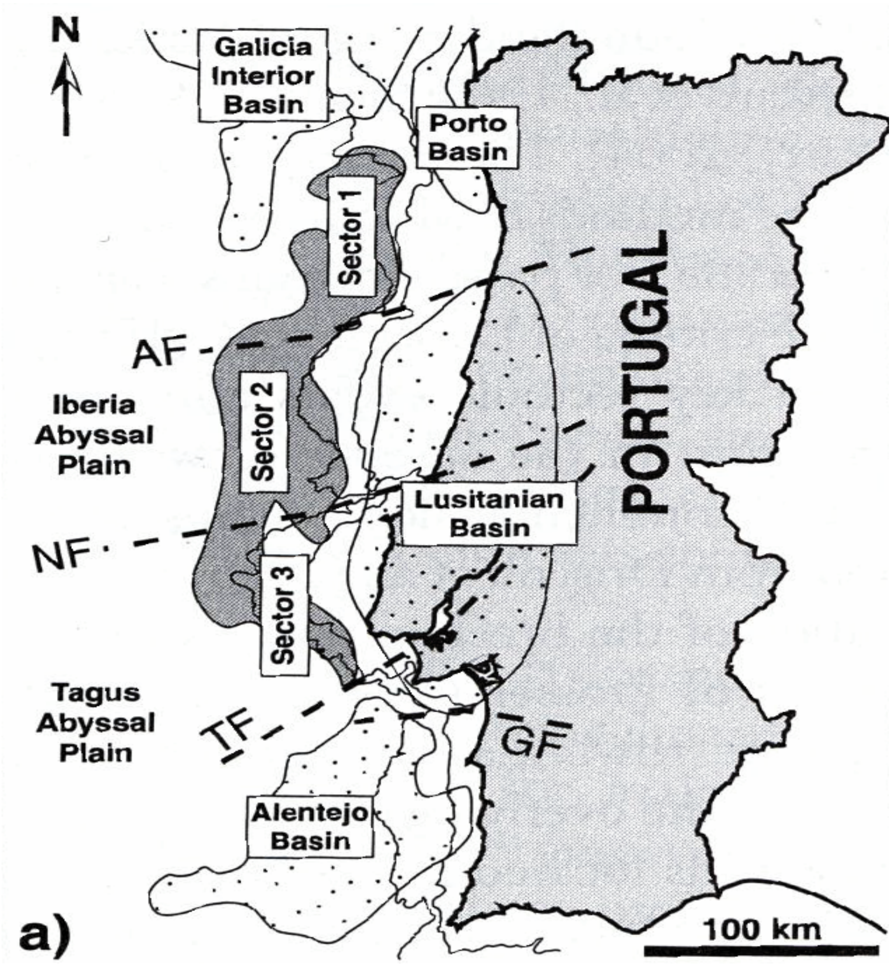



**Fig. 3.3:** Figure shows map of western Iberian continental margin showing the major transfer faults and adjacent offshore basins. The faults are designated AF (Aveiro Fault); NF (Nazare Fault); TF (Tagus Fault) and GF (Gagalo Fault). After Wilson et al. (1989)

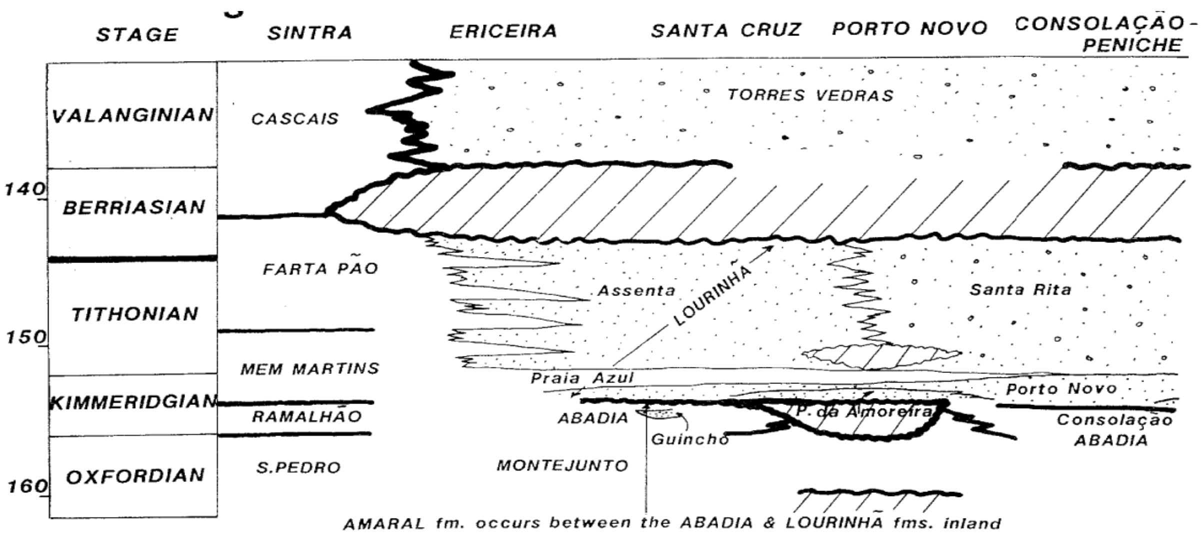
In the upper Jurassic the continuous stretching episodes resulted to the formation of several sub- basins in the Lusitanian basins; and in the south were the Aruda, Bombarral and Tucifal sub-basins. The formation of the sub-basins was probably related to the late Jurassic oceanspreading towards the west in the Tagus Abyssal plain (Wilson et al., 1989). The Aruda and Turcifal sub-basins are half grabens bounded by N-S trending fault blocks, but the Bombarral sub-basin flanked by diapiric structures, basinal subsidence was largely controlled by salt withdrawal (Wilson et al., 1989).

During the Late Oxfordian –Earliest Kimmeridgian rapid subsidence and fault block rotation, resulted to drowning of the sub-basins and a thick marine rift basin infill were deposited (Leinfelder, and Wilson, 1989). In the Aruda sub-basin, the upper part of Kimmeridgian which represented by the Abadia Formation is overlain by fluvial sediments of Kimeridgian-Tithonian age called the Louriñha Formation (Wilson, 1989).

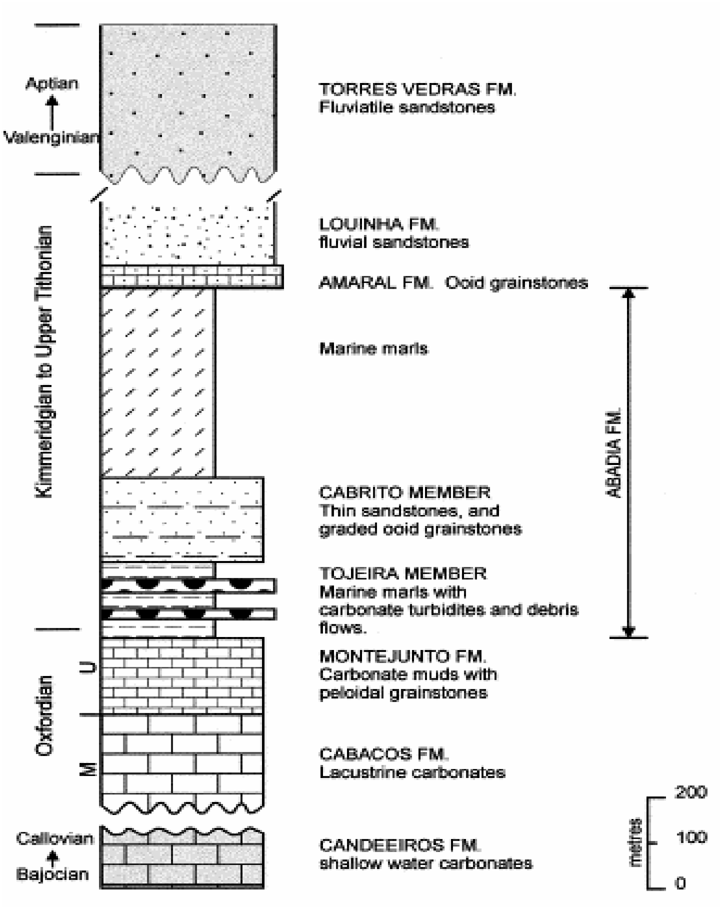
In the Miocene, during the Alpine Orogeny, the central and southern part of the Lusitanian Basin experienced inversion and rifting induced by the orogeny, and the Lower Tagus sub-basin was formed between these two areas (Ribeiro et al., 1996). The Miocene compressional stress was mainly oriented NW-SE shifting to more N-S in the southern part (Rasmussen, E. S. et al., 1998).

### 3.1 Stratigraphy and basin development

As discussed above, the Lusitanian Basin is a marine marginal which started to be established during the opening of the North Atlantic. The stratigraphy discussed in this section is based on Leinfelder and Wilson (1989), (fig.3.4A and B). The Mesozoic basin fill of the Lusitanian Basin can be explained and demonstrated by using the Serra de Montejunto area which exposes sections through five unconformities (Wilson et al., 1989)




(A)




(B)

**Fig. 3.4:** Summary of stratigraphy exposed within the Serra de Muntejunto region based on Leinfelder and Wilson, 1989

Terrestrial red beds, grading into salt and marginal marine carbonate deposits are the first rift sediments of Late Triassic age. These units are bounded by a large sub-aerial unconformity of Late Callovian- Early Oxfordian age. Ramp-like carbonate platforms which occurred at the basin margins and in the shallow parts of the basin centre are of Early to Middle Jurassic age (Leinfelder, 1993). The deposition of the overlying Montejunto Formation of late Oxfordian age on top of the Cabacos Formation marks transition to a fully marine environment. The Montejunto Formation is predominantly made up of fine grained hemipelagic limestones. At the top of the Montejunto Formation, siliciclastic rocks of approximately 800 m in thickness form the Abadia Formation (Curtis, 1998). The Tojeira Member approximately 160 m thick is at the Oxfordian-Kimmeridgian boundary. The Cabrito Member of sandstone and ooid grainstones is approximately 200 m thick and is overlain by 400- 500 m thick grey siltstone, marls and limestones of the Abadia Formation which is 60- 80 m in thickness. On top of the Abadia Formation is the Upper Kimmeridgian Amaral Formation of ooilitic grainstones is

approximately 70 m thick. The dominantly fluvial, siliciclastic Louriñha formation is at the top of the Kimmeridgian to upper Tithonian age. (Leinfelder and Wilson, 1989).
The third succession is formed by the Valanginian to Lower Aptian Torres Vedras Formation which rests unconformably on the Abadia and Montejunto formations. The fourth succession is formed by Almargem Formation of Upper Kimmeridgian age. The firth succession is represented by coarse grained siliciclastic sediments of Oligocene- Miocene age (Wilson et al., 1990).

The Louriñha Formation is made up of four members (fig. 3.4B) and includes, the Praia de Amoreira, Porto Novo, Santa Rita and Assenta members. The Praia de Amoreira Member contains much mudstones with very little channel sandstone deposits (Hill, 1989). Porto Novo Member is dominantly of fluvial meander deposits and the channels are larger relative to those in the Praia Amoreira Member. The Praia Amoreira and the Porto Novo members are thought to have coexisted further east during deposition of distal alluvial fans. The Assenta and Santa Rita Members represent the youngest units. The Assenta Member is fine grained whereas the Santa Rita Member is coarser grained. Development of facies style in the Louriñha Formation is suggested to have been due to diapirism, which might have caused higher slope gradients of the channel system. An increase in relief and slope gradient should have resulted in more slowly subsiding basin thereby promoting river downcutting instead of aggradation (Hill, 1989).

Chapter 4

# 4 Facies

Facies is defined by Turker (2006) as a particular set of sedimentary attributes such as; a characteristic lithology, texture, suite of sedimentary structures, fossil content, colour etc. In the study area fifteen facies were identified based on the definition above and they are summarised in table 4.1.

**Table 4.1:** Summarised description of facies identified in the study area

| Facies | Description | Grain size | Interpretation |
|---|---|---|---|
| **A** | Conglomerates with extra-basinal clasts of possibly basement origin. Matrix supported clasts. Erosional base but upper part may be gradational. Usually interbedded with other facies e.g. facies O | Pebbles in medium to very coarse grained sandstone matrix. Average size of pebble ≤ 1 cm | Basement derived. In fluvial channel with high energy |
| **B** | Intrabasinal conglomerates with mudstone clasts. Matrix supported. Usually occurs at channel base, but often seen scattered in channel sandstone | Medium-very coarse grained mudstone clasts in fine-medium sandstone | Channel lag deposits of shorter transport distances |
| **C** | Trough cross stratified sandstone often associated with facies B. Erosive basal contacts in channel sandstones, upper boundary not usually erosive. | Fine-medium grained sandstone. Facies B may be present in some beds | Migration of 3D-dunes. May be due to bed-load transport |
| **D** | Planar cross stratified sandstone with straight to almost concave foresets | Medium-coarse grained sandstones with both facies A and B | Deposition from migration of 2D-dunes |
| **E** | Massive structureless sandstones, often bioturbated | Fine-coarse grained sandstone | Rapid deposition from suspension within higher energy conditions |
| **F** | Cross stratified sandstones with double and single mud drapes | Fine-coarse grained sandstone. Organic matter of coal is present. | Migration of 3D-dunes within a tidal environment. Estuarine deposits |
| **G** | Plane parallel stratified sandstone. Strata may be horizontal to gently incline | Medium-coarse grained sandstones | Higher flow regime. Bedload traction |
| **H** | Cross laminated sandstone with current and climbing ripples | Very fine-medium grained sandstone | Lower flow regime and unidirectional flow |

| Facies | Description | Grain size | Interpretation |
|---|---|---|---|
| **I** | Plane parallel laminated sandstone. Laminae of mudstone alternating with sandstone in sandstone units | Very fine-medium grained | Deposition from suspension in lower flow regime. Alternating energy conditions |
| **J** | Clay grade mudstone found within sandstone bodies | Mud | Products from deposition in low energy conditions. |
| **K** | Silty mudstone, generally red to greyish brown with obscured bedding | Mud | Sediments deposited during low energy or flooding |
| **L** | Paleosols. Red mudstone often associated with calcrete and root structures | Mud-silt | Sub-aerial exposure for a long period of time. |
| **M** | Soft sediment deformed sandstone consisting of flame structures and convolute lamination | Very fine-medium grained sandstone | Rapid deposition associated with high water content, loading and water escape structures |
| **N** | Shell bank. Siltstone and in some places conglomerates containing shell fragments | Silt-very coarse-grained sandstone. Average size of shell ~2 cm | Indication of brackish environment. Marine flooding surface. |
| **O** | Marine mudstone located below channel sandstone with no identifiable shell fragment | mud | Marine flooding surface |

### 4.1 Conglomerates

Conglomerates form less than 4 % of all rocks encountered in the study area. Their thicknesses are variable, and are not very thick, but generally, bedding is commonly indistinct or absent. More texturally mature conglomerates have well rounded grains and very low matrix as seen in the extra-basinal conglomerates. Immature or intra-basinal conglomerates are poorly sorted with angular and a high matrix content. Matrix supported conglomerates occur in less than 3 % of the study area. Rare clasts up to 6 cm can be found in some conglomeratic units.

Bedding is laterally discontinuous, and is typically found at the bases of sandstone beds which usually grade to sandstones or mudstones .Their top and bottom boundaries are typically irregular or gradational; bottom may be erosive.

### 4.1.1 Facies A: Conglomerate with extrabasinal clasts

*Description*

This conglomerate facies is clasts supported, consisting of very coarse to pebble size clasts, in 50 cm - 1 m thick beds. Individual beds are generally difficult to distinguish from each others. Though, the beds are irregular to lenticular in many occasions. The clasts are sub-angular to round with a diameter range of about 0.5- 2 cm showing evidence of high reworking. Close observation reveals no distinct sedimentary structure. The conglomerate beds are generally poorly sorted. Beds of facies A are recorded at the base (0-3 m) of most of the other facies and do overlie marine mud (facies O) (fig. 4.12). The beds may appear structureless, but initially faint imbrications of clasts may be present in some places.

The lower boundaries of these conglomerate beds are distinctly erosional, whereas the upper boundaries appear to be gradational. Occasionally, reworked mudstone clasts and some few shell fragments can be found between the extrabasinal clasts.

*Interpretation*

The sub angular to almost rounded nature of the extrabasinal clasts, thin bed thickness and clasts supported texture of conglomerate facies A are all together evidence of relatively high reworking and abrasion, and may be due to fluvial channel transportation into their present location. For facies A to overlie marine mudstone suggests regression over this area.

Facies A could have originated from the Hercynian basement rocks in the North West and carried during high discharge.

### 4.1.2 Facies B: Intra-basinal conglomerates

*Description*

Facies B conglomerate beds are apparently reddish brown to grey in colour. They are composed of angular to sub-rounded mudstone clasts in a fine to medium grained matrix. Where the clasts appear grey, they may be termed mud clasts because of their relative softness. The clasts range in size from 0.5 cm to about 3 cm, but average is about 1 cm. The clasts are poorly sorted. Some of these conglomerate beds occur in association with calcretes horizons and conglomerate beds with extra-basinal clasts (facies A).

Bedding is commonly indistinct or absent. However, thickness of less than 1m is commonly recorded in most of the intra-basinal conglomerates encountered. However, bed thickness seems to vary systematically through a very short distance. The facies B conglomerate beds are often in association with sandstone beds. In some beds clasts are arranged in 1–1.5 cm thick layers, though unevenly spaced with a distinct cross stratification having a dip of about 15 degrees.

In some places, as also with conglomerates with extra-basinal clasts, coalified wood fragments and root structures are found within the matrix of facies B. The lower contacts of facies B beds are usually erosive, and in some places ball and pillow structures could be identified.

**Fig. 4.1:** Conglomerate with extrabasinal clasts consisting of subrounded clasts in siltstone matrix. Usually occurs at base of channel sandstone erosive to marine mudstone below. Pencil 15 cm

**Fig**. **4.2:** Grey to brown intrabasinal conglomerate with approximately horizontal bedding. Very coarse grain to pebbly in parts. Pencil 15 cm

*Interpretation:*

The conglomerates beds of facies B being associated with coarse grained sandstone, having limited parallel stratification, reworked wood and coal fragments, together with the presence of marine fauna in some portions, are suggested to have been deposited in an estuarine fluvial dominated channel.

The crudely bedded conglomerate beds represent lag deposits related to migration of the channel thalweg and dunes.

Generally, when gravel is seen above an erosional surface (log-FW-N), may signify a sharp change in the energy in that environment. The rip up clasts (composed of claystone or mudstone clasts) at the base of most channels sandstones were probably derived from the erosion of thin mud layers from the underlying mudstone unit. Calcrete and associated mudstone may have been derived from underlying paleosol horizons.

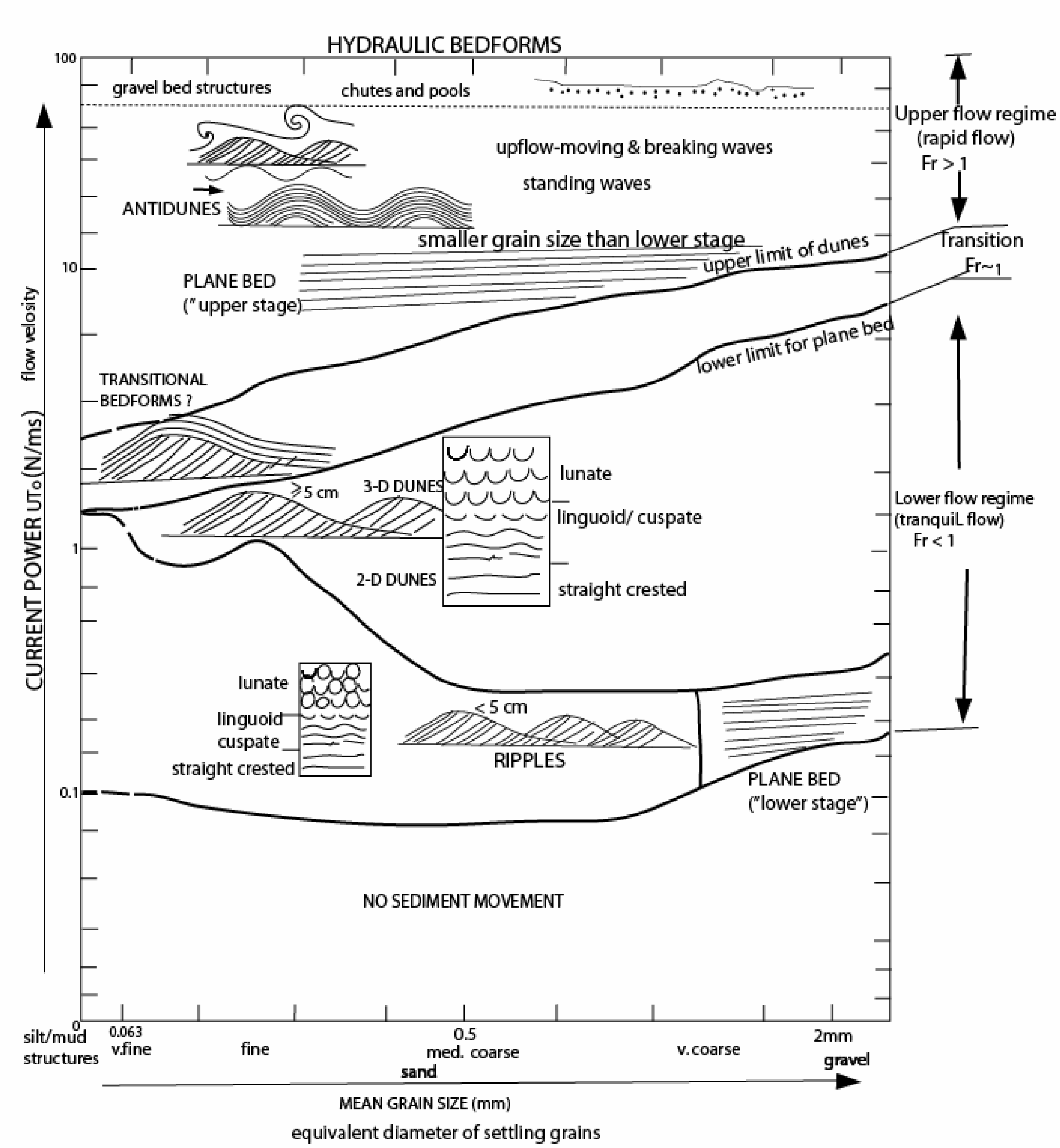



**Fig. 4.3:** Sedimentary structure matrix showing stability fields for subaqueous bedforms in relation to specific velocity/ grain size conditions. Modified from Allen (1968). Variation in flow depth might change the position of the partitions in the various fields.

## 4.2 Sandstones

Sandstones form the second dominant of all lithologies found in the study area, making up at least 30 – 40 %. The sandstones vary from very fine to very coarse grained and from very poor to very well sorted and locally contain scattered mud-clasts. The finer grained sandstones are most commonly horizontal laminated with parallel bedding. Bedding thickness very variable, from thinnest beds and laminae to very thick bedded massive sandstones of less than 3m thick. The beds attain different forms, from tabular to lenticular and their shape is usually determined by their bed form

The lower contacts of the sandstones are usually sharp and planar except for some local, erosional relief into underlying mudstones. The upper boundaries, are usually horizontal to convex up, and may be followed laterally for a few tens of meters. The transverse section are exposed in some while some are embedded in mud or covered by grass.

### 4.2.1 Facies C: Trough cross stratified sandstones

*Description*

Sandstone beds with large and small scale trough cross bedding seem to be more common than other types of facies. Colour ranges from grey to reddish brown. This facies is found as channel sandstone units but few beds of this facies also occur in smaller sandstone bodies, such as crevasse splay or crevasse channels.

The size of individual troughs varies greatly in height; the range in thickness of trough units is 20- 60 cm. The trough length is also variable; some usually die out before the end of the bed, while some troughs traverse from bottom to top of the bed, then representing foreset lengths. But in most of the beds only the foresets of dunes are preserved. The bed thickness ranges from 0.4- 2.8 m, and bed set may reach up to 4.5 m. Grain size ranges from fine to medium grained sand with an erosive base in some occasions. The cross stratification in facies C is basically made by abrupt grain size variation in adjacent strata. The sandstone beds are in general are organically poor, as opposed to facies F where the cross stratification is expressed by mud or enrichment in organic matter.

Angular to sub-rounded mudstone clasts of various size (less than 1.5 cm) are abundant throughout the bed-set in some places (see log FW-B). Mudstone clasts even occur in some foresets. Burrows are mostly vertically oriented, but at times inclined. The sandstone bodies in which burrows occur may run laterally for long distances, beyond which they may be difficult to trace.

The trough cross stratified sandstones beds are usually interbedded with parallel laminated sand (facies I), planar cross stratified sandstones (facies D), mud drapes or drapes of coalified plant debris in cross stratified sandstone (facies F). Large scale troughs cross stratified strata as well as planar cross strata are abundant in the lower part of thick sandstone bodies; whereas small scale cross strata are characteristic of the upper part. Small scale current ripples can also be noticed on the stoss sides of dunes structures. Centimetres to decimetres thick muddy beds commonly are scattered within the sandstone bodies, being reminiscence of originally continuous beds, thus representing surfaces of amalgamation of adjacent sandstone beds. Palaeocurrent measurements taken from cross stratified sets show SE dominating directions, but orientation towards the NW and S were also common. In some places, overturned cross strata are common, but generally in thicker beds (fig. 5.4).

*Interpretation*

The trough cross stratified sandstone facies was formed due to the migration of 3D-dunes. Migration of curved crested ripples and dunes result to trough shaped erosion surfaces upon which new sets of cross strata accumulate, all together resulting in trough cross stratification. 3D-dunes are normally formed during flow velocities in the upper part of the lower flow regime (Harms et al. 1975) in channelised deposits, like trunk river channels and crevasse channels, and also in proximal crevasse splays.

The mud clasts associated with the trough cross stratified beds were deposited during one or several slack water periods. The angular to sub angular nature could be a result of early semi consolidation of the cohesive clay rich mud from partial dewatering prior to rip-up and reworking. 2

The overturned cross-bedding forms (fig. 4.4) as dune foresets collapse in a down flow direction, generally as a result of over-steepening during rapid deposition and frictional drag from high energy flows above (Allen, 1984). Thus, the overturned cross strata mark high bed shear stresses, suggesting high current shear strength (Owen, 1996).

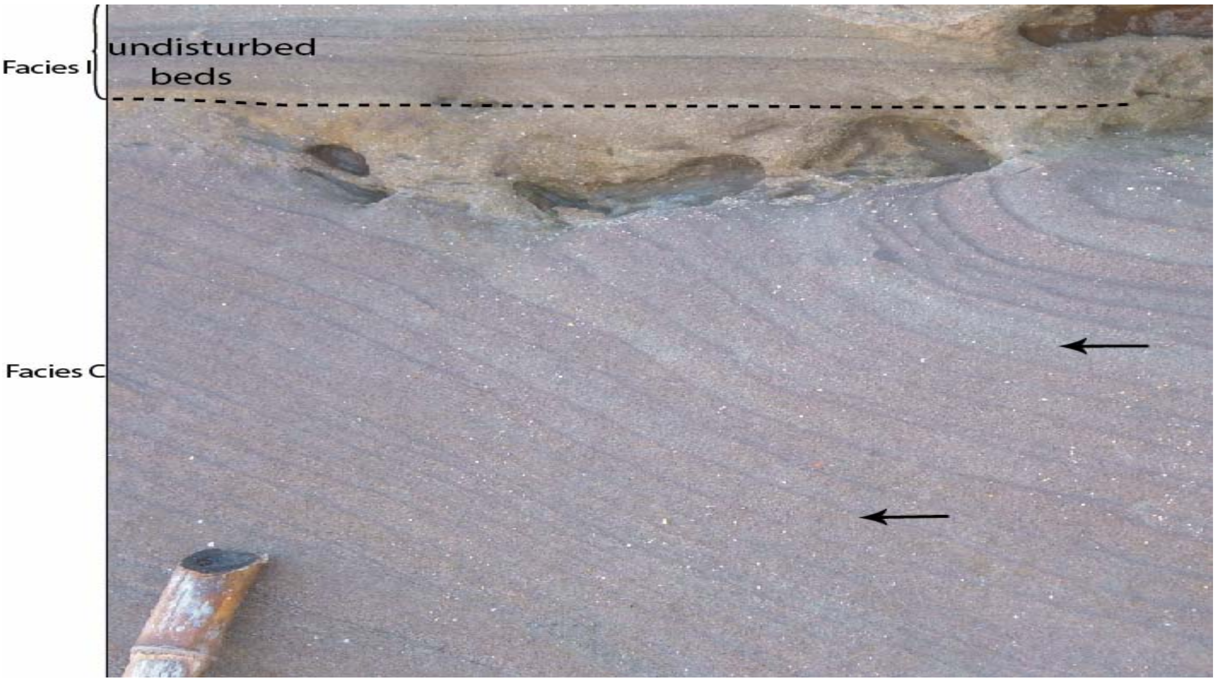



**Fig. 4.4:** Trough cross stratified sandstone showing deformed bedding, black arrows indicate direction of flow. Length of stick 10 cm

### 4.2.2 Facies D: Planar cross stratified sandstone

*Description*

This facies occurs in grey to reddish brown sandstone beds. The sandstone is poorly sorted and is planar stratified, or may appear to be low angle trough stratified sandstones. The facies is recorded in few locations. The individual beds are usually bounded by horizontal or slightly down sloping beds which vary in thickness from 20 - 90 cm, but dune height is hardly up to 1 m. Foresets within the beds are straight to slightly concave upward, toeset is either planar or slightly tangential. Inclinations of the foreset are low and may be less than 15 degrees and maximum azimuth orientation is to SE.

The grain size is dominantly medium to coarse grained and they generally grade down into matrix of intra-basinal conglomerate (facies B).

The lower surface of most of the bedset is non erosional, but in some places the lower boundaries are planar. The beds are laterally not extensive and usually interfere with trough cross stratified sandstones (facies C).

*Interpretation*

Planar cross stratified sandstone could be the result of the migration of 2D-dunes having straight crests and occur at a transitional stage between ripples and 3D-dunes (fig. 4.3) 2D-dunes are thought to represent a transitional stage between ripples and 3D-dunes. When the 2D dunes migrate sand is transported towards the stoss side by traction currents and subsequently deposited at the crest. At the crest the bed shear stress diminishes indicating resulting to avalanching of sand down the foresets.

Facies D sand dune deposits may also be related to the middle or later flood stage of shallow or ephemeral streams; flood stage that is commonly characterized by straight crested dunes, and possible weak to moderate flow causing grain flow on the lee face, hence angular foresets Picard et al. (1973). There are situations where the 3D-dunes may have low sinuous crests and it may be confused for a 2D-dune.

Palaeocurrent displaying bimodality is not frequently found, but the few recorded in this facies may be due to tidal effects.

### 4.2.3 Facies E: Massive sandstone

*Description*

These sandstone units are commonly reddish brown but occasionally grey or mottled. The sandstone bodies are characterised by very few or no visible sedimentary structures. Individual beds are difficult to distinguish, but in one or two locations few layers with crude bedding can be identified. Thicknesses of beds are of decimetre scale and seldom exceed 80 cm. The sorting is poor, grain size ranges from fine –medium, but some coarse grained beds were identified.

The lateral extent was difficult to determine because the beds gradually grade into mudstone, or are being covered by scree material. However, the beds appear forming lenticular geometries, but sheet-like portions of few meters do occur. The sandstones were locally bioturbated and associated with root structures. Facies E is usually found with facies C, D and F then usually located at the base of the other associated facies, though some can be found at higher levels. The lower and upper boundaries are usually conformable or gradational to mudstone.

*Interpretation*

Sandstone beds lacking any visible sedimentary structures may be the product of rapid deposition in a channel, or original stratification may have been obscured by subsequent slumping and dewatering (Nichols et al., 2002). Leclair (2003) reported that deposition of Structureless sandstones lacking evidence of traction and/or dewatering structures is the result of extremely high rates of sediment fallout and bed aggradations associated with a submerged hydraulic jump

### 4.2.4 Facies F: Cross stratified sandstone with mud drape

*Description*

This facies characterises the bulk of section 2, 3 and 4 of the study area. Stratification of the beds is well marked by the presence of single and double mud drapes (forming a couplet in most cases). The sandstone beds are trough cross stratified with mud laminae within bedding planes.

The single mud drapes (fig.4.5) occur in large scale cross bedded units with mud draped bottomsets and foresets. The drapes are composed of mud and sometimes reworked coalified plant fragments. Thickness of individual foreset laminae may range between 0.9 mm and 2 mm. The bottomsets may reach the lowest part of the sandstone beds and truncates directly into mud. The foresets terminates to a near tangential upper lying horizontal to near inclined surface. Bed thickness may be up to a meter.

In case of the double mud drapes, couplets are formed. The paired drapes occur in association with the single drapes and the couplet thickness is almost twice the single drape thickness. The double mud drapes are generally concave upward, and are often seen bundled. The sand is fine to medium grained with some coarse grade. The drapes are composed of reworked coal debris or organic matter, same as for single mud drapes. Range in thickness is between 1mm-3 mm and set thickness is about 13 cm in some. The spacing between groups of drape foreset changes in a haphazard manner along the cross bedding set, varying from 0.1 - 1 cm.


**Fig. 4.5:** Medium scale cross stratified sandstone with tangential, concave up stratification composed of mud draped. Note the well developed bottomsets and the presence of a reactivation surface (blue arrow) as a result of frequent changes in flow direction. Direction of flow from right to left. Finger 9 cm

Convex upward reactivation surfaces are also common in some units of facies F. the surfaces truncate the foresets found on the upper part in most sets, thus forming an erosional surface within the bed set. Small ripples are seen in some units, especially at the lower part of the foresets. Palaeocurrents data measured from most of the drapes are bidirectional

*Interpretation*

The concave up cross stratified sandstones units with mud drapes are formed by the migration of 3D-dunes. The double mud drapes are interpreted to be deposited during first and second slack water stage in a sub-tidal setting and are known to be of the structures which alone represent tidal influence (Allen et al., 1984). The mud drape couplets prove that the environment was subtidal. But Visser (1971) earlier said that mud draped foresets and muddy interbeds suggest slack water interludes, possibly caused by tidal processes.

The lateral bundle successions are arranged in a thickening and thinning cyclic pattern that could be a record of spring and neap tides (fig.4.5). Internal reactivation surfaces and erosional planes bounding each bed set were originated by frequent changes in direction or velocity of flow.

Orientation of inclined reactivation surfaces showing top-lap truncation of foresets indicates the dominant bar migration direction. In tidal deposits, reactivation surfaces have been attributed to reversals in the flow direction such that a bedform`s lee side is changed into the stoss side resulting in substantial bedform modification (Allen et al, 1984).

Generally, double mud drapes are scarce and not very clear relative to the single drapes, they might have been destroyed by the waning effects of currents.

### 4.2.5 Facies G: Plane parallel stratified sandstone

*Description*

This facies is reddish brown or grey plane parallel stratified sandstone. The grain size ranges from medium to coarse grain. The sandstone consists of thin, almost evenly spaced strata. Each stratum, about 0.2 cm -0.6 cm thick occurs in sets 10-15 cm thick in beds up to1 m thick. The strata are parallel to sub-parallel with each other, making the entire set to be approximately horizontal. Where the stratification does not occur in sets the bed thickness may range from 0.4 m- 2 m, in average it may be 1.5m.

Individual stratified beds could be laterally continuous for distances more than 15 m. In the study area, beds of facies G stratigraphically underlies parallel laminated beds (facies I). Some of the beds occur with horizontal and discontinuous stratification. Strata thickness is thicker in horizontal discontinuous than in horizontal parallel bedding. The discontinuous strata occur frequently as inclined beds, and in some places the strata pinch out at distances between 3 – 7 m, forming lensoid structures.

In few occasions, beds of facies G usually have erosional lower contacts but their upper contacts are more conformable to finer sand and occasionally grades to mud.

*Interpretation*

Sandstone beds with horizontal parallel stratification are present in most channel deposits. They are indicative of high flow velocity and conditions within the lower part of the upper flow regime. Miall (1996) proposed that they represent the upper plane bed conditions, at the transition from subcritical flow (fig. 4.3)

This might be proximal deposits or channel fill of the fluvial channel into an estuary.
Those that form in coarse grain sand could have taken place under high velocity conditions, but may also occur under lower velocities in shallow waters.

### 4.2.6 Facies H: Cross laminated sandstone with ripple

*Description*

This sandstone facies is dominantly grey but occasionally brown and is characterised by current ripple cross lamination and climbing ripple cross lamination (fig. 4.7). The grain size is very fine to medium grained

This facies occur in relatively thin beds; beds ranging in size between 0.1 m to 0.8 m with a mean value of about 0.6 m. The ripples are difficult to recognize since only the crests and foresets are preserved and do superpose other facies. The ripples are generally asymmetric with sets which are very thin, rarely exceeding a few centimetres; however, thick sets do occur.

Bed set of this facies usually have conformable lower and upper boundaries and gradually grade into either plane parallel laminated (facies I) or plane parallel stratified bed (Facies G). Some beds are locally burrowed and are composed of angular mudstones clasts, carbonate concretions and scarcely reworked wood fragments.

**Fig. 4.6** A millimetre to centimetre-scale rhythmic plane parallel laminated sandstone. Note the occurrence of the laminae into bundles ~ 2 cm thick. Width of view ~3 m.

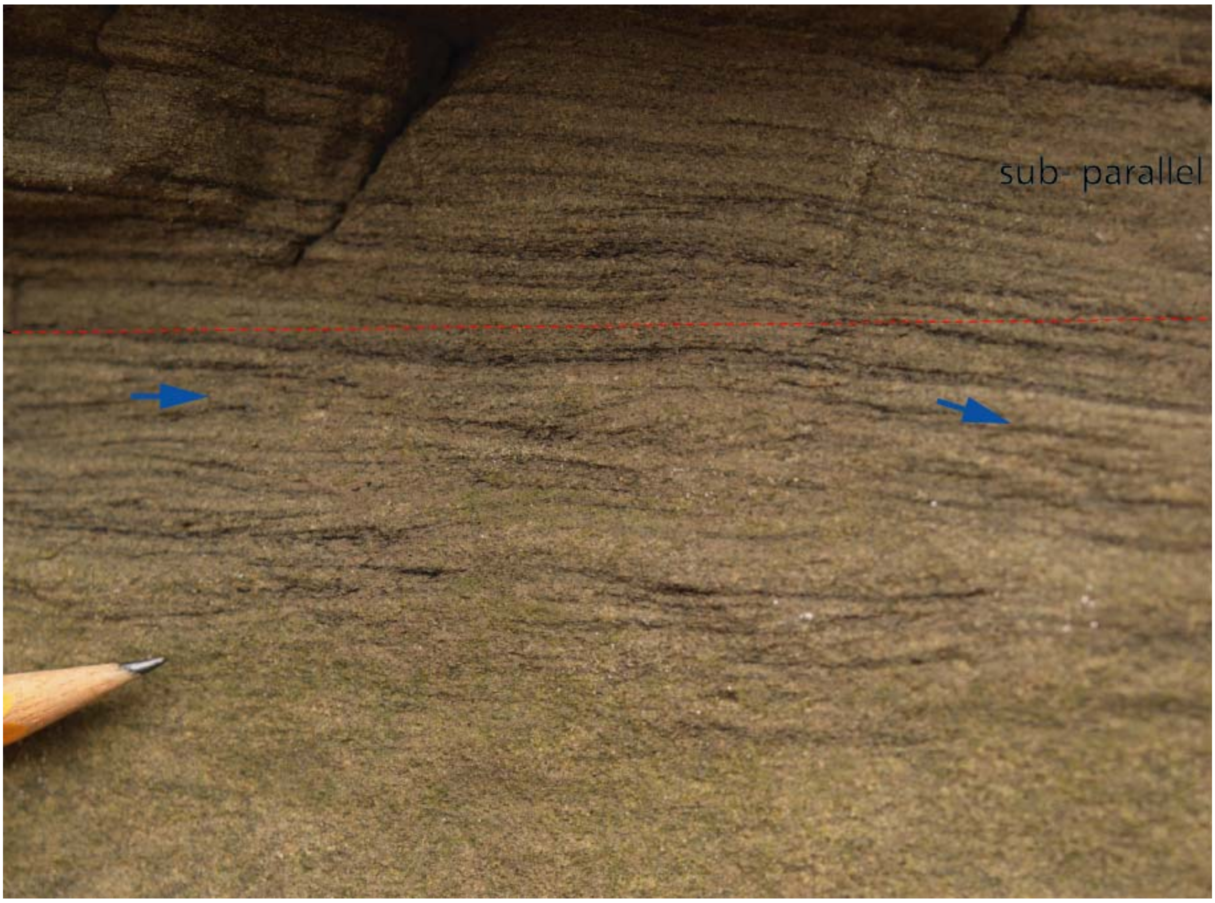



**Fig. 4.7:** Fine to medium grain cross laminated sandstone. Upper part is parallel to sub-parallel laminated while lower part is ripple laminated(blue arrows)

*Interpretation*

Ripples are formed as a result of traction in the lower flow regime and at relatively lower flow velocities compared to dune formation (Boggs, 2001). Ripple cross lamination is formed as a result of migration of ripples coupled with net sediment supply (fig.4.3). When there is no sediment influx, ripples may migrate down stream, and preservation occurs only when movement stop and the ripples get buried. The formation of the ripple structures described above is thought to have been formed in the lower flow regime under unidirectional flow, since the ripples have asymmetric pattern.

Climbing ripple lamination is considered to result from rapid deposition in areas of high sediment input. The presence of climbing ripples and the absence of biogenic structures go further to confirm this assertion. The presence of mudstone clasts, carbonate concretions, and reworked wood suggests temporal changes with decreasing current strength leading to their deposition.

### 4.2.7 Facies I: Plane parallel laminated sandstone

*Description*

This facies is red brown to grey planar laminated sandstone consisting of very fine to medium sand. The sandstone beds generally get finer upwards. Each of the laminae can range in thickness from 0.05 cm – 0.3cm and are made of laminae of mud or reworked coalified plant remains/organic matter intercalated with sandstone laminae. In some localities the laminae are made of grain size contrasts. The laminae set are not equally spaced, spacing about 2mm and seem to occur in sets more than 2 cm thick intervals. The relief is formed by almost perfectly horizontal layers. Some of the horizontal laminae do exhibit parting lineation. They can be found within bedsets of moderate lateral extent, of close to 20 m long but some sections are usually covered by mud.

Beds of this facies gradationally overly beds of facies G in most locations, but there are instances where one can find the plane parallel laminae between parallel stratified (facies G) beds. Locally thin mud films do cover some of the laminae. Like in parallel stratified sandstone, parallel laminated sandstones also possess inclined and discontinuous laminae. Such laminae can be found within bedsets of moderate lateral extent of up to 20 m long but some sections are usually covered by mud.

The lower parts of some facies I sandstone beds do contain mud rip-up clasts. The beds also display soft sediment deformation (facies J) initiated by loading and fluid escape. The lower boundary is usually towards underlying mudstone, but where the lower boundary could be observed, the contact is usually disturbed with erosive bases, but sharp lithological contacts without evident erosion also exist. The upper boundary is gradational and irregular.

*Interpretation*

This type of lamination in fine-grained sandstone is common in low energy environments and usually downstream; where deposition by suspension settling predominates. Alternation of parallel laminae (facies I) and parallel stratified beds (facies G) suggest that the sand has been deposited on tidal flat due to fluctuations in energy levels and sediment supply during tidal cycles. Hence, the parallel laminae of mud origin are deposited from suspension during the quiet periods of the tidal cycle and limited clastic input.

Sand deposit of this facies could also form by traction in the upper flow regime at transition from medium flow velocities to very high velocities (Miall, 1996). Under such conditions ripples and dunes are usually destroyed (fig. 4.3).

## 4.3 Mudstone

Mudstone deposits make up about 55 – 60 % of all lithologies found in the study area, making them the most common of the fine grained facies. They may be grey, brown, red or mottled and locally show small scale rooting and oxidized iron concretions. However, because of their grain size, and their subjectivity to weathering and scree falls, they present more challenge in the field in terms of detail logging of internal stratigraphy and lithological variation. In this study, mudstones are used for the very fine grained siliciclastic rocks, composed mainly of particles less than 63μm.

### 4.3.1 Facies J: Clayey mudstone

*Description*

This facies is not very prominent in the study area and are red to grey and even greyish brown or colour mottling with high clay content. Bed thickness may range from 5 cm to more than 20 cm, but less than 1 m and usually occur within sandstone bodies or as clay plugs above abandoned channels. The primary sedimentary structure is not very visible but faint parallel lamination to structureless can be seen.

This facies is laterally persistent within sandstone bodies, though thickness may vary from one section to another and in some occasions may be truncated by sandstone of trough cross stratified sandstones facies C, planar cross stratified sandstone (facies D) and massive sandstones (facies E). In few instances, facies J could be very short of less than 15 cm, in this case classified as amalgamated surfaces which often show lensoid structures. The lower and upper contact surfaces may be erosive or planar. Bioturbation is uncommon, but root structures and fragments of reworked coalified detritus can be observed. In a few places scouring and soft sediment deformed structures are present.

*Interpretation*

Facies J is products of deposition from very low energy currents. Miall (1996) suggested that mud may have been deposited as clay drapes on top of channel levees, in standing water pools during channel abandonment or water lowstand, or be deposited from suspension in a lacustrine setting. Clay rich mud drapes occur in sandy bedforms, but are more prominent within siltstone packages. In this case, the waters of the abandoned channel may become static, and gradually fills with clay.

The occurrence of facies J in (fig.4.8) may be due to distinct depositional events where the first event must have deposited the one below, followed by a static state for the mud to be deposited and then later another flow stage to deposit the sandstone body above. This is characteristic to ephemeral or crevasse channels where flow is episodic.

### 5.3.2 Facies K: silty mudstone

*Description*

This facies is generally red to greyish brown, but grey to mottled mudstone of clay to silt size and massive mudstone do exist in the logged sections. Mudstone ranges in thickness from a few centimetres to about 35 m. The primary bedding seems to be obscured by loose rocks or is weathered by running water. The main characteristics of this facies are the presence of plant roots, foot imprints and extensive burrows, which range from small excavations dug by invertebrates to relatively large burrows dug by vertebrates. Desiccation cracks were noticed immediately above channel sandstones. Weathered carbonate nodules and colour mottling occur frequently. A good example can be seen from (fig. 6.4). The mud generally grade to silt or very fine sand, especially closer to channel sandstone bodies.

The lateral extent of units of this facies is difficult to determine since silty mudstone frequently interchange with intervals of channel sandstone bodies and/or crevasse deposits; however, generally the depositional units are sheet-like. Some units have irregular, non-erosional basal contacts and horizontal to sub-horizontal sharp upper contacts. A relatively high degree of oxidation with formation of iron oxides can be associated with this facies.

*Interpretation*

The deposition of this facies might have taken place mainly from suspension under low energy conditions, in the low-lying floodplain areas such as in ponds or very shallow lakes. This may happen when the flood water spilled from the main channel onto the nearby low-lying areas. The bedload deposits are suggested to be abandoned lag deposits or channel margin deposits.

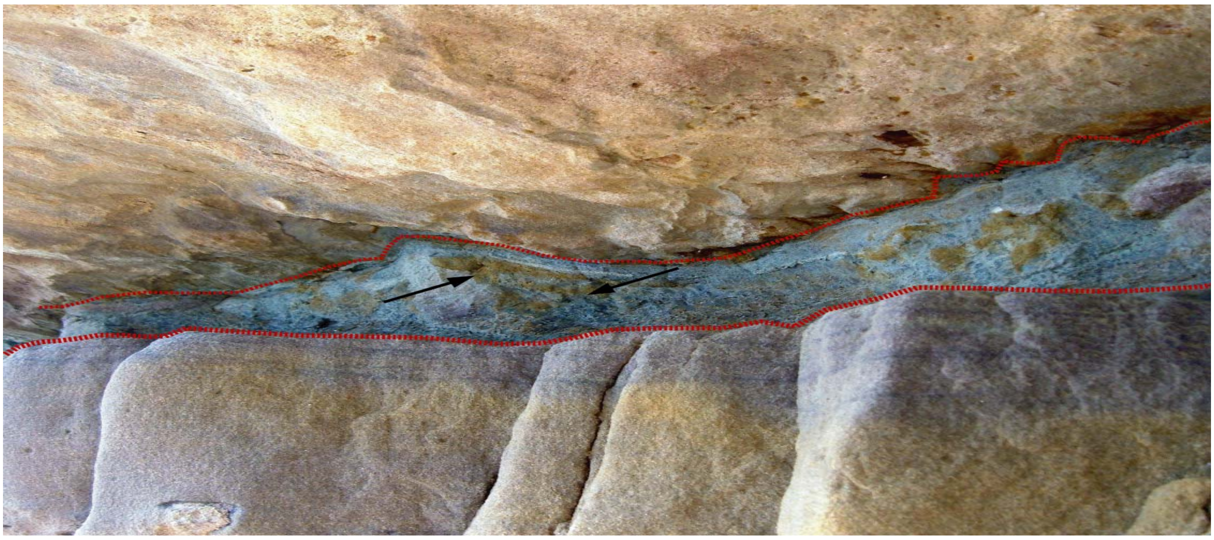


Fig.4.8: Interbed of grey-brown clayey mudstone (dashed red lines) within channel sandstone. The mottle colouration may be due either to oxidation or reduction.

Pedogenic processes, like plant growth and the activities of micro-organisms disturb observation of sedimentary structures in mudstones. Variation in colour may be due either to the action of oxidation/reduction. Mottling is suggested to be due to the action of organisms and especially bioturbation combined with oxidation and reduction.

### 4.3.3 Facies L: Paleosol

*Description*

Paleosol typically show colour mottling and are composed of red, brown, yellow mud and sometimes silty sand with a blocky texture and are usually associated with calcrete and root structures. Paleosol horizons are generally found below channel sandstones and range in thickness from a few centimetres to about a metre. When measured from ground level, the paleosol units are generally buried and true thickness obscured but some occur as continuation of mudstone or flood fines forming a gradational lower contact and a sharp planar or sharp irregular upper contact with the channel sandstone.

This facies can be very extensive, forming sheet-like structures and in some areas may disappear or stop abruptly. In the upper part of units of paleosol facies facies M, calcrete nodules are blockier in shape, and with mudstone clasts may be frequently present and in lower part the units occur as bands alternating with silt grade mudstone (fig.4.9). Some calcrete nodules occur as fine grained infill of cracks.

*Interpretation*

Paleosol is a lithological feature formed by soil-forming processes and not depositional, resulting from long term sub-aerial exposure of greater 10.000 years (e.g. Bridge 2003). The development of calcrete requires a considerable time span for full paleosol maturation (in the range of ten thousand years according to Retallack (1988); dependent on temperature and mineral grain size, but the most important factor is low sediment input.

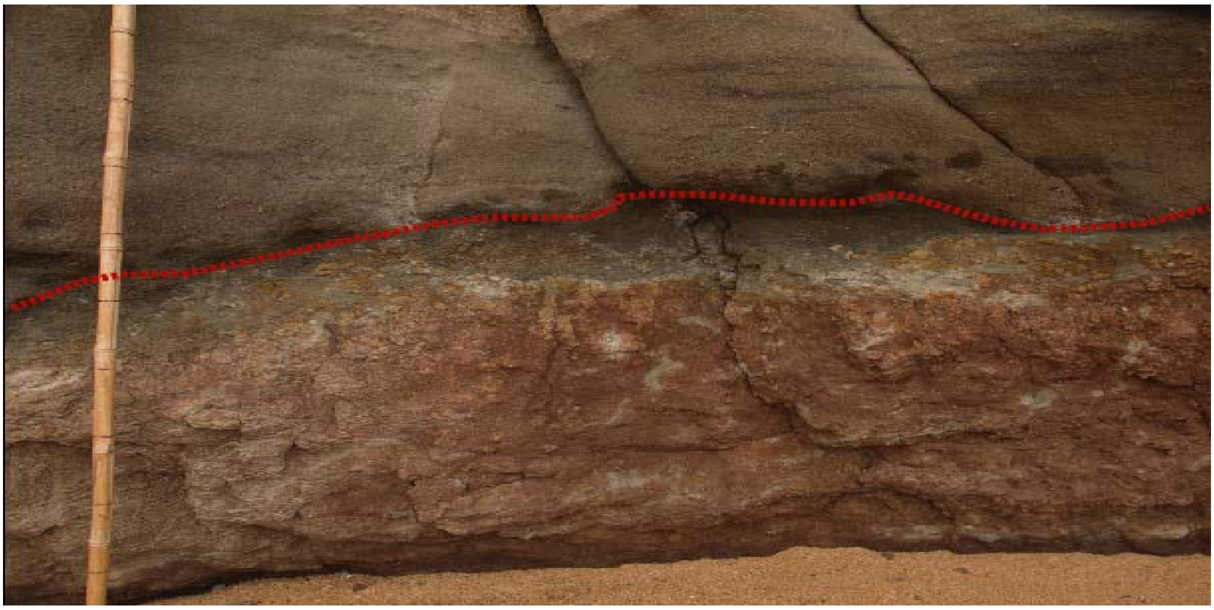


**Fig. 4.9:** Thick paleosol horizon with moderately developed caliche (carbonate) nodules at upper part. Length of stick ~1 m

The concept of pedofacies (Kraus and Bown, 1993) relates paleosol maturity to distance from the active channel, lower maturities being inferred to occur in near channel settings.

Calcrete nodules or massive horizons are common features of paleosol types developed during semi-arid to arid climatic conditions and form in regions with Mediterranean type climates. The blocky calcrete units impregnated with mudstone clast might have formed due to evaporation of carbonate rich groundwater. Above the water table precipitation of $CaCO_3$ takes place in the soil due to evaporation and loss of water, the precipitation binding the sediments to produce calcrete (Zaleha, 1997).

## 4.4 Facies M Soft sediment deformed sandstone

*Description*

Beds associated with soft sediment deformation range from silt to very fine grade, but fine to medium grained are also present and some range up to a metre thick. This facies is generally located beneath coarser beds. Clay to silt size sediments have become penetrated deep upward into coarser overlying sand beds, thus forming something like steep isosceles triangles with a bent top and undulating basal contact. These structures are called flame structures. Water escape structures were also noticed in other localities and do occur in both fine and coarse sandstone. Convolute laminated zones are common and consist of undulated or disturbed strata which are bounded by sharp upper and lower surfaces

*Interpretation*

The water escape structures and associated soft sediment deformation indicate high rate of deposition; this may lead to water escape and bed collapse. When finer sediments situated below coarser sediments become liquefied, there is a tendency of differential sinking of the finer mud into the sand which is coarser, forming flame structures or load casts

The contorted or convolute laminated zone is suggested to have resulted from a sudden deposition of sediment load or when the sediment is dewatered.

**Fig.4.10:** Flame structures (black arrows) along base of very fine sandstone to siltstone. These structures are load induced. Width of view ~50 cm

### 4.5 Facies N Shell bank

*Description*

This facies consists of light grey to dark siltstone containing many shell fragments of sizes ranging from 1cm- 3 cm, with an average about 2 cm. Few coal fragments could be found mixed with this assemblage. The fossils were full-bodied, though reworked shells were also noticed in other locations. There are sections where the shell fragments are found within intra-clast conglomerates and are generally not compacted. The original mineralogy is preserved since the skeletons have not been replaced or been calcified. The grain size generally decreases upward and the fossil content is higher in the lower part and gradually decreases up. The fossil assemblage seems to belong to oyster species, but gastropod shells were also noticed in the upper part of some beds.

Laterally, this facies could be traced in some beds for long distances while some are very short. Towards the north the bed could be traceable for a few hundred metres, but since it is at a cliff side it becomes difficult to access. The fossil assemblage is identical throughout the whole length. The lower part is erosional but the upper boundary is gradational.

**Fig.4.11:** Shell fragment rich siltstone. Shell fragments possibly derived from flooding under brackish conditions.

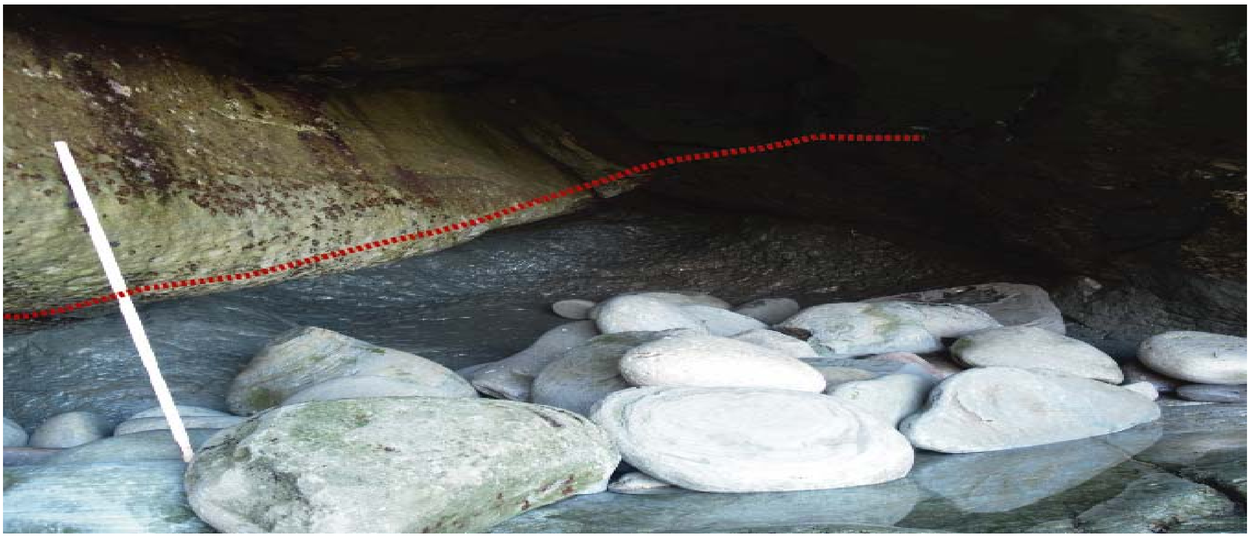


**Fig.4.12:** Dark grey marine mudstone. Above fluvial sandstone channel has an erosive basal contact with this mudstone which is suggested to represent a sequence boundary. Stick 1 m

*Interpretation*

The mudstone/siltstone dominated lithology with shell of oyster mixed with coalified plant debris is an indication of marine flooding surface suggesting that the winnowing action of weaker currents remove the finer sediments and skeletal grains to be deposited. This should be a possible explanation for the impersistent nature of the shell bank deposit.

Abundance of the oyster species is an indication that normal marine salinity prevailed during the possibly marine transgression. The broken shell fragments in certain areas are suggested to be associated with storm-dominated deposits and they are reasonably interpreted to be deposits above the wave base. Reworking and redeposition then took place during sea level rise. As a result, there is a clear increase in the degree of shell fragmentation towards the south and southeast, since it is thought the sea was in that direction.

## 4.6 Facies O Marine mudstone

*Description*

The marine mudstones appears greyish black to grey with silt size mudstone. This facies is seen in only one location and consist of up to 20 – 60 cm thick units of structureless mudstones. The thickness of facies O generally increases to the south, which suddenly disappears. Laterally, facies O may be up to 8 m long. Bioturbation is hardly seen but faint

The lower boundary is difficult to see since it is buried, but the upper boundary is usually erosive to extra-basinal conglomerates (facies A), but a few distance to the left the upper boundary appears to be more planar and slightly erosive to channel sandstone (fig. 4.12).

*Interpretation*

Structureless mudstones of silt grade indicate deposition took place below storm wave base. The grey to black colour may signify high organic content which is due to anoxic conditions in the bottom water. This could be confirmed with the absence of bioturbation.
They are thought to represent maximum flooding surface that occurred during the Late Oxfordian in which the sub-basin was relatively starved of sediments (e.g. Leinfelder, R.R et al., 1998)

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# 5. Facies Association (FA)

Facies association is a complex arrangement of different facies, depositional cycles, mesosequences, and elements grouped together in distinct units, characteristic of a particular depositional setting (Stow, 2005)

**Table 5.1:** Table of facies association and their respective facies and architectural elements

| Facies association | Description | Architectural elements | Facies |
|---|---|---|---|
| **FA 1** | Channel infill | Distal braid-plain sandstone sheets | C, D, H |
| | | Lateral accretionary bar/ point bar | C, D, E, F, H |
| | | Downstream | |
| | | Distributary channels/bay fill | D, G, H, I |
| **FA 2** | Overbank deposits | Levee | G, H, I |
| | | Crevasse channel | C, G, H |
| **FA 3** | Floodplain | Crevasse splay | D, G, H |
| | | Mudstone (continental) | K, L |
| | | Abandoned channel/lake | D, G, H |
| **FA 4** | Heterolithic stratification | Inclined heterolithic (IHS) | C, D, H, I |
| | | Horizontal heterolithic | C, G, H, I |
| **FA 5** | Marine deposits | Marine sandstone | N, H, I |
| | | Transgressive lag | N, B |
| | | Marine mudstone | |

### 6.1 Facies association 1 (FA 1): Channel infill

Facies association 1 (FA1) constitutes relatively higher energy deposits and are principally channelised deposits. A simplified version of channel geometry can be seen in fig. 5.1. The sand to mud ratio here is generally higher compared to the other deposits. The architectural elements defined under this heading are discussed below.

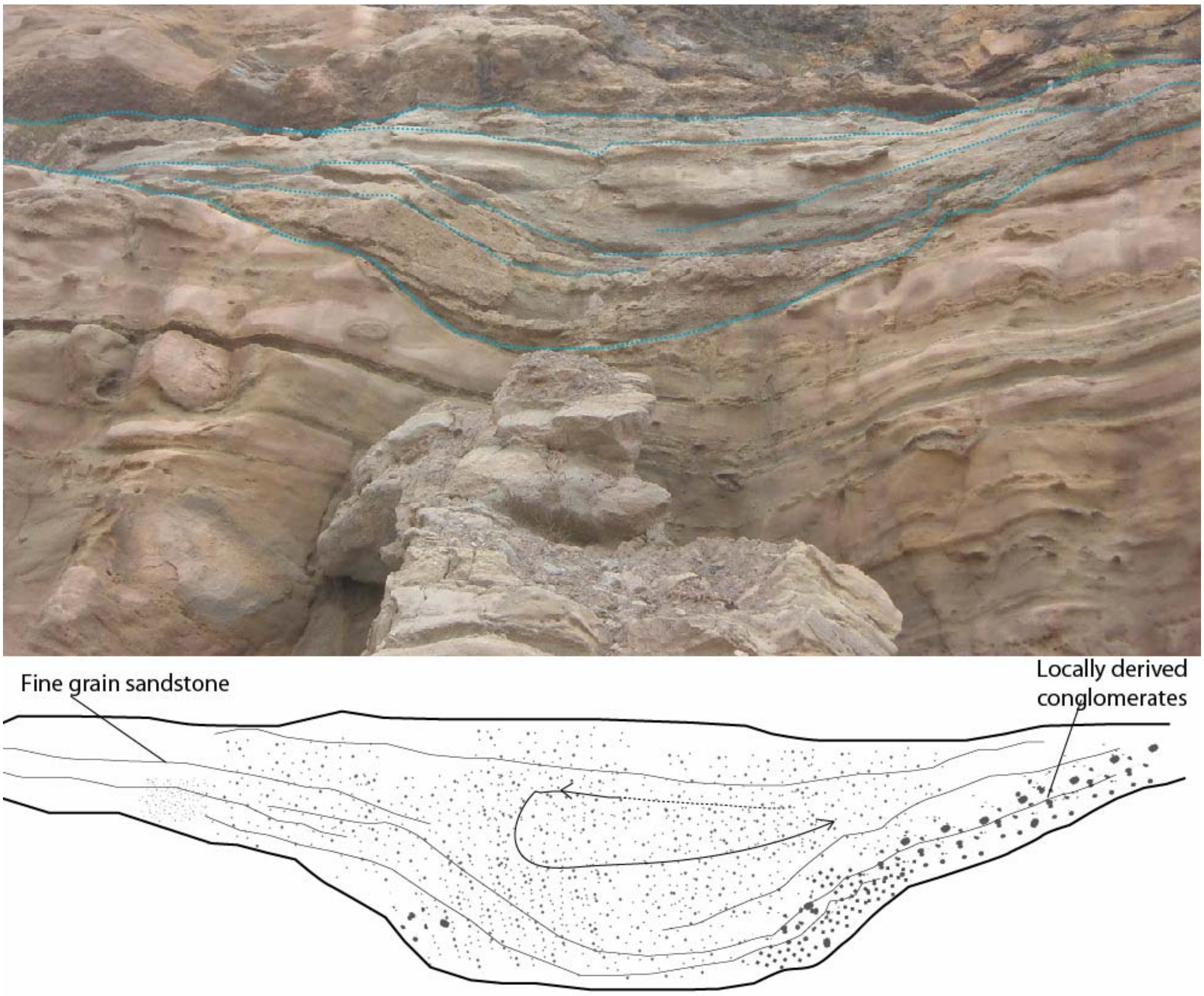



**Fig. 5.1:** Small channel cutting down into previously deposited channel sandstone, note the early stage of accretionary pattern. The cliff is perpendicular to the long axis of the channel.

#### 5.1.1 (FA.1.1) Distal braid plain sandstone sheets

*Description*

The low sinuosity or braided channel deposits relatively contain the coarsest deposits in the study area. This facies association is composed of trough cross stratified sandstones (facies C), planar cross stratified sandstones (facies D) and ripple laminated sandstone (facies H). Sandstone bodies tend to erosively overly paleosol units and abrupt lithologic boundary in most locations. Fining upward with bed thickness reducing upward can be seen in some sandstone bodies .Average bed thickness may range between 30- 40 cm and consist of coarse sand to mud with medium sand as the mean. The mudstone deposits can be up to a few metres in thickness, be located as mudstone remnants along amalgamated surfaces or be localised as mud clast in intra-basinal conglomerates. This facies association is devoid of sand - mud couplet. Bioturbation is common in the lower parts of individual sandstone bodies.

Channel dimension ranges between 7- 11 m in thickness. Laterally some of the sandstone bodies can be traced for long distances, range 100- 150 m and may disappear in mud or they pinch out abruptly. The beds are typically poorly sorted to well sorted. In most examples the upper section is characterised by alternating finer and coarse grain units. The finer intervals are between 2-5 cm thick, mottled (red/grey) and bioturbated. Palaeocurrent measurements gave values of 95, 140,165 and 275 degrees, which show most of the flow was in the SE direction. Root structures, wood and plant fragment and even gastropod shell can be found occasionally in some of the sandstone bodies.

*Interpretation*

The absence of tide – mud couplets suggests that this facies association occurs upstream from the tidal limits. The presence of gastropod shell, finer grain sizes and smaller bed thickness upward (Log FWB-2) also supports that this FA should be fluvial dominated depositional units, as generally suggested by Allen (1964) for these types of sandstone bodies. The large width to thickness ratios, the lack of lateral accretion beds, the medium to coarse grain size of the channel fills, the low abundance of overbank deposits and relatively low palaeocurrent variability suggest that the channels had relatively low sinuosity (Bridge et al., 2000).

Vertical stacking of different bedform types indicates long or short term changes in flow regime. Amalgamated contacts between distinct beds may signify that two depositional episodes were closely spaced in time, and that the energy available for the second depositional event partly reworked the beds that had been deposited before it could achieve its existence. This usually occurs during flash floods and during changes in flooding seasons. Plant structures or coal may represent presence of vegetation within the depositional environment, whereas the paleosol horizon beneath the sandstone bodies signify sub-aerial exposure for a long time before the channel could come to existence

### 5.1.2 (FA. 1.2) Lateral accretionary elements and point bar deposits

*Description*

This FA 1.2 is composed of trough cross stratified sandstone (facies C), plane parallel to sub parallel stratified sandstone (facies G), massive sandstones (facies G), planar cross stratified sandstone (facies D), trough cross stratified sandstone with mud drape (facies F) and cross laminated sandstone (facies H).

The grain size is fine to medium grade, with a generally fining upward trend. The sand bodies are made up of complex compound association of sandstone beds having cross stratification and smaller sets of cross bedding. The beds are generally inclined with sigmoidal reactivation surfaces. Total bed thickness ranges between 30 cm – 1.5 m in its deepest part, forming clinotherms, each 3 – 6 m wide (but rarely above 6 m). Beds have a dip of 10 -15 degrees with maximum dips toward SE. The inclined beds are seen to be curved or bent inward with near horizontal to slightly convex upward surfaces The bed set thickness ranges between 8 m- 12 m, but toward the flanks, the sets become smaller and finally turn into mud. The total length of the eastern extent of the point bar can be measured to ~65 m and the western flank to ~ 40 m; hence the point bar is approximately 105 m long all together. The surfaces of most of the dipping beds and bedsets are erosional to non erosional, being separated by facies J (fig. 5.2 ), but the surfaces become unconformable when they are near horizontal to slightly convex at the bottom of the bar unit (e.g. fig. 6.4)

In the most southern part of the point bar unit, this FA 1.2 is grey, inclined to near horizontal and widest and consists of facies E, structureless sandstone with coalified wood fragments and very coarse grain matrix supported sand amalgamated with finer sandstone located above a distinct erosional contact.. Higher up in the section there are several minor unconformable surfaces formed within individual sigmoidal shaped clinothems (fig. 5.3).

Toward the northern flank of the point bar unit the clinothems have more unconformable surfaces, amalgamated surfaces of mud in sand, and within the clinothems facies B, facies C, and facies H can be identified. In some places facies H show tidal bundles not up to 5 cm thick that run laterally ~50 cm – 1 m; in this case the bundles consist of sandstone-shale couplets. The channel sandstone is immediately underlain by marine mudstone of facies O.

Moving to the south of the point bar unit, the bar element consist of massive sandstone, smaller sets of cross stratified sandstones associated with millimetre scale mud drapes. A bit up faint asymmetric ripples can be identified on low angle stratified beds, together with some amalgamation surfaces. Other structures are wood fragments, desiccation cracks, and carbonate concretions which are attached to the sandstone bodies. Mudstone clasts can be identified in the mottled silty mudstone in the basal section (Log FW I). Palaeocurrent data recorded from cross stratification show a SE and SW directions.

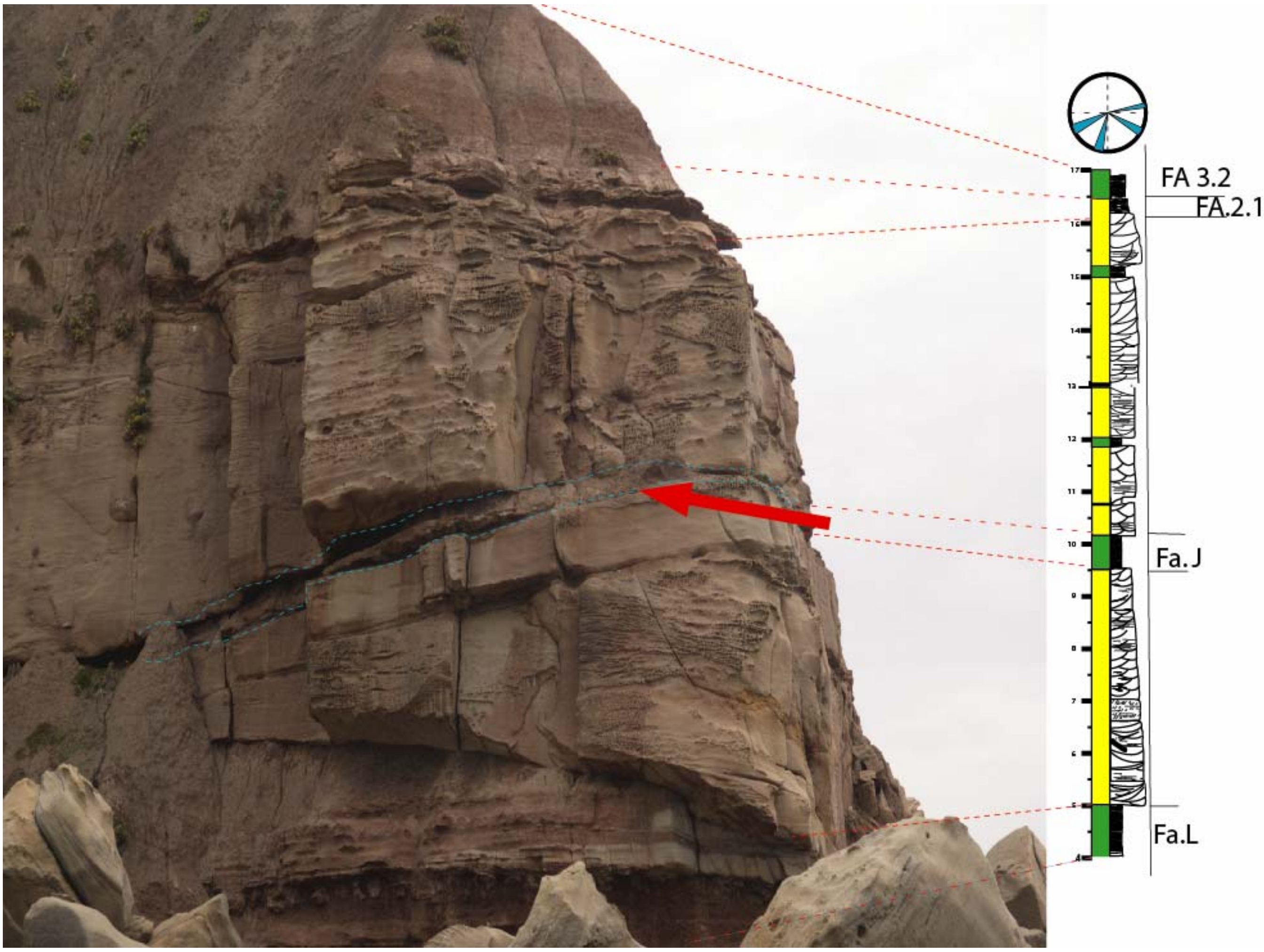



**Fig. 5.2:** Section through point bar element showing erosional surfaces (red arrow) filled with facies J. The rose diagram demonstrates the measured palaeoflow direction.

*Interpretation*

High sinuosity channels develop in areas of very low slope gradients and where the rivers are commonly dominated by high suspended load relative to bedload ratio (Leopold and Wolman, 1957).Lateral accretion bedding or low angle compound cross stratification is a characteristic depositional feature of point bar successions (Allen, 1970; Collinson, 1978). Flow in high sinuosity streams leads to erosion of the concave banks and deposition on the convex bank, this may be the cause of the lateral accretion bars (fig. 5.6).The smaller scale structures within the larger clinothems may be due to flow sub-parallel to the strike of the inclined surface described above. The lower erosional surface and curved pattern of the inclined beds further suggest a curved channel side, probably in a meandering stream bend.

The shape of the channel, long inclined cross stratification, the presence of mud drapes and parallel laminated mud-sand couplet (especially toward south) suggest deposition in a channel oriented in a southerly direction. However, bipolarity of mud drapes indicates reversing currents of almost equal strength and the presence of dominant and subordinate currents.

The presence of marine mudstone immediately below the channel sandstone in the middle section suggests that the channel has been cut into sediment very close to the shoreline. Allen (1980) and Levell (1980) interpreted similar complex and compound cross bedded sandstone bodies as formed within tidally influenced rivers.

### 5.1.3 (FA. 1.3) Distributary channel/ bay fill

*Description*

This facies association (FA. 1.3) is composed of parallel laminated sandstones (facies I), parallel stratified sandstones (facies G), and some sandstone beds with cross lamination and climbing ripple lamination (facies H). Some few sandstone beds with low angle planar and trough cross bedding (facies D) also occur. Mudstone laminae are common. Coal fragments (plant debris) could be seen in the beds. Heterolithic strata may be 10- 60 cm thick and occur interbedded with sandstone beds and form together with these inclined heterolithic stratification.

The bed thickness ranges from 10 cm – 50 cm, and bed sets may be up to 4 m in thickness. Fining upward sandstone successions with internal deformed bedding, associated with facies I and even facies F occur in some places.

The heterolithic facies is succeeded about 2 m to the south by a stacked sandstone of about 4 m thick, with mud-clast conglomerate at base, forming a seemingly erosive base, generally fining upward into heterolithic levee deposit and finally into mudstone.

Laterally the sandstone units are not extensive and may reach 30 m. As in Log FW k-3 and Log FW J-1) mudstone clasts and coal fragments are locally abundant especially in the middle part and somewhere below. Also root traces, burrows, minor clinothems and soft sediment deformation occur in the lower part of these facies association unit; the sandy beds grade gradationally into mudstone, thus forming a non-erosive lower contact in some places. At other sites, intrabasinal clast conglomerate beds are present above an erosive base. The sandstone bodies of this type are usually capped by channel sandstone bodies.

*Interpretation*

Those sandstone bodies of this facies association having faint coarsening upward trend, non-erosive lower boundary, and ripples cross lamination with abundant mudstone laminae may have been formed as prograding bay head deltas. The flanks could have been slightly or intensely modified by waves.

Sandstone bodies having fining upward sandstone bed sets, dominated by cross bedding and ripple cross lamination with abundant mudstone laminae, are interpreted as formed as point bar deposits in a distributary channel, for example in an estuary (cf. Reineck & Wunderlich, 1968). The stack of upward fining sandstone beds and a lower erosional contact covered with a mudstone clast conglomerate could be interpreted as the fill of such a minor distributary channel

### 5.1.4 (FA 1.4) Downstream/midstream accretionary bars

*Description*

This facies association is composed of trough cross stratified sandstones (facies C), cross stratified sandstones with mud drapes (facies F), planar stratified sandstones (facies D) parallel stratified sandstones (facies G), parallel laminated sandstones (facies I) and few containing cross laminated sandstones (facies H). The grain size of this facies association is fine to medium grain, and range in thickness between 2 and 10 m. The bed thickness ranges from 30 cm- 1 m which may be sub horizontal to horizontal in geometry. The bed sets range in thickness from 1 – 4 m and is characterized by numerous reactivation surfaces and is mostly associated with facies F. Bedset units show a fining upward pattern. Elements of facies F occur in bundles.

The lower boundary of this facies association is usually erosive to floodplain fines (FA 3), while the upper contact of the sandstone units is usually sharp, but may grade to levee deposits. The lower surface is generally concave up and at some places the upper boundaries of the sandstone units are convex up. Some of the channel bars are very extensive, and may be above 100 m. The channelised sandstone bodies usually occurs as isolated units, but few are connected laterally to other channel sandstones

*Interpretation*

According to Bristow (1987) these channels are second order channels which may amalgamate to form compound bars that persist for many years and become well vegetated. In the study area the channels described as midstream accretionary bars are characteristically ~5-20 m with numerous amalgamated and sandstone bounding surfaces dipping in the down current direction. Also within the sandstone units are coalified wood fragments indicating the area was well vegetated. Miall (1996) described downstream accretionary elements as having several cosets of downstream- oriented bounding surfaces.

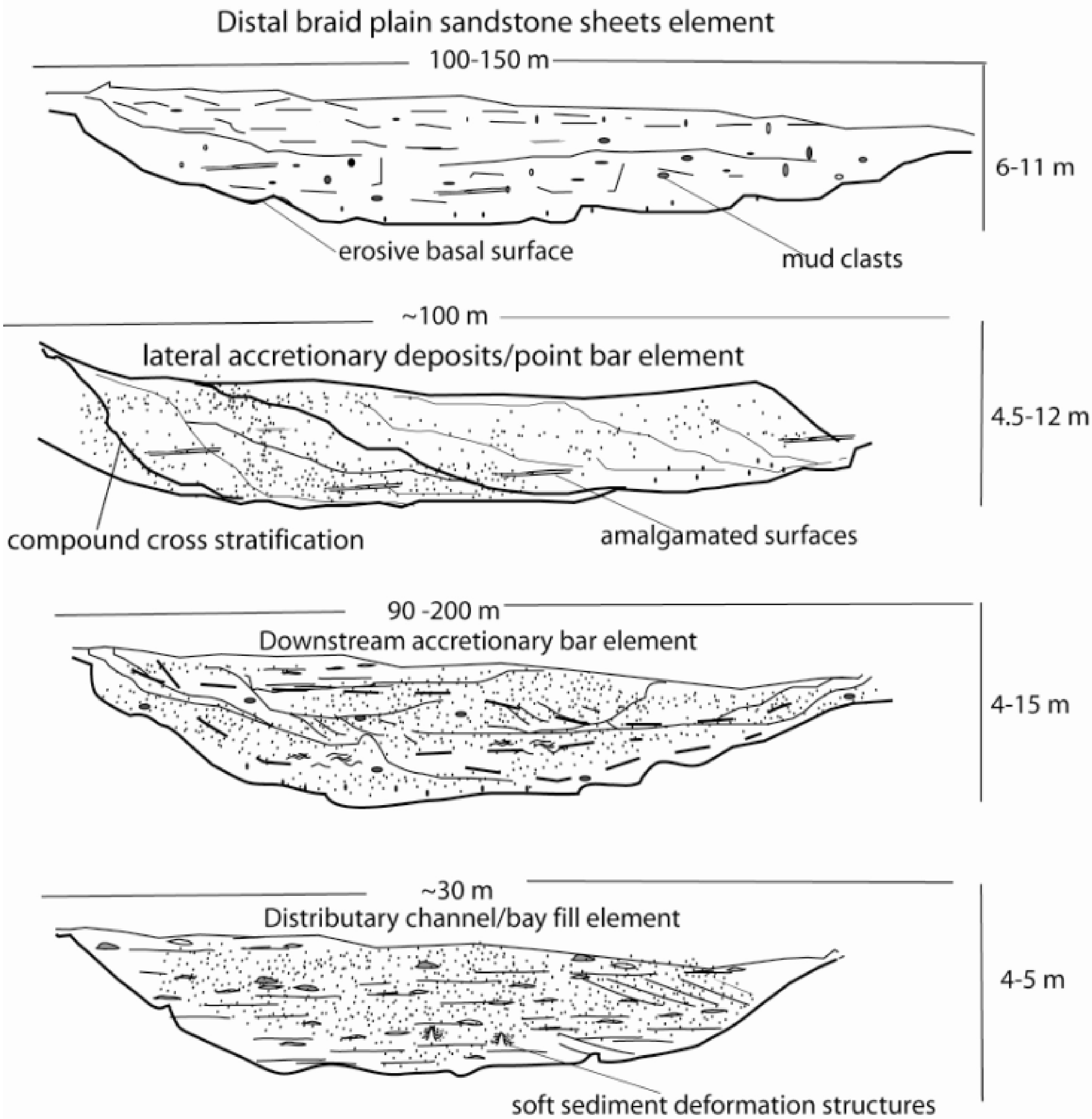



Fig. 5.3: schematic representation of the various channel fill architectural elements showing some of the external geometries which might influence fluid flow internally

**Table 5.2:** Summary description and interpretation of the various channel fill elements.

| *Channel fill elements* | *Grain size/basal contacts* | *Bed and bedset thickness/bed continuity* | *Interpretation* |
|---|---|---|---|
| Distal braid plain sandstone sheets | Fining up succession Fine to coarse grain Lower part contains rip up clasts consisting of facies B | 30 to 40 cm thick in lower part and 2-5 cm thick in upper part. Amalgamated beds may be up to 3 cm thick. Bedset may range up to 5 m | Fluvial dominated, relatively low sinuosity channel characterised by long or short term flow regime |
| Lateral accretionary elements or point bar | Fine to medium grain but generally fining up succession. Erosive basal contact | Bed thickness 30 cm -1.5 m forming clinotherms. Each clinotherm 3 to 6 m wide. Bedset thickness ranges to 12 m | Lateral bedding or low angle compound cross stratification which is characteristic of a point bar. |
| Distributary channel/ point bar | Alternating fining up and faint coarsening up succession. Usually non erosional contact, but in some places facies B present forming erosional basal contacts | Bed thickness may be 10 to 50 m and bedset may range up to 4 m. presence of minor clinotherms usually associated with soft sediment deform ation structures | Fining upward succession associated with faint coarsening up successions and non erosive basal contact may result from progradation of a bay head delta |
| Downstream/Midstream accretionary elements | Fining up succession consisting of fine to medium grain sandstones. Basal surfaces are usually erosive to floodplain deposits | Beds range from 30 cm to 1 m in thick. Bed and bedset may extend for more than 20 cm to 10 m laterally | Forming complex downstream dipping surfaces and numerous amalgamate and surfaces |

## 5.2 Facies association 2: FA. 2 Overbank

FA 2 consists mainly of sediments deposited at the sides or proximal to the main channel. They constitute finer grained deposits, relative to the main channel sandstone bodies. This FA is sub-divided into levee and crevasse channel.

### 5.2.1 (FA. 2.1) Levee deposits

*Description*

Facies association II.1 is generally narrower and seen embedded in flood plain fines or in some cases they gradually grade into mudrock. Colour is grey to greyish red and even mottling can be seen in some parts, but most have the same colour as the main channel sandstone bodies. Bed thickness ranges from 15 cm to 2 m, but average is about 90 cm. Their grain size ranges from silt to very fine sand with an upward fining succession; but few cases of coarsening upward trends were recorded. The grain size increases toward channel sandstone facies. Faint current ripples sandstones (facies H), planar laminated sandstones (facies I) and facies G are more frequent than in the initial channel sandstone body, but most of the sedimentary structures in the levee are destroyed by bioturbation.

Bioturbation concentrates more in the upper part, but some vertical burrows can extend to the lower part. The lower boundaries of levees units are usually transitional to channel sandstone bodies and are conformable. The architectural elements of this facies association may appear concave upward to near planar. Most point bar sandstone units in the study area are overlain, either gradationally or abruptly by finer grained sediments of levee and floodplain facies association. Log $I_4$.show the ideal succession of stratification with thin sandstone beds, representing flood sheets, and in some locations beds with abundant coalified plant fragments are present. Levees are wedge-like in geometry and seem to thin away from channels (fig.5.4), laterally may be up to 50 m from the associated channel sandstone body. Other features associated with levees are desiccation cracks in the muddy and pedogenic structures

*Interpretation*

As water leaves a main river channel at bank full top during a flood, the water spreads out and loses velocity very quickly. The drop in velocity prompts the deposition of all sandy and silty suspended loads, leaving only clay in suspension (Hughes and Lewin, 1982). The sand and silt are deposited as a thin sheet over the flood plain.

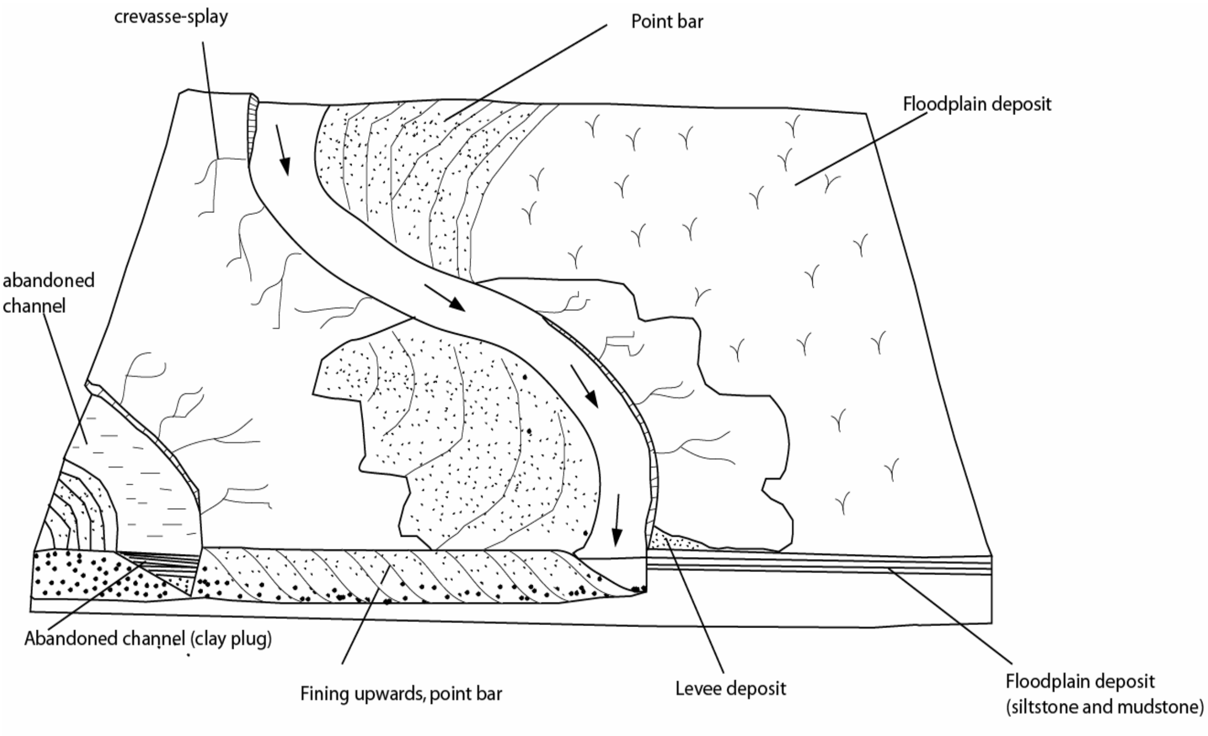



**Fig**.5.4: Depositional environment within a meandering fluvial channel. Modified from LeBlanc 1972

The sedimentary sand facies with current ripple (facies H) and planar lamination (facies I) are formed as flow velocity decreases. The planar stratification (facies G) was produced during relatively high flow rates. The sand or silt sheets are thickest near the channel bank because the coarsest suspended load is dumped quickly as soon as the sediment-loaded flood water starts to flow away from the main channel.

### 5.2.2 (FA.2.2) Crevasse channel

*Description*

Few successions suggested to represent crevasse channel deposits were recognised in the studied area. These facies are composed of trough cross stratified sandstones (facies C), plane parallel stratified sandstones (facies G) and ripple cross laminated sandstone (facies H)

The channel successions have concave up erosional bases with an almost horizontal upper contact (fig. 5.1).The units can be up to 3 m thick and are fine to medium-grained. The channel succession in fig. 5.1 is seen embedded between major channels. The lateral extent of the channel beds were difficult to evaluate since only a two dimensional view could be accessed.

Usually, the uppermost parts of this facies association associated with root structures, desiccation cracks and burrows.

*Interpretation*

Crevasse channels are formed when the main channel cuts through its levee during floods, and a new smaller channel forms in the levee crevasse and adjacent flood plain.

Crevasse channels within crevasse splay lobes may be recognised by its distinct erosive base, fining upward trend, and content of trough and ripple lamination (Flores et al, 1985). Grain size is fine to medium grained, whereas crevasse channels are slightly coarser than the crevasse splay sheets (Flores et al, 1985). These characteristic features correspond well with observation obtained from the actual facies association of the study area.

The presence of desiccation cracks, interbedded sand- mud laminae at interval, root structures and burrows is an indication of periodic cessation of discharge.

**5.3 Facies association 3: (FA. 3) Floodplain fines**

Elements of this facies are generally clay to fine grain and they constitute more than 55 % of all lithologies found in the study area. Mudstone, and crevasse splay dominate the floodplain facies association. In the description, they are considered as distal deposits relative to over bank deposits described above. Paleosols are grouped under mudstones since they do develop from them.

**5.3.1(FA. 3.1) Crevasse splays**

*Description*

Crevasse splay bodies are seen as near lenticular to sheet-like units within floodplain fines, composed of planar cross stratified sandstone (facies D), plane parallel stratified sandstone ( facies G) and cross laminated sandstone (facies H) to plane beds. Individual beds are very thin whereas bedsets generally have a range of 20 cm- 30 cm in thickness, with a maximum up to 2 m. The grain size is very fine to fine grained sand, but the sand gets coarser toward the crevasse channel. A slight coarsening upward trend is noticed in most cases, but occasionally fining upward trend is also recorded.

Lower surfaces are usually planar, erosive to gradational, the tops are usually gradational. In some occasions it is difficult to differentiate between the crevasse channel and the crevasse splay. The sheet-like sandstone bodies laterally have been observed at an extent of 1-2 m. However, the real lateral extent of this facies association likely must be in order of several tens of metres.

*Interpretation*

As the crevasse channel proceeds away from the main channel, hence stream power gradually reduces. This affects the grain size both vertically and horizontally due to waning flows.

The sediment is introduced into by a crevasse channel and is deposited as a result of low flow expansion and loss of flow power as discharge leaves the confines of the channel and spreads out as sheet floods (Miall, 1996). Bridge (2003) proposed that crevasse splay deposits at times become difficult to distinguish from levee but are generally coarser. These characteristics also correspond well with observation obtained from the actual facies association of the study area.

### 5.3.2 (FA. 3.2) Mudstone

*Description*

Facies association 3.2 generally consists of clayey mudstone (facies J) and silty mudstone (facies K).

The geometry of mudstone units is hard to predict, since this lithological units occurs in patchy exposures. However, mudstone units cover most of the whole section. Vertically the thickness of mudstone units are variable, the thickest units may be above 35 m, while the thinnest can be very thin of centimetre scale. The shape of mudstone units could be described as tabular with abrupt and/or gradual terminations. The lower and upper boundaries are planar or erosive, but the upper boundaries are usually gradational. Within the mudstone units there may be portions of coarse pebbly and sandy mudstones

*Interpretation*

Presence of significant thicknesses of mudrock indicates the existence of considerable periods of low energy sedimentation dominated by fallout from suspension (Hill, 1989). The coarse pebbly and sandy mud rocks could be products of debris or mud flows.

The difficulty to observe sedimentary structures could be due to the small grain size contrast or due to bioturbation and soil forming processes. The mottling and root structures correspond to the alteration of the sediments by pedogenic processes after deposition (fig 5.5).

The grey to black mudrocks are relatively richer in organic matter, this could be plant material or material of other biogenic sources. Mudrock units and their paleosols may indicate climatic conditions that existed during formation by their clay mineralogy, but this has not been studied. Generally, the environment was more oxidizing, seen from the coloration and the presence of carbonate nodules especially on top of mudstones.

**Fig. 5.5:** Grey to brown mudstone deposit showing colour mottling (red circles). Weathering and scree falls usually obscure most of their sedimentary structures.

### 5.3.3 (FA.3.3) Abandoned channel/lake

*Description*

This facies association can actually be found in few locations in the study area. The channel units are dominated by sandstone in the lower part and heterolithic and finally clay-rich in the upper part. The claystone is slightly laminated (facies J). The sandstone lenses may range between 0.9 – 5 cm in thickness, but the sandstone/silt units in the top of these successions may be up to 1.5 m thick. The sandstone bed sets below are composed of facies D, facies G and facies H. Facies G may be irregular or slightly discontinuous but inclined.

The sandstone is laterally not extensive and may not be more than 50 m. Other features found here are plants roots, burrows and foot imprints. The depositional bodies of this facies association have concave up lower erosive base, but the upper boundary is usually gradational to mud.

*Interpretation*

The sediments are thought to have been deposited in ponds or shallow lakes, or may represent infill of abandoned channels (fig. 5.6). The lateral migration of meandering channels results to the change in direction of the channel and when this continues for some time, the meander will finally cut off (fig. 5.4). A body of standing water or pond will result in the abandon channel segment. As a result, abandoned channels or oxbow lakes may become gradually filled with fine grained sediments and predominantly clay. Allen (1965) suggested that an abrupt transition from channel sandstone to laminated mud is an indication of channel abandonment, due to neck cut of in a high sinuosity channel.

The presence of the foot imprints and some coal fragments derived from plant debris is an indication that this was entirely continental.

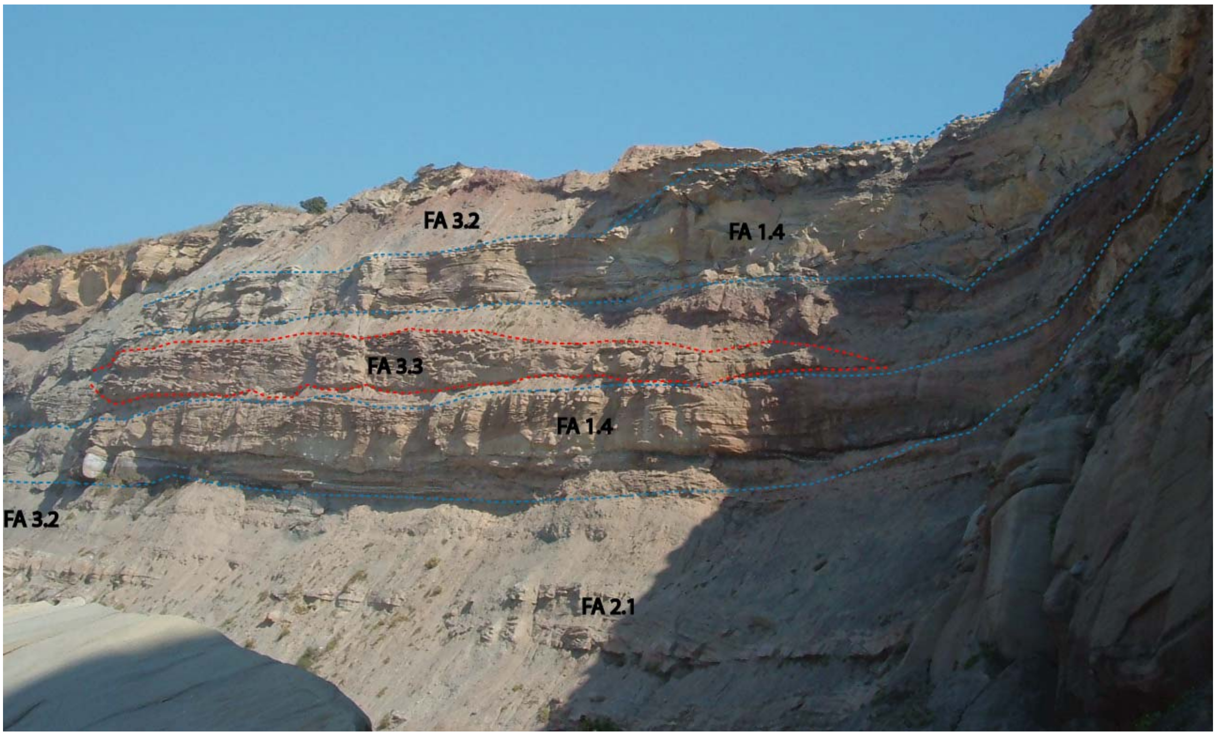



**Fig. 5.6** A characteristic abandoned channel fill (FA 3.3) between channel fill elements. The abandoned channel consist of abundant mudstone especially in the upper parts.

### 5.4 Facies association 4 :( FA.4) Heterolithic stratified sandstones

### 5.4.1 (FA. 4.1) Inclined heterolithic bedsets

*Description*

Facies association IV.1 is composed of repetitions of alternating layers (15cm – 2.5 m) of fine sand and silt/mud couples with numerous internal erosion surfaces. The sand – mud couplets are generally inclined within a larger bed form with a dip ranging between 5 – 10 degrees. The alternating layers are not equally spaced. This FA is divided into two sections: a mudstone dominating part with sandstone lenses and a sandstone dominant part with lenses of mudstone. There is a general fining upward trend in units of this facies association. The FA units are characterised by laterally extensive strata of more than 50 m.

The sandy part of the couplets occurs in 15 cm – 40 cm thick sets; coset thickness is in the range of 1m – 4 m. (see log FW-H). Units of planar cross stratified sandstones (facies D) to plane parallel laminated sandstones (facies I), with an interval of trough cross stratified sandstone bed (facies C), fining up to ripple cross stratified sandstones (facies H). In some places, there are double drapes separating thicker sandstone laminae. The sandstone units are

are normally grey to brown and locally bioturbated. Majority of the inclined sandstone strata have erosional lower contacts and contain multiple internal erosive surfaces, with a very low relief. Mudstone clasts can be seen in some locations. Abundant plant remains, flame structures, water escape structures, carbonate concretions; siderite bands mixed with coal are frequently seen.

The muddy part is red to greyish brown in colour with high clay content and minor sandstone lenses between, couplets range in thickness of 10 – 50 cm and cosets from 50 cm to 3 m. The bedding is not very clear, consisting of indistinct patches of sandstones in mudstones.

*Interpretation*

The interbedded sandstone and mudstone facies association is interpreted as representing inclined heterolithic stratification (Thomas et al., 1987). Inclined heterolithic stratification has most commonly been documented from tidally influenced point bars (Thomas et al., 1987). The double mud drapes documented within some of the flood-ebb cycles, separating a thicker sand lamina from a thinner one, imply a strong asymmetric tide or weak subordinate current (Nio and Yang, 1991).

The environment of deposition of the heterolithic facies association was likely 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 as generally suggested by Hill (1989) for this type of facies association. The cyclicity, or rhythmicity, of sand separated by mud bands or mud separated by sand bands may represent episodic depositional events created by dominant and subordinate currents.

### 6.4.2(FA. 4.2) Horizontal heterolithic bedsets

*Description*

Horizontal heterolithic stratification is composed basically of plane parallel stratified sandstone (facies G) and plane parallel laminated sandstone (facies I), though there are some ripple laminated sandstone (facies H) and trough cross stratified sandstones (facies C) in few locations. This facies forms 40 cm – 4 m thick fine to medium grained successions (fig. 4.6), and shows faint fining upward succession.

This sub facies association is characterised by thinly interbedded sandstone and mud stone which are parallel to gently inclined. However, towards the top of some individual channel bodies the amount of sand beds increases. The sandbodies are laterally not persistent as a unit and may not be more than 10 m in extent.

*Interpretation*

The horizontal to gently inclined heterolithic stratification is interpreted as shallow subtidal to intertidal flat deposits. Where there are alternations of laminae of mudstone and sandstone in a horizontal or slightly inclined manner this may suggest a tidal flat; the facies association formed by fluctuations in energy and sediment supply during tidal cycles. In this case mud is deposited during quiet periods of the cycle, and the sandy portion during relatively stronger currents.

## 5.5 Facies association 5: (FA. 5) Marine deposits

The presence of tidal signatures and some marine trace fossils is an indication of a near shore environment in many of the lithologies. Most of the facies described above are not excluded from marine activity, but they are not grouped under this heading because of certain characteristics in which they possess. Marine sandstones units, marine mudstone units and the shell bank deposits that are described below are placed under this heading because of their peculiar characteristics of abundant marine fossils, marine trace fossils and the grey coloration.

### 5.5.1 (FA. 5.1) Marine sandstone

*Description*

Few marine types of sandstone were located in the study area. A highly bioturbated dark grey, very fine to medium grain poorly sorted sheet-like sandstone is a candidate for fully marine origin. This unit is located about 35- 50 cm below a conglomerate bed with few shell fragments and ca 2 m below the silt to muddy sand shell bank (facies N). The sandstone also consists of several small indistinct beds and faint ripples (facies H), whereas crude parallel laminated sandstone (facies I) could be located especially in the upper part in some places.

The bioturbation is mostly concentrated in the upper portion (fig. 5.7) and was typical of *zoophycos* and *skolithos*. Laterally, the sandstone unit covers a length of at least 11 m and may be up to 4 m thick. The lower boundary is difficult to locate, but in few exposures it seem undulating and erosive. Paleocurrent measurements obtained from this sandstone bed is dominantly south – east and south- west (135, 225, 150, 240, 160).

**Fig. 5.7:** Highly bioturbated dark grey marine sandstone; bioturbation more evident in the upper part (black arrows), crudely laminated fine to medium grain in parts.

*Interpretation*

The presence of shell bank (facies N) coupled with the conglomerate bed with moderate shell content, the high bioturbation and high organic nature of this bed, suggests these sediments were deposited in a shoreface or shallow shelf environment during transgression. The burrowing could have taken place during more quiet periods hence, obliterating most of the sedimentary structures.

The presence also of the ripple laminated sandstones (facies I) suggests also a relatively low energy environment. The conglomerates were locally derived during high energy periods. The diverse paleocurrent direction may be due to waning currents. Also, the ~4 m thick sandstone unit with a sheet-like geometry, has no evidence of channelization or scour at the base, is most likely considered of shallow marine of upper shoreface origin (e.g. Nichols et al., 2002). It may represent post rift lowstand submarine fan deposit (e.g. Leinfelder et al., 1998)

### 5.5.2 (FA. 5.2) Transgressive lag

*Description*

This facies association is composed of the shell bank (facies N) and marine mudstone (facies O) as shown in figures 4.11 and 4.12 respectively. However, there are some sections with intra-clasts conglomerates to medium grained muddy sandstones with moderate amount of shell content. This facies association does occur in about three to four locations and some of them are made almost entirely of shell fragments with very little matrix.

Laterally, one of such units can be traced for distances of about 200 m before the unit finally disappear beneath scree cover, or is inaccessible in the cliff side. Such a layer could have been an excellent stratigraphic datum surface but since it is laterally not persistent, it is problematic to apply in this way.

*Interpretation*

The presence of shell fragments of marine origin in clay to silt grey coloured matrix suggests marine flooding surfaces or transgression. The coal fragments could have been derived from adjacent terrestrial areas as charcoal or coalified plant remains. The abundance of the oyster species is an indication that normal marine salinity was evident during the possibly marine transgression. This is a general characteristic for brackish environments.

# 6 Depositional environment

### 6.1 Depositional setting

The depositional environment of the study area (Louriñha Formation) is based on the sediment distribution, description of the architectural elements (facies associations) and the tectonic history of the Lusitanian Basin. From the panel (fig. 2.2), the depositional facies of this area record the evolution from a fluvial dominating environment or a tidal river, to an estuary, and finally to a mixed environment (estuarine-deltaic). From petrographic studies it is realised that the clastic sediments were derived from Hercynian basement rocks in the west or northwest.

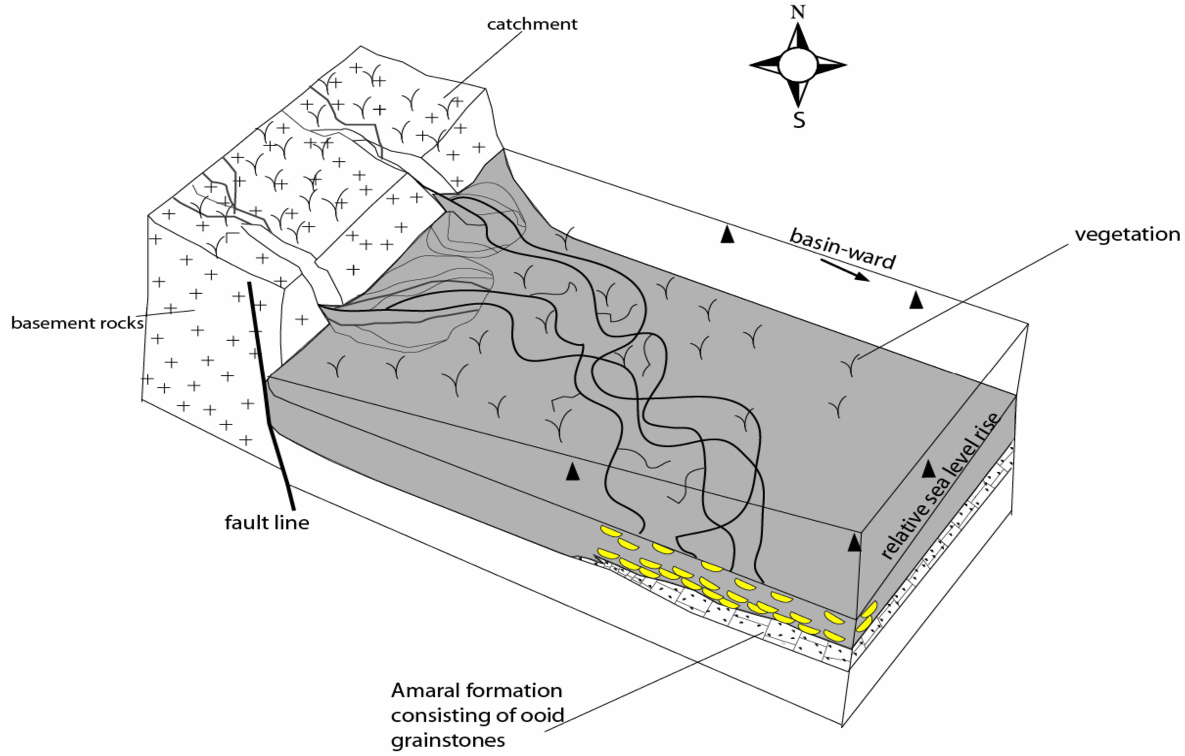



**Fig.6.1:** Schematic summary diagram showing the depositional model of the Lourinha Fm

The most distinct fluvial dominant channel units are referred to as units 1-7 (panel). Units 1-7 is characterised with isolated channels, crevasses embedded in floodplain fines. For example, channel sandstone unit 2 in the panel (fig. 2.2) is a fining upwards ~7- 10 m thick channel sandstone. This sandstone body is a typical example of the description and interpretation made in section (5.1.1). The fluvial dominant section is composed mainly of fining upward successions of poorly sorted fine to coarse sand containing intra-basinal conglomerates (facies B), molluscs shell, and shell fragments. However, some tidal signatures could be found in unit

1 (described as FA 1.4) and some few others; this suggests they may have been connected downstream to areas influenced by tidal activity. Most of these sandstone units have undulating erosional contacts with the underlying floodplain deposits, which in some instances are paleosols or mudstone.

According to section 5.1.2, most of the channel sandstones have tidal signatures. This may suggest that they were deposited in an estuary dominant environment or a tidal delta. The dominating facies associated with the sandstones are trough cross stratified sandstones (facies C), bipolar double and single mud drapes (facies F), ripple cross lamination (facies H), and seemingly flaser and lenticular bedding, indicating a tidal environment. Some of the beds within the succession have coal fragments associated with root horizons. This is an indication of periods with vegetation cover. The fining upwards successions may represent minor tidal channels. The heterolithic beds may suggest tidal flat deposits. In section two the sandstone 59bodies change from isolated sheet-like and single storey bodies to multi-storey (fig. 2.2) or aggradational. The upper part of section two and the majority of section three are overall represented by fine grained sandstone with abundance of heterolithic stratification (FA 4). This may indicate a shift in facies as the sea was drowning the fluvial system, with tidal flats located at the top. The finer uppermost part of the succession could also have resulted from autogenic shift due to avulsion, leading to abandonment.

Sandstone, coal, and mudstone coarsening upward heterolithic succession may have been deposited as a bay head or tidal delta or as tidal channel bars. The combination of an overall tidal setting involving dominantly fine to coarse-grained or heterolithic sediments, with only minor channel sandstones as seen above, is an indication of deposition in the outer part of an estuary according to Dalrymple et al., (1992). A closer look at the sediment distribution in the study area reveals that there is a relative fining towards section 4 (Ariea Branca).

### *6.2* Depositional models and controls

The complex architectural pattern of basin fill in the study area can be explained to be influenced by global or regional sea level changes, climatic processes and syn-sedimentary tectonic activity. There may be several possible explanations for the origin of the sandstones in the distal mud dominating Louriñha Formation. Interpreting which variables controlled the fluvial deposition for a given non marine deposit is complicated by problems in isolating individual variables, and by the fact that fluctuations in each of the main controls can produce similar results (Schumm et al., 1991). A recognizable variation of the accommodation space influenced by both allogenic and autogenic processes can be seen from the geometry of the sandstone bodies.

### 6.2.*1 Autogenic controls*

The main autogenic controlling factor in the study area is avulsion. River avulsion represents the quick transfer or shift of a river out of its channel into a newly created channel in the floodplain (Slingerland and Smith, 2004). Avulsion is supposed to have played a great part in the distribution and stacking pattern of the sandstone bodies in this area.

Levee slope and the rate of deposition control the rate of avulsion in a fluvial system. From fig 5.3 it is seen that levees act as barriers to channel load, preventing them from spilling above the surrounding flood plain. In this case a certain amount of energy is required for the river to cut through the levee and into the floodplain. The amount of suspension and bedload and consequently deposition rate near the channel being higher than adjacent floodplain increases the rate of avulsion. Studies of modern avulsions show that newly avulsed rivers are strongly attracted toward pre-existing channels, the beds of which are preserved as lows on the floodplain (Morig et al., 2000).

In the study area, field observations indicate that very little avulsion occurred in section one, but in sections two and three avulsion could be higher, seen from the multi-storey channel pattern and less floodplain mud and levee. Section 1 is generally fluvially dominated and coupled with the relatively higher flow velocity a reduced deposition of floodplain deposits could have been favoured; but instead more flood plain deposits are noticed. An explanation could be that a reduced rate of subsidence, with a corresponding reduced accommodation space within the main channel segments resulted in deposition of more floodplain deposits in the system. But in the other sections avulsion rate appears to have been higher, caused by relatively higher sea level variations and the opposing forces caused by tides. Another explanation could be that, in section two, avulsion could possibly be caused by a sudden influx of sand into the estuary resulting to higher deposition rates.

### 6.2.2 *Allogenic controls*

*Tectonics*

The study area as a whole lacks evidence of significant tectonic activity like faults and folds, but an igneous intrusion which is thought to be post depositional does exist in section one. Tectonics discussed here is more of a regional scale.

During the Oxfordian and Kimmeridgian transition, there was maximum subsidence and the trend was similar in all parts of the basin though little variation did occur from one region to another. In the Tithonian and early Berriasian transition there was a decrease in the subsidence rate with a corresponding long term eustatic fall in sea level. The subsidence resulted in progradation of the paralic channel sandstones (Leinfelder 1987).

During the tectonic active periods, the catchment could have been uplifted with a corresponding increased gradient of the palaeoslope. This may have affected the climatic condition of the area towards higher precipitation in high-relief hinterland, leading to increase in discharge. In this case, the accommodation space created during the active rifting period is filled by sediments of the flashy flow system. The increased discharge and sediment supply then tries to keep the base level in equilibrium.

*Relative sea level changes*

Eustatic sea level changes become the major controlling factor when there is no tectonic activity in a particular area. But the Louriñha Formation was deposited during the Tithonian when tectonics was still active; hence deposition was syn-tectonic. During the Late Tithonian, there was widespread retrogradation of the clastic input into the basin (Leinfelder 1987). The cause is suggested to be syn-sedimentary tectonics resulting in local differential subsidence and uplifts, thereby causing relative rise and fall in sea level, respectively. This might be the mechanism of the flooding surfaces identified in the study area.

*Climate*

The paleosol found in section 1 is slightly less developed, but not very significant to those in section two, three and four. Paleosol development and type are strongly controlled by climate. Hence, the slight difference in paleosol development suggests that a major shift in climate was not a significant factor in changing the depositional style in the study area. But paleosol develop in arid to sub humid conditions during periodic precipitations. A slight increase in precipitation during the depositional time of the successions represented by the sections two, three and four could probably be a candidate factor for the difference.

However, climate might have an effect on the depositional pattern in the Louriñha Formation on a regional scale. Climate may have had an effect on channel sinuosity. In such a case, with increased discharge fluctuations coupled with frequent floods may have provoked erosion of the banks, leading to high channel mobility and sinuosity.

### 6.2.3 *Palaeocurrent measurements*

In the study area a total of about sixty-seven Palaeocurrent measurements were made and were mainly collected from trough cross stratified beds (facies C), planar cross stratified beds (facies D) and rarely from cross laminated sandstones (facies H). The cross stratification are well developed on the lower to middle portions of most of the channel sandstones.

Several measurements were made and the mean value calculated. There were significant variations in the paleocurrent measurements in all the sections. For example Log FWA-1 in section one gave values of $140^0$, $132^0$ and $176^0$ with a mean value of $150^0$, also FW-I in section two gave values of $149^0$, $240^0$, $142^0$, and $260^0$ and a mean value of $200^0$. The paleocurrent measurement in section one shows a strong south east flow direction, while paleocurrent measurements in section two shows a southwest flow direction. The ambiguity of the results could be due to one section being more sinuous than the other.

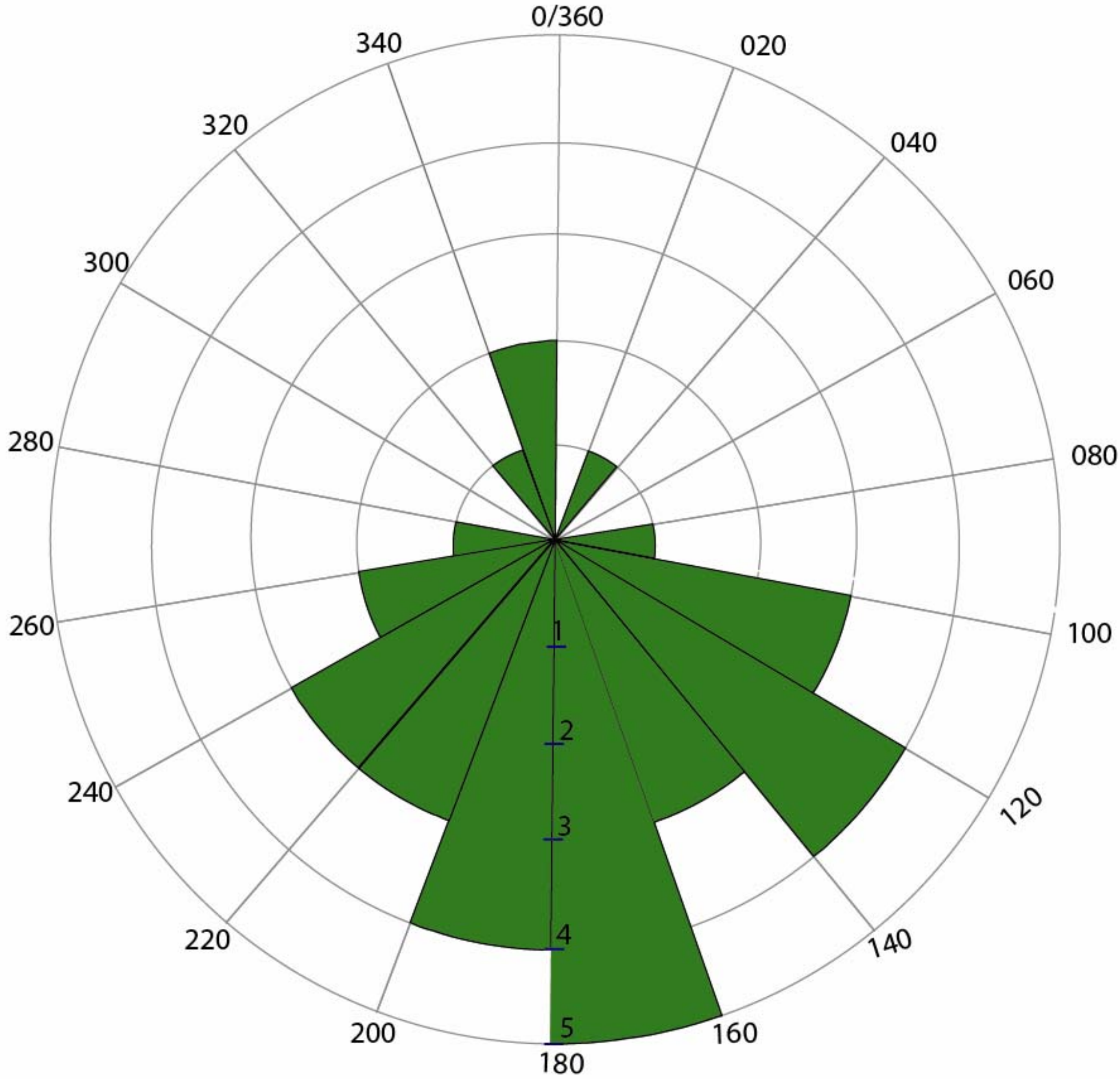



**Fig. 6.2:** Palaeocurrent rose diagram that summarises the flow directions for the measured sections in the study area. The azimuthal dispersion may be due to sinuosity in some parts of the channel

The distribution of the paleocurrent measurements in a rose diagram (fig. 6.2) indicates a more south easterly flow direction. Other factors may contribute to the other flow directions indicated in the rose diagram. For example measurements from overbank and floodplain deposits like crevasse channel (FA 2. 2) and crevasse splay strata (FA 3. 1) may be candidate factors, since they often give values perpendicular to channel flow direction.

**2.4 *Channel sinuosity***

Schumm (1993) proposed that through time, rivers will attempt to establish a stable graded profile for a given discharge of water and sediment. Disturbances in this system, such as sea level fluctuations, climatic change and tectonics, cause the stream to re-establish a new state of equilibrium by changing the external and internal attributes of the river. He further proposed that channel morphology evolves downstream in response to changes in valley slope, sediment load, bank materials, climate and tectonics.

The assessment of palaeochannels pattern or sinuosity needs to incorporate factors that include palaeochannels dimension reconstruction and paleocurrent variance (Bridge, 1985). Paleocurrent data from sedimentary structures were used to indicate channel sinuosity in the study area. According to Bridge (1985), a low dispersion in palaeocurrent values is consistent with a low sinuosity or braided stream. A higher dispersion of values can indicate deposition in higher sinuosity or meandering stream setting (Brett et al., 2006), though the above methodology have been considered by other authors to have potential errors (e.g. Miall, 1976; Le Roux, 1992). Variation of paleocurrent values in terms of channel sinuosity is discussed in section 6.2.3.

The sinuosity of a river may decrease with a decrease in the gradient. In case of the study area, changes in base level may be a dominant factor and the base level change will impose gradient changes. The river will respond to these changes by regulating its internal characteristics.

Increased sediment supply and yield may also affect channel sinuosity in the study area. A combination of factors could have been responsible for the sinuosity; including the nature of bed rock, tectonics and the frequency and types of sediment load. However, the degrees of sinuosity of most of the sandstone bodies are difficult to evaluate since the three dimensional geometry is not exposed.

According to Schumm (1981) high bedload content leads to the development of bars in rivers. When these bars are well developed, the river tends to change its course and especially when the overbank deposits are easily eroded. Differential weathering of the bedrock or floodplain deposits by wind or coastal currents may result in topographic irregularities and by the changing topography influencing sinuosity of the rivers on the coastal plain.

Intrabasinal tectonics exerts the dominant control on the mid-stream portion of the fluvial system by affecting the location and types of drainage patterns on the alluvial plain (Miall, 1981). Tectonic tilting associated with extension or thrust-related asymmetric subsidence superimposes a new slope on the pre-existing flood-plain topography (Wells and Dorr, 1987). Rifting in the Lusitanian Basin was obviously not symmetrical, thereby affecting the graded profile of the channel, this causes channel deflection and the channels tend to follow topographic lows.

Bank materials or floodplain deposits may also affect channel sinuosity. According to Allen (1984) cohesive floodplain deposits favour increased sinuosity, decreased meander wavelength and deep, narrow channels. The floodplain deposits of the study area are more of silty mudstone than clayey mudstone. But clays are more cohesive than silt. So high bank stability is suggested to have been present in the study area, though some variations caused for example by changes in base level may have influenced bank stability, and thereby increasing sinuosity. This is mostly seen in section two and three where increase in base level might have increased the cohesiveness of the silty mudstone resulting to more meandering.

## 6.3 Architecture and sequence stratigraphy

Transition from section one in the panel (fig.2.2) of the Louriñha Formation of relatively fluvial dominance through section two of estuarine influence and sections three and four of mixed estuarine and deltaic, represents a typical transgressive and retrogressive succession. The general decrease in grain size in the upper beds and towards the south in section four in most of the sandstone bodies may represent an increased distance from source.

### *6.3.1 Stratigraphically significant surfaces*

Recognising sequence stratigraphic surfaces was difficult since the study area is relatively small and the sandstone bodies are tilted southward. However, there were three flooding surfaces recognised in the study area. Key surfaces such as maximum flooding surface (MFS) and transgressive surfaces (TS) do not extend landward beyond the bay line ( e.g. Posamentier & Vail, 1988), but terrestrial equivalents to this flooding surfaces are suggested to occur within the flood plain deposits (FA 3). These flooding surface equivalents in terrestrial settings may result from variation in the ground water table in the coastal plain due to relative sea level fluctuations.

The erosional bases of the channel sandstone bodies and the presence of clayey mudstone (facies J) separating the beds within these sandstone bodies, may be due to autogenic facies changes, but more probably due to regional variation in base level. The erosive channel base and the interstitial mudstone horizons can be treated as sequence stratigraphic surfaces.

A basin-ward shift in facies should also be observed as sequence boundaries at the base of incised valleys (Van Wagoner et al., 1990). The presence of marine mudstone (Facies O) at the base of the lateral accretionary bar (FA1.2) could be taken as a sequence boundary, since a basal erosional surface has been formed at the base of FA1.2; the erosion has been into the marine mudstone beneath. Also the unconformity that separates the relatively more alluvial deposits from the estuary deposits is interpreted as a candidate sequence boundary. The surface indicates a basinward shift in facies. Sequence boundaries are usually represented by regionally extensive surfaces or unconformities. The erosional boundary on top of the marine mudstone and below the accretionary bar of fluvial origin (fig. 4.12) has only been traced for a very short distance of ~3 m. The lateral extent of the marine mudstone is not known. However, it is very probable that the marine mud must have had a rather wide extent in the basin, and thus also the erosional unconformity on top of it. By these reasons it is suggested that this erosional surface is a sequence boundary, though the hierarchical magnitude of the sequence boundary must remain unsolved.

In the study area, channel sandstone bodies that stratigraphically lie above sequence boundaries may represent lowstand system tracts (LST) or the lowermost part of the transgressive system tracts (TST) (fig.2.2). Three surfaces were recognised as flooding surfaces and taken as datum.

### *6.3.2 Channel infill with sea level changes*

The channel fill pattern of the study area is dominated by single channel fill units, though multi- storey channel sandstone bodies occur in few places. Single storey channel sandstone bodies are indicative of rapid channel infill, while multi-storey channel sandstone bodies with numerous erosional surfaces may indicate repeated erosion while infilling continues (Shanley & McCabe, 1993). The spatial distribution, geometry and stacking pattern of the sandstone bodies of the study area was largely controlled by changes in accommodation space caused by sea level fluctuations, tectonism and sediment infill (e.g. Leinfelder et al., 1998). The

Lourinha Formation was deposited during Mid-Tithonian and Earliest Berriasian (fig. 3.4a) a time interval dominated by the post-rift thermal subsidence. Therefore, it implies that the major controlling factor in the basin fill was eustacy followed by tectonism and climate.

Fall in base level might have created an incised valley and deposition of channel sandstone bodies, but this is unclear. Andsbjerg (2003) suggested that channel development probably is initiated during falling or static base level, and that lateral migration of channels during early base level rise may have caused erosion of a significant proportion of the lowstand deposits (e.g. fig. 6.4). Thus the channel sandstones described in sections 5.1.1 and 5.1.2 may represent lowstand deposits and subsequent transgression. The fining upward trend as seen in FA 1.1 (log-FW-B) and decreasing sandstone/ mudstone ratio is interpreted to be due to sea level rise. Slowly rising sea level enhances the development of multi-storey channel sandstone bodies. The multi-storey channel sandstone bodies, for example unit 8 in section two indicate that the channel was stable for a long time, for several erosion and infill events to take place.

**Fig. 6.3:** Large-scale progradation geometry (dashed lines) in section one. Also abundant floodplain deposits separating the main channel sandstones.

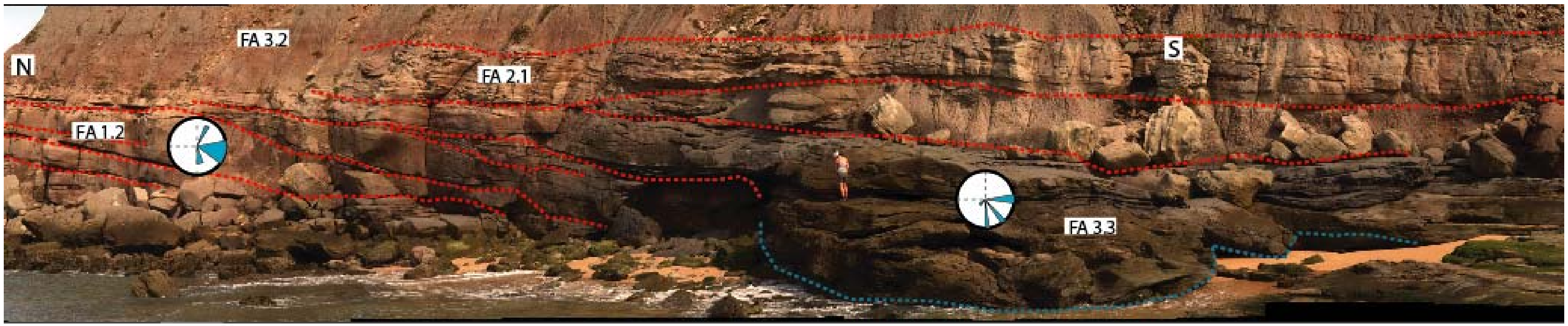



Fig.6.4: Minor progradational geometry (dashed red lines), characterised by lateral accretionary elements with multistorey channel fill. The channel (dashed blue line) is inetrpreted as an abandoned channel linked to the point bar. The rose diagrams demonstrate that both the LA element and abandoned channel are oriented normal to palaeoflow direction.

The increasing rate of creation of accommodation space during rising base level favoured high levels of storage of floodplain sediments and more isolated channel bodies, according to the general model proposed by Shanley & McCabe (1993). The very few multi-storey channel sandstone bodies and the dominance of single storey channel infill units in the upper part of the section suggests a relatively rapid sea level rise and the development of more floodplain fine (FA. 3), and overbank deposits (FA 2). Base level rise may result to changes of sediment-water load within the channel belt and consequently may lead to FA 3 and FA 2. Variation in the rate of discharge and sediment load from the source may also lead to more FA 3 and FA 2 When the rate of base level rise decreased during late transgression and early highstand, the decline in accommodation space caused decrease in the floodplain deposits and increase in the multi-storey, multilateral stacking of channel sandstones. In the study area highstand deposits could represent sandstone bodies in the upper portion of section two (above facies N), section three and four. The heterolithic facies located especially in the upper portion of sandstone bodies in this section may be due to differential sea level fluctuations during highstand.

### *6.3.3 Hypothetical depositional model*

From the results above, an idealised model for the evolution of the fluvial dominated Louriñha Formation is as seen in fig 6.1. The Lourinha formation could be characterised as a tide dominated estuary or delta seen from the characteristics of the deposits. However, majority of the sandstone deposits in this environment were fluvial in origin with limited marine deposits.

The major controlling factor for the deposition of the sandstone bodies is base-level fluctuations that may be related to eustacy, local tectonics or climate change. Mackey and Bridge (1995) proposed that during rising base level associated with wetter climate aggradation rate and avulsion frequency would be relatively high and greatest in the distal areas. Channels with large sizes may be indicative of increasing flood discharge. This may be one of the causes of the increased sinuosity as described in section 6.2.4.

Another hypothesis for the deposition of the successions in the study area could be changes in sediment supply from the hinterlands, which could be influenced by tectonism and/or climate. When the supply of sediment increases, aggradation rate and avulsion frequency would be relatively high and resulting in larger sinuosity pattern and variability in channel-belt orientations (Mackey and Bridge 1995). The increase in channel sizes may be related to climatically and/or tectonically induced increase in sediment discharge from the hinterlands, or due to distributive river systems (e.g. Legarreta and Uliana 1998)

# 7 Heterogeneity

Heterogeneity of reservoir sandstones in the study area is basically geared towards fluid flow. According to Thomas et al. (1999), several studies have been attempted to characterize reservoir heterogeneity, but the most descriptive classifications have been proposed by e.g. Weber (1986) and Dreyer (1993). The key to reservoir heterogeneity is the basic understanding of the processes and the various scales of sedimentary building materials from the microscale (e.g. particles and pores) to the gigascale (e.g. sedimentary environment and basin fill). Knowledge of the processes and the various scales of sedimentary building materials will also enhance the ability to characterise and predict the architecture of reservoir rocks (e.g. Cross et al., 1993).

### 7.1 Fluvial architectural style

Fluvial architecture can be defined as the three dimensional geometry and interrelationship of the channel deposits, and deposits from levee, crevasse splay and floodplain and other sub-environments of the fluvial system (Emery and Myers, 1996)

Understanding how the different architectural elements interact with each other both laterally and vertically is the basis of architectural style analysis in the study area. Also the stacking pattern, i.e. whether the architectural elements are isolated, laterally and vertically stacked gives a good assessment of architectural style. Fluvial architectural style can be viewed in 2D from outcrops, which may further be used to predict architectural style of sandstone bodies in 3D. Knowledge of the 3D architecture of the sandstone bodies could make it easier to determine the degree of connectivity within the reservoir sandstone and also with adjacent sandstone bodies. However, the architectural style, geometry, dimensions and thickness of reservoir sandstones is controlled by many factors but the most important are base level change and rate of avulsion.

### 7.2 Connectivity of sandstone bodies

According to Heller et al (1996), avulsion related stacking is difficult to observe because large scale avulsion is a poorly understood process. By relating avulsion to other factors like sedimentation rate Heller (1996. pp.305) suggested that ‘‘for a constant avulsion frequency, the average density and interconnectedness of sandy amalgamated channel belt deposits is inversely proportional to sedimentation rate”. This means the higher the sedimentation rate, the thicker the sediment deposited in the channel belt and floodplain between avulsion events. The resulting sandstone deposited will have minor amalgamated surfaces within channelised sandstone bodies and they are liable to be isolated from each other by floodplain deposits. Such a situation is analogous to FA1.1 in section 5.1.1. When the sedimentation rate is low, low amount of sediment is deposited between avulsion events, resulting to a higher rate of interconnectedness.

The internal and external geometry of the channel fills, coupled with the facies assemblage led to the recognition of the different channel fill elements as discussed in chapter 5, (fig. 5.3). Table 7.1 show estimates of the sand: gross ratio in the study area.

At the beginning of the study area, i.e. closest to Paimogo, there is a multi-storey channel fill element of FA 1.2 and above, the sand:gross ratio changes drastically. Section one is characterised by numerous channel and crevasse splays within a high proportion of mudstone deposit. However, there is one distinct channel sandstone element of ~12 m thick. The presence of many crevasse deposits and abundant mud will lead to poor communication between the various sandstone deposits.

Channel sandstone bodies in section 2 display both vertical and lateral amalgamation. Above the point bar (unit 8, from panel) there is a mudstone covering in which a flooding surface passes through laterally. The thickness of this mudstone unit is ~5 m thick, though the thickness is not uniform laterally. This mud can act as a vertical barrier to fluid flow.

Section 3 is characterised by thick channel complexes and a considerable levee development. The sinuosity here is difficult to assess, but the channel sandstone display a general vertical and lateral stacking pattern. Getting to section 4, the architectural style changes, i.e. from vertical and lateral stacking to more isolated channels, but the geometry of the sandstone bodies can not be compared to those in section 1. The sandstones are generally heterolithic with sand–mud laminae and the grain sizes are relatively smaller compared to those of the other sections.

**Table 7.1:** Approximate values of net:gross ratio in all the sections

| section | Approximate net: gross ratio (%) |
|---|---|
| 1 | 35 |
| 2 | 45 |
| 3 | 65 |
| 4 | 40 |

### 7.3 Levels of Heterogeneity

There are several levels of sandstone heterogeneity as classified by Thomas et al. (1998) and Weber (1986) and include gigascale, megascale, and macroscale, mesoscale, and microscale heterogeneity. Haldorsen and Lake (1984) in Miall (1988) subdivided the scales slightly differently, and came out with a scale of four levels. The various heterogeneity scales used in this project is as shown in fig 7.1

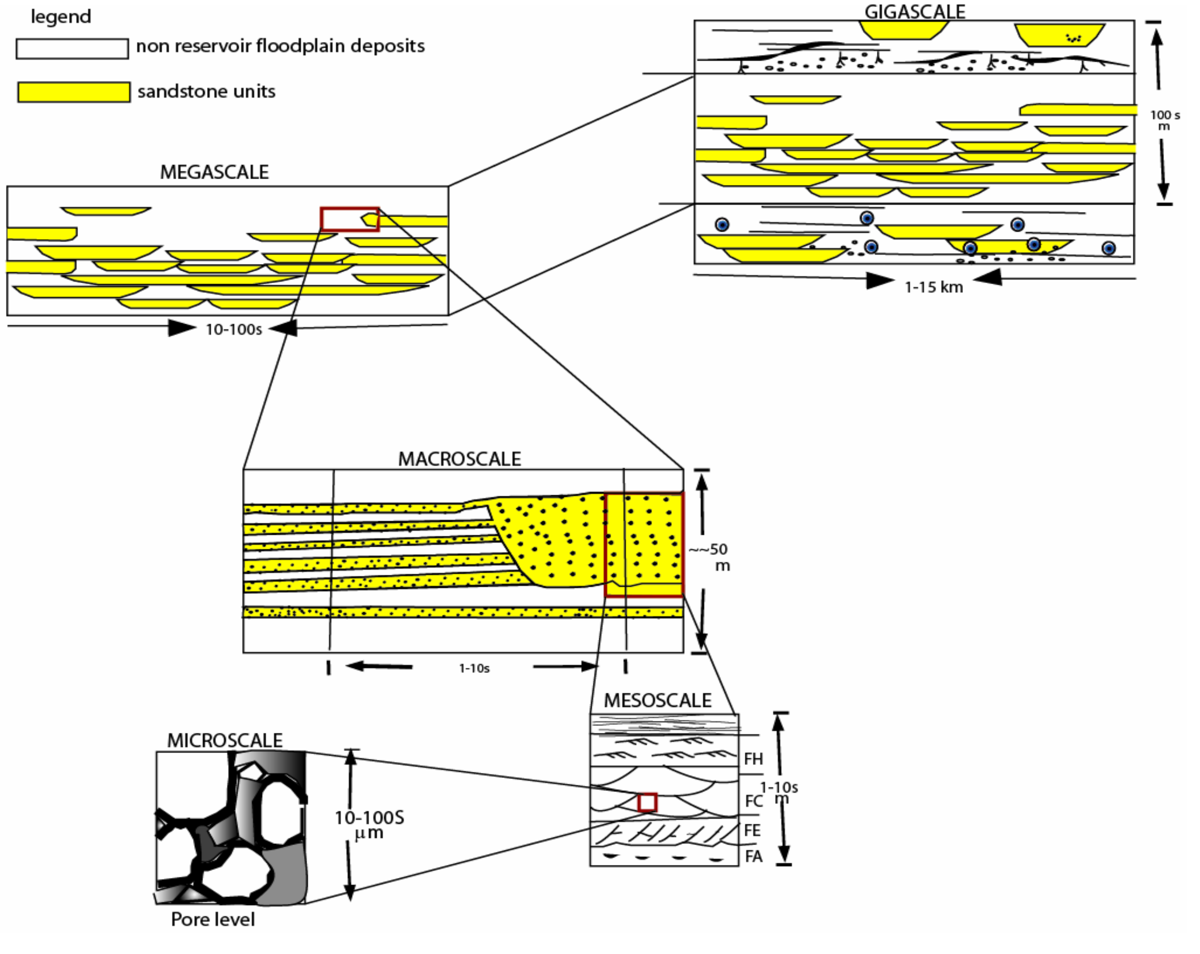



**Fig. 7.1:** Heterogeneity scales of fluvial sedimentary deposits. Modified from Nystuen and Fält, (1995).

Gigascale heterogeneities are influenced by the stacking of sandstone bodies within a period of regional and large-scale base level fall and rise (“sedimentary cycle”) and are controlled by long term base level fluctuations. Thickness may be up to decimetre scale. Megascale heterogeneities are based on a combination of architectural elements which may group to form a sedimentary cycle. The scale is usually of several metres in thickness, but decimetres to metre scale have also been recorded. Macroscale heterogeneities are at the level of architectural elements (e.g. channels and bars) and may be based on the lithology, clay content, geometry, permeability and porosity of these architectural elements. Mesoscale heterogeneities are on centimetre to decimetre scale and are based on sedimentary structures and stratification types. Microscale heterogeneities are based on particle texture, compositions as well as pore characteristics.

As already mentioned above gigascale heterogeneities are related to sedimentary cycles and in the context of this study can be referred to the Lusitanian Basin. Also megascale heterogeneities may represent the entire Louriñha Formation, but the study is concentrated on a small portion (~ 2.2 km) of this formation. So discussion will basically be from macroscale heterogeneity, though reservoir engineers and geologists are more interested with heterogeneities at intermediate scales.

### 7.3.1 Macroscale heterogeneity

Macroscale heterogeneity in relation to the study area is based on the succession between channel sandstones, overbank deposits (FA 2) and floodplain fines (FA 3). FA 2 and FA 3 deposits are supposed to have relatively lower permeability compared to channel sandstone bodies, and these fine grained facies then tend to disturb the vertical communication between channel sandstones and lateral communication within sandstone bodies. Sandstone body dimensions and geometry, orientation, sand/gross and net/gross ratios determine reservoir connectivity (Nystuen and Fält, 1995). The size, geometry and dimension of reservoir sandstones are controlled by the degree of amalgamation and connectivity of the channel sandstones, coupled with the possible elimination of floodplain deposits.

A change in the A/S ratio (rate of creation of accommodation space versus rate of sedimentation) may have effects on the heterogeneity, geometry and lateral continuity of fluvial to shallow marine sandstone reservoirs. Amalgamated and laterally extensive sandstones may have been deposited in low A/S conditions, as depicted from the sequence stratigraphic model of Shanley and McCabe (1993). In such a case, mudstone that usually occupies the upper parts of channel infill successions is not preserved. Sandstones deposited in low A/S conditions e.g. FA.1.2 are more homogenous, porosity is relatively high and reduced mudstone content promotes high permeability and connectivity is enhanced (fig. 7.2). Also channel sandstone bodies toward the base of section 3 tend to be more amalgamated and interconnected, being associated with few floodplain deposits, whereas toward the top of this section floodplain deposits increases in frequency and the channel sandstones are relatively thinner.

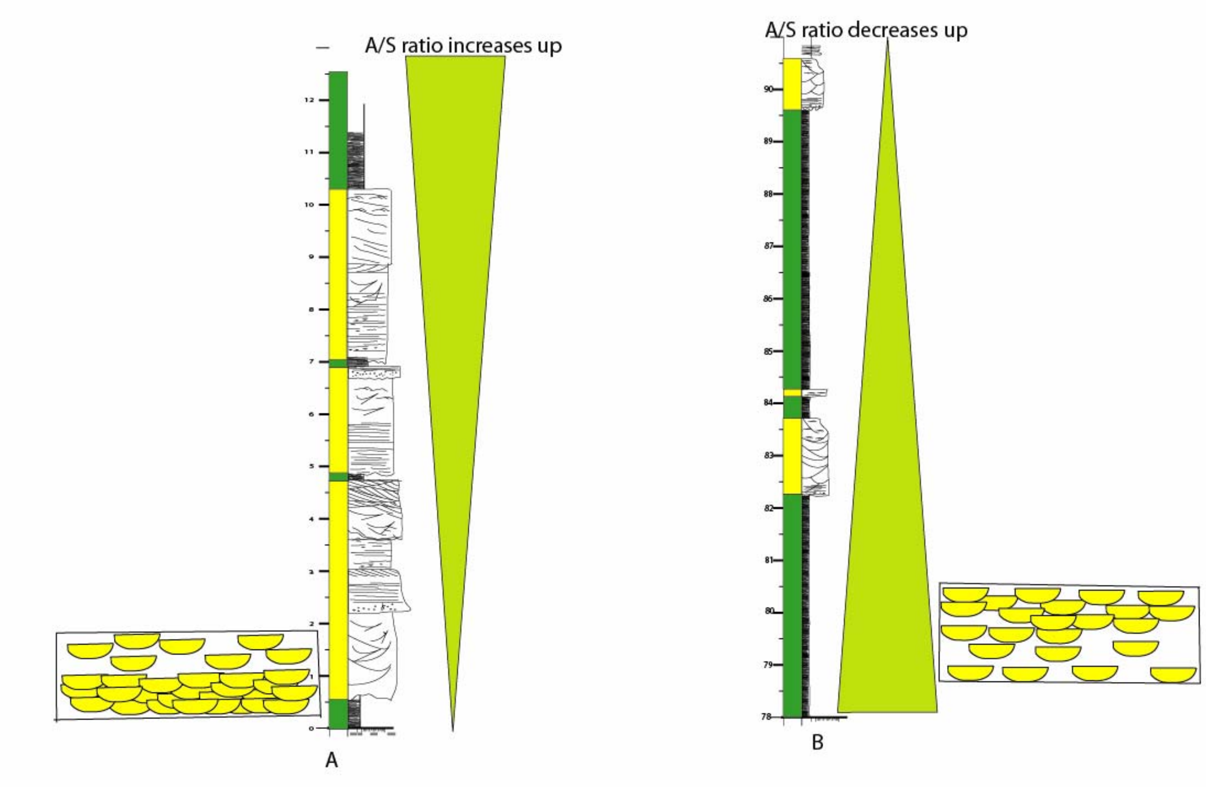



**Fig. 7.2:** Diagram showing relative preservation of channel sandstones (A) low A/s conditions and (B) high A/S conditions, increased amalgamation and multi-storey channel enhances connectivity in (A) while isolated channels in (B) reduces connectivity.

In contrast, high A/S ratios give rise to lower amalgamation rate of channel sandstones, and increased preservation of floodplain deposits (fig.2.2). Channel sandstones become single storied and some isolated sandstone bodies are embedded in floodplain deposits. In the fluvial dominating area, e.g. FA 1.1, porosity and permeability generally decrease upward, since the upper part becomes heterolithic containing more mud.

Due to their different sedimentologic attributes, the various channel fill elements (fig.5.3) possess particular reservoir characteristics different from other deposits e.g. FA 2.2 and FA 3.I. The lateral and vertical thickness of each of the sandstone units, the degree of amalgamation within each other and the stacking pattern may control the connectivity of the sandstone bodies, as seen in section two and three. Channel sandstone represented as unit 12 from the panel, has a considerable vertical thickness (> 15 m in the medial portion) of channel sandstone, and is also laterally extensive > 150 m with good continuity. Since this sandstone unit 12 has erosive basal contacts its degree of incision into the underlying laterally extensive channel bar sandstone deposit unit11 controls its connectivity. But horizontally channel sandstone unit 11 is not as extensive as unit 12; this might reduce their vertical connectivity. However, in this same channel fill sandstone units, mudstone laminae and mudstone clasts are scattered within the bed units. This might affect the total reservoir volume in the sandstone, although their patchy distribution does not affect the connectivity within the sandstones seriously.

The accretionary/point bar element unit 8 from the panel, having lateral extent ~105 m and thickness ~ 8 m consists of inclined strata made dominantly of mud (fig. 5.2). The strata are made of sandstone rich package that become amalgamated downdip with channel element unit 9 and above is covered by mudstone and facies N. The sandstone packages consist of clinothems and each clinothem unit has low lateral continuity (range 2- 6 m). The inclined sandstone clinothem units are separated by continuous and discontinuous thin bedded intervals consisting of mudstones, forming local permeability barriers, hence reducing vertical permeability.

As discussed in section 6.2.1 (under autogenic factors), avulsion influences the distribution and stacking pattern of channel sandstone bodies. Allen (1978) suggested that fluvial channel sandstones deposited during high frequency of avulsion relative to subsidence rate form vertically and laterally interconnected blanket-like sandstone bodies that could function as single compartment reservoirs. By contrast, channel sandstone bodies may become more isolated if deposited during low frequency of avulsion (Cross et al., 1997). Transition between the situations above can be observed in section 3 and 4. Fig. 2.2 shows a transition of relatively more interconnected, amalgamated sandstone bodies and less floodplain deposits in section 3 to less interconnected sandstone bodies with more floodplain deposits in section 4.

### 7.3.2 Mesoscale heterogeneity

Mesoscale heterogeneity generally concerns heterogeneity within individual sandstone bodies that form internal barriers to fluid flow such as sedimentary structures, bedding, and accretionary surfaces. Internal fluid flow barriers are especially mudstone plugs formed as abandoned channel infill, discontinuous mudstone horizons, and lag conglomerates with mud matrix occurring along bottom of channel sandstone beds. Most of the literature discussed below is sourced from Cross et al. (1997).

From the logged section (see appendix B), the sandstones reveal good vertical stratifications, some due to grain size variation, but some are filled with mud. The presence of sandstone beds of fine grain size alternating with medium grain size sandstone in the same sandstone but in an overall upward fining trend may result in decreasing vertical permeability. This type of channel fill succession, being generally characterised from bottom to top by trough cross stratified sandstone (facies C) and planar cross stratified sandstone (facies D), coupled with ripple laminated sandstones (facies H), parallel laminate sandstones (facies I), and mud drape sandstones (facies F), with thin mudstone laminae may have high horizontal permeability within bedsets of high porosity

Variation in the A/S ratio may also have an effect on the proportion of mesoscale heterogeneity in a sandstone body. In trough cross stratified sandstones (facies C), permeability changes may be coincident with variations in sedimentary facies attributes which change as a function of stratigraphic position (Cross et al., 1997). In case of channel sandstone, e.g. unit 4 which is thought to be deposited during high A/S conditions, permeability will generally be highest at the channel base and gradually decrease upward, as a function of a general decrease in grain size. When the A/S conditions become higher, there will be decreased cannibalism of the previously deposited sandstone and floodplain deposits, and a corresponding decreased amalgamation of sediments existing in the active channel, results in increased facies diversity (e.g. facies J), decreased channel sandstone continuity, increased preservation of finer grain sediments and poor sorting.

Figure 7.2 A and B illustrates how permeability varies within a channel sandstone body with increased and decreased A/S conditions. An increased higher proportion of ripples and parallel lamination in the upper part of most sandstone bodies in the study area suggests that the sandstone bodies may have been deposited during an upward increase in the A/S ratio. Thus, also the permeability can be anticipated to be reduced upward.

### 7.3.3 Microscale heterogeneity

Microscale heterogeneity plays a vital role in predicting the diagenetic changes and reservoir properties of sandstone bodies. The fluid flow in reservoir sandstones is closely associated not only with constituent grains, but often more significantly with the pore structures at microscale, which are formed during the coalescence of constituent grains and subjected to a series of modifications due to factors such as compactions, cementations, etc. (Shuwen et al., 2007).

Microscale heterogeneity includes variation of the sandstone properties caused by grain size variation, sandstone composition, texture and carbonate cementation. Microscale heterogeneity is related to petrographic studies which are based on examination in thin sections. During petrographic analysis attention is based on compositional, textural and porosity variations and how these properties could assist in provenance estimation.

## 7.4 Sandstone petrography

The sandstones encountered are moderately mature representing first generation sands. The sandstones generally show intergranular porosity and rare intragranular porosity.

Quartz occurs as colourless in plane-polarised light, with white, grey to dark grey colours under polarised light and is the dominant mineral observed in all the samples. A few proportion of the quartz mineral could be observed as sub-rounded, but sub-angular grains constitute the majority. A faint dark overgrowth surrounds some of the quartz grains. The overgrowth may probably represent clay or organic matter.

The feldspars have relatively low relief and appear to have a grey to white interference in polarised light associated with few cleavages. The dominating feldspars observed include microcline or K-feldspar, and plagioclase feldspars. Their grain size and roundness are similar to quartz grains in most of the samples. The K-feldspar or microcline could be observed as having cross hatches in polarised light. Plagioclase feldspars are characterised by albite twin, having parallel narrow zones which appear as dark and white bands and are relatively not as many as microcline grains in all the samples examined (table 7.2).

Micas include both biotite and muscovite (or light mica). Biotite is dark brown to green in non-polarised light, whereas the light mica is colourless to faint green. The presence of cleavage is very diagnostic for the micas. The distinguishing factor between them is the presence of few cleavages on the muscovite grain. Biotite also appears brown to dark brown when the stage of the microscope is rotated. Quantitatively mica fragments are the smallest in relation to quartz and feldspars. The mica grains are tabular in shape with variable sizes.

In most samples, calcite cement partly or wholly fills the pore spaces. They were recognised as having the highest birefringence with few cross-like cleavages. The clay of mud observed occur as intraclasts, and some could be seen as coats surrounding grains.

In one sample designated as ‘FS’ or flooding surface, there is abundance of calcite cements and presence of calcareous clasts. Quartz could also be identified associated with some larger shell portions, but were difficult to distinguish with smaller shell fragments. In this sample, porosity is completely lost by the calcite cement.

**Table. 7.2:** Values determined from point counting of samples representing fluvial, tidal and crevasse sandstones in terms of dominating architectural elements. It is noticed that the sample representing fluvial architectural element has the highest pore space and corresponding low calcite cement whereas sandstone derived from tidal and crevasse splay architectural elements have low pore space with high calcite cement.

| Sample name | 13 | Im | L1 |
|---|---|---|---|
| Total No. of points | 506 | 507 | 509 |
| Architectural elements | Fluvial | Tidal | Crevasse |
| Minerals | | | |
| Quartz | 41,3 | 28,4 | 37,4 |
| P- feldspar | 2,5 | 2,5 | 1,7 |
| K - feldspar | 11,1 | 22 | 22,9 |
| Leached feldspar | 3,1 | 2,1 | 0,2 |
| Mica | 8,3 | 8,2 | 9,4 |
| Calcite | 6,8 | 26,7 | 19,5 |
| Mudstone | 0,7 | 0,7 | 0,1 |
| Organic matter | 1,3 | 1,9 | 1,5 |
| Pore space | 24,9 | 7,5 | 7,3 |

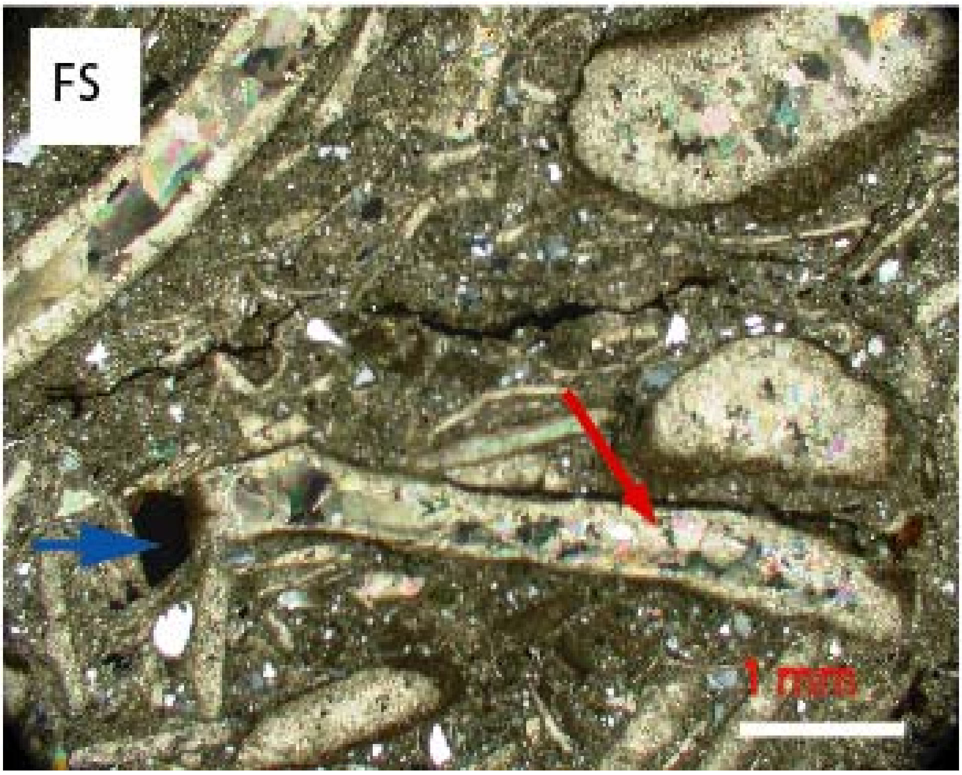



**Fig. 7.3:** Thin section photomicrographs showing coal fragments (blue arrow) and calcareous clasts (red arrow) within bioclasts and micro calcite cement filling all the pores. This might be a typical barrier to fluid flow.

There is no major difference in the mineral composition of the samples from the fluvial dominant, tide dominant and the crevasse splay architectural elements. Quartz and feldspar grains constitute the major components. However, variation occurs in the grain size, texture, and degree of porosity and cementation in the different samples. Thin-section photo micrographs of each of the tide dominated, fluvial dominated and crevasse sandstone are represented in figures. 7.4 to 7.6

The tide dominant sample shows a close packing structure with more filled intergranular pores by calcite cements and no quartz overgrowths (fig. 7.5). The presence of high calcite cement may be a contributing factor for the low porosity value. The sample from crevasse deposit equally has a close structure with smaller grain sizes (fig. 7.4). The low porosity values could have been enhanced by mechanical compression coupled with the presence of calcite cement which ranges up to 19, 5 % of the grain volume.

The fluvial sample has an open structure, the grains are relatively larger with larger size variation and few cementations occur mostly on the grain surfaces (fig. 7.6). This sample (sample 13) contains relatively smaller amount of mud ranging up to 7% of the grain volume. The pores seem to be partly filled by mud. Some charcoals were equally identified and occur within the pore spaces.

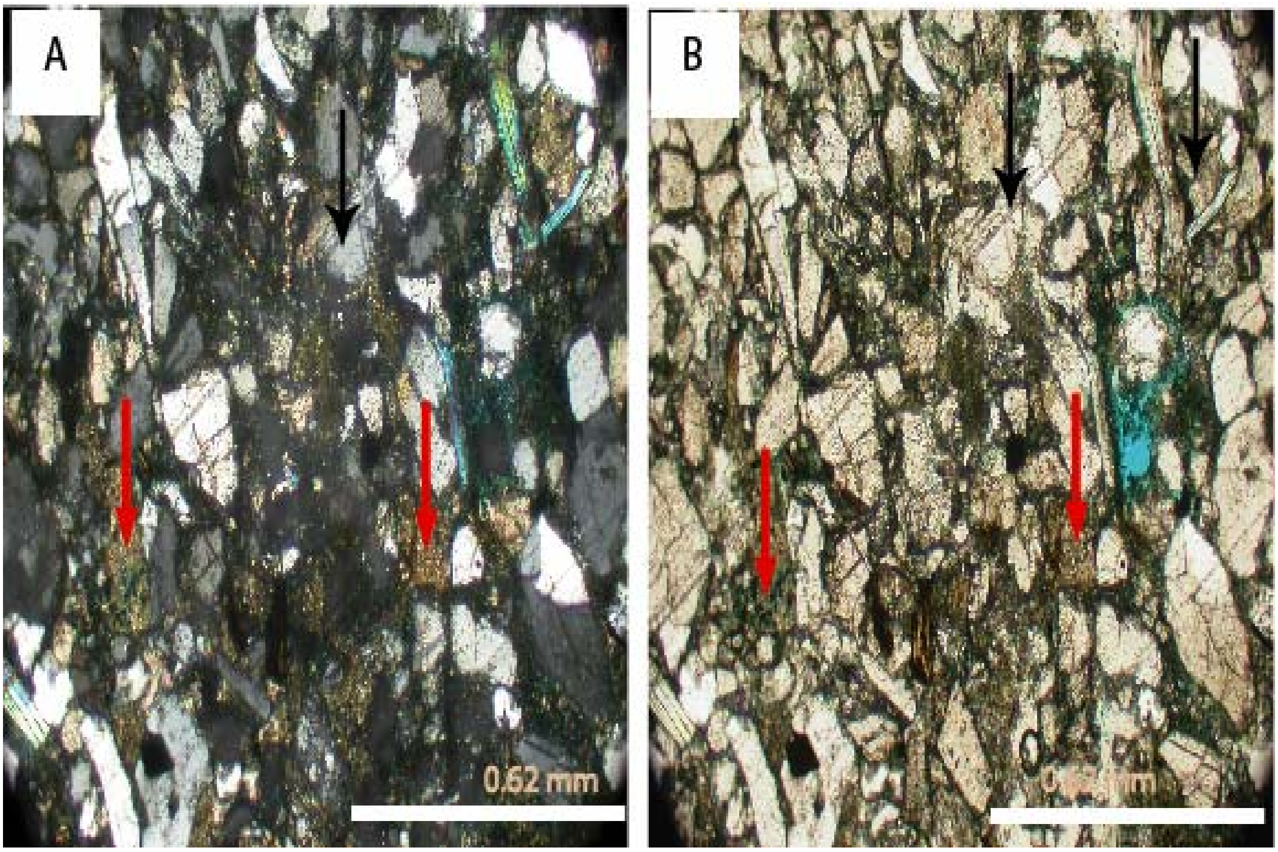



**Fig. 7.4:** Thin section photomicrographs showing sample L1 (crevasse splay architectural element (A) polarised light and (B) non polarised light, calcite (red arrows) and feldspars (black arrows) filling pores of tightly packed grain framework. The blue epoxy represents pore space. Magnification= 10x

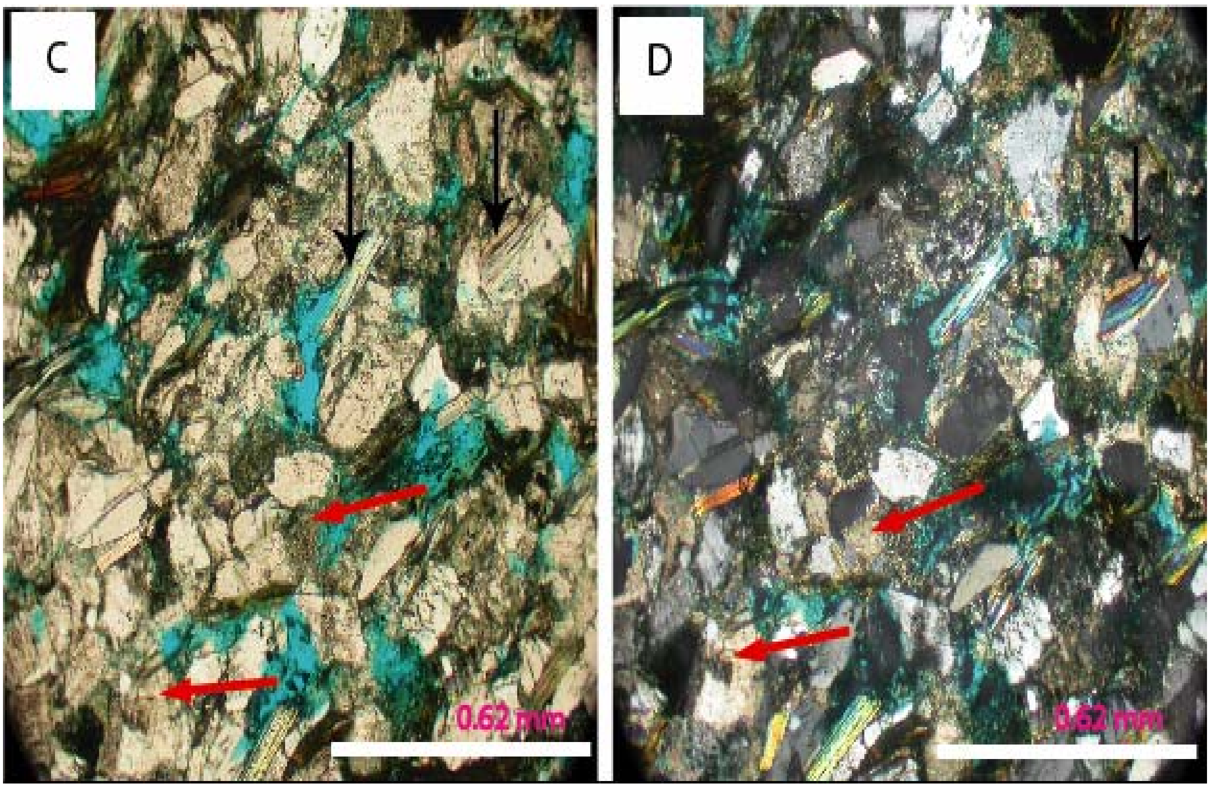



**Fig. 7.5:** Thin section photomicrographs showing sandstone sample Im (tide dominating architectural element) showing pore distribution, (C) non polarised light and (D) polarised light. Calcite (red arrows) occurs both as grains and cement, mica (black arrows) occur as tabular dark brown to faint green under polarised light. The blue epoxy represents pore space. Magnification10x

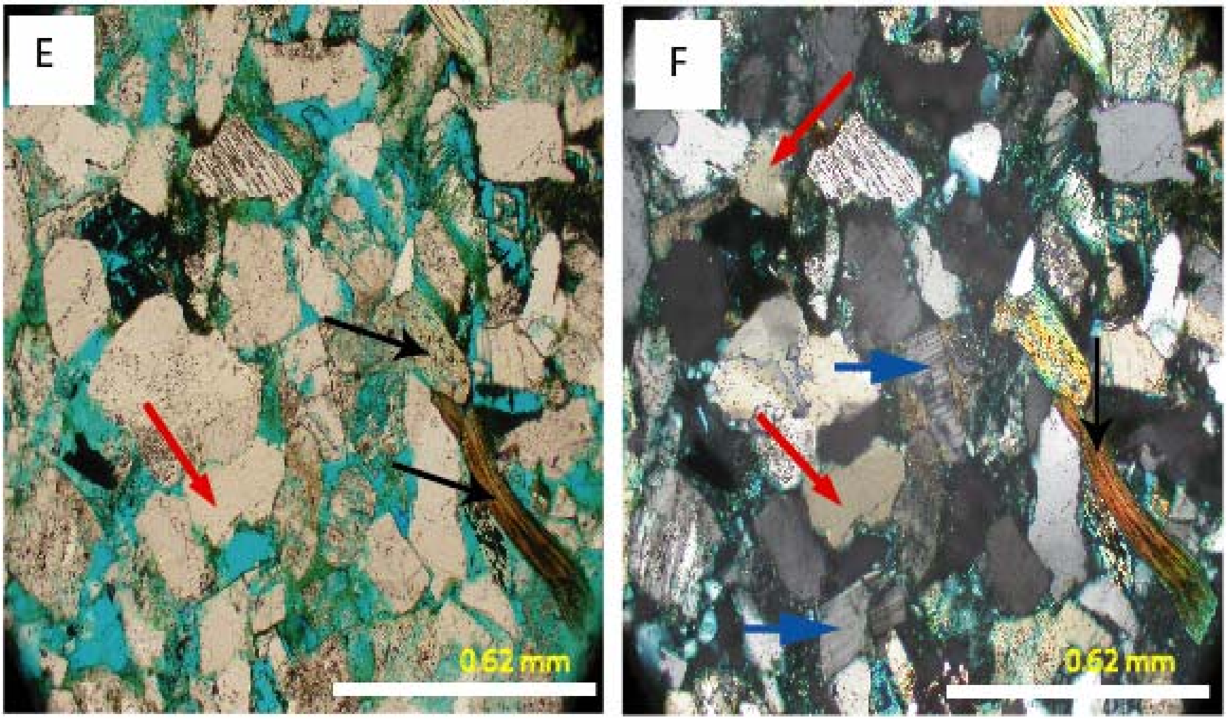



**Fig. 7.6:** Thin section photomicrographs showing good intergranular porosity in sandstone sample 13 (fluvial dominating architectural element), (E) non polarised light and (F) polarised light with medium grained angular to sub-angular grains. Few calcite overgrowths (red arrows) could be identified, albite (blue arrows) with dark and white bands could be identified. The blue epoxy represents pore space. Magnification = 10x

### 7.4.1 Possible errors

limited Due to knowledge in identifying minerals in a microscale there could be possible mistakes in mineral identification. There were possibilities for minerals like quartz and leached feldspars to be confused with each other. The above uncertainties can affect the overall percentage calculation after point counting. However, uncertainties for the pore volume calculation was limited since the blue epoxy was clearly visible in plane polarised light.

## 7.5 Overview of reservoir character

The main porosity identified in all the samples was intergranular porosity with very little intra granular porosity. Individual intergranular porosity is polyhedral in shape. The potential reservoir sandstone among the samples shows varying degrees of pore filling calcite. Some samples are easy to disintegrate while others are hard, e.g. “FS” (fig. 8.3) is very hard and well cemented. This might be due to diagenesis of the shell rich siltstone. There is no evidence of deep burial of the samples analysed, since no quartz cement could be identified Mudstone, especially facies J occur in varying proportions in all the samples and as pore filling materials, grain coating and in few occasions as pore lining. The mudstones are suggested to have been produced from secondary processes, such as leaching of the dominant mineral grains.  In the three samples in which point counting was performed, the occurrence of mudstone were few and scattered. This quantity cannot significantly affect reservoir quality.

In a location were there are several channel sandstones which are interconnected e.g. in section 3 changes in the rock properties such as cementation, and mineralisation may disturb fluid flow within sandstone bodies. Also increased amounts of finer grain particles and mudstone reduces fluid flow

# 8 Reservoir modelling

Geological reservoir modelling involves the characterisation and spatial distribution of reservoir features that influence fluid flow. Understanding the 3D geometry, architectural style and heterogeneity in relation to the interconnectedness and fluid flow properties of the sandstone bodies in the study area were the prime goal in the previous chapters. The model to be created is a combination of information from logs, sandstone geometry, architectural style and the correlation efficiency of the reservoir units in the depositional system (e.g. Purvis et al., 2002). Uncertainties exist in the distribution of sandstone bodies in the subsurface; hence stochastic reservoir modelling is used to remedy the crisis. For purpose of this study, three zones have been created, and each of the zones represents surfaces/horizons that cut across significant distances in the logged area. The zones are divided as defined below in fig. 8.1

The motive of this modelling is to construct an architectural element model of the study area. Special attention is focused on macroscale heterogeneity of the outcrop. The model is built using stochastic methods in which parameters controlling the modelling algorithms are based on observations from the field.

### 8.1 Parameters

Modelling outcrops where data s available provide important information for modelling of subsurface depositional systems of specific petroleum reservoir of type for which the Louriñha Formation might act as an analogue formation (e.g. Statfjord Formation in the North Sea). In reservoir modelling, representing the geology of a subsurface system, it is important for the input parameters to be representative of the depositional system under study (Brandsæter et al. 2005)

However, the data collected in the field are 2 dimensional, though 3D parameters such as wavelength, amplitude, and widths are needed in order to carry out the modelling. According to Nami and Leeder (1978) the dimensions of fluvial sandstone bodies derived from outcrops are relatively uncommon and are derived from widely different depositional environments. Also errors in measurements during data collection might affect the input data. Another uncertainty could be due to the fact that most analogue outcrops are not well exposed and only a partial view of the sandstone bodies is available.

*Channel depth and width*

Attempts have previously been made for the assessment of palaeochannel depth and width as well as channel belt dimensions (e.g. Bridge, 1985; Bridge and Mackey, 1993). The dimensions are essential to assist in determining the interconnectivity and continuity of subsurface sandstone bodies. However, the parameters measured from the outcrop may not be a true representation of the actual dimensions of the original channel sandstone. Due to the above discrepancies, empirical equations have been derived, in which for example, mean bankfull channel depth could be used to calculate the bankfull channel width

$Wc = 8.88d_m^{1.82}$ Bridge and Mackey (1993)

Wc is the bankfull channel width, and $d_m$ is the mean bankfull depth. However, care should be taken while using these empirical equations since they are derived from modern fluvial channels, hence they should only be used as guidance when applying the parameters to ancient fluvial channels.

Other parameters, like wavelength especially for sinuous channels was estimated by Leeder (1973) and is related to the bankfull channel width as seen in the equation below.

$L = 10.9\ w^{1.01}$ Leeder (1973)

The parameters used for the modelling is as seen in table 8.1

*Orientation*

To carry out the modelling, it was assumed that the orientations of the sandstone bodies were NW-SE, and that there is no other orientation pattern of channel sandstones. However, there is no indication of any possible local tectonic movements that might have affected the orientation of the sandstone bodies.

**Table 8.1:** Input parameters used for the object modelling

| zone | Thickness (m) | Channel width (m) | Channel-belt width (m) | Amplitude | Wavelength | recommendations |
|---|---|---|---|---|---|---|
| 1 | 6 | 200 | 1500 | 4000 | 17000 | Stacked channel low A/S ratio |
| | 1 | 35 | - | 200 | 900 | Abundant cr. splay with few ch. High A/S ratio |
| | 4 | 200 | | 800 | 5000 | |
| | 5 | 250 | | 3700 | 15000 | |
| 2 | 8 | 80 | - | 9500 | 7000 | More sinuous channels and low A/S ratio. Amalgamated channels |
| | 11 | 280 | - | 11000 | 20100 | |
| | 13 | 1800 | - | 13500 | 35000 | |
| 3 | 4 | 60 | - | 1500 | 950 | Relatively higher A/S ratio compared to section 2. lower sinuosity. Stacked but amalgamation is lower |
| | 6 | 250 | - | 1050 | 2900 | |
| | 9 | 1000 | - | 2050 | 7000 | |
| | 3 | 400 | - | 2500 | 4000 | Lower sinuosity with few cr. Splay. |
| | 8 | 1500 | - | 6500 | 20000 | |

### 8.2 Modelling approach

The reservoir modelling package Petrel ™ is used to build an object based stochastic model using data of sand/gross ratio collected in the field (table 8.1) and published sandstone body width/thickness that may seem likely.

Data used for creating the model include: stratigraphic logs mapped in the field (see appendix) and panel (fig.2.2) characterising the architectural style and correlation of the sandstone deposits in the study area

The total area modelled has a vertical range of ~145 m and a width of 2200 m. The sedimentary log data were converted into a text file, that the Petrel program could utilise and each of the logs were designated, its name, stratigraphic thickness and position in the study area. Well-tops were tabulated separately and were based on laterally correlatable surfaces in

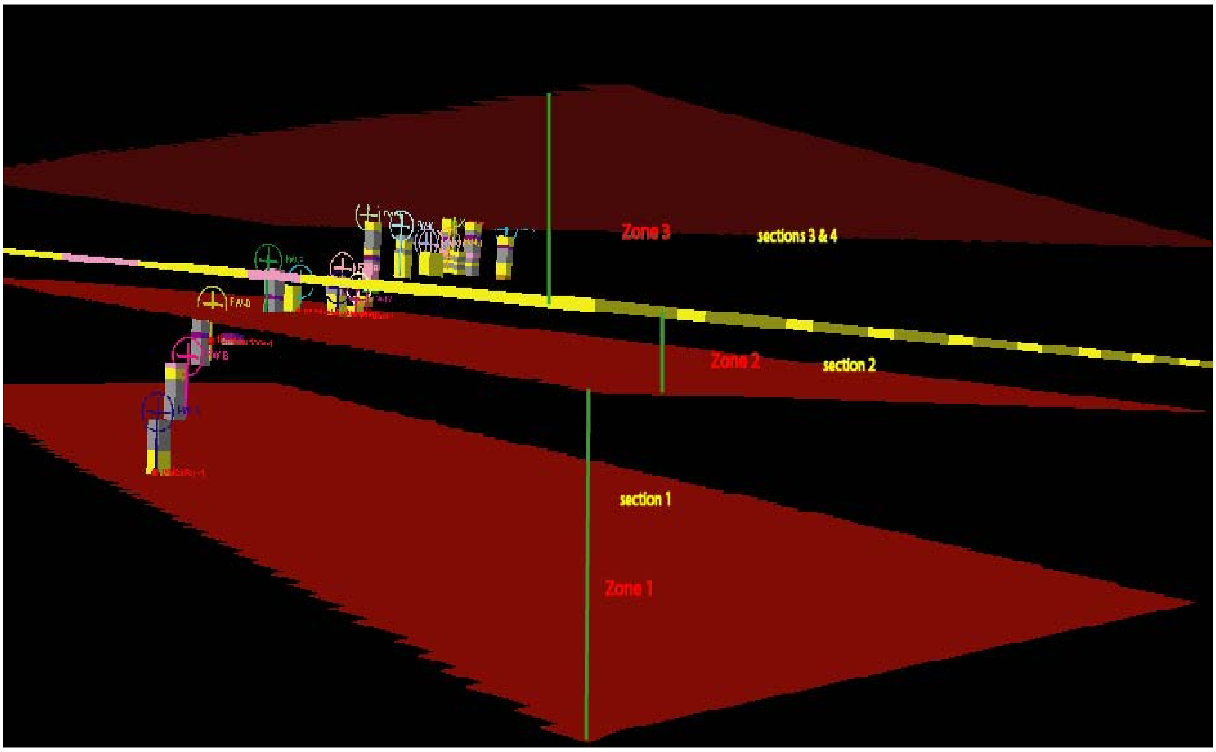



**Fig. 8.1:** Diagram illustrating the various zones used in the modelling. The cylindrical blocks represents upscaled logs. Note: The sections are not exact the zones since the sections run vertically and the zones run horizontally (see panel, fig. 2.2 for clarification)

relation to the datum. Building the framework of the model started with importing the logs (pseudo-wells) and well tops. The pseudo-wells represent the various facies and architectural elements encountered in the study area. Surfaces and horizons were created from the well tops including the top and lower boundary. Lateral correlation of the wells is based on the well tops. The horizons and surfaces serve as the framework in the model in which the architectural element distribution can be modelled. Gridding of the volume between the horizons (zones) then followed to create cells filling the zones.

The next step is to manually correlate the facies (architectural elements) into the pseudo-wells. The pseudo-well data are further upscaled on the basis of the dominant architectural element (fig.8.1)

**8.3 Stochastic object modelling**

Three runs were made, and on the basis of suggested channel dimension measurements (Table 8.1), a value is assigned as input to each of the architectural elements based on geometry and measurements made in the study area. The models created are as shown in figures 8.2 to 8.4

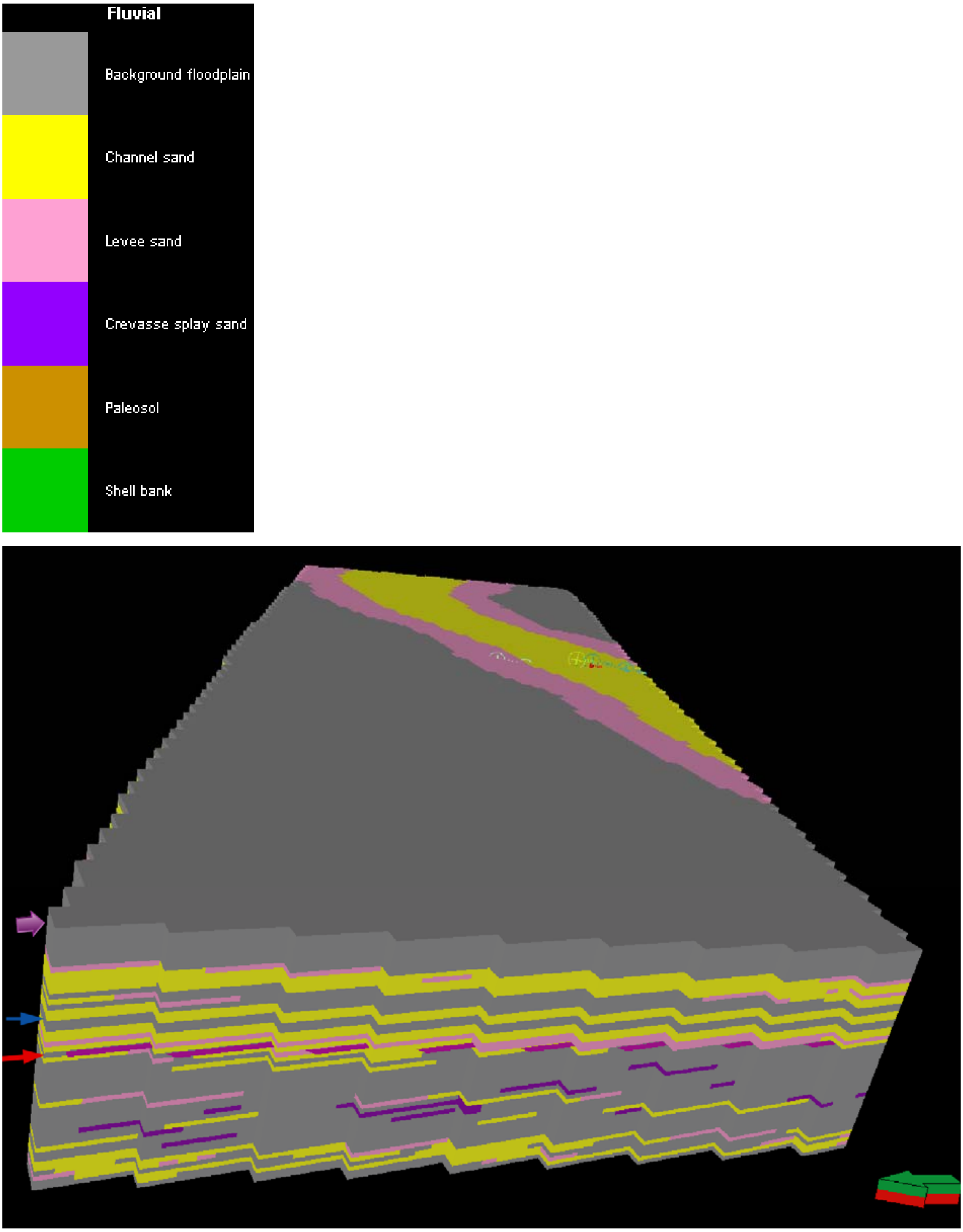



**Fig. 8.2:** Stochastic object Model created after run 1. Arrow points North (green= top and red = bottom). Fluvial facies (above); grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand. Red arrow points to top of Zone 1, blue arrow points to top of Zone 2 and purple arrow points to top of Zone 3. Vertical extent of profile ~145 m and width ~2200 m.

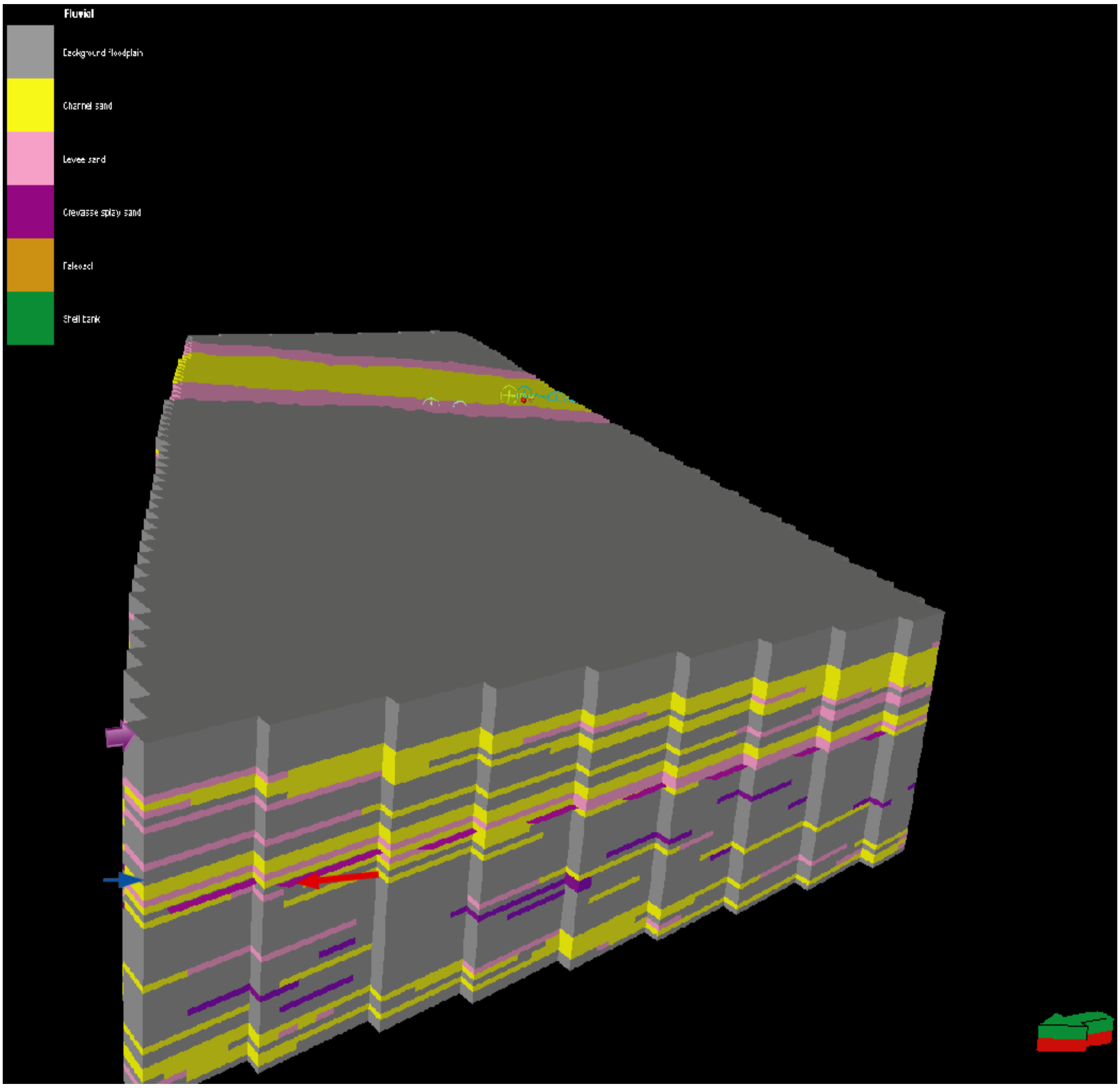



**Fig. 8.3:** Model created after run 2. Facies; grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand. Red arrow points to top of Zone 1, blue arrow points to top of Zone 2 and purple arrow points to top of Zone 3

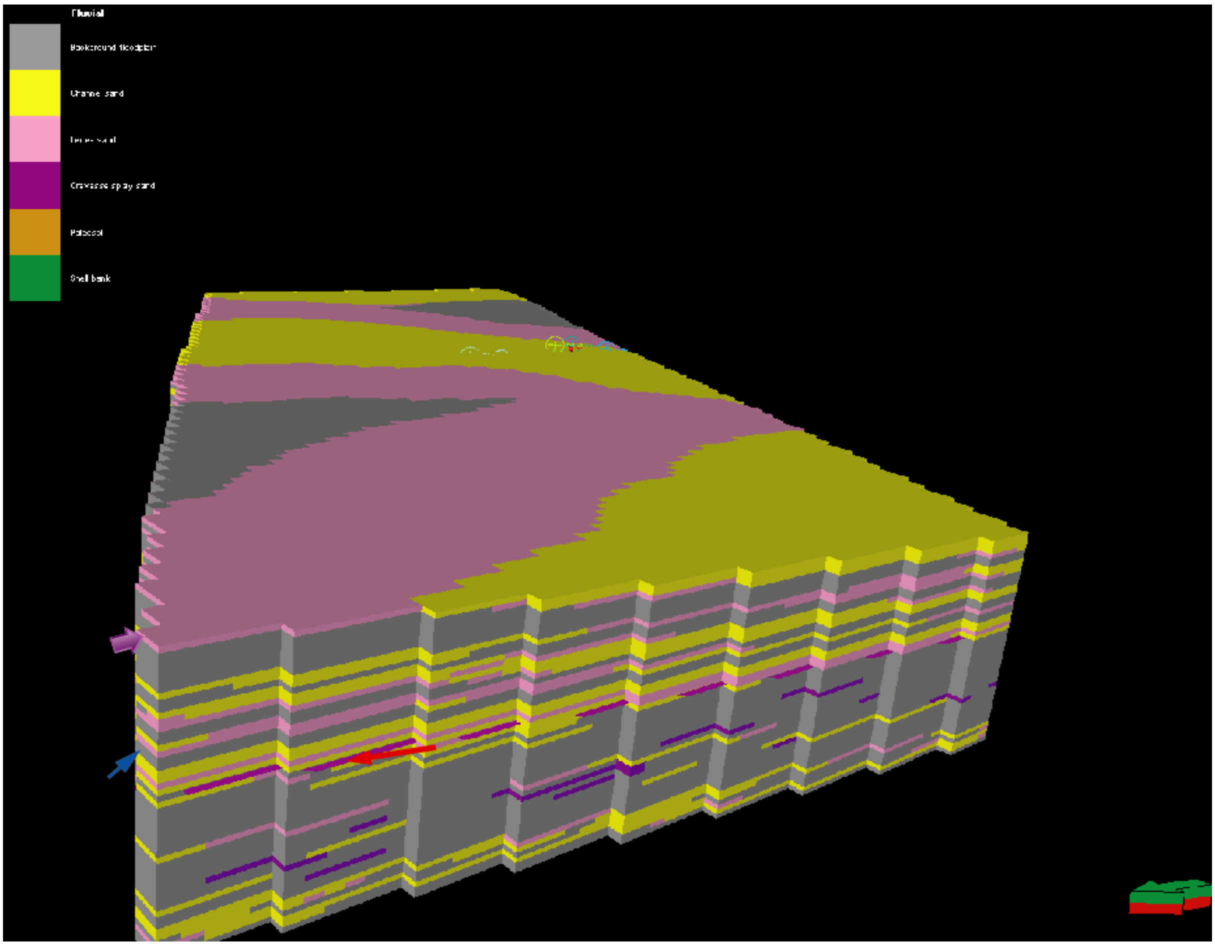
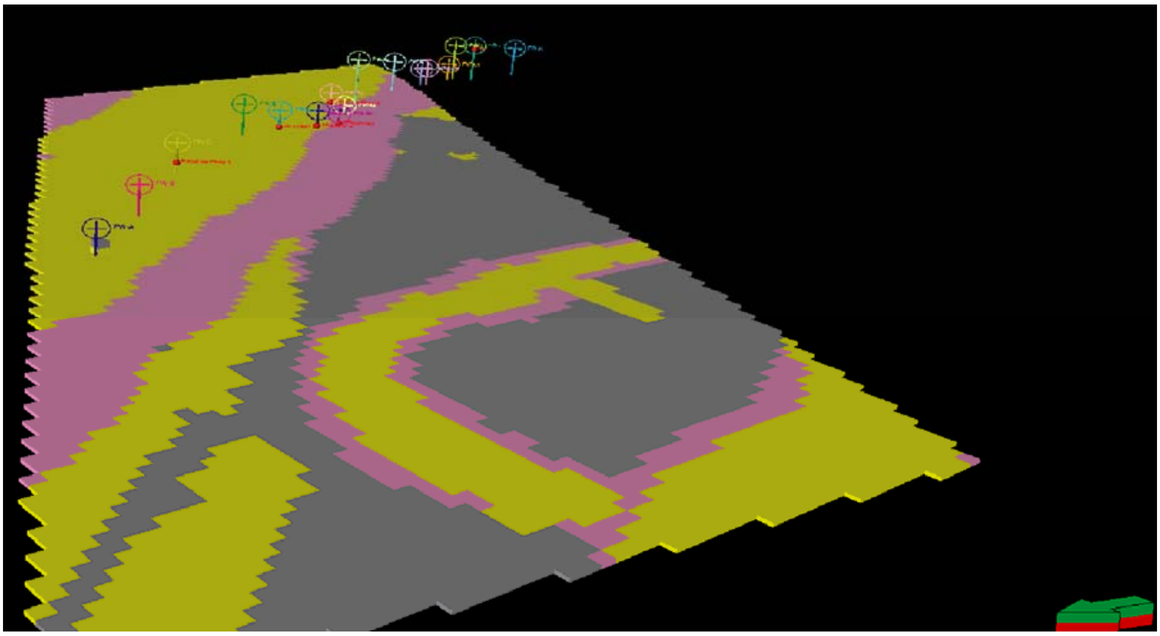



**Fig. 8.4:** Model created after run 3. Facies; grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand. Red arrow points to top of Zone 1, blue arrow points to top of Zone 2 and purple arrow points to top of Zone 3

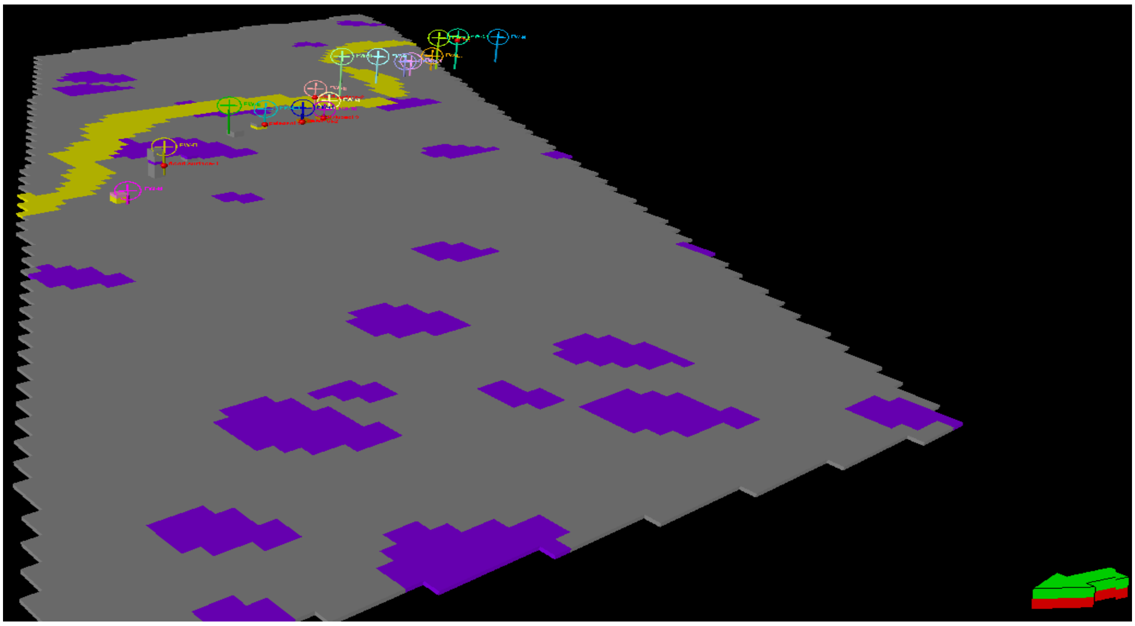
## 8.4 Results

The three models better explain channel sandstone body interconnectedness in the three zones. The upper part of Zone 1 is dominantly made of floodplain fines and crevasse deposits having little interconnectedness both vertically and horizontally. The lower part of Zone 1 constitutes some sandstone bodies which are interconnected and amalgamated at some points. These sandstone units are relatively thin and may not be of good reservoir quality.

Zones 2 and 3 display both good sandstone interconnectedness, though between Zone 2 and 3 there is a thin barrier or flooding surface (see Ch.8, section 8.4.1) which may act as a seal for vertical movement of fluid. Also within each of the zones, the net/gross ratio varies and since we can only see the upper part of each slice, having control of the stacking architecture of the channel sandstone bodies is difficult. The figures below show some horizontal slices within each of the zones as modelled using the parameters in table 9.2.

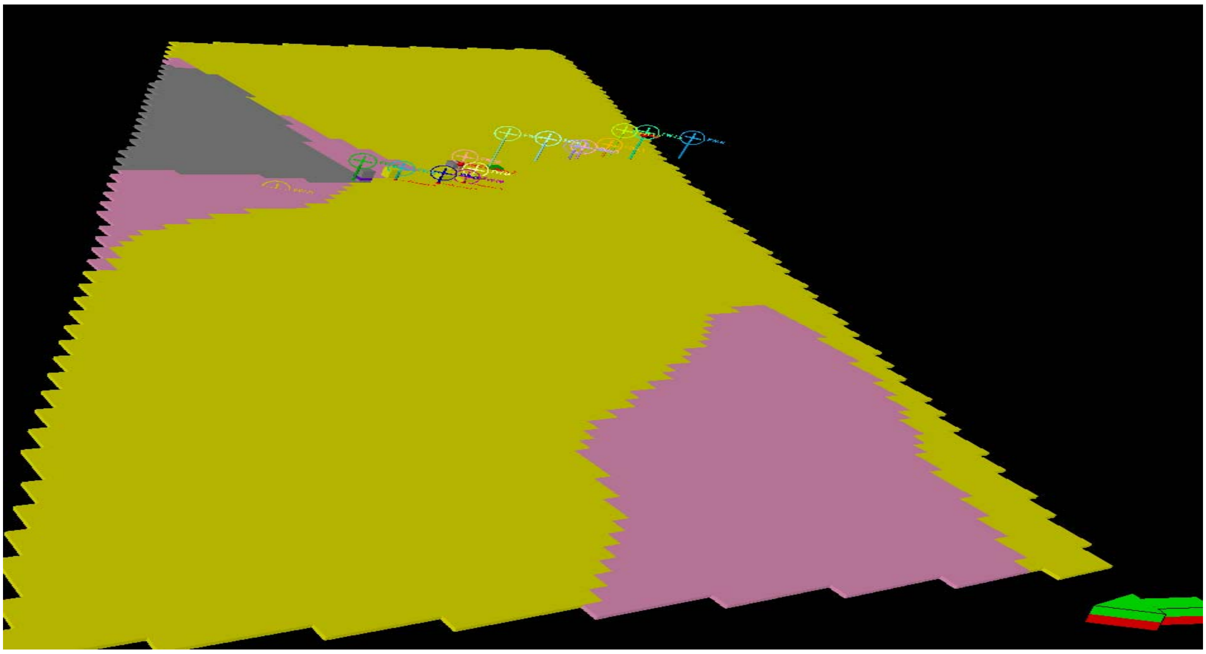


(A)

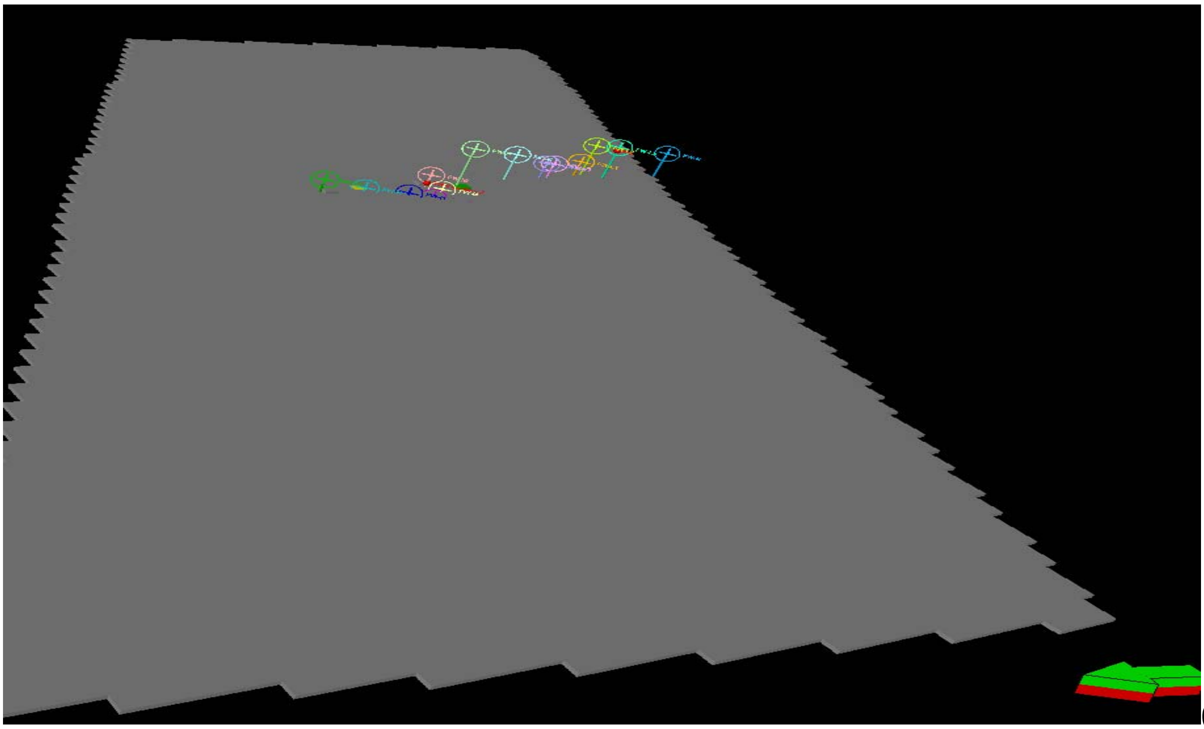


(B)

**Fig. 8.5:** Two horizontal slices within zone 1. Figure (A) is stratigraphically lower than figure (B). Notice the high net/gross ratio in (A) than (B). Arrow points North (green= top and red = bottom) Facies; grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand.



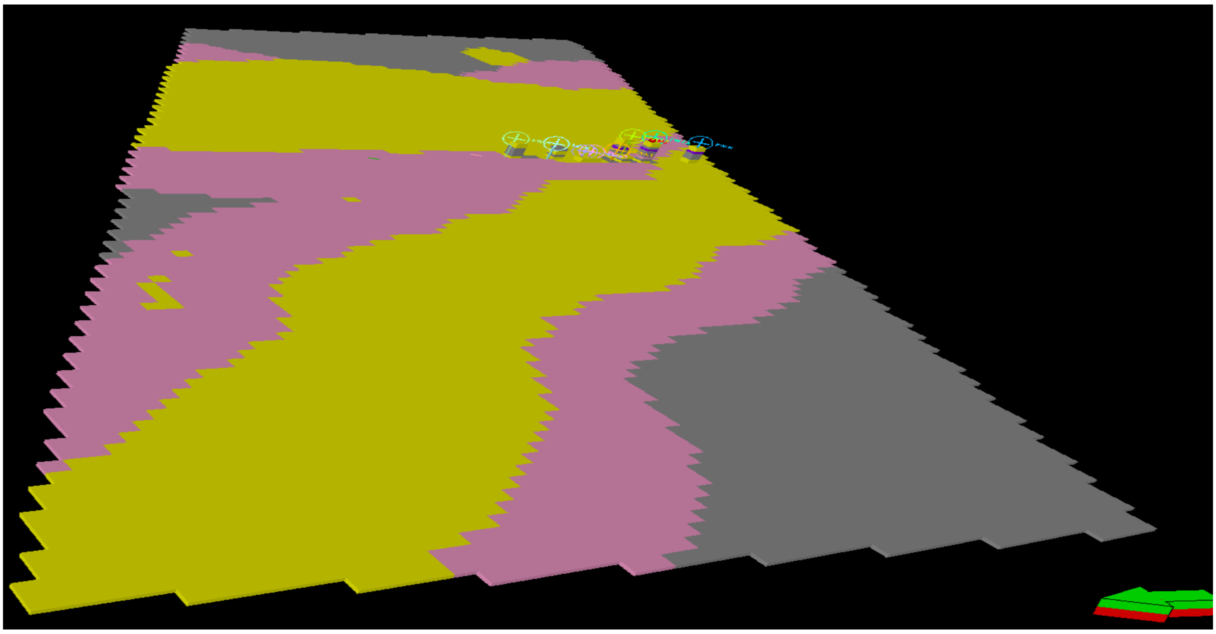
(A)



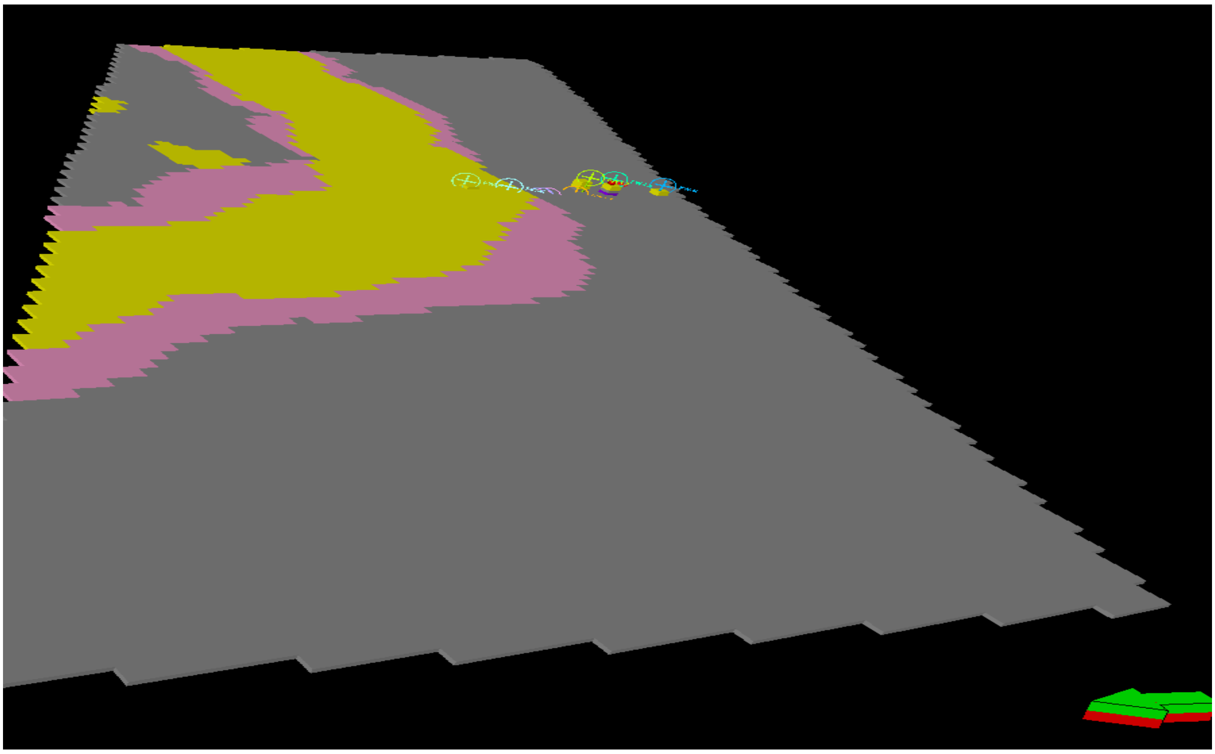
(

B)

**Fig. 8.6:** Horizontal slices within zone two. A) Shows large stacked channel with considerable levee development, B) floodplain deposit and separates zone 2 and zone 3, the flooding surface passes through and may function as a barrier. Arrow points North (green= top and red = bottom). Facies; grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand.



(A)



(B)

**Fig.8.7:** Horizon slices of Zone 3. A) is stratigraphically lower than B). A represents the stacked sandstones in section 3 and B) resents the low sinuous channels in section 4. Arrow points North (green= top and red = bottom). Facies; grey= background floodplain, yellow= channel sandstone, pink= levee sand, and purple= crevasse splay sand.

### *8.4.1 Possible errors*

As mensioned earlier (e.g. Ch.2), the architectural elements are only partially exposed, hence having a proper control on the geometry of sandstone bodies was difficult. However, some of the dimensions used could be quite extreme in an occasion whereby the equations above may be misused. E.g. a braided channel may be mistaken for sinuous channel and such situations might affect the connectivity of the sandstone bodies and may result to uncertainties.

# 9 Discussions

The combined results of facies analysis, petrography, palaeocurrent measurements, and the channel architectural style suggest that sedimentation within the studied interval of the Louriñha Formation was brought about by a fluvial system of rivers flowing from the west to northwest to the east to southeast. From palaeocurrent and log measurements, it could be thought that the stream channels were developed by increasing sinuosity and decreased depth towards the southeast.

A petrographic result shows that the sandstones may have been derived from granites not far from the area. Provenance candidate is the Hercynian granitic basement rocks to the northwest. The low plagioclase content compared to that of the potassium feldspar confirms their provenance from granite *sensu strictu*. The high quartz content compared to other minerals in many of the sandstone samples analysed suggests that they may have been derived through enrichment by chemical weathering. The recorded plagioclase content may also reflect a preference of weathering of plagioclase relative to that of potassium feldspar

Both allogenic and autogenic processes are thought to have played a major role in the sedimentation pattern within the rift basin. Uplift in the Hercynian basement rocks coupled with localized intra-basinal tectonic movements could be responsible for the development of the facies style in the Louriñha Formation. The localised uplifts could be due to diapirism (Alves et al., 2002), causing topographic undulations, hence resulting to sinuous channels.

It is likely that the uplift of basin shoulders could have give rise humid conditions in the hinterland highland areas, which together with high fluvial discharge and transport capacity may explain progradation of channelised sandstone bodies in the basin. Humid climate could have promoted chemical weathering and production of mud and clay to the fluvial systems and hence the production of mudstone deposits (FA. 3.2). FA. 3.2 deposits are abundantly preserved because of the decreased channel depth and are stabilised by the presence of high vegetation cover.

When there was a change in base level there was a subsequent change in accommodation space, and these changes regulate the sandstone body geometry and distribution. Since the study area was paralic, a change in base level was represented by changes in the relative sea level. Therefore, changes in relative sea level, the combined effect of subsidence and eustacy, controlled the accommodation potential of the sediments and distribution of facies within the study area.

Also the rate at which the fluvial systems could adjust to the base level fluctuations, controls the variations in facies architectural style within the depositional area. An increase in the relative sea level could have created more accommodation space for preservation of mud and mudstone deposits.

However, it seems unclear if the increase in relative sea level was due to allogenic or autogenic factors, or both. The beds of the flooding surfaces are relatively thin and laterally not extensive, as well as those of the paleosol surfaces. In this light it might be thought that autogenic factors might have a part to play. There is a general fining upward succession in the studied interval of the basin deposits though interludes of coarsening upward succession were identified within certain channel sandstone bodies. This could be due to local transgression and regression episodes.

Determining the magnitude of landward and/or basinward displacement of facies belts is confusing since the surface identified as marine mud is limited laterally. Also, there is no great distinction of facies across the whole section which might be an indication of a basinward or landward direction of shoreline migration. According to Willis and Gabel (2001), carefulness needs to be taken while interpreting such an environment to represent an overall transgression, since changes in tidal current strength may be due to a change in the morphology of the basin resulting from variation of sediment supply and accommodation space. It is also unclear if the section was incised by the prograding streams, since there was no correlatable extensive surface, such as flooding surface or paleosol.

The best quality sandstone reservoir rocks in the study area could be within section two, three, and probably section four (e.g. from panel) all with multi-storey and more amalgamated channel sandstones characterised by general fining upward succession. Internal connectivity could be disturbed by intra-basinal mud clasts conglomerates at erosive surfaces within the sandstone units and local calcite cementation (e.g. Reynolds et al., 1998).

Majority of the isolated fining upward succession in section 1 shows a lower degree of amalgamation in the sandstone units, and consequently, internal communication is reduced. Heterogeneity is also created by lateral continuous mudstone deposits (FA.3.2) that could cause vertical and horizontal barriers to fluid flow. Hence, it could be deduced that heterogeneity in this scale is influenced by the depositional environment. From the stochastic model runs it could be deduced that the lower part of Zone 1 could represent a good reservoir unlike the upper part with abundant background floodplain deposits. Hence the upper part of zone 1 could function as a vertical barrier to fluid flow between the lower parts and Zone 2 and 3. Possible horizontal barriers in the study area could be the flooding surfaces.

In the mesoscale, heterogeneity appears to increase towards the upper part of the stratigraphy. This could be seen with increased intervals of multi-storey channels having more tidal signatures like reactivation surfaces, mud drapes, ripples and heterolithic sandstone units. Sandstones with tidal signature are more prominent in sections 2, 3 and 4. This implies that majority of the sandstone units higher in the stratigraphy could have been deposited in the lower limit of plane beds, as generally proposed by Allen (1968) (fig. 5.3). These sandstone units could have been deposited in a relatively low energy environment, hence allowing the mud to persist in the sandstone bodies and even in the grain framework facies. Heterogeneity in such sandstone units will be higher compared to those deposited in the higher plane bed conditions such as dunes characterised by grain size variation.

Petrographic analysis shows that porosity gets reduced upwards in the stratigraphy due to the presence of calcite and in some cases also mud in the intergranular spaces. Also very few intragranular porosity was recorded in all the sandstone samples analysed. The highest porosity value is 24.9 % from sample 13 located in Zone 1 (table 8.1). The porosity values are very contrasting within a very short interval. In this light, it could be suggested that textural heterogeneities in the sandstone units higher in the stratigraphy are mainly influenced by calcite precipitation which might be post-depositional. Moreover, despite the higher amalgamation, interconnectedness and net/gross ratio of sandstone units, for example, in section 2 and 3, their reservoir properties are reduced by the presence of calcite and mud in the intergranular spaces.

# 10 Conclusion

The Louriñha Formation is deposited in the Lusitanian Basin which is a Mesozoic rift basin associated with the break up of Pangea and the opening of the Atlantic Ocean. The Lusitanian Basin consists of several sub-basins which are bounded by faults and salt structures.

The sandstone bodies are dominated by channels, crevasse deposits, levees consisting of trough crosss stratified sandstones (facies C), planar stratified sandstones (facies D), massive sandstones (facies E) and sometimes horizontally stratified sandstones (facies G) interbedded with parallel laminated sandstones (facies I) to cross laminated sandstones and mudstones. Shell banks interpreted as marine flooding surfaces were also recorded.

Detailed analysis of the facies and architectural elements of the deposits gave a thorough knowledge of the controls on deposition of the sandstone and mudstone successions in the study area. Five facies associations were recognised. The Louriñha Formation is thought to have been deposited in a warm and humid alluvial to coastal plain palaeoenvironment.

The medial areas are characterized by amalgamated sandy channel deposits with little preservation of overbank or floodplain facies association. The channel sandstone body dimensions are generally small. The size ranges between 2 m and 20 m in thickness and laterally some may be greater than 200 m in width. Five types of channel fill architectural elements were identified. Each of the channel fill elements consists of unique assemblage of facies, internal geometry and bounding surfaces. The channel fills are basically sandstones and may pass upwards into mudstones. Channel abandonment due to channel avulsion or migration gave origin to overbank and floodplain deposits dominated by mudstone. The amount of floodplain deposits in the study area decreases from north to south with sheet-like sandstone bodies deposited as crevasse splays or crevasse channels.

Based on the above analysis, the study outcrop is subdivided into four stratigraphic sections characterized by sandstone architecture, in accordance to facies stacking pattern, sandstone/mudstone ratios, and channel morphology. Though the changes are not very distinct, architecture and stacking pattern in the study area were controlled by a combination of allogenic and autogenic factors. The most significant allogenic factor is thought to be eustacy followed by tectonics and climate.

The presence of coalified wood fragments in most of the sandstone deposits coupled with the presence of calcretes may indicate shift in climate from humid to arid or semi-arid. This climatic shift may be a candidate factor in establishing the contrasting depositional style and architectural pattern witnessed in this area.

Sandstone heterogeneity shows vertical and horizontal variations across the study area, and due to the low net:gross ratio, sandstone connectivity and stacking pattern were determined using the Petrel ™ stochastic software modelling program.

From petrographic analysis it can be concluded that the sandstones were derived from the Hercynian Basement Horsts in the north western margin of the Lusitanian Basin, but is presently exposed in the Berlengas and Ferilhões Islands. The sandstones are fine to medium grained and poorly to moderately sorted. The sandstone recorded variable amounts of calcite cement. Values of porosity recorded after point counting ranges between 7, 3 % to 24, and 9 % with an average of 13 %. The porosity values are however not of reservoir scale and the high calcite present in intergranular spaces are thought to be post-depositional. This may be as a result of influx of carbonate rich water into the pore spaces and subsequent precipitation during warm climates. The phenomenon of carbonate precipitation is common in coastal deposits. However, if the reservoir sandstones were possibly in the subsurface, below the limit of meteoric water, leaching of minerals like feldspar and precipitation of carbonates from sea water will eventually be reduced.

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# A. Paralic Sedimentary Environment

### A.1 Introduction

The paralic environment is a broad zone between the landward limit of marine processes and the seaward limit of alluvial and shoreline processes. According to Nichols, G. (1999), where the land meets the sea, the surface processes which act on the continent interact with marine processes, resulting in the most complex depositional environments

Waves, tides and ocean currents act on sediments brought down by rivers, redistributing material in estuaries and deltas at the river mouth. Away from deltas and estuaries are stretches of coastline which may be sites of either erosion or sediment accumulation. Erosional coasts may be backed by cliffs or consolidated rock, or by loose sediment. Depositional coasts may be sites of accumulation of clastic carbonate deposits or evaporates. Terrigenous clastic sediment is supplied to the coast by wave, storm and tide activity transporting detritus along the coast and then reworking it on beaches and associated environments. Estuaries and lagoons are particularly characteristic of transgressive coasts; deltas are features of prograding coasts. The main paralic environments are (Fig. A.1) coastal to shoreline shelf systems, delta, strand plain to open shelf, estuaries etc (Emery and Myers, 1996). But also includes all those parts of the coastal plain affected by the proximity of a shoreline( sub environments) such as: upper delta plain, lower delta plain, lower coastal plain, shoreline shelf, estuary shelf, estuary mouth shoal, central bay, bay-head delta, estuary head, levees, estuary mouth, lagoons, lakes etc.

Paralic sandstone body types (Fig. A.2) include deposits from a great variety of environments: valley, fluvial channel, distributary channel, distributary mouth bar, crevasse channels, crevasse splays, shoreline-shelf sand, flood tidal delta, chenier, tidal flats, tidal inlet, tidal creek barrier islands etc (Reynolds, T 1994). At a large scale, paralic depositional systems respond sensitively to sea level change, often by large lateral shift in facies belts. As a result, deltas, estuaries, and shoreline-shelf systems, and their sub-environments, are commonly interleaved forming highly layered successions. At an intermediate scale the discrete layers display a wide range of sandstone body types (such as channels, splays and shoreline shelf sands) with each sandstone body type having distinct dimensions (width, length, thickness, and shape) and internal properties (for example, coarsening or fining upward).

Consequently, paralic sandstones reservoirs can range from thick, extensive sheets to thin laterally restricted sandstone, as well as sandstone bodies having low width (W/T) ratios, as in fluvial distributary channel sandstones. Thick, extensive sandstone bodies typically have structurally defined spill points, and good reservoir properties. Such sandstone reservoirs usually have high recovery, good reservoir performance, and form primary completion targets. Thin, laterally restricted, isolated sandstones have lower recovery factors and poor performance (Reynolds, T. 1994).

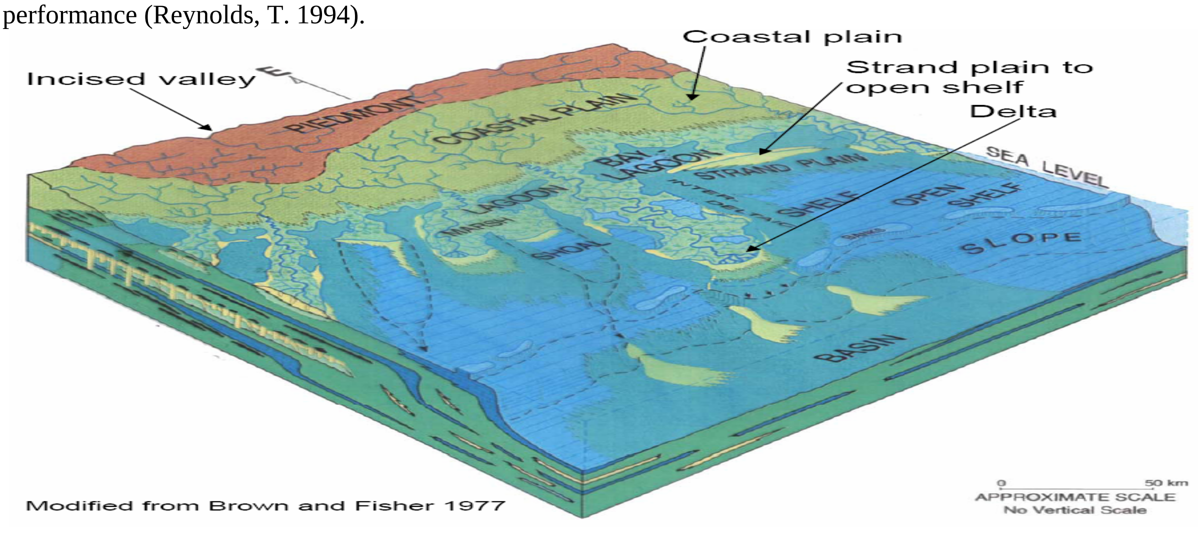



**Fig. A.1:** The main Paralic environments

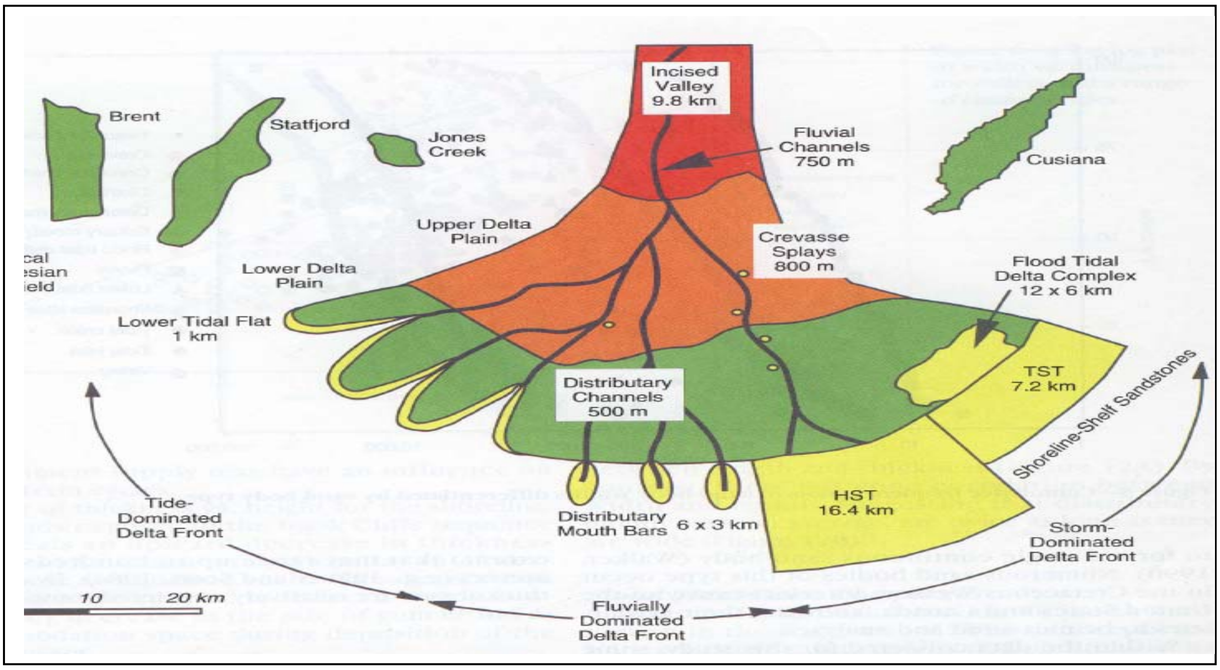



**Fig. A.2:** A schematic delta showing a range of sand-body types at their average dimensions, together with a number of paralic oil and gas fields at the same scale. The delta front is divided into three segments that are storm, fluvial, and tide dominated, respectively. The delta itself and the division into an upper (fluvial-dominated) delta plain and lower (tide-dominated) delta plain are not to scale. Modified from Reynolds (1994)

### A.2: Deltaic systems and sediment characteristics

A delta is the prism of sediments that accumulates where a river enters a standing body of water. Moore and Asquith (1971) defined a delta as “the subaerial and submerged contiguous sediment mass deposited in a body of water (ocean or lake) primarily by the action of a river”. The deltaic environment occurs downstream from the river environment and is directly adjacent to, and up dip (landward) from, the marine environment. Most large deltas occur on the margins of marine basins, but smaller deltas also form in inland lakes, seas, and coastal lagoons and estuaries.

All deltas are fed by a river and there is inevitably a transition from the channel which is considered part of the fluvial environment to the channel which occurs on the delta plain (Fig. 3). Delta channels may be in the form of a single course to the delta front or may be distributary in form. The coarsest delta top facies are found in the channels, where the flow is strong enough to transport and deposit bed load material. Adjacent to the channels are subaerial overbank areas which are sites of sedimentation of suspended load when the river floods. These may be vegetated under appropriate conditions and may be sites for the accumulation of peat and coal, and if there is frequent overbank flow from the channel the deposit may contain enough clastic material to form carbonaceous floodplain mud. Crevasse splay may result in lens shaped sandy deposits on the delta top (Nichols, G. 1999)

High rate of deposition may take place in two general environments, in subaerial areas and in subaqueous areas.

*The subaerial component* of deltas is generally larger than the subaqueous component and is divided into an upper, and a lower delta plain. Although these two areas are comprised of different components, both are dominated by fluvial processes

*The upper delta plain* or delta top lies mainly above tidal influence and is little affected by marine processes. In the upper delta plain deposition comes from both within channels and between channels. Four main sources of sediment deposition are represented by braided channels, meandering channels, lacustrine delta-fill, and floodplain deposits. Sedimentation on the upper delta is dominated by migration of distributary-channel and associated fluvial sedimentation processes such as channel and point bar deposition, overbank flooding, and crevassing.

*The lower delta plain* lies between the low-tide mark and the upper limit of the tidal influence. It is typically a brackish to saline environment and is the scene of more active deposition, mostly by crevassing or over banking from the deltaic distributaries, modified by the input of marine sediments from storms. Crevassing is more common than the occurrence of natural levees because the levees are generally more poorly developed. Vegetation is necessarily tolerant towards saline water and is confined to those plants that can withstand prolonged inundation (Reading and Collinson, 1996).

*The subaqueous delta plain* constitutes that area of a delta that lies seaward of low tide level and actively receives fluvial sediments. These are best developed on shallow continental shelves. Deposits of the subaqueous delta thus form the base over which subaerial delta builds seawards. The delta front immediately forward of the channel mouth is the site of deposition of bedload material as subaqueous distributary mouth bars that consists in part of sands, and possibly gravels (Fig. A.3). Mouth bar deposits typically coarsen upwards, since coarser sediment can be transported and deposited in higher energy environment shallower water. Mouth bar deposits may be extensively reworked by wave and tide action. The finest silts and clays are transported still farther seaward and settle in the prodelta environment (Nichols, G., 1999 )

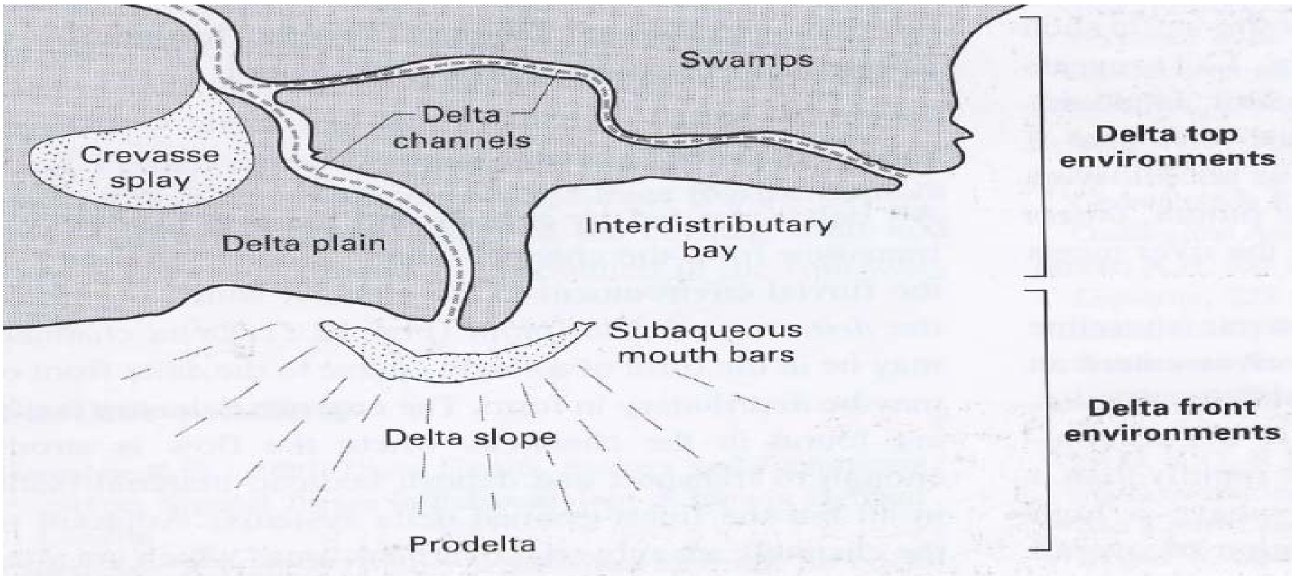



**Fig. A.3:** Delta sub-environments (Nichols, G., 1999. Fig.12.1, pp150)

### A.2.1 Source and transportation of sediments

Sediments deposited in large deltas are generally derived from extensive continental regions that may be composed of rock types of varied compositions and geological ages. Thus, the composition of deltaic deposits could be quite varied.

The sediment load of rivers consists of two parts:

(1) the clays and fine silts transported in suspension, and
(2) the coarser silts and sands, and in some cases gravels, transported as bed load

According to Reading and Collinson (1996), the ratio of suspended load to bed load varies considerably, depending upon lithology and climatic conditions of the sediment source areas. The suspended load is generally much greater than the bed load. The transportation of sediment to a delta is an intermittent process. Most rivers transport the bulk of their sediments during flood stages. During extended periods of low discharge, rivers contribute very little sediments to their deltas. The extent to which deltaic sediments are dispersed into the marine environment is dependent upon the magnitude of the marine processes during the period that a river is in flood stage. Maximum sediment dispersal occurs when a river with a large suspended load reaches flood stage at the time the marine environment is most active. Minimum dispersal occurs when a river with a small suspended load (high bed-load) reaches flood stage at a time when the marine environment is relatively calm. Figure A4 shows the main controls in a deltaic system.

According Reading and Collinson (1996) delta size is dependent upon several factors, but the three most important are (i) the sediment load of the river (ii) the intensity of marine currents, waves and tides and (iii) the rate of subsidence. For a given rate of subsidence, the ideal condition for the construction of a large delta is the sudden large influx of sediments in a calm body of water with a small tidal range. An equally large sediment influx into a highly disturbed body of water with a high tidal range results in the formation of a smaller delta, because a large amount of sediment is dispersed beyond the limits of what can reasonably be recognised as a delta. Rapid subsidence enhances the possibility for a large fluvial system to construct a large delta

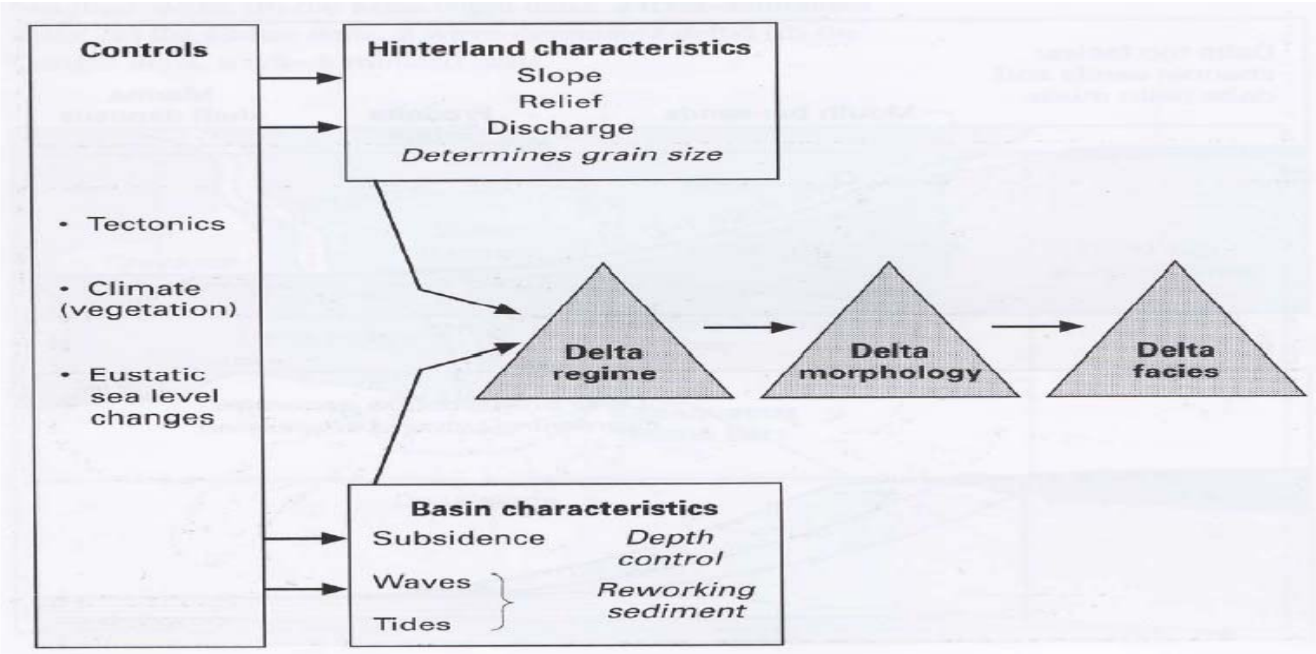



**Fig. A.4:** Controls factors on delta environments and facies (After Elliot 1986)

### A.3 Classification of deltas and sedimentary processes

A popular way of classifying deltas in terms of processes was proposed by Galloway, (1975). The classification was based on the relative influence of wave, tide and river processes and does not take into account water depth or grain size.

#### A.3.1 Sedimentary processes of river dominated deltas

A delta is regarded as river-dominated where the effects of tides and waves are minor, resulting from high sediment input and as a consequence of relatively low-energy shelf processes. This requires a microtidal regime and a very gently sloping shelf where wave energy is effectively dissipated before the wave reach the coastline. The direction of river flow is unidirectional and the channel formed is maintained with well defined subaqueous levees and over bank areas (fig. A.5). Deposition is mainly on the subaqueous levees building up a characteristic "bird's foot" pattern of river-dominated delta.

Most of the sands and coarser silts brought into a body of marine water in a basin are deposited in the immediate delta front environment as river mouth bars and slightly beyond the bar front zone. The degree of sand dispersal is controlled by the level of marine energy. However, in most bird foot deltas, sand is not transported beyond (15 m) water depth (Nichols, G. 1999). The finer sediments, which are transported in suspension, are dispersed over a much wider area. The degree of dispersal is controlled by current intensity and behaviour. Accumulations of clays seaward (distally) are referred to as prodelta or distal clays

Some river dominated deltas also show distinct sediment dispersion patterns produced by the equilibrium between fluvial and marine energies. The sediment dispersion through multiple terminal distributary channel result in an overall lobate shape of the river-dominated delta that is opposite to the bird's foot, river dominated Mississippi type, but similar with the deltas described as wave-dominated.

In strict river-dominated deltas, the channels build out to form the toes of a bird's foot, between which there are large interdistributary bays. Channels supply sediments into these bays that gradually fill up to become the vegetated part of the delta plain.

Many river-dominated deltas (e.g. Mississippi) show numerous "terminal" distributary channels caused by channel instability due to the very low gradient on the delta plain causing frequent avulsion of the major and minor channels.

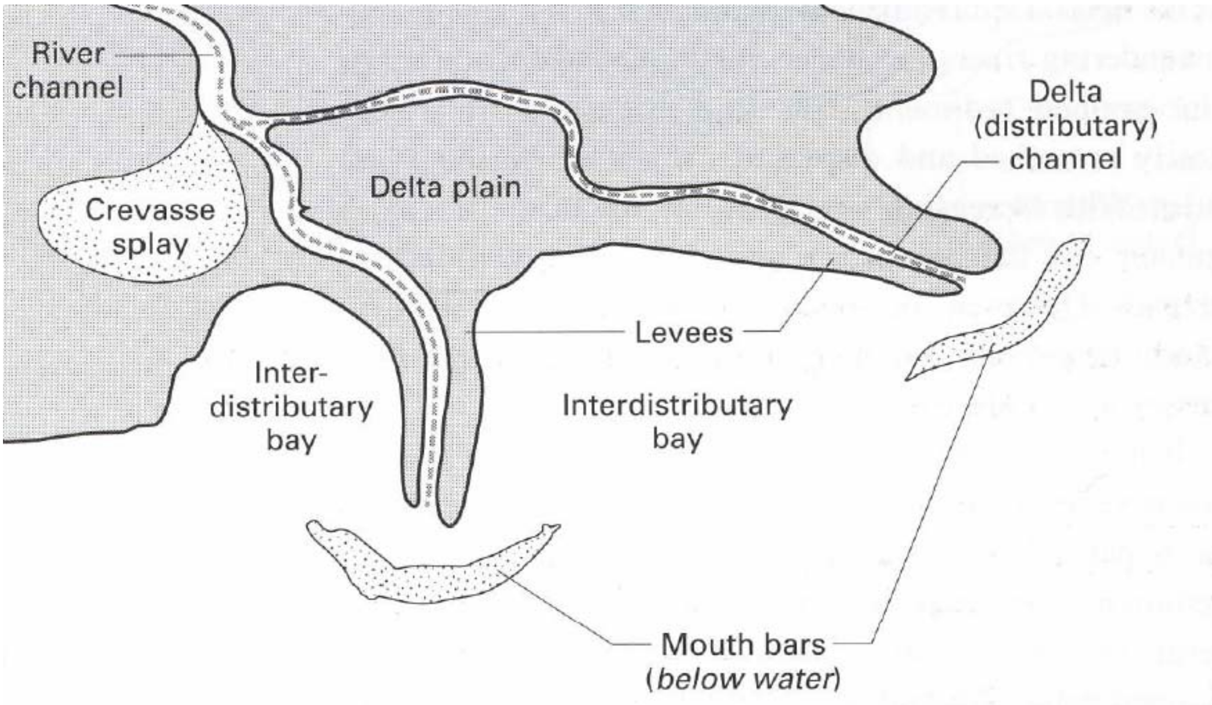



**Fig. A.5:** River dominated delta environments (Nichols, G., 1999. Fig. 12.7, pp 153)

**A.3.2 Sedimentary processes of wave influenced deltas**

The river mouth and mouth bar areas of a delta are susceptible to the action of waves, resulting in a modification of the patterns seen in river-dominated deltas. Progradation of the channels outwards is limited because the subaqueous levees do not form and bedload is acted upon by waves as quickly as it is deposited. Hence, deltas characterised by strong wave energy have a relatively straight coastline, and shore parallel sandstone that are oriented in the direction of prevailing long shore currents.

A net supply of bed loads by the river results in a series of shore parallel sand ridges forming as mouth bars build up and out to form a new beach (fig. A.6)
During the later stages of deltaic evolution (or sediment infilling), the connectivity between the river channel and tidal inlet increases. This results in more efficient delivery of fluvial sediments to the ocean, and the bypassing of remnant central basin features, which slowly infill and become swamp areas. As a consequence, the gross morphology of wave-dominated deltas is relatively stable, and may persist over long periods of time with little change (Nichols, G. 1999). In areas with increased sediment supply, such bypassing to the coast can allow the barrier to prograde, which may result in the formation of a coastal protuberance adjacent to the mouth of the river (Roy et al., 1980)

In the stratigraphic record wave-dominated delta deposits display well-developed mouth bars and beach sediments, occurring as elongate coarse sediment bodies approximately perpendicular to the orientation of the delta river channel. The Nile delta is a

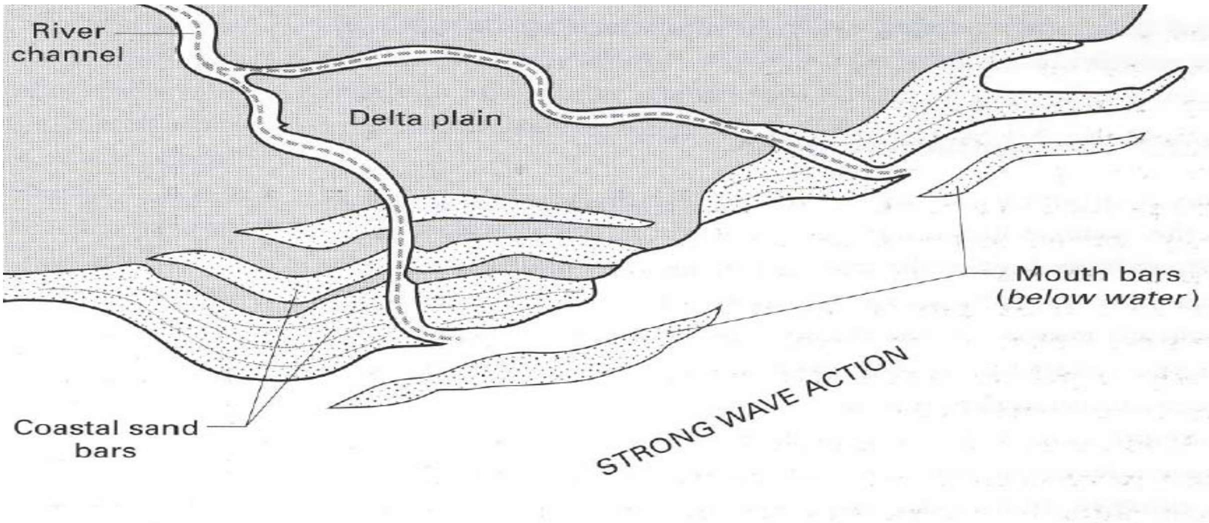



**Fig. A.6:** Wave influenced delta environments (Nichols, G., 1999. Fig.12.10, pp. 155)

good example of a wave-dominated delta. This is in contrast to the river-dominated delta deposits, which would be expected to show less continuous mouth bars and a high proportion of channel and overbank deposits forming the delta lobes.

**A.3.3 Sedimentary processes of tide influenced deltas**

A delta building out into a region with strong tides will be modified into a pattern distinct from both river and wave dominated deltas (fig. A7). Coastlines with high tidal ranges experience onshore and offshore tidal currents which move both bedload and suspended load, resulting to numerous coeval channels at the delta front and result in a finger-like irregular sandstone isopach (Coleman and Prior, 1982). First, the delta top channels are subject to tidal influence and the channel is subject to either reverses of flow or periods of stagnation as a flood tide balances the fluvial discharge. This may be seen in strata as reversals of paleocurrent indicated by cross stratification, and layers of mud deposited in channel due to flow stagnation. The deposits of tide-influenced deltas can be distinguished from other deltas by the presence of sedimentary structures and facies association which indicate that tidal processes were active (reversal of paleoflow, mud drapes, etc.). However, cross-bedding with opposite directions may easily develop in a wave dominated environment. Rivers in areas with high tidal ranges broaden markedly as they approach the seas because of the increasing greater volumes of water to be transported out and in. At the outermost point, broad estuary forms, with clay and sand banks cut by channels which transport tidal water.

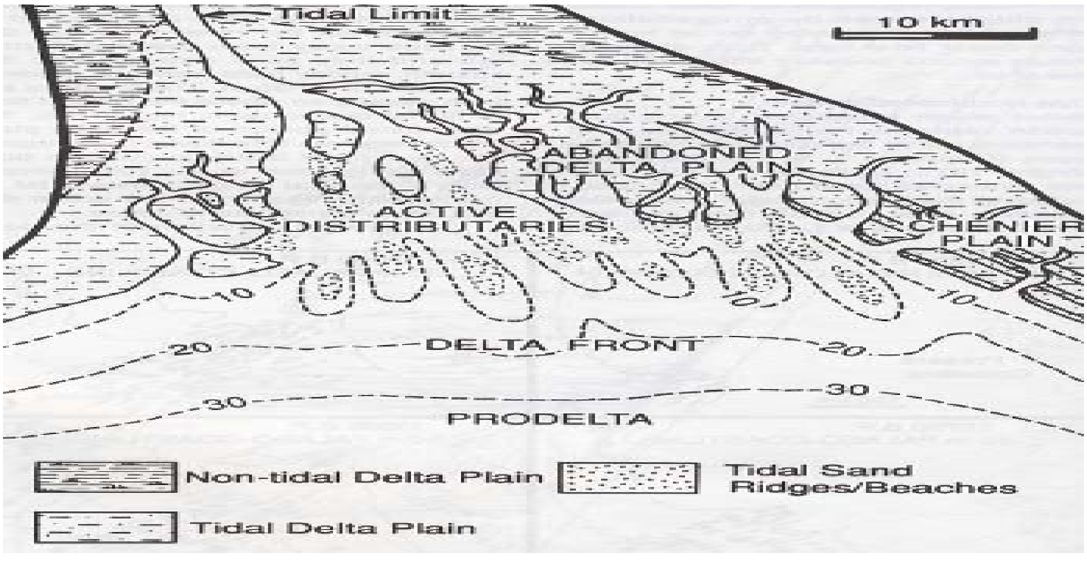



**Fig. A.7** The morphology of tide-influenced delta. From Dalrymple (1992)

### A.4 Delta sedimentation

During the early stages of delta development, the sediments which are transported to the receiving basin exhibit a lateral decrease in the grain size from river mouth seaward. As more sediments are deposited, the shoreline progrades and grain size range is shifted both laterally and vertically. Delta sediments are deposited in successive successions of sedimentary environments. Each of these environments is dominated by a set of characteristic sedimentary structures. When the shoreline progrades, these environments will shift laterally, superimposing different environments on one another. The lateral shift in grain size range and environments results in the deposition of coarser over finer sediment and produces a unique vertical succession of sedimentary structures belonging to different environments.

If influx of sediments diminishes, or the channel becomes inactive at that site, deposition terminates marking the end of one deltaic cycle. Basin subsidence or silt and clay compaction allow marine transgression over a delta deposit resulting in the deposition of finer sediments over the deltaic complex. Several cycles may develop in one site as local marine transgression and regression are repeated.

### A**2.5 Estuaries**

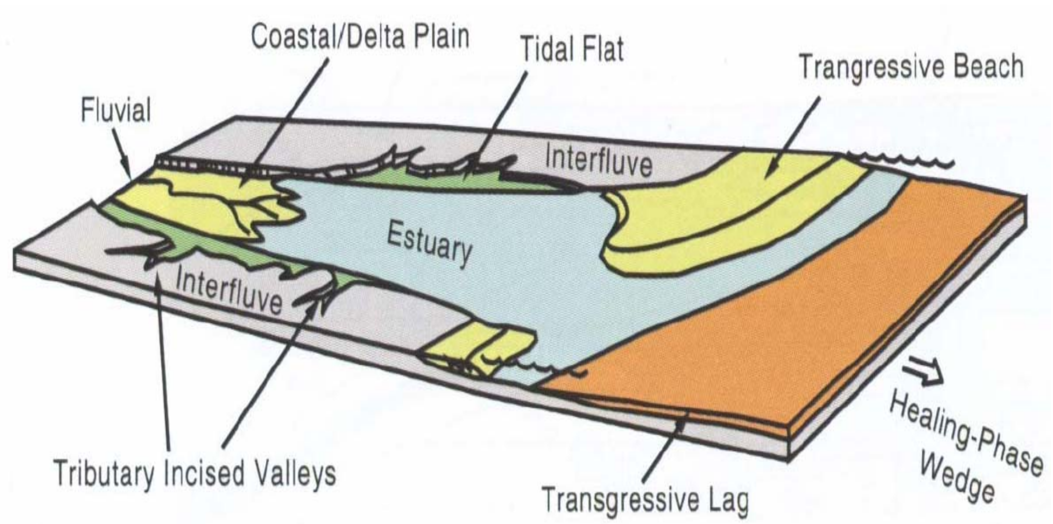
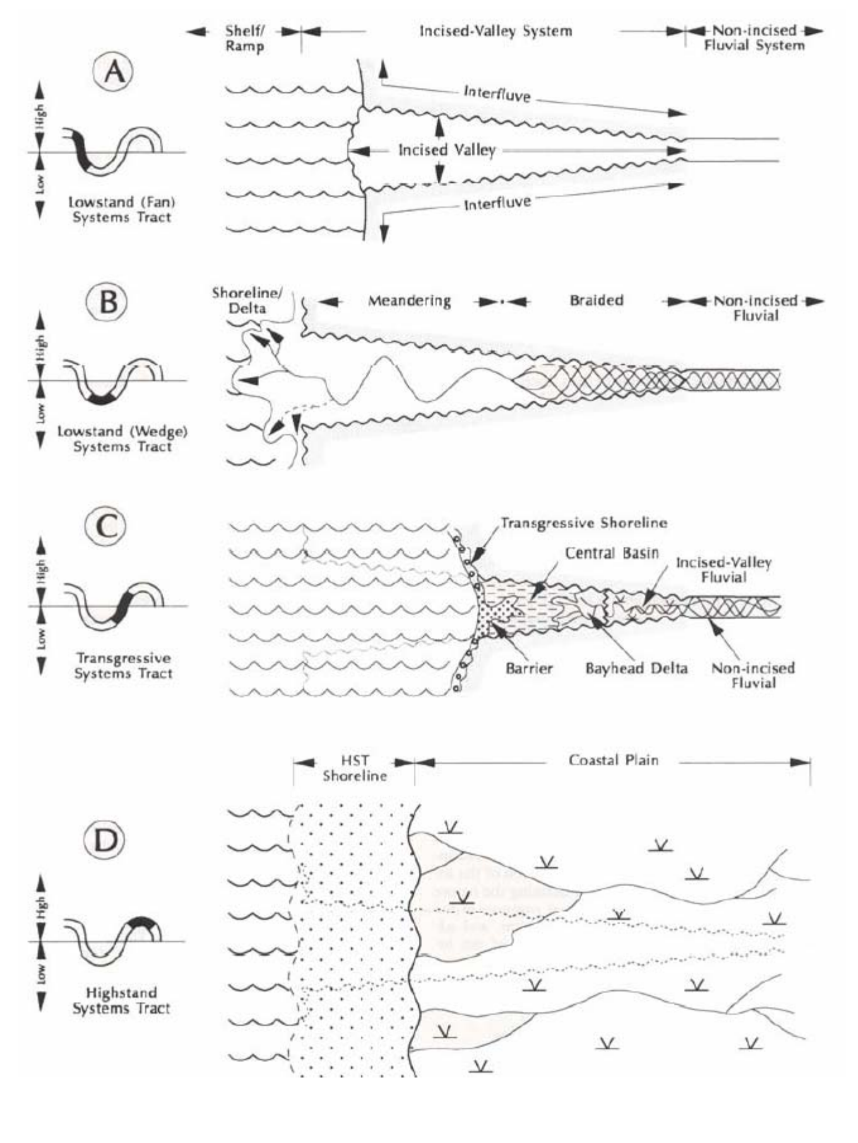



**Fig. A.8:** The diagram shows the main physiographic features of an estuary, or a drowned incised valley, during early stages of transgression. Posamentier and Allen, 1999, fig. 2.36

Estuaries are partially enclosed bodies of water where fresh water meets salty water. Bays, inlets and ocean-flooded river valleys are all examples of estuaries. They are characterised by sediment input from both fluvial and marine sources (Fig. A.8). Where fluvial input is strong, bay-head delta may develop. This consist of delta top facies (channel and delta plain) which may be similar to those found in other deltaic environments but are confined by the incised valley of the estuary. Where fluvial input is weak and tidal currents relatively strong, fluvial channels become progressively tidally influenced to form tidal channels downstream




A) Falling sea level causing valley incision.

B. Lowstand with fluvial drainage and deposition of lowstand wedge delta and associated paralic facies.

C. Rising sea level creating increasing accommodation and filling of estuary formed in the incised valley.

D. Highstand and progradation of coastal plain and shoreface facies

**Fig. A.9:** Idealized model of formation and infill of an incised valley (Zaitlin et al. 1994)

The incised valley extends completely across the shelf and coastal plain due to sea level fall that fully exposes the shelf as seen in (fig.A.9) above. The shelf slope gradient is steeper than the elongated graded stream profile. Incised valley develops only at the inner and outer-shelf with shelf slope gradient steeper than the graded stream profile; lowstand sediments are transported across the shelf through lowstand bypass channels where shelf slope gradient equals graded stream profile

**A.5.1 Wave dominated estuaries**

In wave-dominated estuaries, along-shore and onshore sand transport results in accumulation of a sand plug at the head of the estuary (Fig.A.10).

The sand barrier has the characteristic of a strand plain or barrier island complex with associated washover deposits. An inlet in the barrier allows water to pass through into the lagoon, a low energy zone characterised by muddy facies. Alternatively, the barrier may close the estuary entirely at times to produce a blind estuary or coastal lake. At the mouth of the river a bay head delta forms.

**A.5.2 Tide dominated Estuaries**

In tide dominated estuaries, tidal-current energy exceeds wave energy at the mouth of the estuary, creating energy conditions in the estuary higher than those typical of wave dominated estuaries. The strong tides ensure an active exchange between the estuary and the open sea, preventing the development of bay head deltas, a muddy central basin or a discrete sand plug (fig. A11)

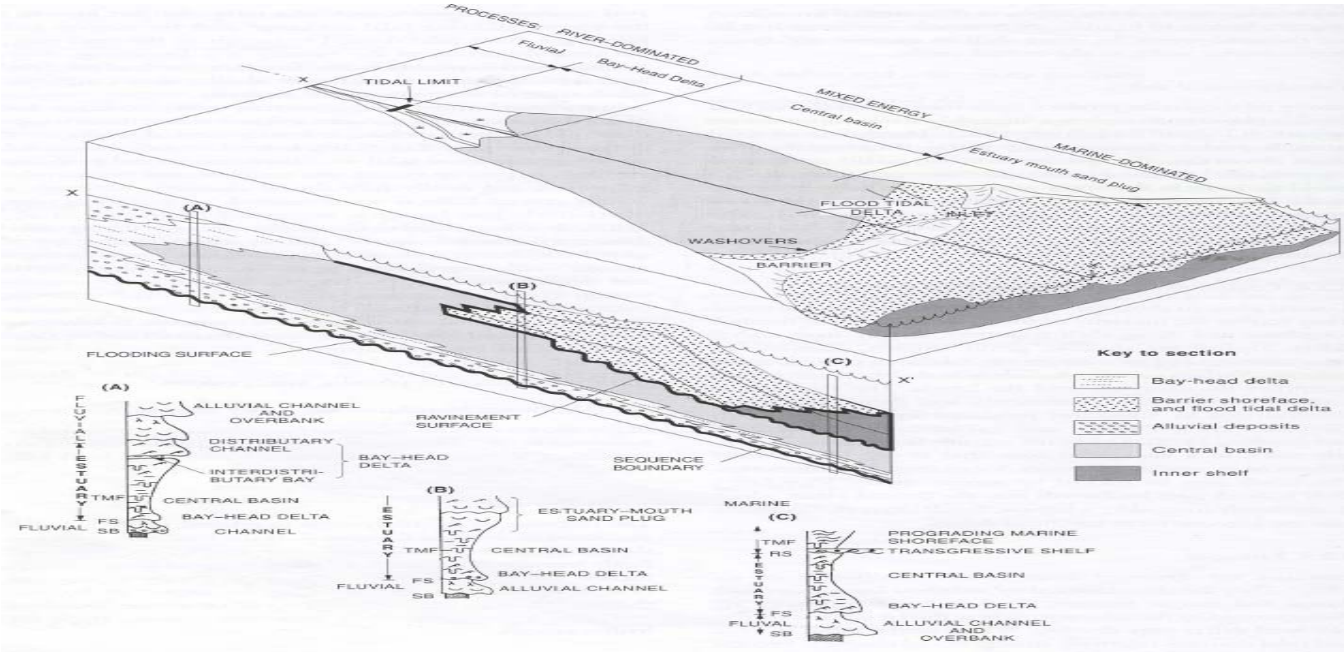


**Fig. A.10:** Fluvial dominated bay-head delta with sand in innermost part. Central basin with mixed fluvial-tidal energy and mud Marine dominated outer part with estuary mouth and sand plug formed as barrier sand, with tidal inlet, flood tidal delta and washover fans. (Emery and Myers 1996, fig. 8.3)

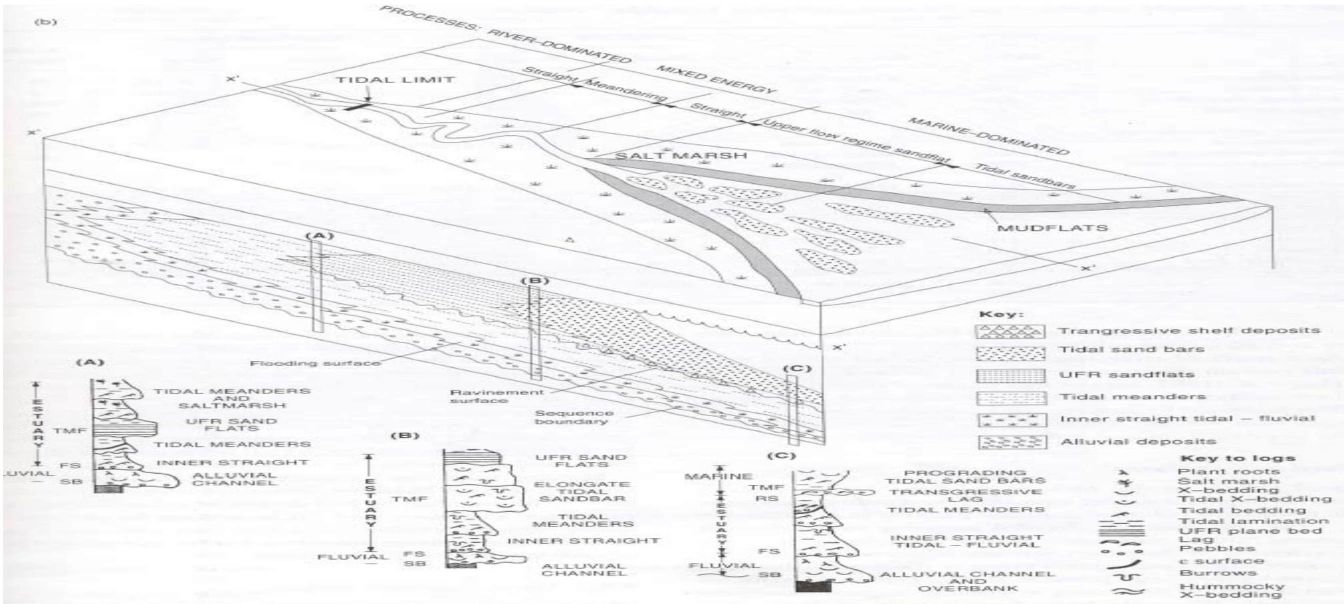


**Fig. A.11:** In tide dominated estuaries, elongate sand bars develop parallel to the length of the estuary from sand carried into the estuary from marine sources. Although a muddy central basin is not developed, an analogous zone of relatively low energy is recognised (Dalrymple et al., 1992).

**A.6 Coastal inter-deltaic sedimentation**

The ideal interdeltaic deposit, as the name implies, occurs along the coast between deltas and comprises mud flats and cheniers (abandoned beach ridges) of the chenier plain complex and the barrier island –lagoon-tidal channel complex (fig. A12).

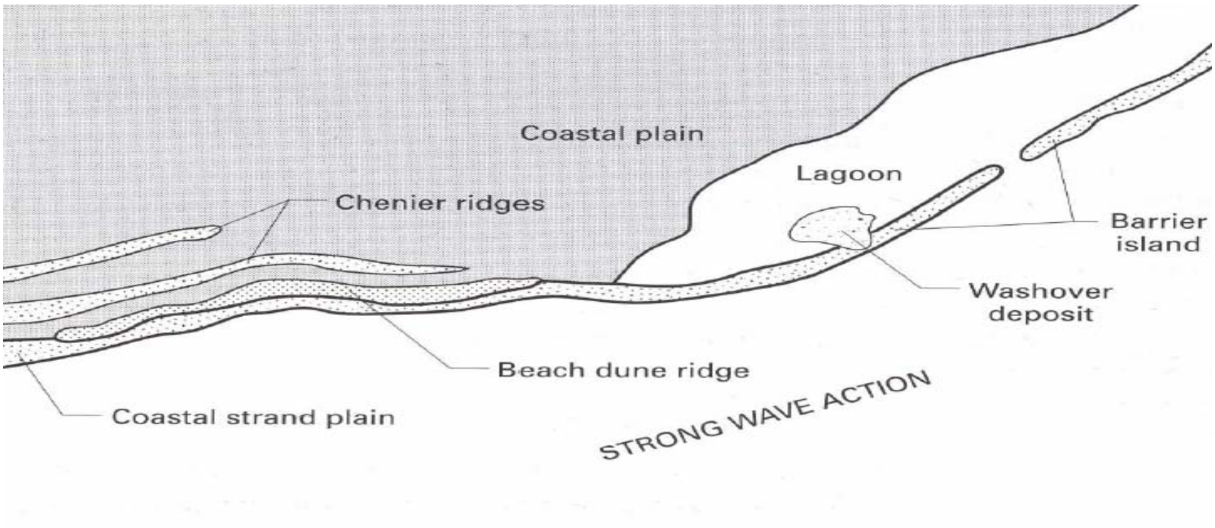



**Fig. A.12** Morphological features of a coastline. Nichols, G., 1999. Fig.13.6, pp.171

### A.6.1 Source and transportation of sediments

Most of the sediments deposited in the paralic environment are derived from land, but minor amounts come from the marine environment and from adjacent continental shelf areas if erosion occurs in the marine environment. A portion of the sediment transported to the marine shoreline by rivers and smaller streams is dispersed laterally by marine currents for great distances along the coast. Clays and fine silt are carried in suspension. Sand is transported mainly as bed load or by wave action in the beach and near shore zone. The suspended silt and clay load is dispersed at rapid rate and is most significant in the development of mud flats of the chenier plain. Lateral movement of sand bed load occurs at a relatively slow rate and is most significant in the development of chenier and barrier island complex.

## A.7 Reservoir properties of paralic sandstone-bodies

The way in which channel sand bodies are stacked in relation to one another and to associated over bank sediments controls their degree of interconnectedness and reservoir character (Johnson et al, 1985). There are three basic types of deltaic sandstone bodies: delta-fringe, abandoned distributary channel, and point bar sands. The geometry and quantity of sandstone deposit depends on the type of delta. Coalescing point bar sands can actually form a “blanket-like” sand body of very large regional extent. Such a coalescent sheet is more heterogeneous than comparable marine shelf sheet sand. The continuity of sand is interrupted only by “clay plugs” which occur in abandoned meander loops, or in the last channel position of meander belts which have been abandoned abruptly. There exists strong asymmetry; with upward decrease in grain size and bed thickness, possibly with conglomerate near base. Larger channel fill sandstone bodies tend to be coarser grained than smaller ones. Vertically, there are overlying silty shales, commonly of alluvial origin. Possibly peat and coal may be present in the sandstones. The basal contacts are commonly sharply disconformable. The successions show both multi-storey and multilateral channel sandstone bodies (fig. A13). Laterally, there exists silty shale and siltstone commonly with abundant carbonaceous material as well as roots, leaves, and stems.

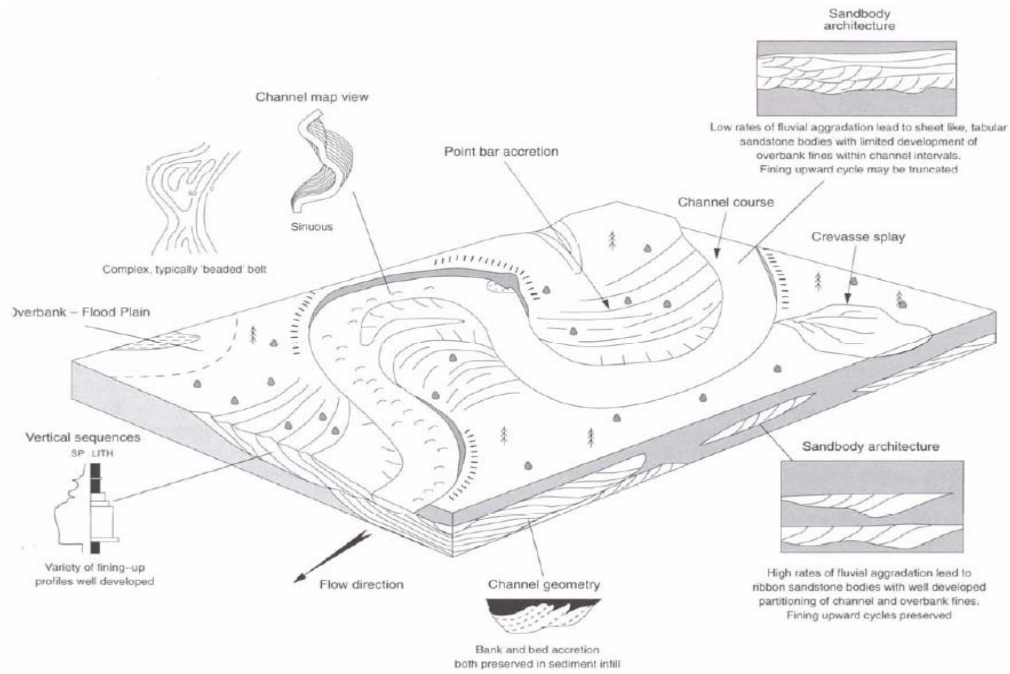



**Fig. A.13** Setting and general characteristics of meandering stream model of clastic sedimentation. The channel may migrate laterally and downstream giving rise to single or multi-storey channel sandstone body of relative high W/T ratio. (Modified after Miall, 1992)

In the bird-foot type delta, the most common sands are those of the delta fringe environment. These sands occur as relatively thin, wide-spread sheets, and they contain a substantial amount of clays and silts. Abandoned distributary channels contain varied amounts of sand, probably composing less than 20% of the total delta sand content. These sand bodies are long and narrow, are only slightly sinuous, and are encased in the delta fringe sands or prodelta clays; depending upon channel depth and distance that delta has prograded seaward.

In cuspate-arcuate type of deltas, the delta fringe sand complexes are wide, though individual sand bodies are relatively narrow, and are generally much cleaner than delta fringe sands of the bird foot-type delta. Distributary channel sands and point bar sands are much more common than in bird foot deltas and can constitute up to 50% of the total sand content of the delta. These two types of sands are encased in delta fringe and prodelta sediments.

Sand bodies of tidal environment appear to be of two principal types: larger ones in estuaries and passes between barrier islands and smaller ones in creeks and inlets that transgress the intertidal zone along the strand. Available information suggests that both types share many of the characteristics of alluvial sand bodies.

Most tidal sand bodies, like fluvial ones are elongate. Orientation of tidal sand bodies which develop in inlets of intertidal zone is at right angles to the beach. Such sand bodies are usually thin, and may exceed 3 m in some places; width is small, probably exceeding 15.2 m-30.4 m in some places. (Potter, P. E, 1967)

In estuarine type deltas, the delta fringe sands appear to be much more common than distributary and point bar sands. Elongate tidal sand bodies in the estuaries probably extend parallel for great distances within the marine environment in front of the delta. The cross bedding is parallel with the elongation; the principal mode may point seaward as well as landward. Thickness may be more than 6.1 m-7.6 m and width more than 152.4 m-304.8 m. (Potter, P. E, 1967). The basal contact is disconformable and commonly marked by shelly gravel. Internal organisation is strong; grain size decreases upward, commonly from a basal gravel. Bed thickness shows a similar decline upward. Tidal sands have both chemical and detrital cements; associated with interlaminated shale and sandstone of tidal origin.

The coastal interdeltaic or shoreline shelf environment is characterised by six different types of deposits but all are related with each other: mud flat, barrier island, lagoon, tidal channel, and tidal delta. Sand bodies of barrier island origin are elongate and parallel to the coast line and separate marine from non-marine environments, giving maximum lithologic contrast. Multilateral sandstone bodies are common. Petrologically, the sands of most barrier island complexes are mature. Good sorting, less clay and strong abrasion promotes roundness and eliminates many rock fragments. Cements are mostly chemical. Data indicate a vertical increase in grain size and bed thickness, especially in regressive successions. The widths may range from a few meters to more than several kilometres, and thicknesses are of few meters. The shape of the sandstone bodies is fairly straight to gently curving, and there are some weakly branching patterns. Tidal channel sandstone bodies are oriented perpendicular to the barrier sands, and their thickness can vary considerably, depending on the depth of tidal channels.

Valleys are generally wider than channels, and it is critical to distinguish between them. Channels have average widths of less than 1 km and aspect ratio of 1: 100. By contrast valleys have an average width approximately 10 km and aspect ratio of 1:1000 (Table A .1)**.** Crevasse channels are the narrowest and shallowest; distributary channels are the widest and deepest. The bar finger sand-bodies (for example Mississippi delta) are elongate down dip, have unidirectional cross-bedding parallel with the long axis, and are fine grained (Reynolds, T. 2005). The sandstone bodies thin and narrow upstream, and coarsen upward. In an alluvial-delta complex, such sand stone bodies may be intercalated with ordinary sediments of flood-plain origin.

According to Reynolds (1994), the width of distributary mouth bars range from 1.1 to 14 km with and average of 2.8 km, and the length ranges from 2.6 to 9.6 km with an average of 6.5 km. Flood tidal delta complexes range between 1.7 km and 13.7 km in width, with an average width of 6.2 km, and between 25.7 and 2.9 km in length, with an average length of 12.3 km. correlating between the length and width suggests that flood tidal delta complexes tend to be twice as long as they are wide. Crevasse splays range in width from 7.7 km to18 m, with a mean width of 787, and range in length from 160 m to 11.7 km, with a mean length of 5577 m.

**Table A .1** Summary statistics of paralic sandstone after Reynolds (1994)

| | Widths (m) | | lengths (m) | | Thickness (m) | |
|---|---|---|---|---|---|---|
| | Mean | Range | mean | Range | Mean | Range |
| Incised valleys | 9 850 | 63 000-500 | | | 30.3 | 2-152 |
| Fluvial channels | 750 | 1400-60 | | | 9 | 24-2.5 |
| Distributary channels | 520 | 5900-20 | | | 7.8 | 40-1 |
| Crevasse channels | 60 | 400-5 | | | 2.4 | 17-0.2 |
| Shoreline-shelf sandstones | | | | | | |
| Undifferentiated | 25 000 | 106 000-1600 | 93 200 | 190 000-47 000 | 19.1 | 49-2.7 |
| Highstand system tract | 16 400 | 43 000-16 000 | | | | |
| Transgressive system tract | 7200 | 20 000-3300 | | | | |
| Distributary mouth bars | 2900 | 14 000-1100 | 6500 | 9600-2400 | 9.7 | 42-1.2 |
| Tidal ridges | 6400 | 29 300-750 | 17 500 | 61 000-3250 | 11.2 | 22.9-3 |
| Crevasse splays | 800 | 7700-20 | 5600 | 11 700-160 | 1.4 | 12-0.3 |
| Flood tidal delta complex | 6200 | 13 700-1700 | 12 300 | 25 700-2900 | 6.7 | 23-1.8 |
| Lower tidal flat | 1000 | 1550-400 | | | 4.6 | 9-2.0 |
| Tidal creeks | 810 | 1550-160 | | | 5.2 | 18-1 |
| Tidal inlet | 1850 | 2550-700 | 4300 | 4300 | 4.8 | 3-7 |
| Estuary mouth shoal | 2400 | 2900-1700 | 3750 | 4700-2200 | 10 | 10-35 |
| Chenier | 3650 | 6400-900 | 21 800 | 38 600-49 000 | 5.8 | 7-4.6 |
| All sand | 5050 | 106 000-5 | 35 300 | 190 000-160 | ? | ? |

### A.7.1 Controls in development of paralic sandstone bodies

The fundamental controls that affect the development of paralic sandstone bodies combine in different ways in different paralic systems. The downstream reaches of the fluvial system are affected most by changes in relative sea level. By definition, valley fills form in lowstand systems tracts. Commonly, the gradient of the lower reaches of the river is less than the gradient of the coeval shelf and a relative sea level fall will impose a steeper gradient on the graded profile of the river.

The broad links between a relative sea level cycle and its impact on fluvial and shoreface architectural patterns are schematically represented in (Fig. A.14).The impact of base level rise is equally complex. In these cases the river responds by depositing sediment. Rapid base level rise may be considered analogous to damning down stream portions of the river valley. In these cases the rate of rise will be greater than the rate of deposition, and the lower course of the fluvial system will be flooded. Significant fluvial aggradation will mainly occur as the shoreline progrades basinward by deltaic extension; that is when the rate of sediment supply exceeds the rate of base level rise. The net result may be a narrowing of the river channel belt, a change in channel style and a decrease in sand body connectivity. Where the rate of transgression is lower, the fluvial record of a transgressive system tract may include the increased partitioning of channel and flood plain deposits in vertical section, the decreasing connectivity of channel meander-belt sands, and the increasing influence of tidal processes in the lower reaches of the fluvial system (Allen, 1991; Shanley and McCabe, 1993). Fig. A.15 shows the infill pattern of the incised valley in terms of changing accommodation/ space (A/S) and sedimentary processes.

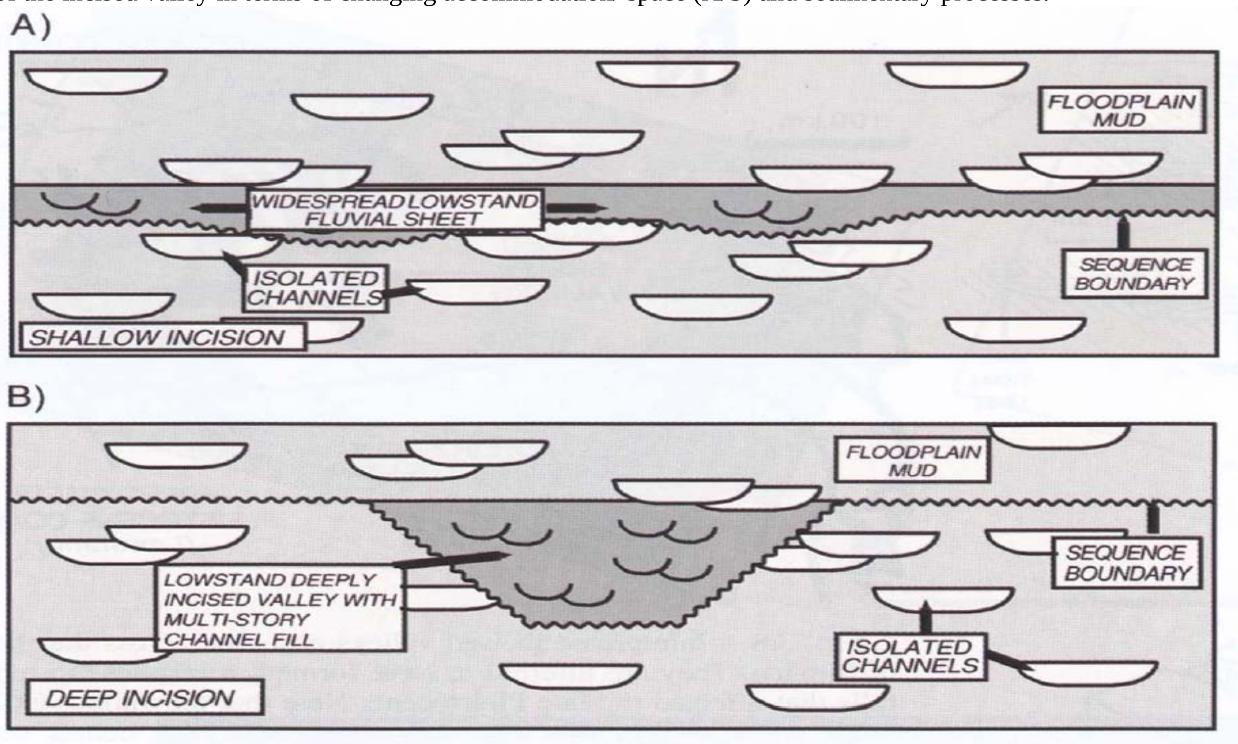



**Fig. A.14:** Incision in shelf during fall in relative sea level and related sandstone body architecture. **(A)** widespread shallow fluvial incision forming sandstone sheets that can extend over tens or hundreds of kilometres along strike and dip. **(B)** Deep narrow incision forming incised valley with multi-storey channel sandstone bodies (Posamentier and Allen1999, Fig. 3.9)

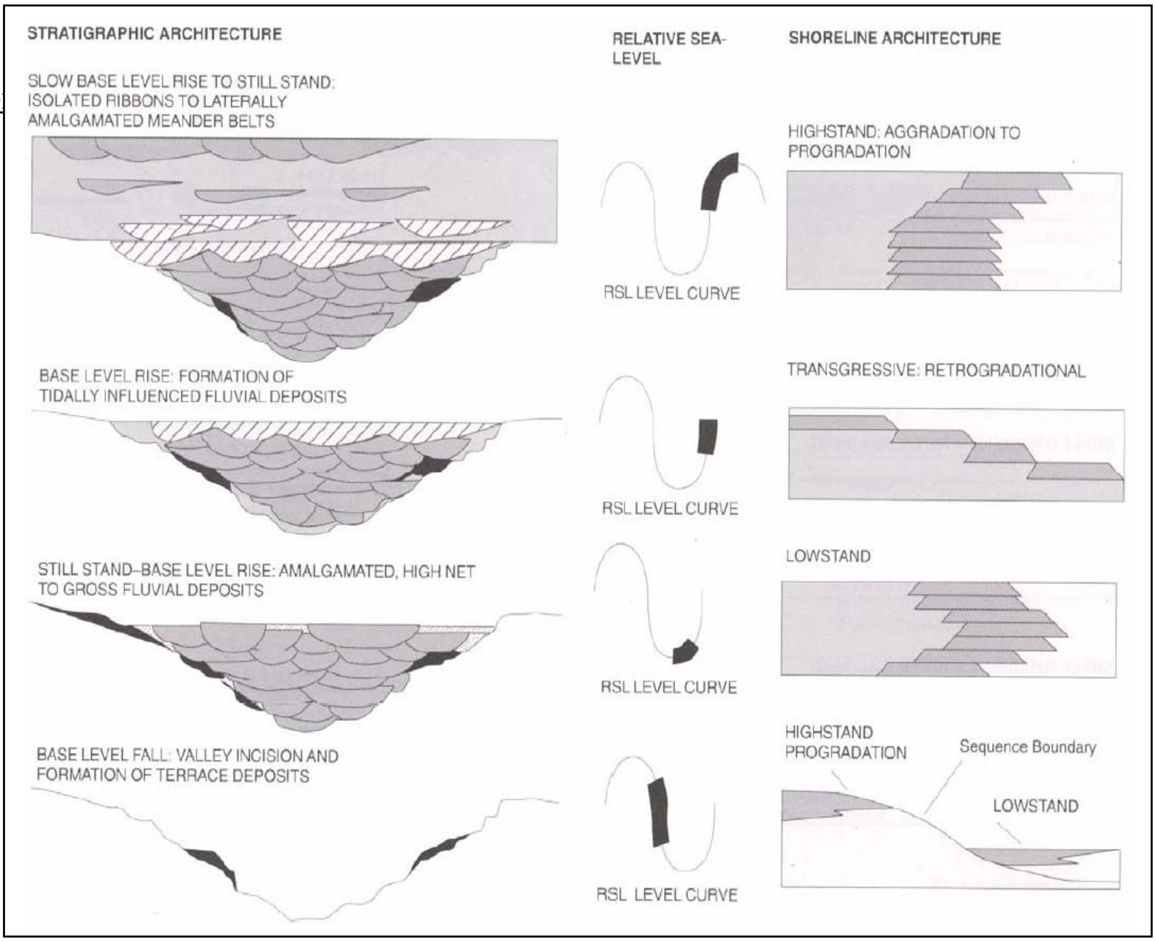



**Fig. A.15** Evolution of shoreface and fluvial architecture during fall and rise in relative sea level; Model by Shanley and McCabe, (1994). The development starts with the bottom figure; as seen from the sea level curve.

Detailed correlations suggest the timing of the incised-valley-fill occurred after the initial transgressive surface. During the late highstand, reductions in the flood basin accommodation may result in the increased potential for lateral, rather than vertical accretion and the formation of laterally interconnected and amalgamated channel and meandering belt systems with poorly preserved flood-plain deposits (Shanley and McCabe, 1993; (Fig. A.15))

Rhythmic stacking pattern is thought to reflect low-frequency changes in rate of generation of accommodation space and sediment supply. Changes of this type are likely to have a progressive influence on sandstone body dimensions. Given constant sediment supply, shoreline-shelf sandstone body width would be expected to increase through a highstand set of strata because the amount of accommodation space generated during each short-term fluctuation in relative sea level progressively decreases.

It has been analysed that distinct sandstone body types occur preferentially in specific system tracts. Majority of shoreline–shelf sandstone bodies occur in highstand system tracts

The overwhelming majority of sand bodies deposited landward of the shoreline (e.g. distributary channels, crevasse splay, tidal creeks, and flood tidal deltas) occur in transgressive system tracts, suggesting that accommodation space in the delta plain and coastal plain is generated largely during transgression. As well as being preferentially preserved during transgression, certain sandstone body types are favoured when base level is rising rapidly. For example, flood tidal deltas develop in lagoons behind barriers, which are favoured during transgression. Similarly, base level rise may also favour crevassing.

Mouth bars commonly characterise strong prograding conditions and often are developed in late highstand and early lowstand (Shanley and McCabe, 1993).

In basins with high wave and storm energy (Fig.A.16), the transition from shelfal to non marine conditions commonly is recorded by a cleaning and coarsening upward succession. At their base, these successions are characterised by discrete shelfal storm beds that become thicker and increasingly amalgamated upward before passing gradationally into shoreline sands (Walker, 1990; Reynolds, 1992). In high energy settings shoreline and shelf environments are linked, sand is supplied from the shoreline to the inner shelf to form a single continuous sand-body (Walker, 1990). Larger sandstone bodies are assumed to be parasequence sets rather than discrete sand bodies.

Also it has been revealed that sandstones that occur in high stand system tracts (HST) tend to be wider with a mean width of 16.4 km, over twice that of similar sandstones deposited in transgressive system tract (TST) that have mean widths of 7.2 km. similarly, the mean width of shoreline shelf sands deposited in progradational parasequence sets (17.3 km) is over three times that of similar sandstones deposited in retrogradational parasequence sets (5.3 km). Hence, shoreline-shelf sands are huge sheets with widths of kilometres to tens of kilometres and lengths that may range up to hundreds of kilometres (e.g. Palmer and Scott, 1984). But in some cases thicknesses are relatively restricted that may range only for some few meters.

Laterally, barrier sandstone bodies may coalesce, a condition depending on stability of shoreline. The vertical sequence changes according to regression and transgression. The barrier sandstone body may be overlapped by continental deposits, if it is part of a regressive cycle, or by marine sediment, if it is part of a transgressive cycle. Multi-storey sandstone bodies are relatively common and may be of mixed origin-alluvial or tidal sandstone bodies superimposed at right angles on barrier island sand bodies

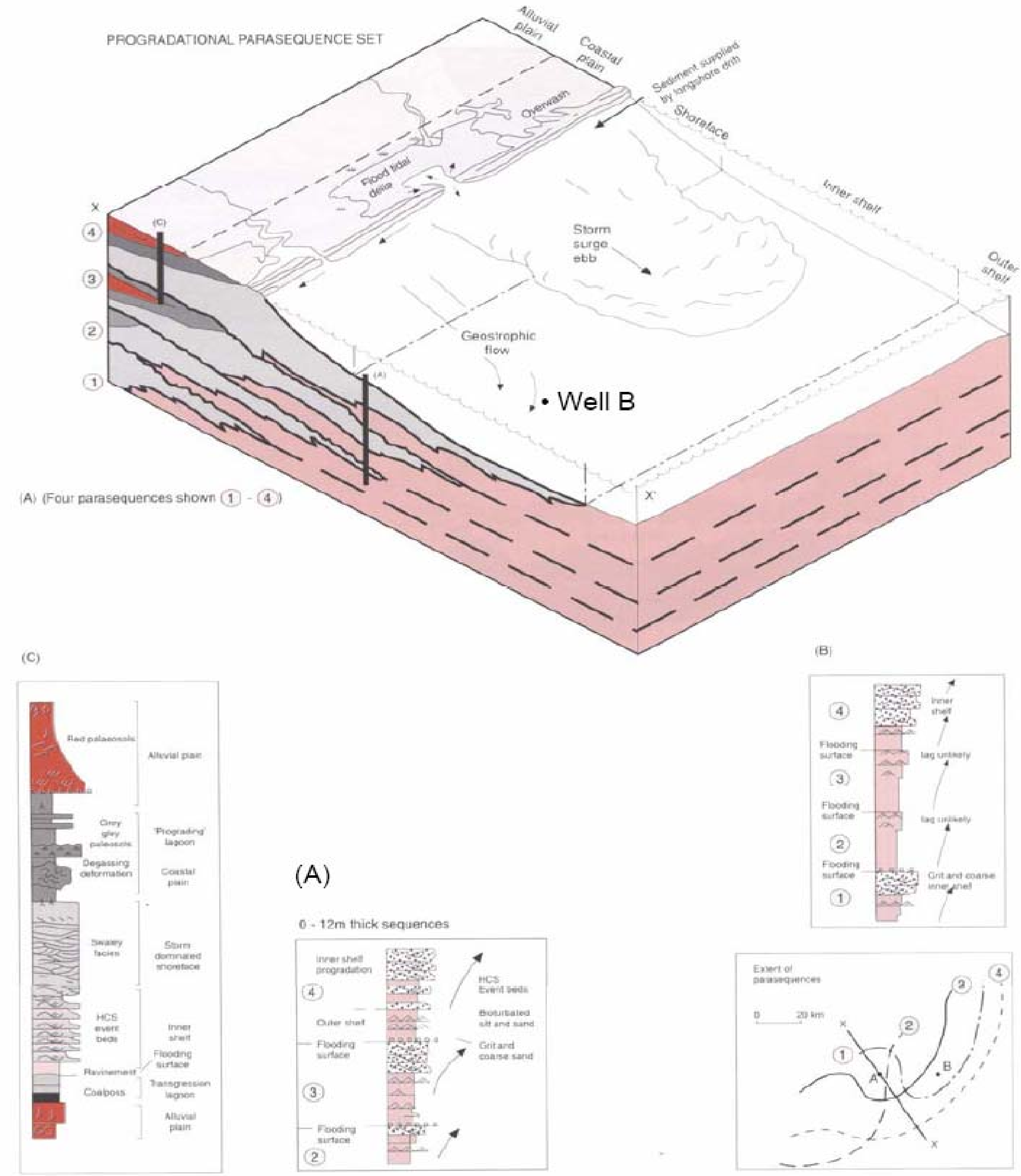


**Fig. A.16:** Progradational parasequence set in a storm dominated shoreline shelf succession. A, B, and C are wells. The map shows the extent of shoreface sands in each parasequence 1-4.
Well A with parasequences 2, 3, and 4. Notice the progradational vertical stacking pattern of the three parasequences. Also notice the position of well B relative to pinchout lines for sand in the various parasequence (Devine, (1991) and Van Wagooner et al., 1990)

**A.8 Summary: Paralic sandstone reservoir**
The most common paralic sandstone reservoirs are of deltaic origin. They are laterally equivalent to fluvial sands and prodelta and marine clays, and they consist of two types: delta front or fringe sands and abandoned distributary channel sands. Fringe sands are sheet like, and their landward margins are abrupt against organic clays of the deltaic plain. Seaward, these sands grade into the finer prodelta marine sediments. Distributary channel sandstone bodies are narrow, they have abrupt basal contacts, and they decrease in grain size upward. They cut into, or completely through, the fringe sands, and also connect with the upstream fluvial sands or braided or meandering streams.

Some of the more porous and permeable sandstone reservoirs are deposited in coastal interdeltaic realm of sedimentation. They consist of well sorted beach and shoreface sands associated with barrier islands and tidal channels which occur between barriers. Barrier sand bodies are long and narrow, are aligned parallel with the coastline, and are characterised by an upward increase in grain size. They are flanked on the landward side by lagoonal clays and on the opposite side by marine clays. Tidal channel sand bodies have abrupt basal contacts and range in grain size from coarse at the base to fine at the top. Laterally, they merge with barrier sands and grade into finer sediments of tidal deltas and mud flats.

The most porous and permeable reservoir sandstones are products of wind activity in coastal regions. Wind laid (aeolian) sands are typically very well sorted and highly cross bedded, and they occur as extensive sheets. Marine sandstones are those associated with normal marine processes of the continental shoreline and shelf system.

The dimensions of paralic sandstone bodies vary widely. The main control on these variations is sandstone body type, with different sandstone body types having distinct ranges of dimensions. For some sandstone body types a robust correlation exists between sandstone body thickness and sandstone body width. Systematic relationships exist between sandstone body dimensions and sequence stratigraphic cyclicity

From the mean dimensions of paralic sandstone (Table A1), it can be deduced that storm-dominated shoreline-shelf sands and incised valleys occur on the same scale as all but the largest fields; mouth bars have areas comparable to small fields or segments of large fields; and fluvial channels, crevasse splays are small aerially.

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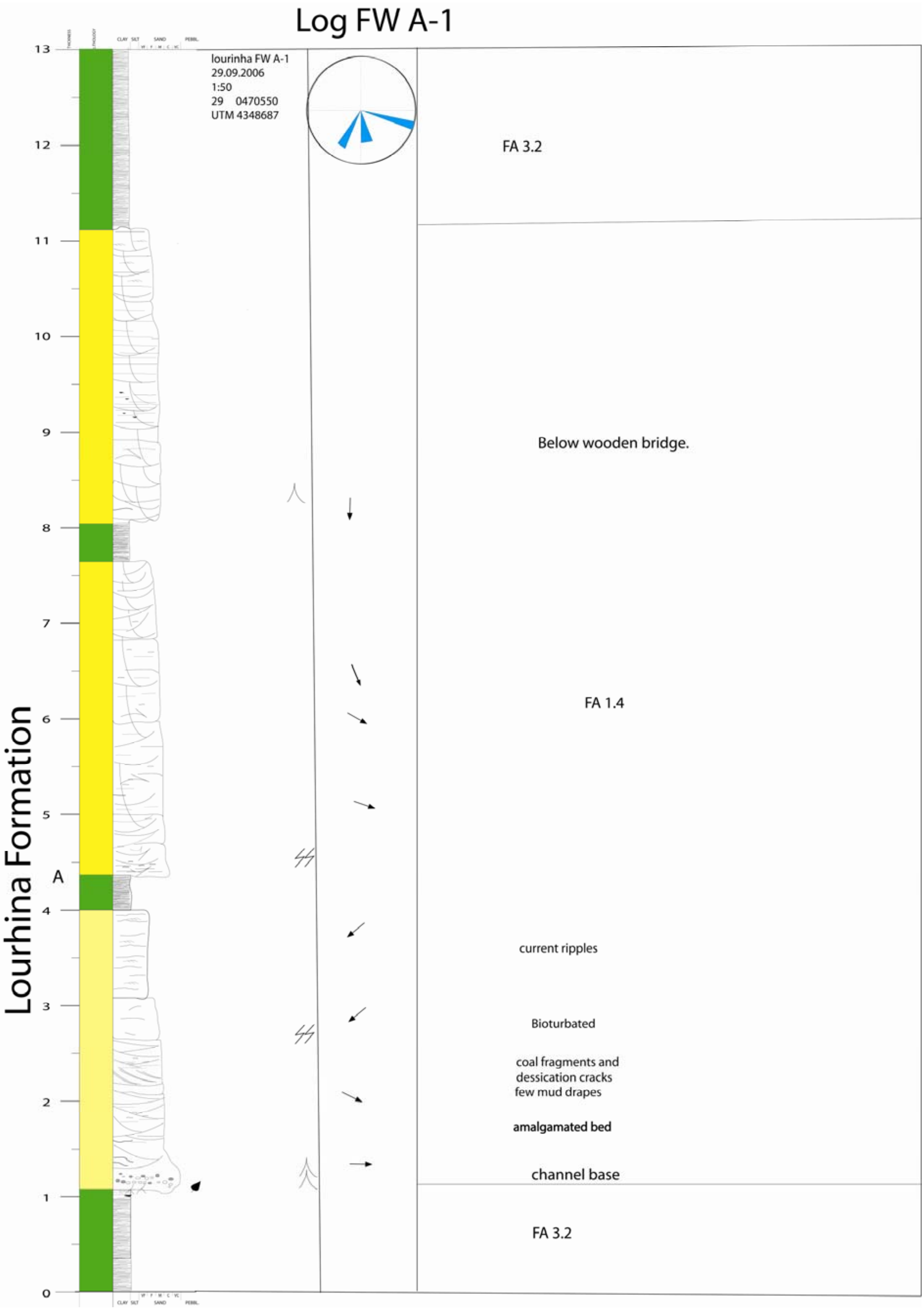
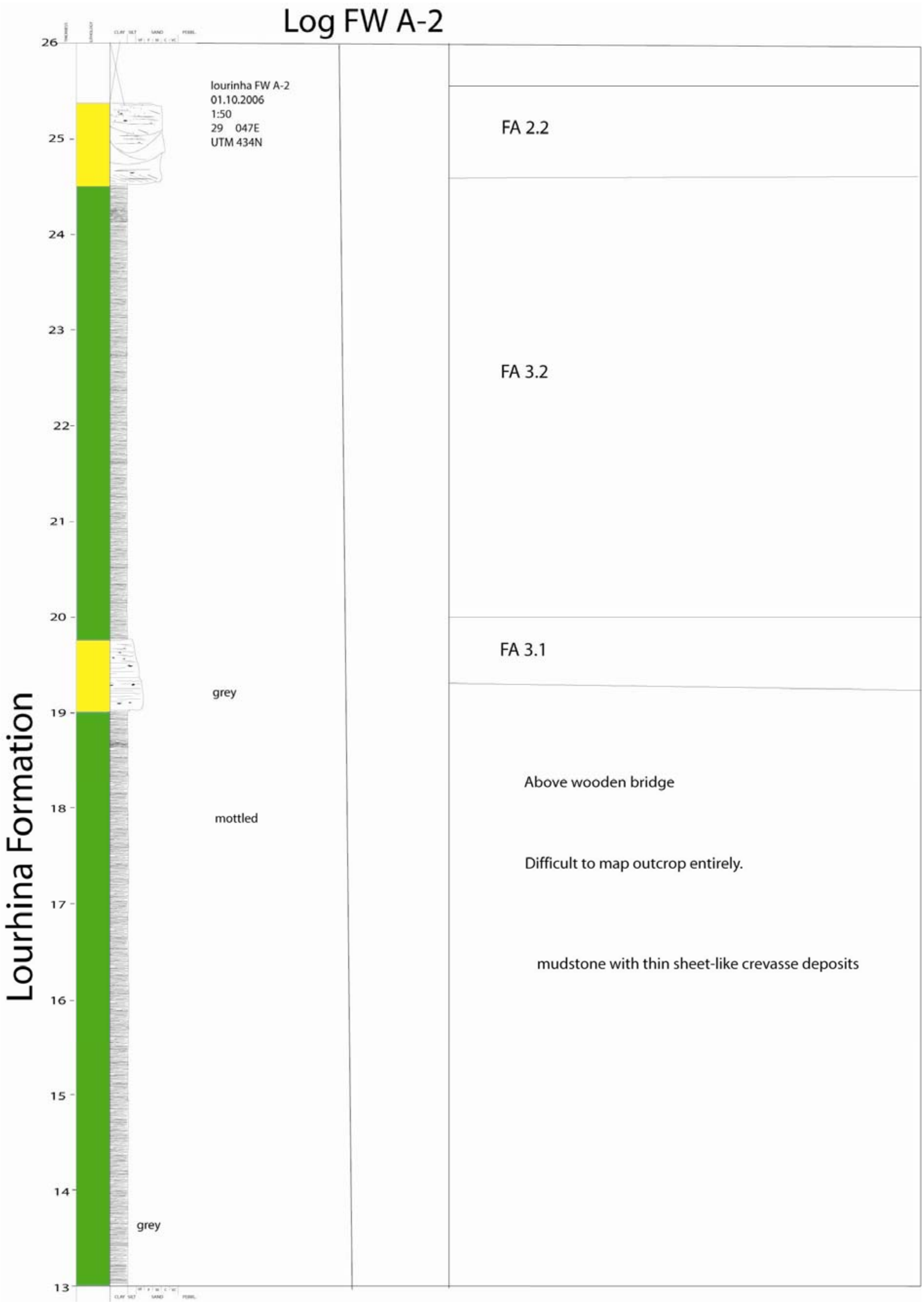
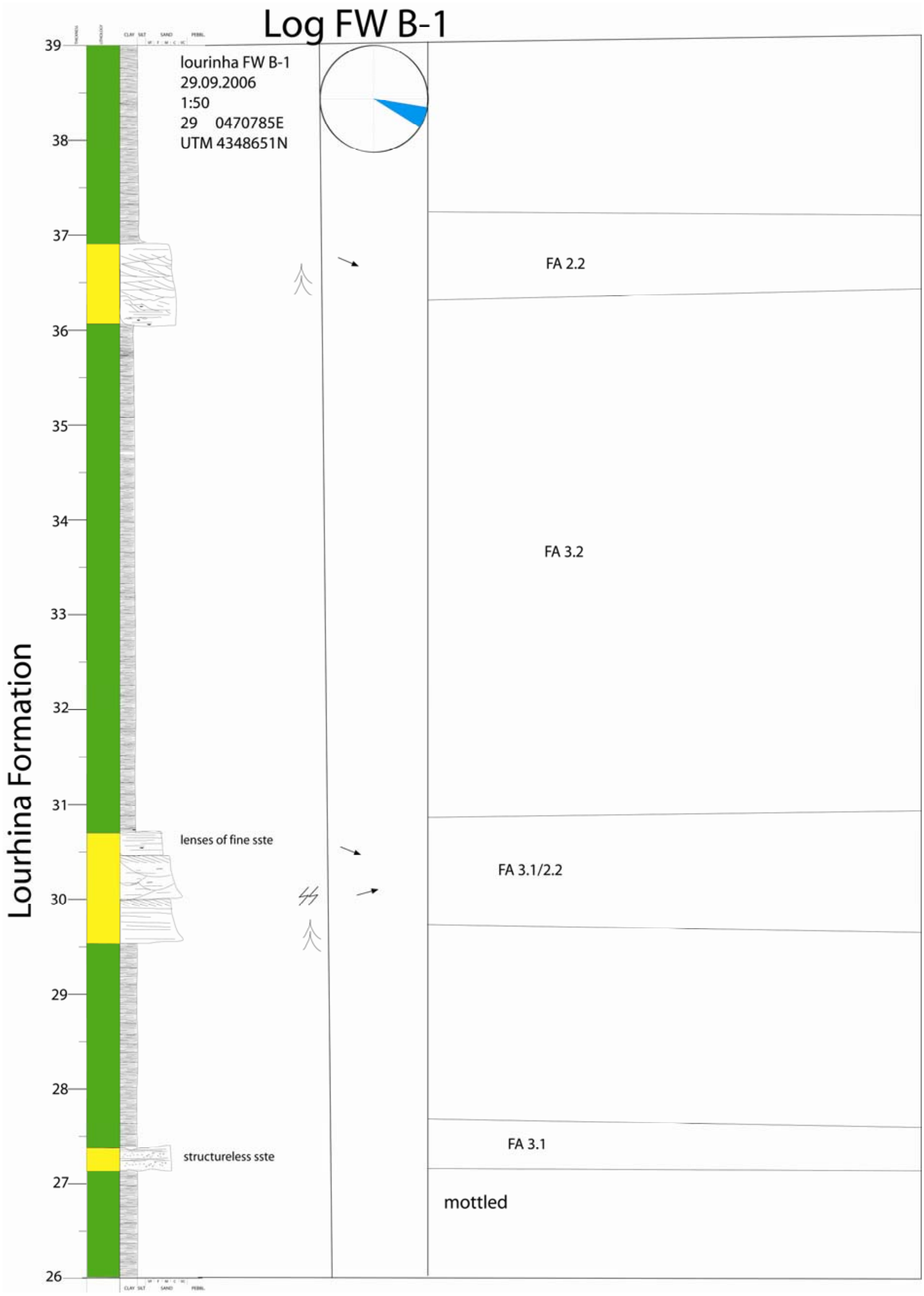
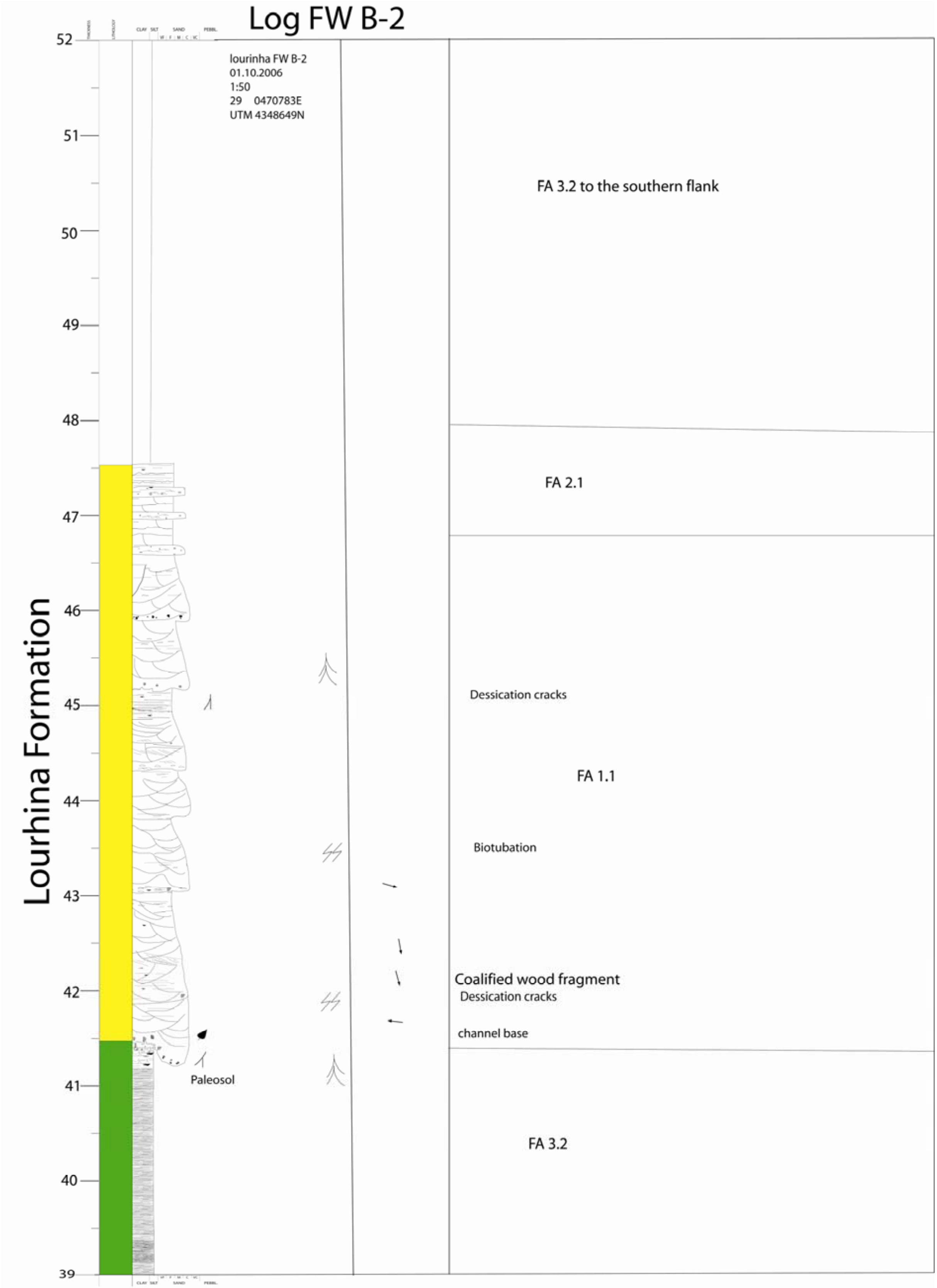
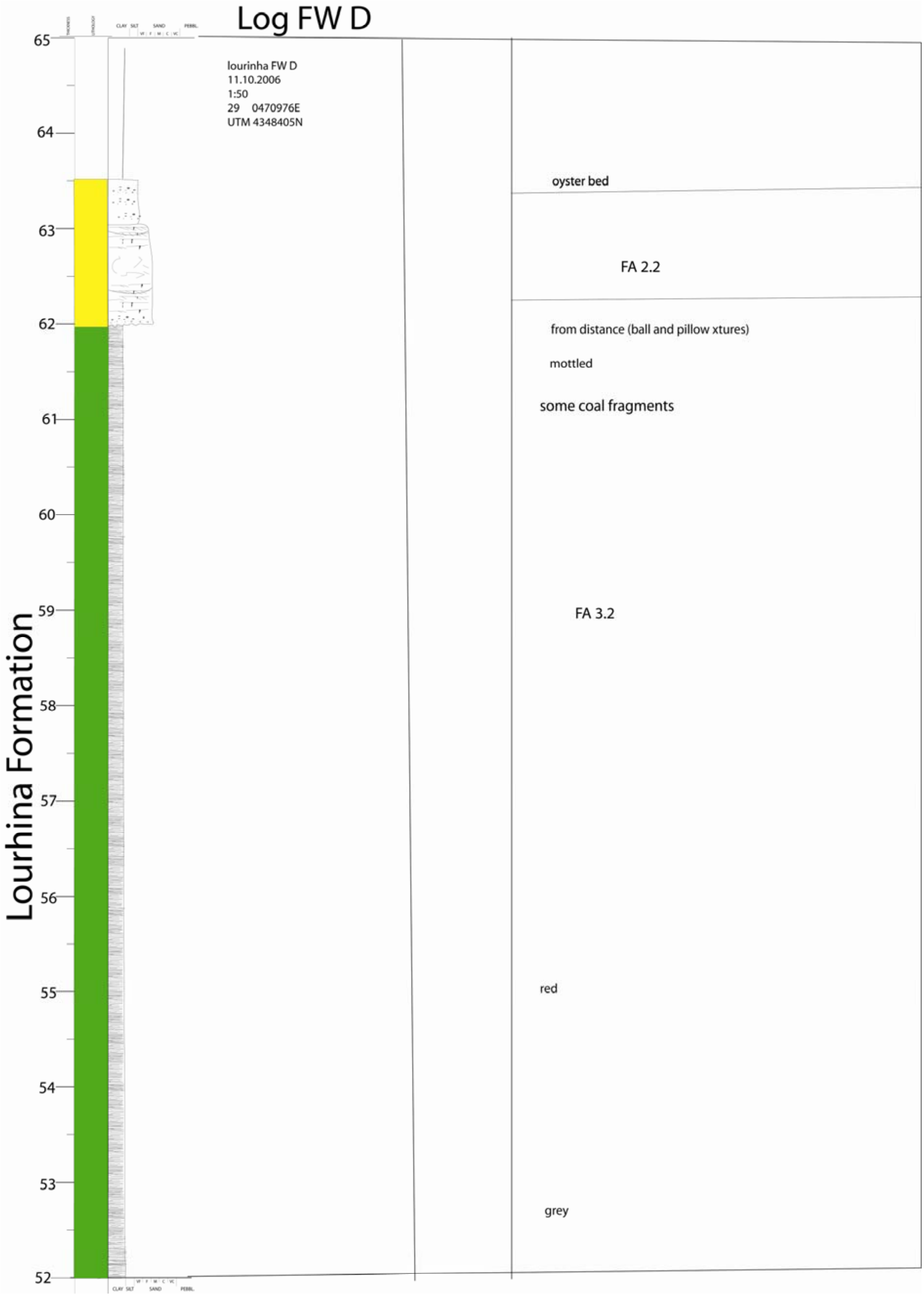


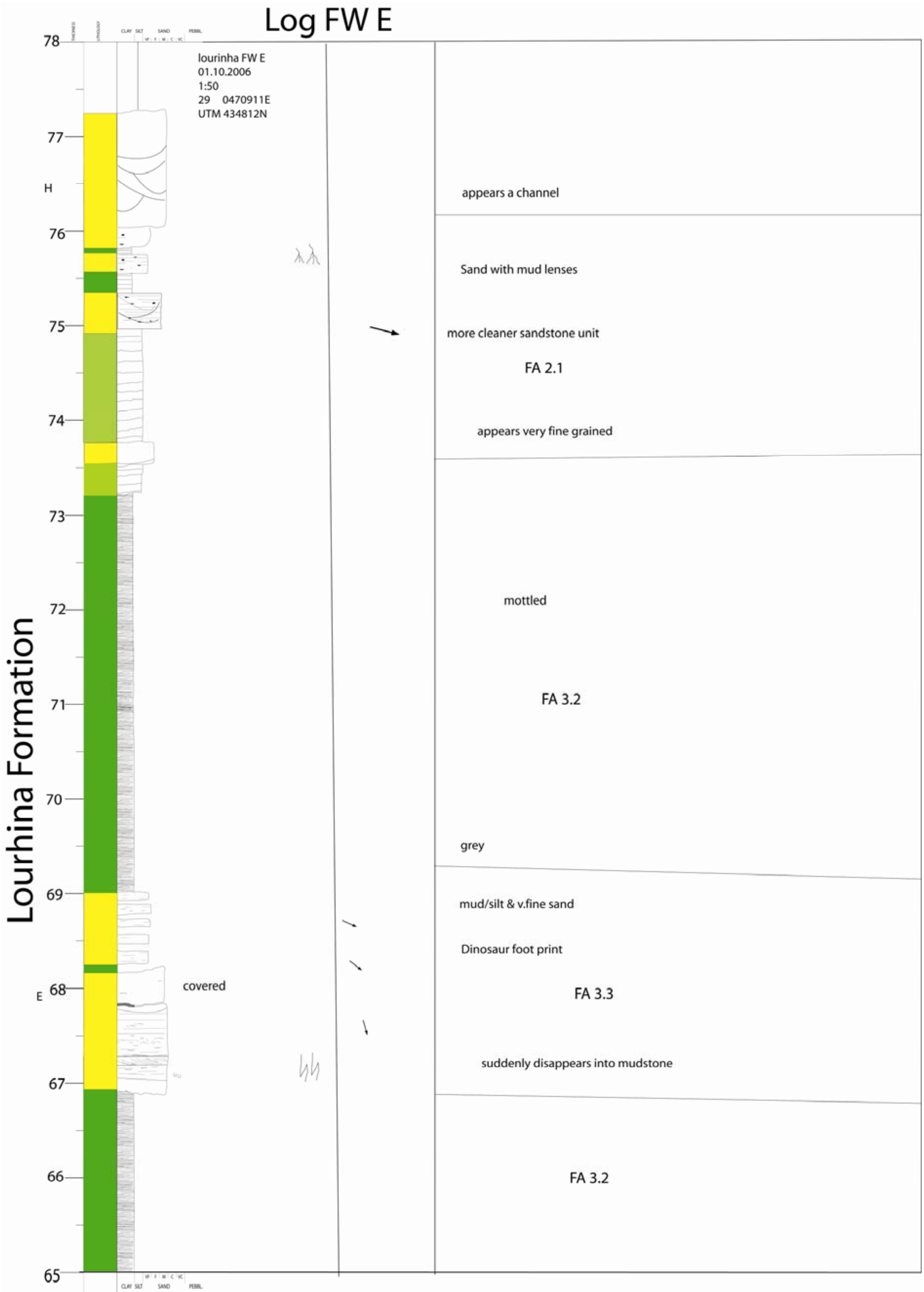
Log FW A-1
lourinha FW A-1
29.09.2006
1:50
29 0470550
UTM 4348687
FA 3.2
Below wooden bridge.
FA 1.4
current ripples
Bioturbated
coal fragments and
dessication cracks
few mud drapes
amalgamated bed
channel base
FA 3.2
Lourhina Formation
A
13
12
11
10
9
8
7
6
5
4
3
2
1
0

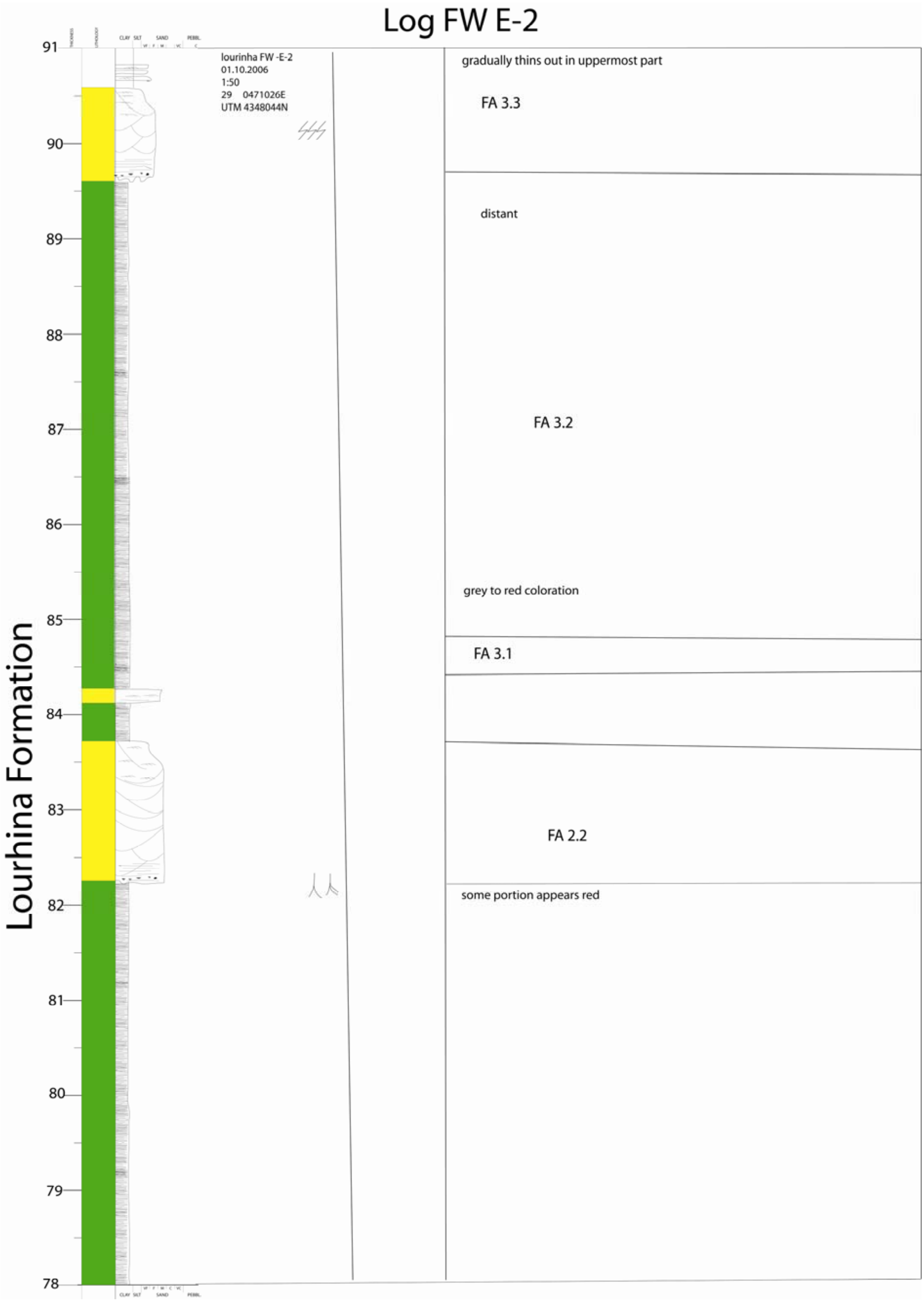

Log FW A-2
lourinha FW A-2
01.10.2006
1:50
29 047E
UTM 434N
FA 2.2
FA 3.2
FA 3.1
grey
mottled
grey
Above wooden bridge
Difficult to map outcrop entirely.
mudstone with thin sheet-like crevasse deposits
Lourhina Formation
26
25
24
23
22
21
20
19
18
17
16
15
14
13

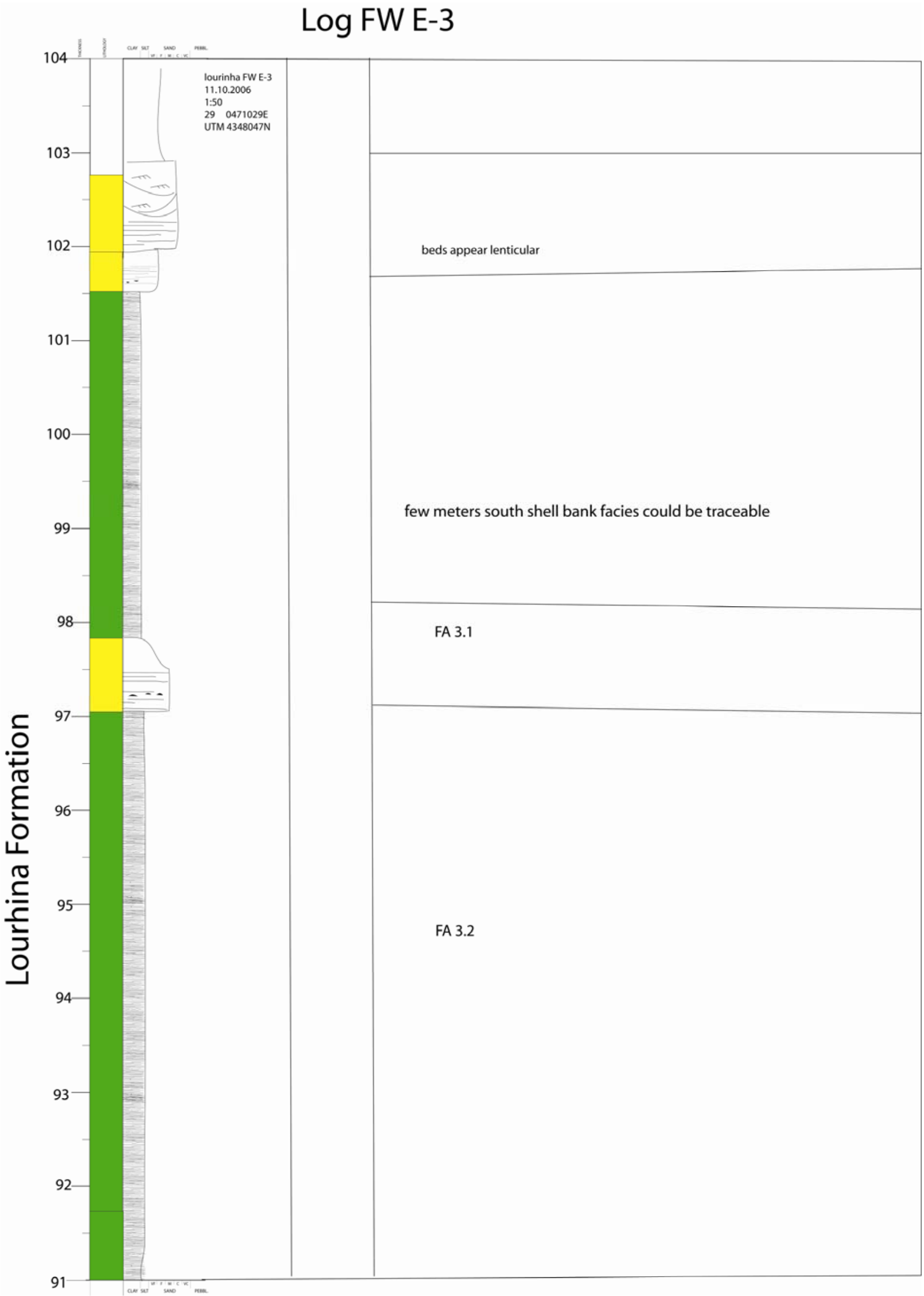

Log FW B-1
lourinha FW B-1
29.09.2006
1:50
29 0470785E
UTM 4348651N
lenses of fine sste
structureless sste
FA 2.2
FA 3.2
FA 3.1/2.2
FA 3.1
mottled
Lourhina Formation
39
38
37
36
35
34
33
32
31
30
29
28
27
26

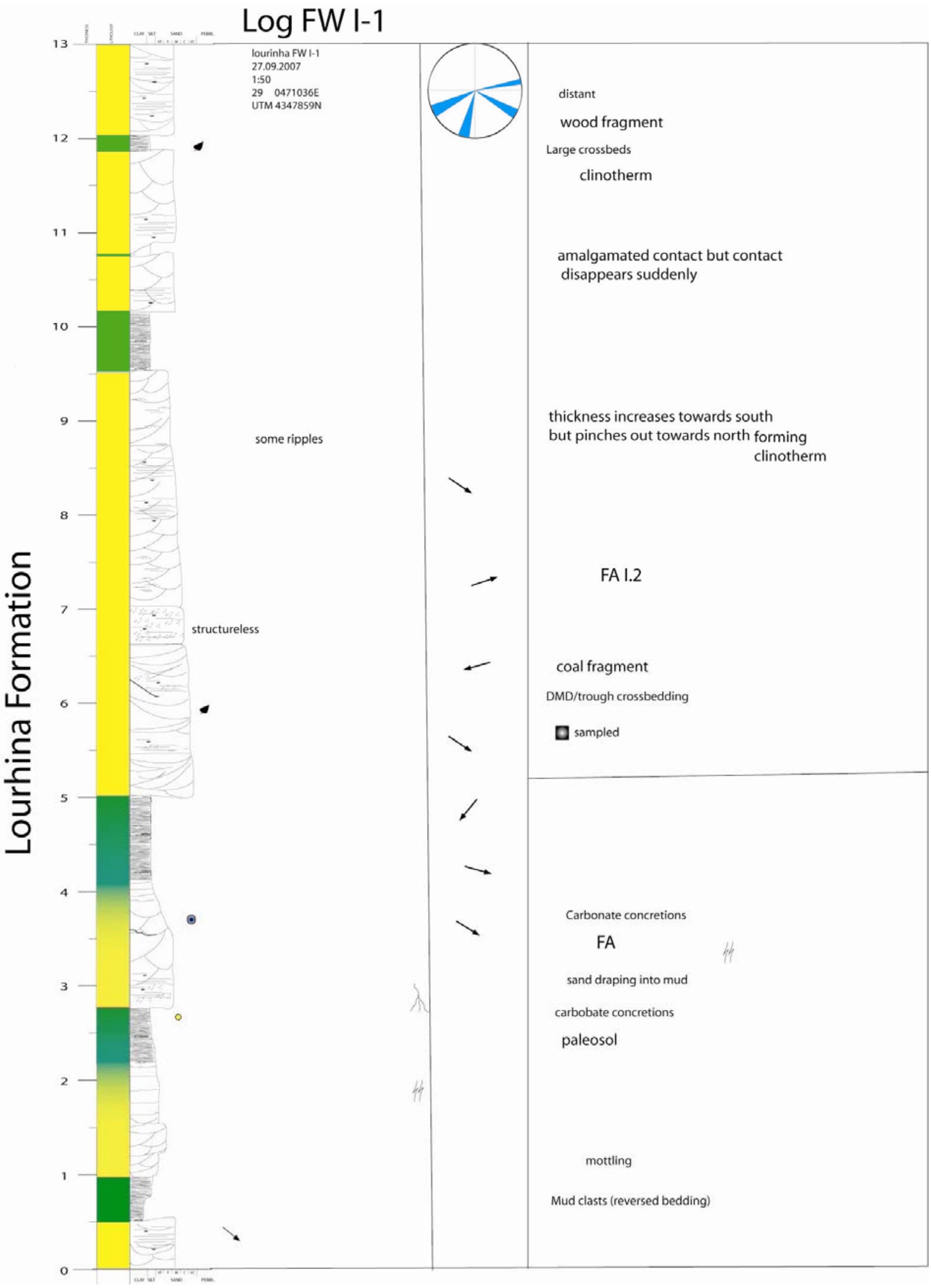

Log FW B-2
lourinha FW B-2
01.10.2006
1:50
29 0470783E
UTM 4348649N
Lourhina Formation
52
51
50
49
48
47
46
45
44
43
42
41
40
39
Paleosol
FA 3.2 to the southern flank
FA 2.1
Dessication cracks
FA 1.1
Biotubation
Coalified wood fragment
Dessication cracks
channel base
FA 3.2

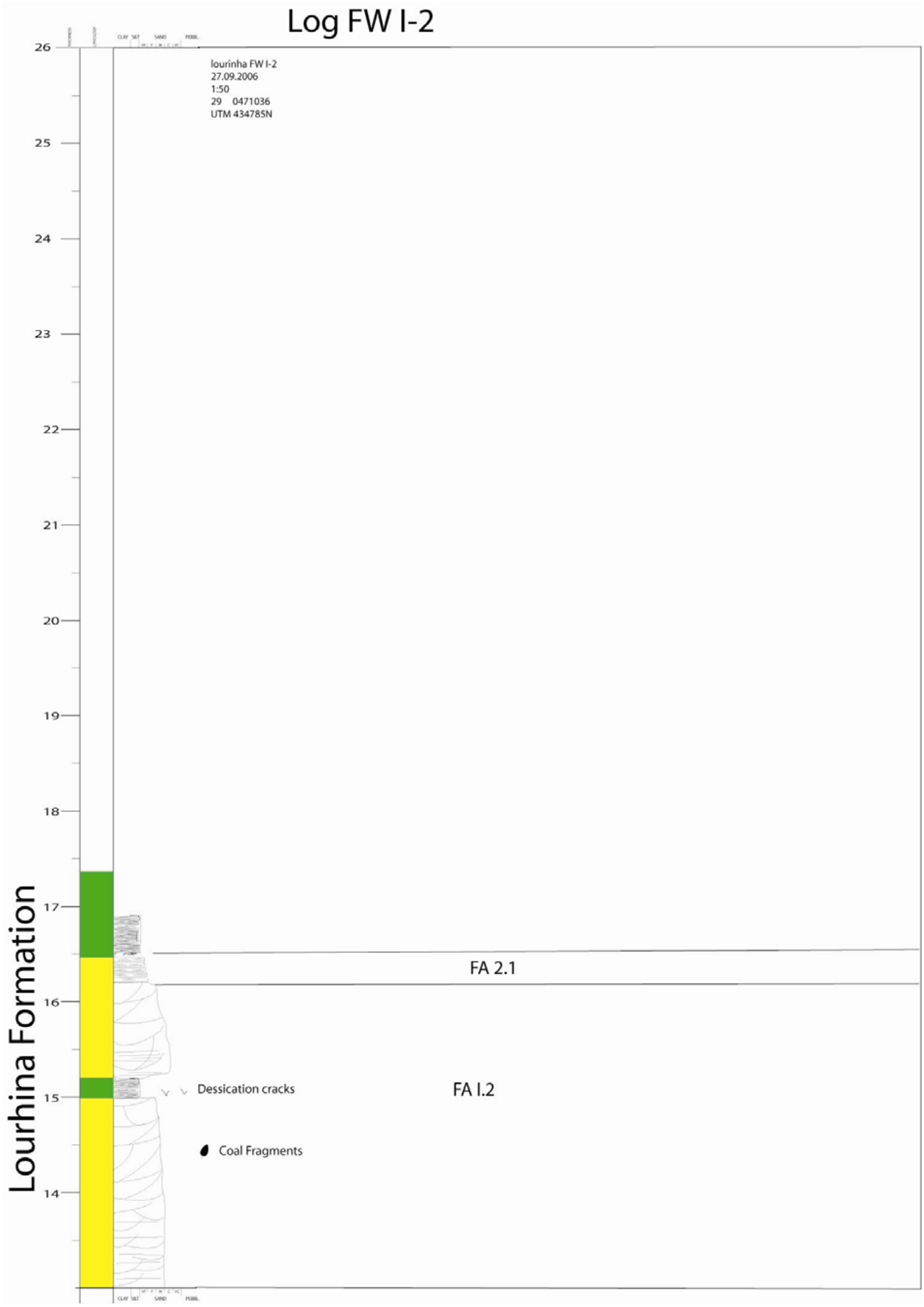

Log FW D
lourinha FW D
11.10.2006
1:50
29 0470976E
UTM 4348405N
65
64
63
62
61
60
59
58
57
56
55
54
53
52
Lourhina Formation
CLAY SILT SAND PEBBL.
oyster bed
FA 2.2
from distance (ball and pillow xtures)
mottled
some coal fragments
FA 3.2
red
grey

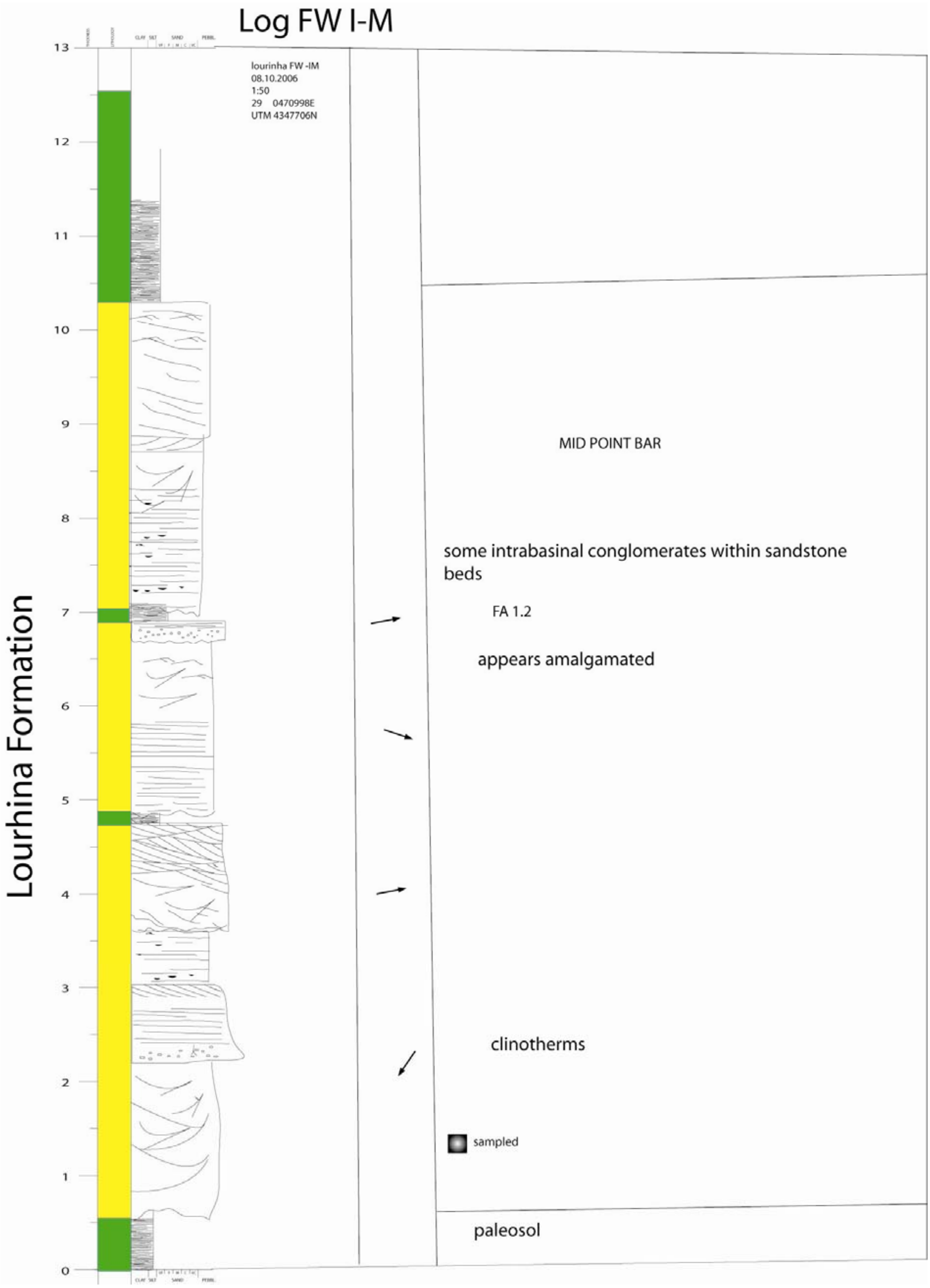

Log FW E
lourinha FW E
01.10.2006
1:50
29 0470911E
UTM 434812N
Lourhina Formation
78
77
H
76
75
74
73
72
71
70
69
E 68
67
66
65
appears a channel
Sand with mud lenses
more cleaner sandstone unit
FA 2.1
appears very fine grained
mottled
FA 3.2
grey
mud/silt & v.fine sand
Dinosaur foot print
covered
FA 3.3
suddenly disappears into mudstone
FA 3.2

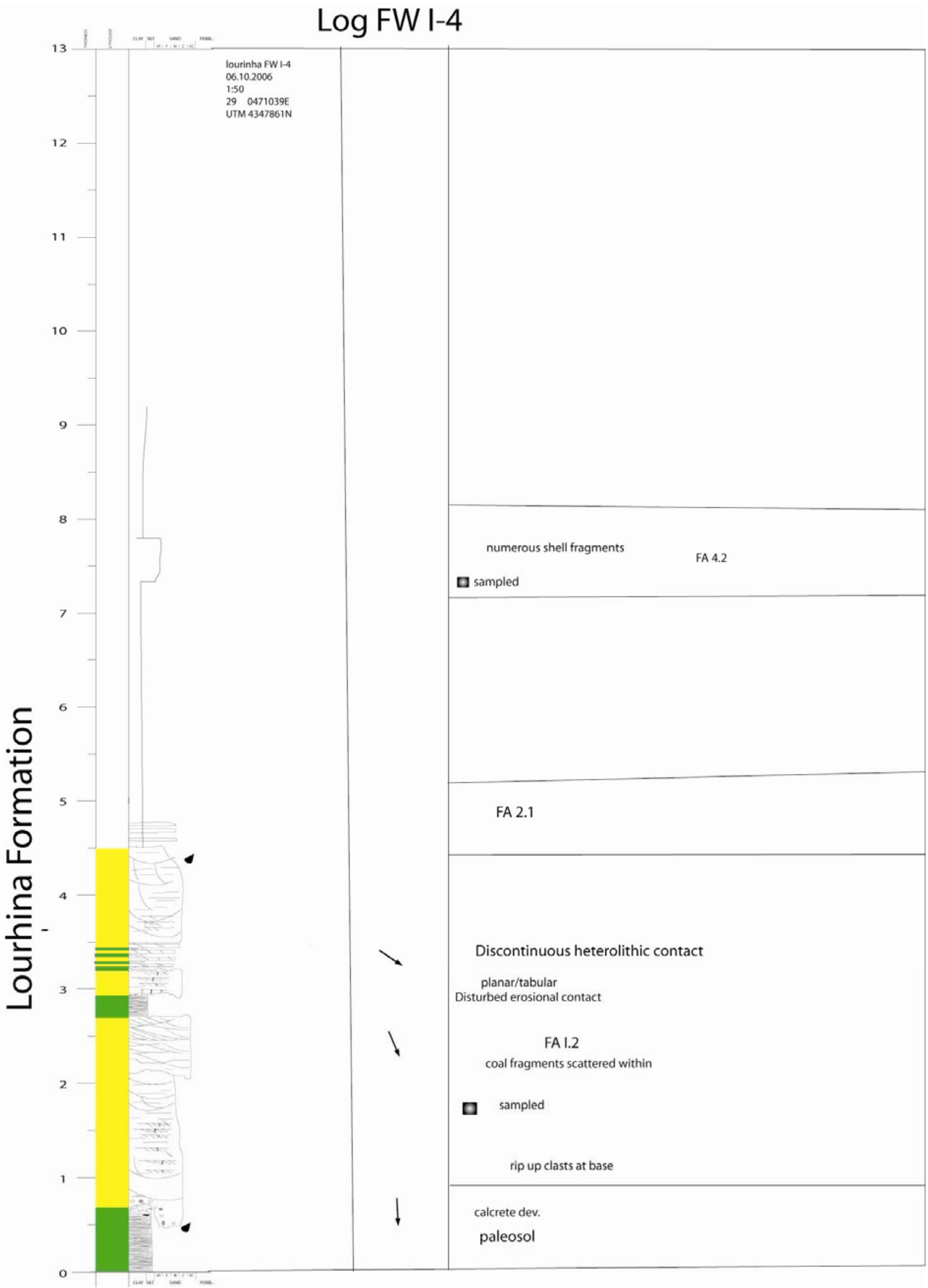

Log FW E-2
Lourhina Formation
91
90
89
88
87
86
85
84
83
82
81
80
79
78
lourinha FW -E-2
01.10.2006
1:50
29 0471026E
UTM 4348044N
gradually thins out in uppermost part
FA 3.3
distant
FA 3.2
grey to red coloration
FA 3.1
FA 2.2
some portion appears red


Log FW E-3
lourinha FW E-3
11.10.2006
1:50
29 0471029E
UTM 4348047N
104
103
102
101
100
99
98
97
96
95
94
93
92
91
Lourhina Formation
beds appear lenticular
few meters south shell bank facies could be traceable
FA 3.1
FA 3.2

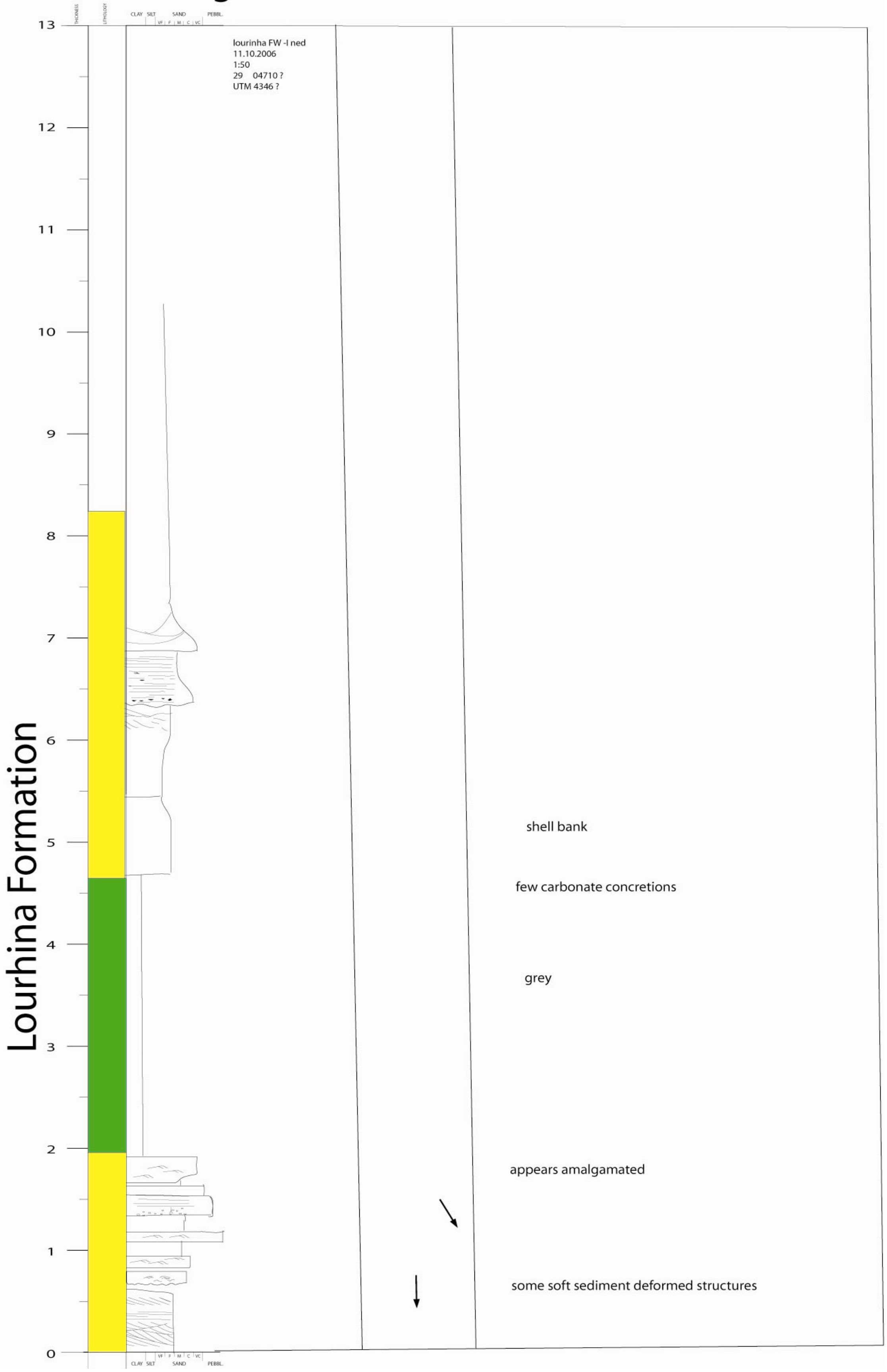

Log FW I-1
lourinha FW I-1
27.09.2007
1:50
29 0471036E
UTM 4347859N
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0
distant
wood fragment
Large crossbeds
clinotherm
amalgamated contact but contact
disappears suddenly
some ripples
thickness increases towards south
but pinches out towards north forming
clinotherm
FA I.2
structureless
coal fragment
DMD/trough crossbedding
sampled
Carbonate concretions
FA
sand draping into mud
carbobate concretions
paleosol
mottling
Mud clasts (reversed bedding)

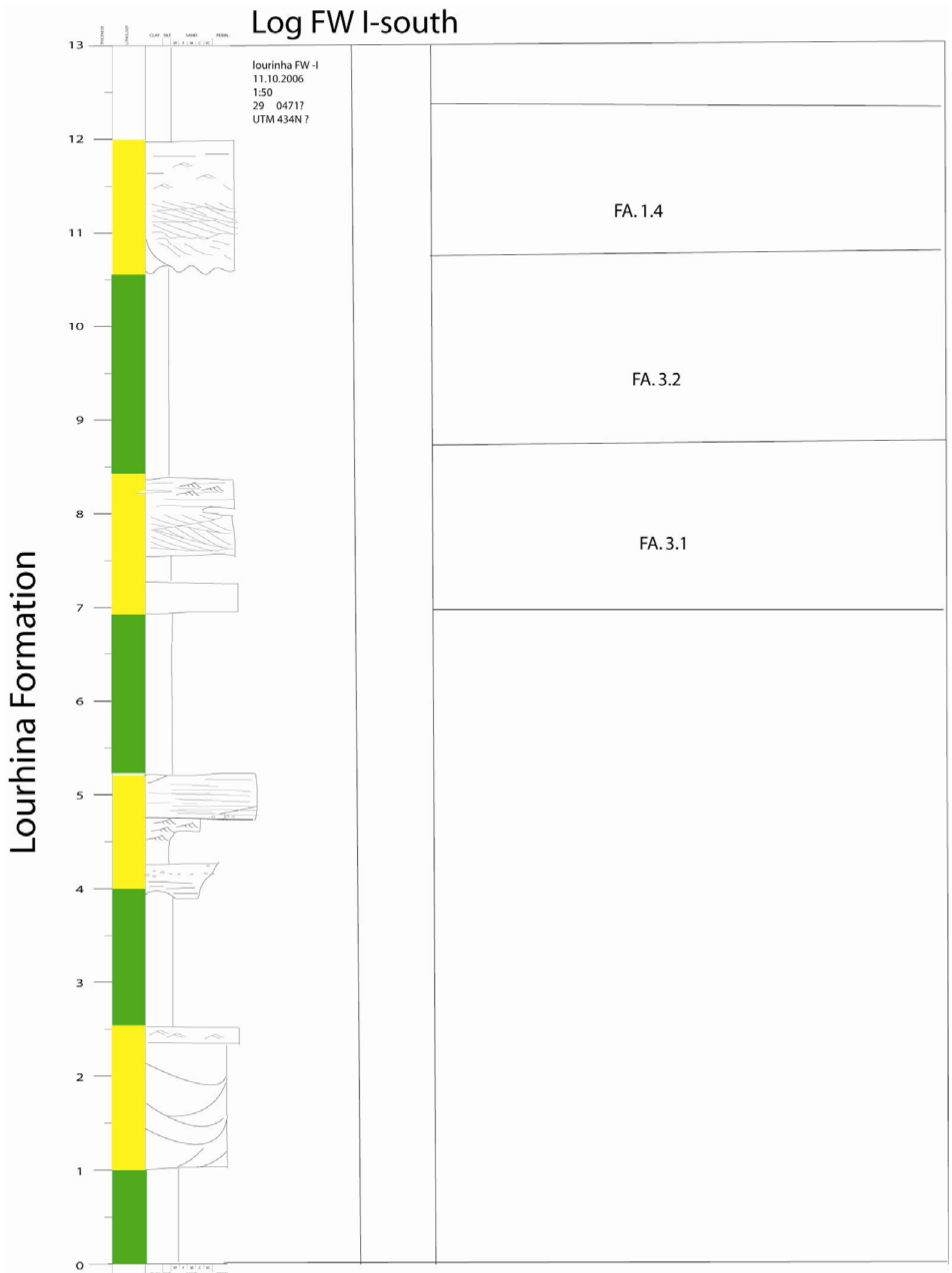

Log FW I-2
lourinha FW I-2
27.09.2006
1:50
29 0471036
UTM 434785N
Lourhina Formation
FA 2.1
FA I.2
Dessication cracks
Coal Fragments
26
25
24
23
22
21
20
19
18
17
16
15
14
CLAY SILT SAND PEBBL.

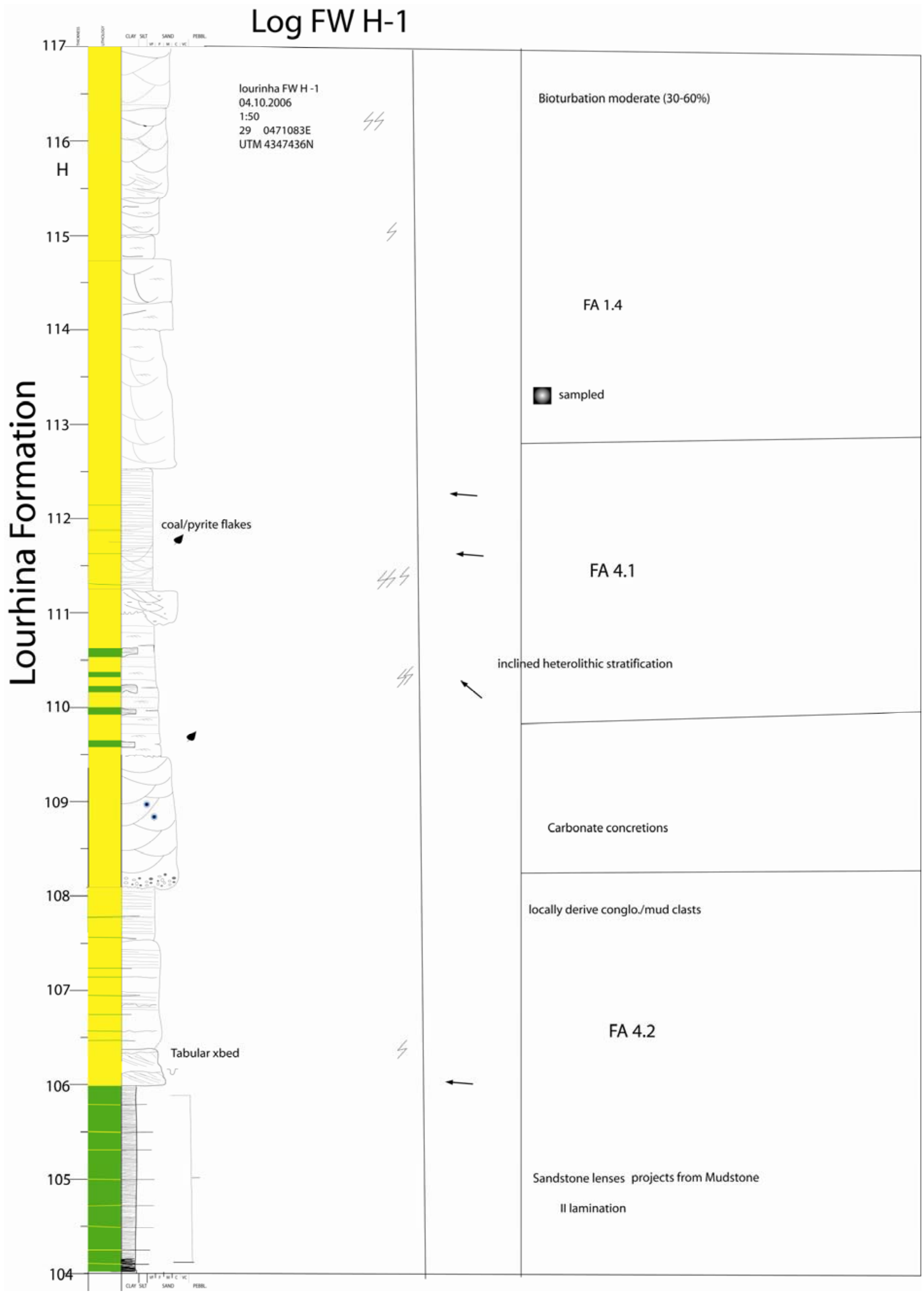

Log FW I-M
lourinha FW -IM
08.10.2006
1:50
29 0470998E
UTM 4347706N
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0
MID POINT BAR
some intrabasinal conglomerates within sandstone beds
FA 1.2
appears amalgamated
clinotherms
sampled
paleosol

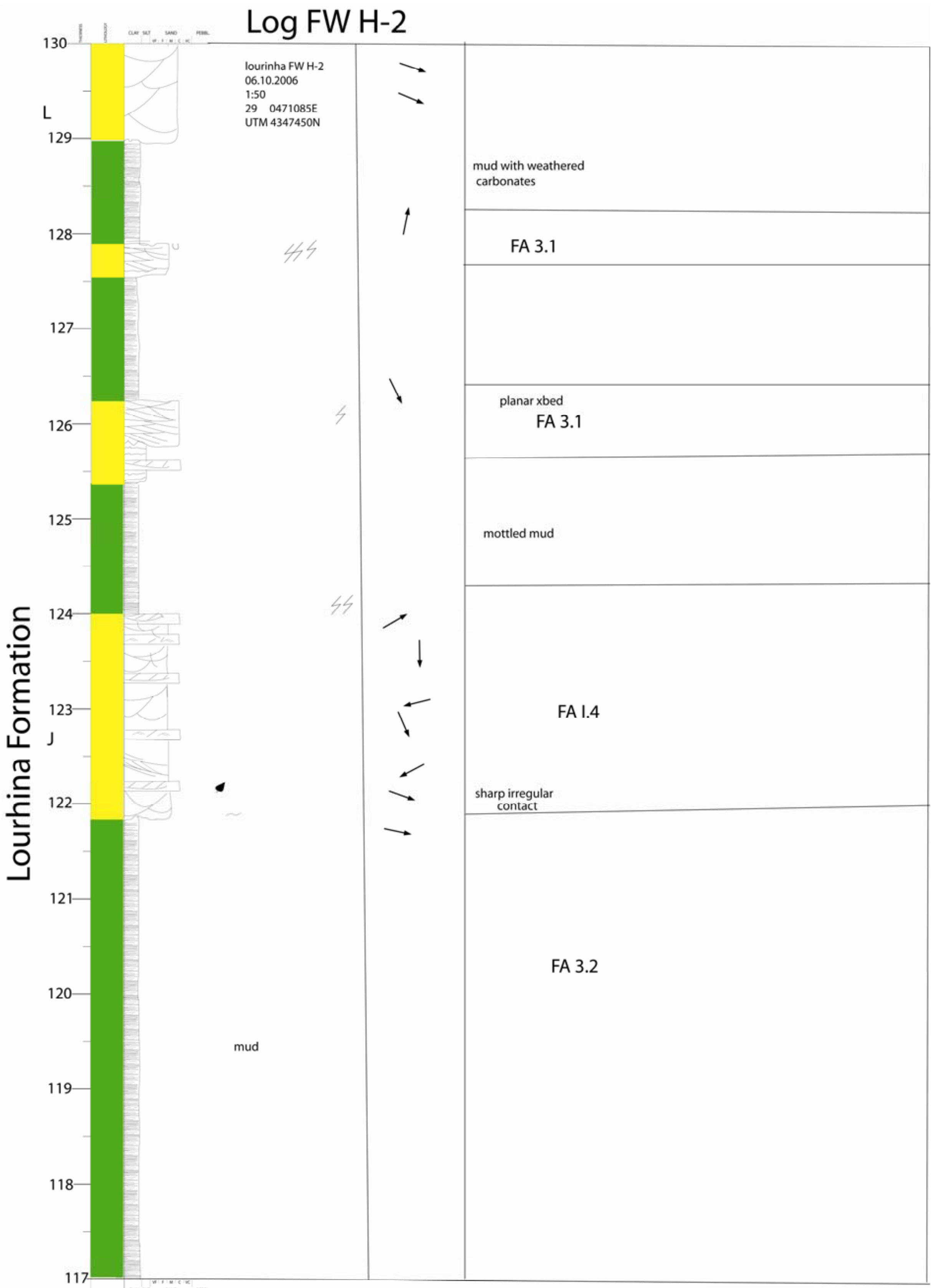

Log FW I-4
lourinha FW I-4
06.10.2006
1:50
29 0471039E
UTM 4347861N
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0
numerous shell fragments
FA 4.2
sampled
FA 2.1
Discontinuous heterolithic contact
planar/tabular
Disturbed erosional contact
FA I.2
coal fragments scattered within
sampled
rip up clasts at base
calcrete dev.
paleosol

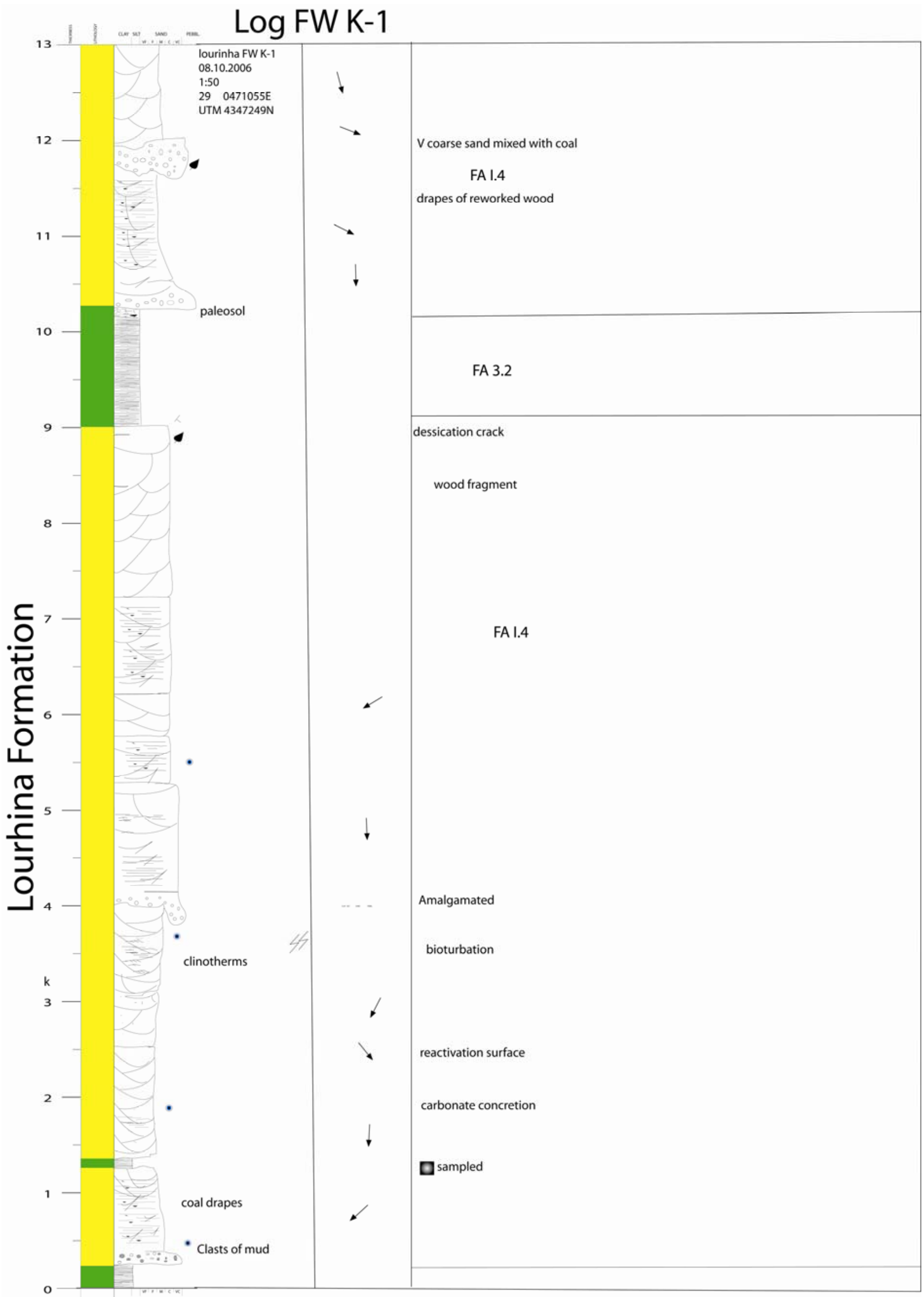
# Log FW I-ned

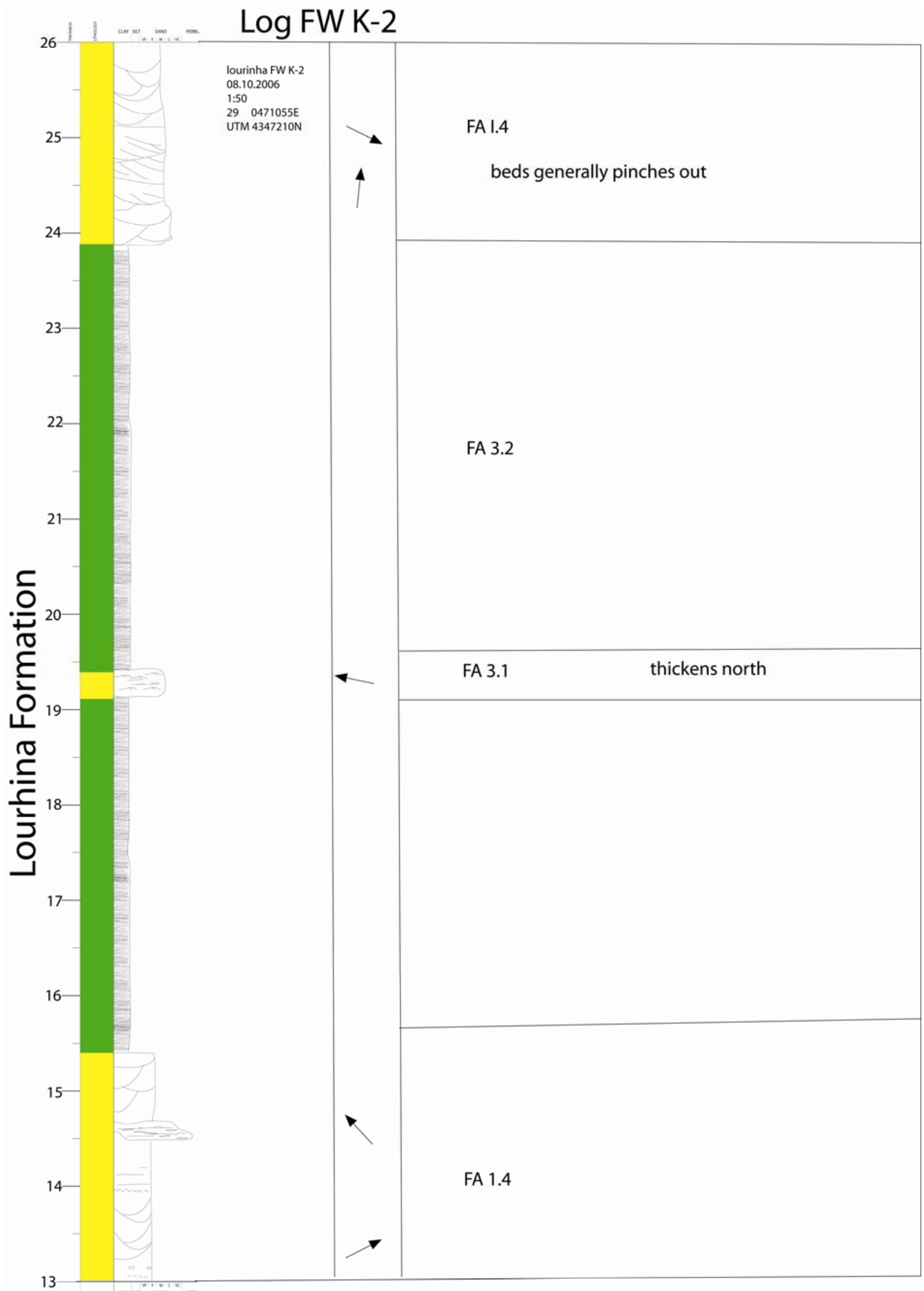

THICKNESS
LITHOLOGY
CLAY SILT SAND PEBBL.
VF F M C VC
lourinha FW -I ned
11.10.2006
1:50
29 04710 ?
UTM 4346 ?
13
12
11
10
9
8
7
6
5
4
3
2
1
0
Lourhina Formation
shell bank
few carbonate concretions
grey
appears amalgamated
some soft sediment deformed structures

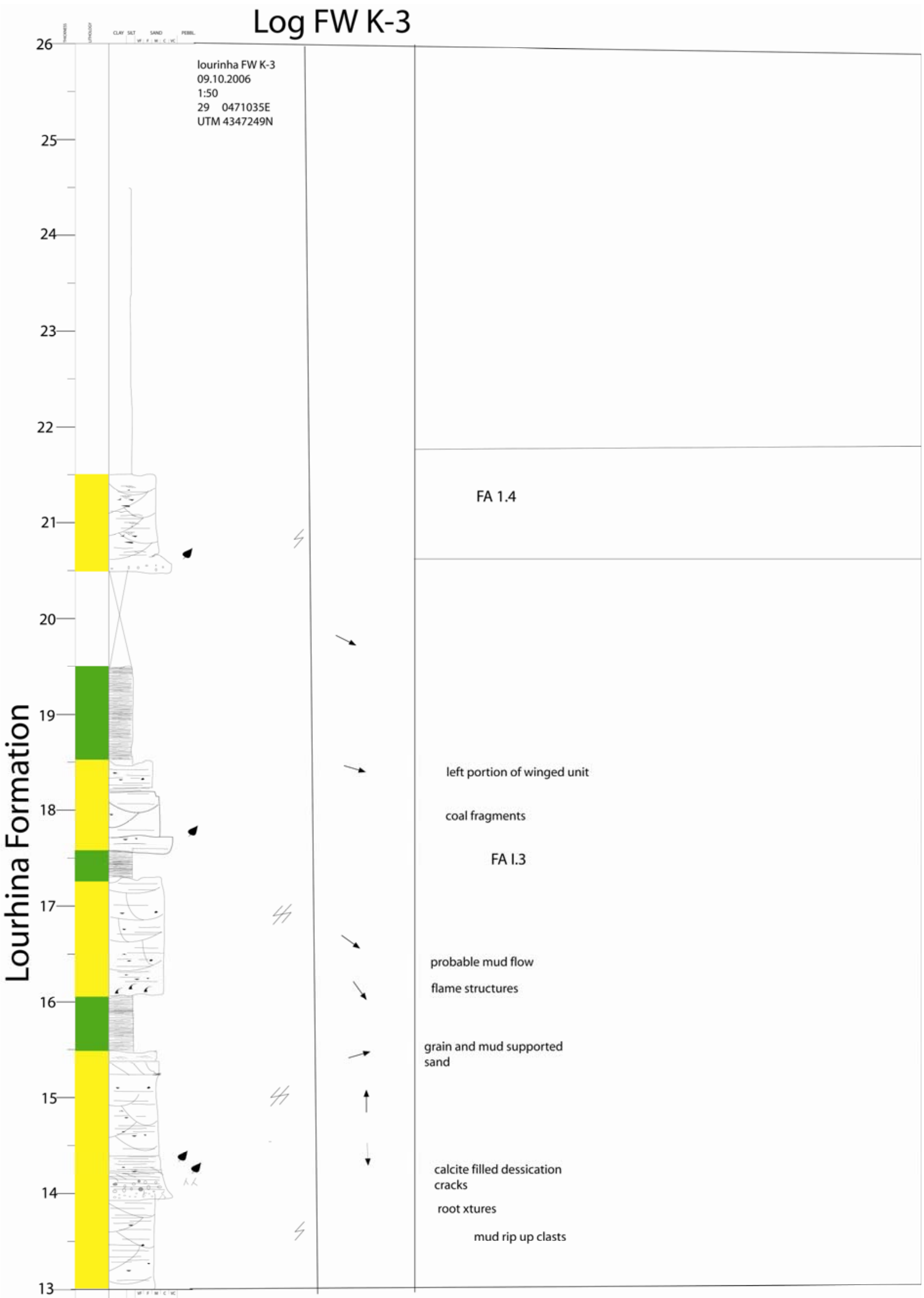

Log FW I-south
lourinha FW -I
11.10.2006
1:50
29 0471?
UTM 434N ?
FA. 1.4
FA. 3.2
FA. 3.1
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0

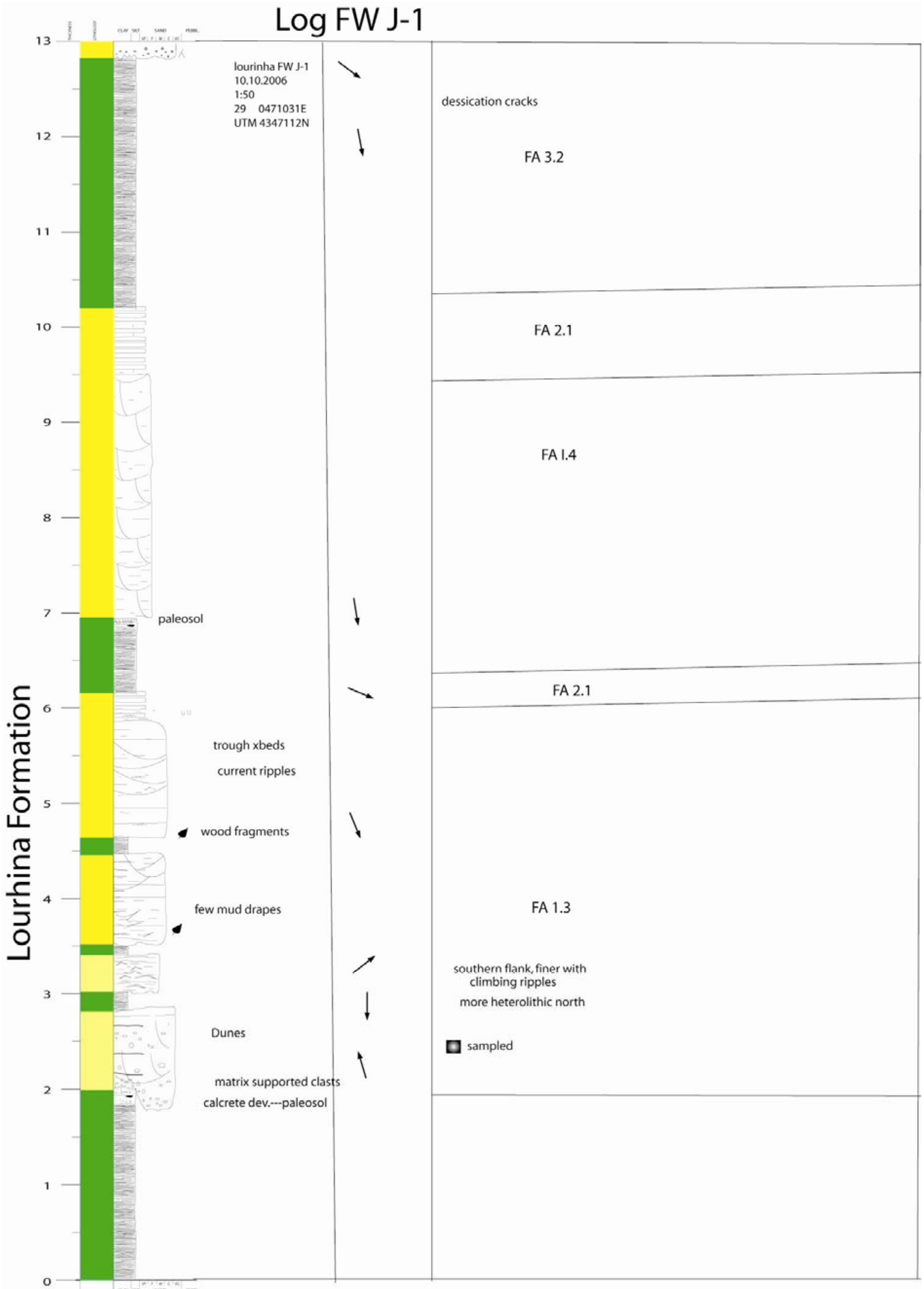

Log FW H-1
Lourhina Formation
H
CLAY SILT SAND PEBBL.
117
116
115
114
113
112
111
110
109
108
107
106
105
104
lourinha FW H -1
04.10.2006
1:50
29 0471083E
UTM 4347436N
coal/pyrite flakes
Tabular xbed
Bioturbation moderate (30-60%)
FA 1.4
sampled
FA 4.1
inclined heterolithic stratification
Carbonate concretions
locally derive conglo./mud clasts
FA 4.2
Sandstone lenses projects from Mudstone
II lamination

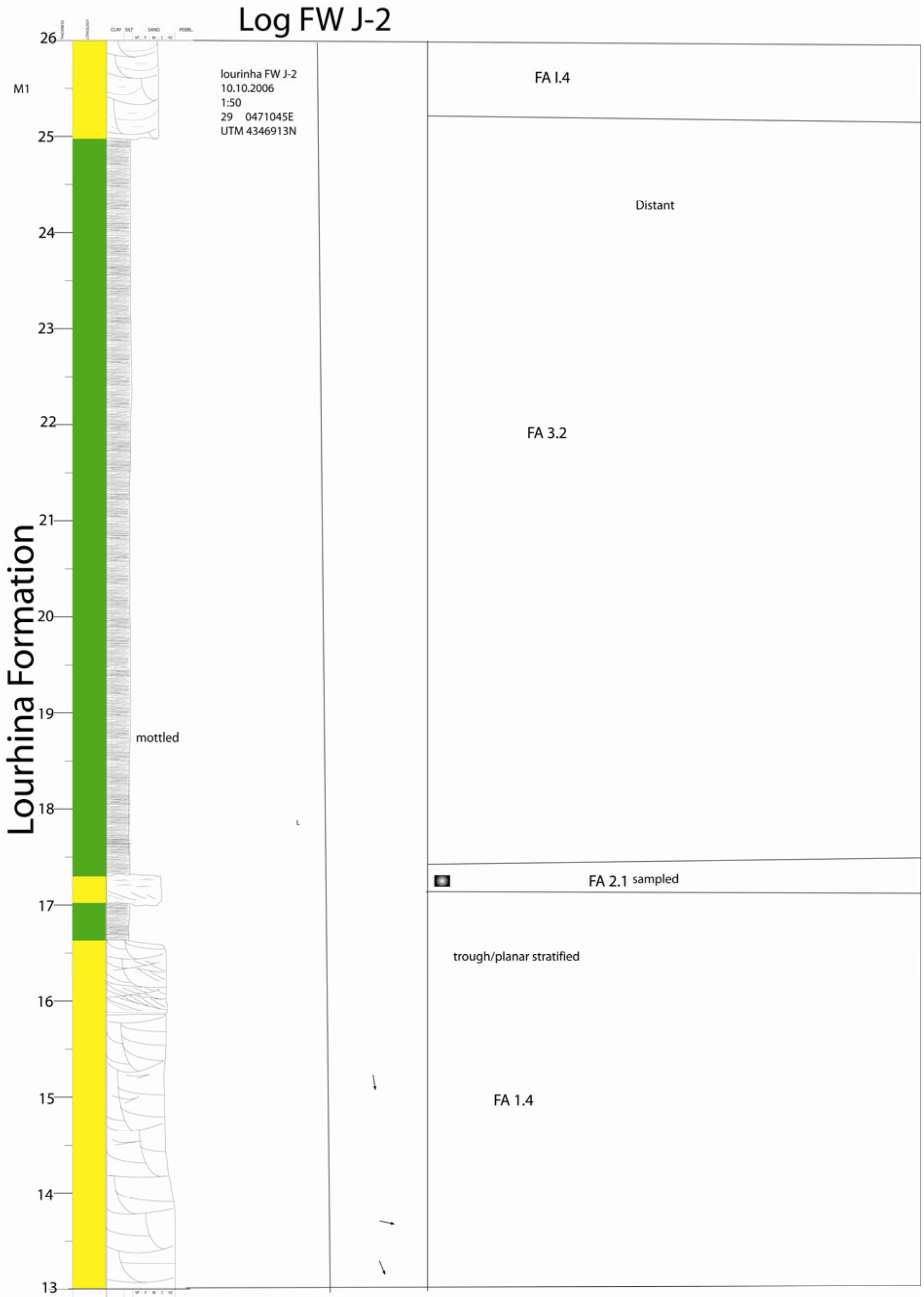

Log FW H-2
lourinha FW H-2
06.10.2006
1:50
29 0471085E
UTM 4347450N
Lourhina Formation
L
J
130
129
128
127
126
125
124
123
122
121
120
119
118
117
CLAY SILT SAND PEBBL.
mud with weathered carbonates
FA 3.1
planar xbed
FA 3.1
mottled mud
FA I.4
sharp irregular contact
FA 3.2
mud

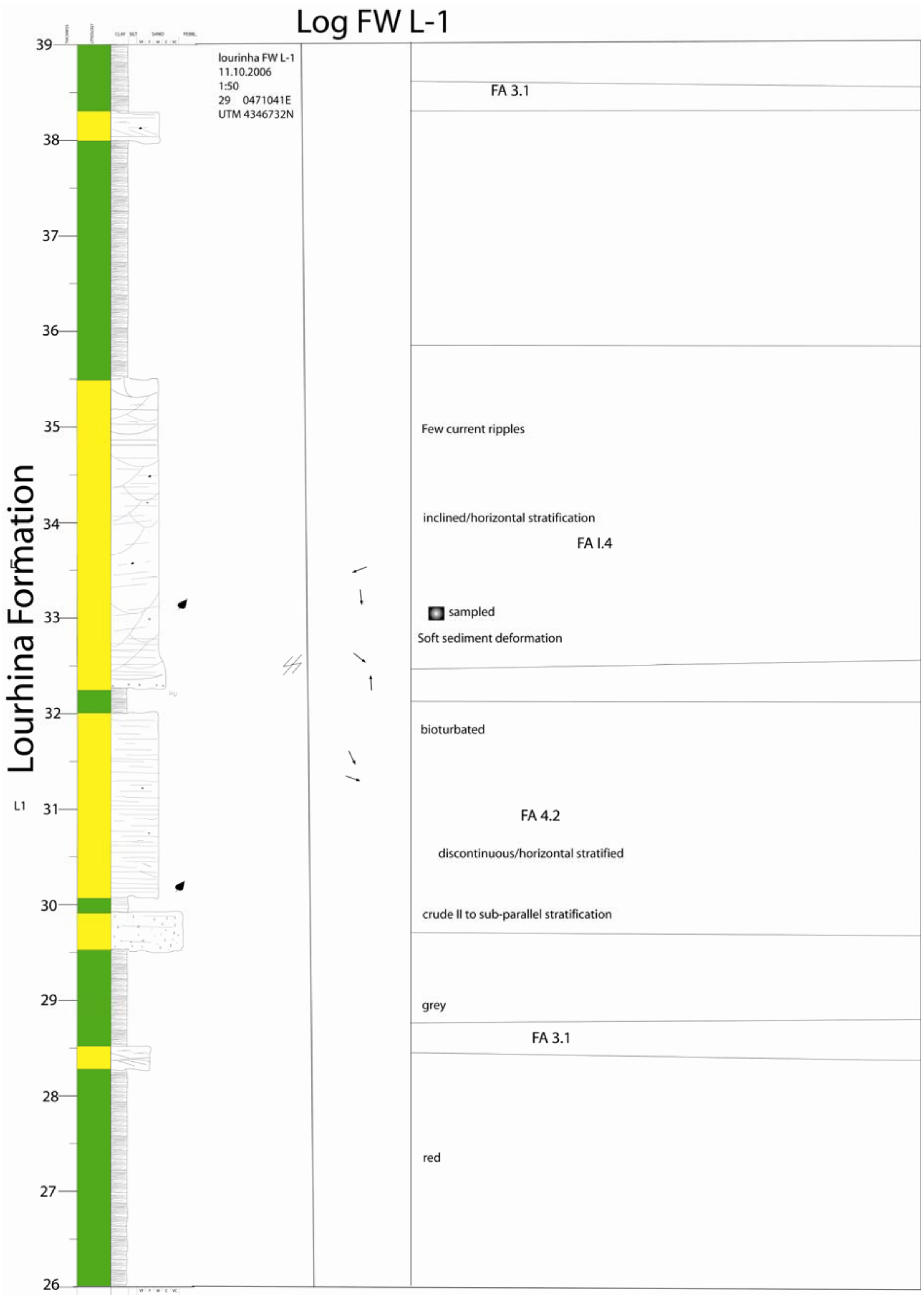

Log FW K-1
lourinha FW K-1
08.10.2006
1:50
29 0471055E
UTM 4347249N
Lourhina Formation
V coarse sand mixed with coal
FA I.4
drapes of reworked wood
paleosol
FA 3.2
dessication crack
wood fragment
FA I.4
Amalgamated
clinotherms
bioturbation
reactivation surface
carbonate concretion
sampled
coal drapes
Clasts of mud
13
12
11
10
9
8
7
6
5
4
3
2
1
0

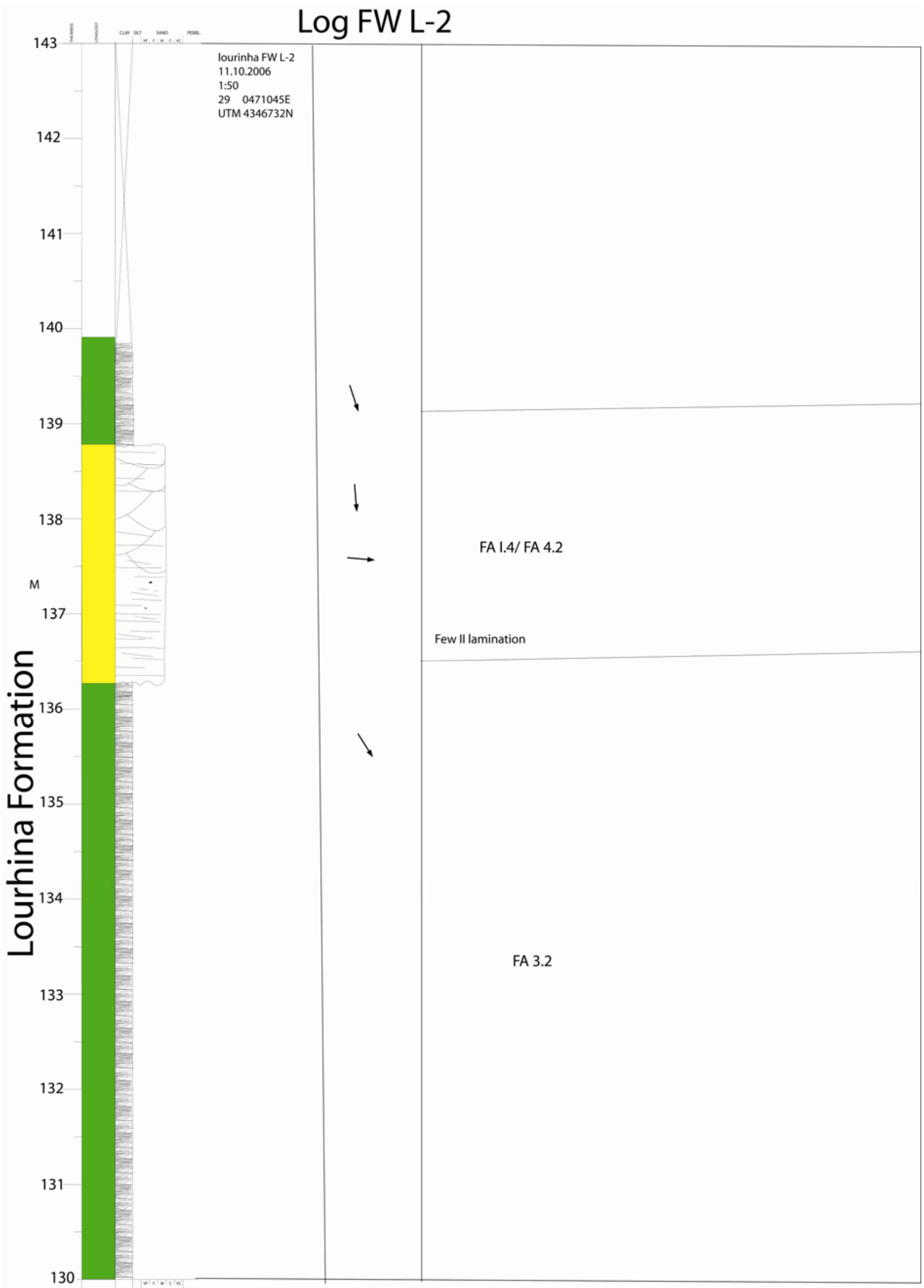

Log FW K-2
lourinha FW K-2
08.10.2006
1:50
29 0471055E
UTM 4347210N
FA I.4
beds generally pinches out
FA 3.2
FA 3.1
thickens north
FA 1.4
Lourhina Formation
26
25
24
23
22
21
20
19
18
17
16
15
14
13

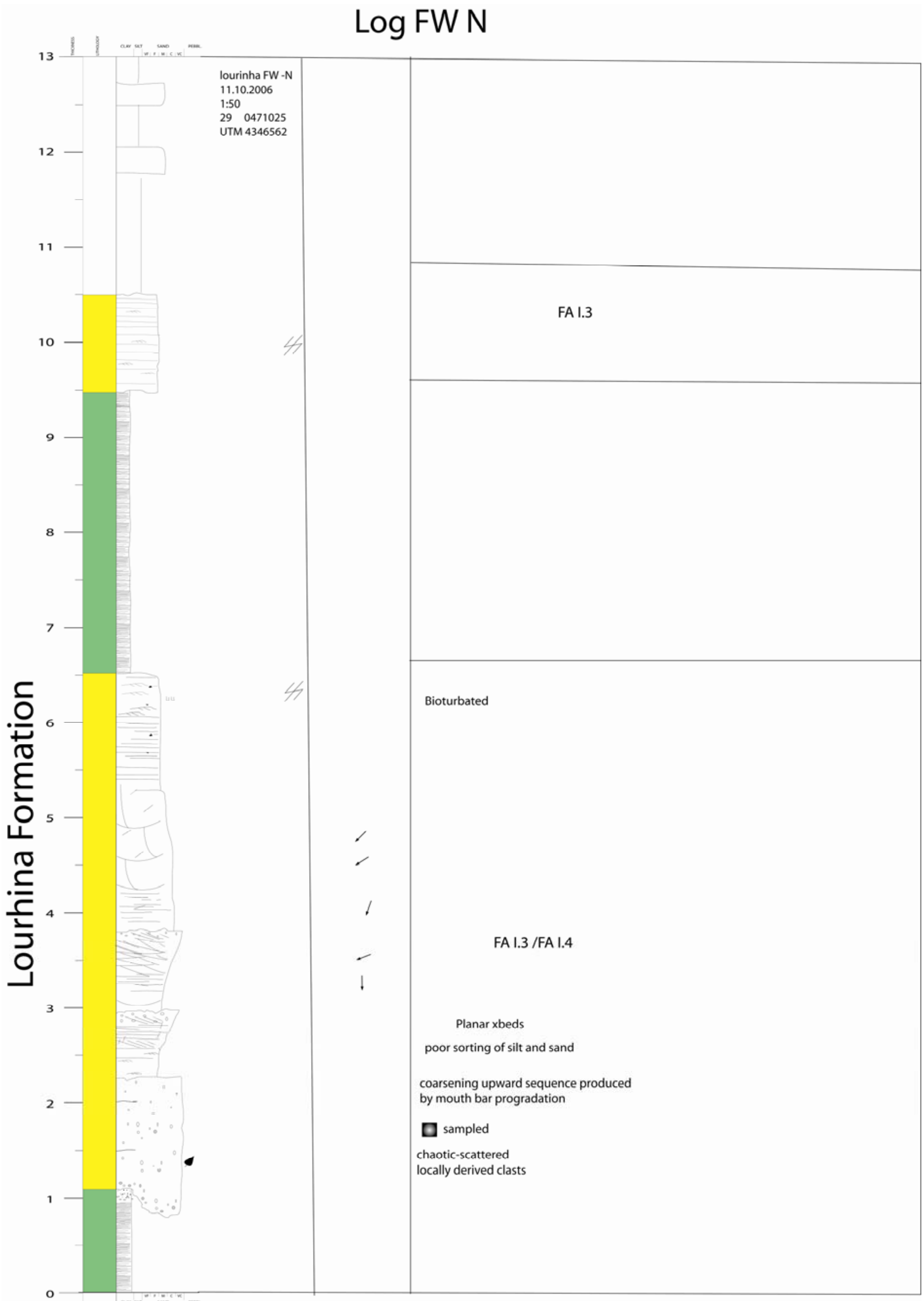

Log FW K-3
lourinha FW K-3
09.10.2006
1:50
29 0471035E
UTM 4347249N
Lourhina Formation
26
25
24
23
22
21
20
19
18
17
16
15
14
13
FA 1.4
left portion of winged unit
coal fragments
FA I.3
probable mud flow
flame structures
grain and mud supported
sand
calcite filled dessication
cracks
root xtures
mud rip up clasts

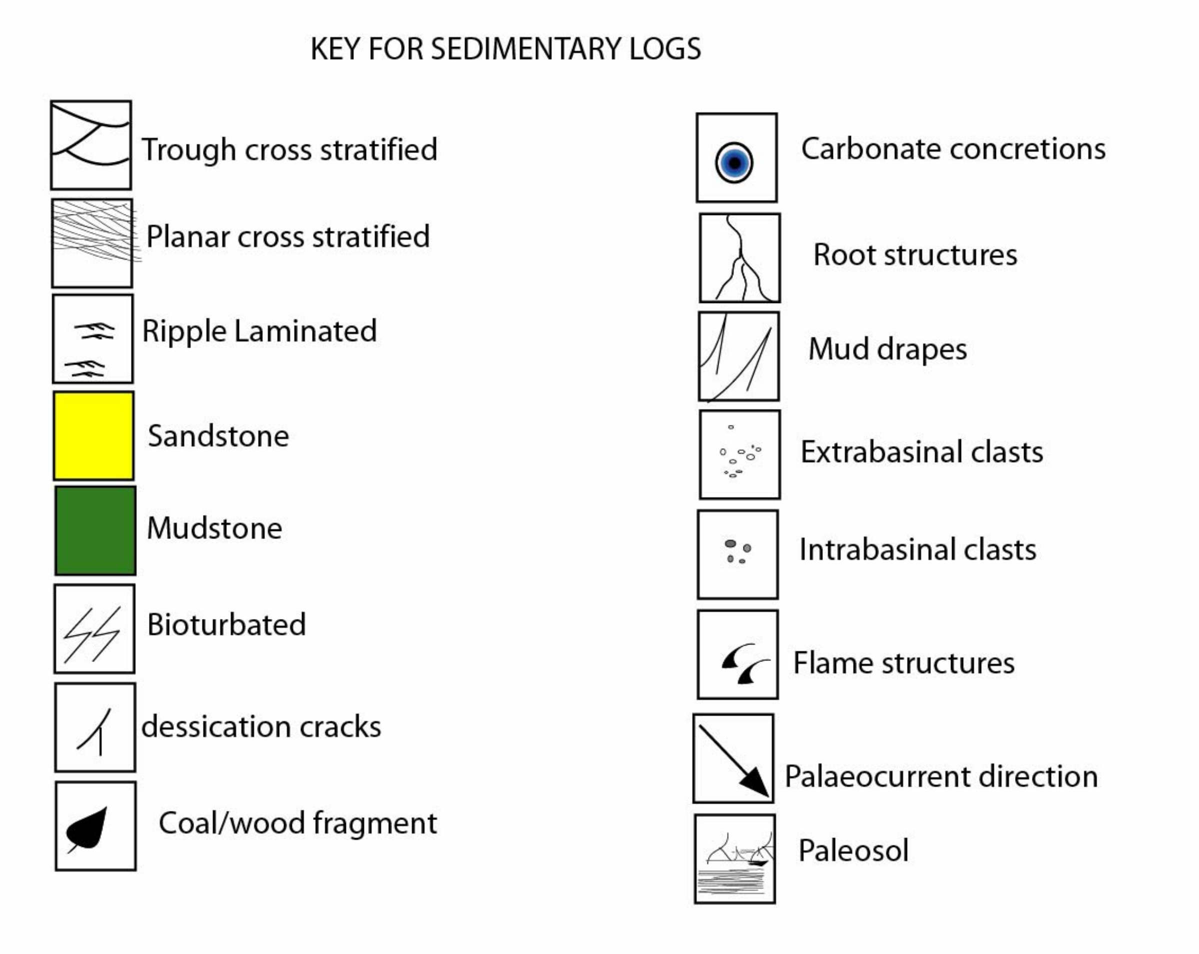

Log FW J-1
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0
lourinha FW J-1
10.10.2006
1:50
29 0471031E
UTM 4347112N
dessication cracks
FA 3.2
FA 2.1
FA I.4
paleosol
FA 2.1
trough xbeds
current ripples
wood fragments
few mud drapes
FA 1.3
southern flank, finer with
climbing ripples
more heterolithic north
Dunes
sampled
matrix supported clasts
calcrete dev.---paleosol

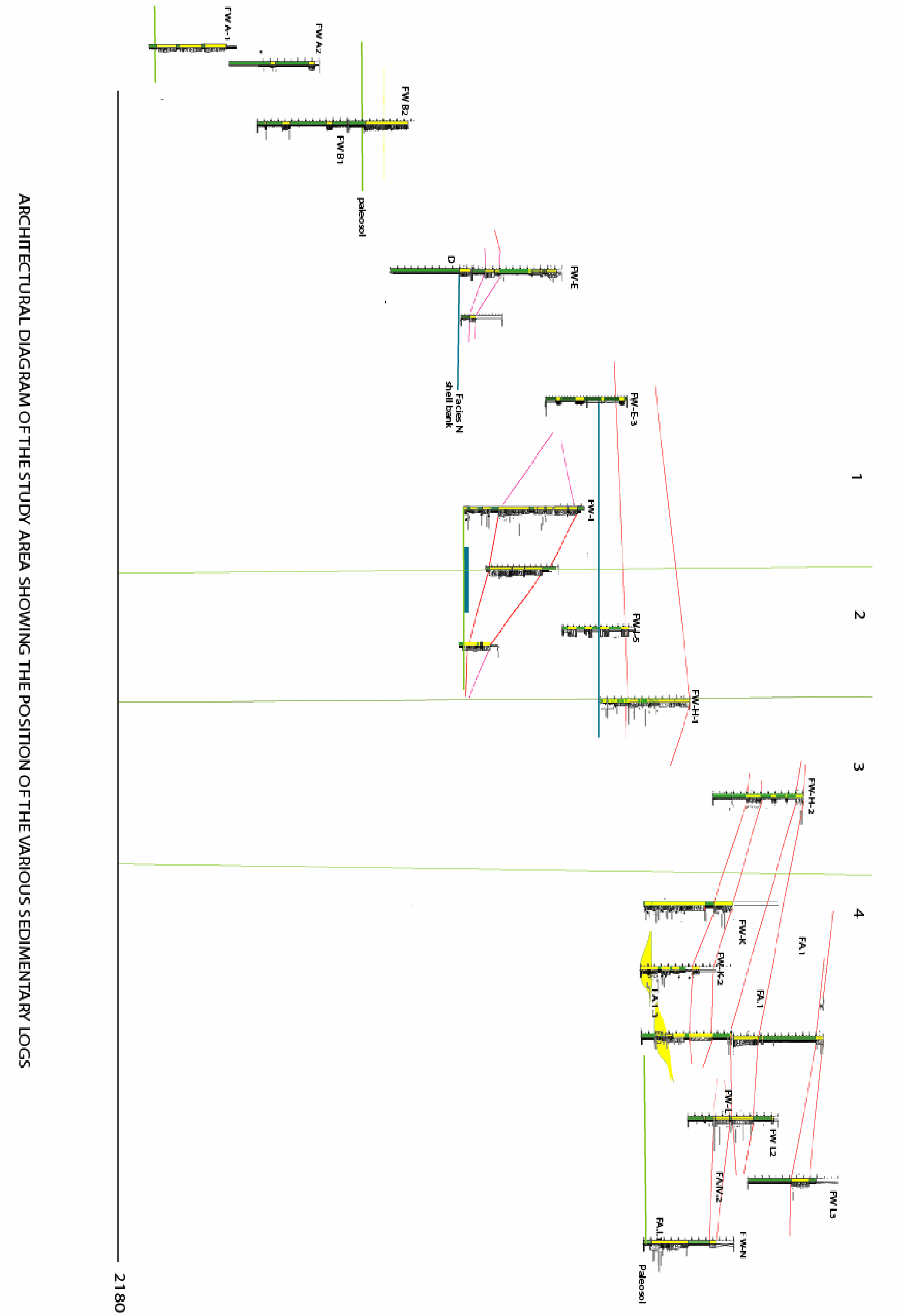

Log FW J-2
lourinha FW J-2
10.10.2006
1:50
29 0471045E
UTM 4346913N
Lourhina Formation
M1
CLAY SILT SAND PEBBL.
FA I.4
Distant
FA 3.2
mottled
FA 2.1 sampled
trough/planar stratified
FA 1.4
26
25
24
23
22
21
20
19
18
17
16
15
14
13


Log FW L-1
lourinha FW L-1
11.10.2006
1:50
29 0471041E
UTM 4346732N
Lourhina Formation
L1
39
38
37
36
35
34
33
32
31
30
29
28
27
26
CLAY SILT SAND PEBBL.
FA 3.1
Few current ripples
inclined/horizontal stratification
FA I.4
sampled
Soft sediment deformation
bioturbated
FA 4.2
discontinuous/horizontal stratified
crude II to sub-parallel stratification
grey
FA 3.1
red


Log FW L-2
lourinha FW L-2
11.10.2006
1:50
29 0471045E
UTM 4346732N
CLAY SILT SAND PEBBL.
VF F M C VC
143
142
141
140
139
138
M
137
136
135
134
133
132
131
130
Lourhina Formation
FA I.4/ FA 4.2
Few ll lamination
FA 3.2


Log FW N
lourinha FW -N
11.10.2006
1:50
29 0471025
UTM 4346562
Lourhina Formation
13
12
11
10
9
8
7
6
5
4
3
2
1
0
FA I.3
Bioturbated
FA I.3 /FA I.4
Planar xbeds
poor sorting of silt and sand
coarsening upward sequence produced
by mouth bar progradation
sampled
chaotic-scattered
locally derived clasts


KEY FOR SEDIMENTARY LOGS
Trough cross stratified
Planar cross stratified
Ripple Laminated
Sandstone
Mudstone
Bioturbated
dessication cracks
Coal/wood fragment
Carbonate concretions
Root structures
Mud drapes
Extrabasinal clasts
Intrabasinal clasts
Flame structures
Palaeocurrent direction
Paleosol


ARCHITECTURAL DIAGRAM OF THE STUDY AREA SHOWING THE POSITION OF THE VARIOUS SEDIMENTARY LOGS
FW A-1
FW A2
FW B2
FW B1
paleosol
FW-E
D
Facies N
shell bank
FW-E3
FW-I
FW-J-5
FW-H-1
FW-H-2
FW-K
FW-K-2
FA.1
FA.I.3
FW-L
FW L2
FA.IV.2
FW L3
FA.I.1
FW-N
Paleosol
1
2
3
4
2180