ASPECTS OF PALAEOZOIC CRATONIC SEDIMENTATION IN SOUTHWEST LIBYA AND SAUDI ARABIA

by

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ABSTRACT

The facies of four Palaeozoic formations of the ancestral stable cratonic margin of Gondwanaland have been described and are interpreted as fluvial, estuarine, deltaic, shallow-marine and glacial deposits. In Southwest Libya the sediments of the *Tadrart Formation* (Lower Devonian) may be divided into four facies which are present in fining-upwards transgressive sequences of sandstones which, together, form a widespread blanket sandstone unit, the External Tassilian Sandstones. Thickly-bedded successions of multi-storey cross-stratified and channelled sandstones (*facies MSd and CSdSt)* are characteristic of the lower beds of the transgressive sequences of these sheet sandstones. The term "Nubian-type facies" has been applied to these sediments which have been interpreted, in this thesis, as the deposits of braided-fluvial and estuarine environments. Thinly-bedded successions of rippled, bioturbated and vertically-burrowed sandstones and siltstones *(facies RBSdSt and VBSd)* are present in the upper beds of the transgressive sequences. These are interpreted as littoral and shallow-marine deposits, and some evidence exists to suggest that they may have been laid down on tidal flats and in sub-tidal environments. These four facies are characterised by differences in their trace fossil assemblages, facies RBSdSt having the richest ichnofauna, and facies MSd the poorest. The *Arthrophycus* assemblage of facies CSdSt and MSd is thought to be evidence for fresh-water/brackish environments. The uppermost transgressive sequence of the External Tassilian Sandstone is capped by shales and subordinate sandstones of the *Ouan Kasa Formation* (Middle Devonian) which bear witness to open-marine deposition. The transgressive sequences are thought to be retrogradational deposits of a fluvially-dominated large-scale delta.

In Northwest Saudi Arabia the facies of the *Saq and Tabuk formations* (Cambrian to Lower Devonian) bear witness to a sedimentary history for this area that is very similar to that of North Africa. The Cambro-Ordovician succession comprises fluvial and estuarine deposits *(facies 1 and 2)* which pass up to deltaic, ?tidal flat and some extremely interesting late Ordovician fluvio-glacial deposits that lie in deeply incised palaeovalleys *(facies 3, 4 and 5)*. The Silurian succession comprises shallow-marine deposits *(facies 3)* which are overlain by Lower Devonian fluvial sediments *(facies 1)*. The relative importance of the eustatic and epeirogenic controls on

this Palaeozoic succession may determine the style of stratigraphic succession preserved, i.e. whether dominated by regressive or retrogradational deltaic (transgressive) sequences. The latter, which are characteristic of the Cambrian and Lower Devonian successions of both countries, are thought to have been deposited in stable cratonic conditions of very limited epeirogenesis.

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QUOTATIONS

"La paléogéographie de séries à 90% détritiques gréseuses, ne pouvait progresser qu'en recherchant les lois régissant la répartition géographique des grès, c'est-à-dire les mécanismes de transport et de dépôt des sables correspondants. Cela impliquait de connaître la forme, le volume, l'orientation, l'arrangement et la distribution des corps sableux qui constituent les réservoirs."

Y. Gubler (1966)
('Essai de Nomenclature et Caractérisation des Principales Structures Sédimentaires'; on the Nubian-type sandstones of the Tassili-n-Ajjer.)

"To see a World in a Grain of Sand,
And a Heaven in a Wild Flower,
Hold Infinity in the palm of your hand,
And Eternity in an hour."

William Blake
('Auguries of Innocence')

"That man, I think, has had a liberal education who has been so trained in youth that his body is the ready servant of his will, and does with ease and pleasure all the work that, as a mechanism, it is capable of; whose intellect is a clear, cold, logic engine, with all its parts of equal strength, and in smooth working order; ready like a steam engine, to be turned to any kind of work, and spin the gossamers as well as forge the anchors of the mind; whose mind is stored with knowledge of the great and fundamental truths of Nature, and of the laws of her operations; and who, no stunted ascetic, is full of life and fire, but whose passions are trained to come to heel by a vigorous will, the servant of a tender conscience; who has learned to love all beauty, whether of Nature or of art, to hate all vileness, and to respect others as himself."

T. H. Huxley
('A Liberal Education and Where to Find It')

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PREFACE

Scientific exploration, documentation and cartography were undertaken in Libya, prior to independence in 1951, predominantly by American cartographers and by French and Italian geologists and cartographers. Since then Western European and American oil company geologists have published in at least four different languages, and the geological mapping of the country has been undertaken by both the United States Geological Survey and, more recently, by the geologists of the Industrial Research Centre, Tripoli, which has been staffed to a large extent by Czech geologists through a co-operative agreement. The literature is, consequently, full of different spellings of place names and formation names owing to the differing transliterations of Arabic names adopted by geologists and cartographers of various nationalities. The publication in 1960 of the Stratigraphic Lexicon of Libya by the Petroleum Exploration Society of Libya provided a basis for subsequent refinement and standardisation of stratigraphic nomenclature. Recognising the need for a major revision of this work a National Committee for the Stratigraphic Nomenclature of Libya was formed in 1975 under the auspices of the Industrial Research Centre, Tripoli. Two years later the Industrial Research Centre published the second edition of the Geological Map of Libya (Conant & Goudarzi, 1964; Maghrabi & Cheshitev, 1977) on which the spelling of place names was adjusted to be in line with the recommendations of the Survey Department, Secretariat of Planning. The spelling of formation names was also adjusted and was "based on the correct English translation of the original Arabic names of the localities after which the formations were named". The adjustment of formation names in this manner is discouraged in guides to stratigraphic procedure (e.g. Holland *et al.*, 1978) and in its revised Stratigraphic Lexicon of Libya (Banerjee, 1980) the Industrial Research Centre has reverted to the old formation name spellings, thereby showing a greater regard for the Rule of Priority in stratigraphic procedure. Their maps, however, retain the new revised formation names and there appears to be considerable confusion over what usage should be adopted (Salem & Busrewil, 1980, preface). The publication of the three-volume work "The Geology of Libya" *(ibid.)* has greatly advanced the standardisation of geographic place names but the editors have been obliged to accept the present confusion over stratigraphic nomenclature. In

this thesis the geographic names adopted are those recommended by the Survey Department of the Secretariat of Planning of the Socialist People's Libyan Arab Jamahiriya, as shown on the geological map of Libya (Maghrabi & Cheshitev, 1977) and in "The Geology of Libya" *(op. cit.)*. In cases where place names do not appear in either of these sources but in earlier maps and publications, the Italian spelling has been adopted where possible. This is because the most detailed geographical map of the Jabals Akakus and Tadrart is by an Italian (Mori, 1965), and because the Italian transliteration of Arabic names is more in sympathy with that adopted by the Survey Department than that of the French, the other nationality to produce detailed maps of the area. The stratigraphic nomenclature adopted in this thesis is that of the revised Stratigraphic Lexicon of Libya. The geographical place names and stratigraphic nomenclature adopted for the work undertaken in Saudi Arabia follows that of the United States Geological Survey Overseas Mapping Division (Powers *et al.*, 1966) in which the English transliteration of Arabic names is broadly the same as that used in Libya.

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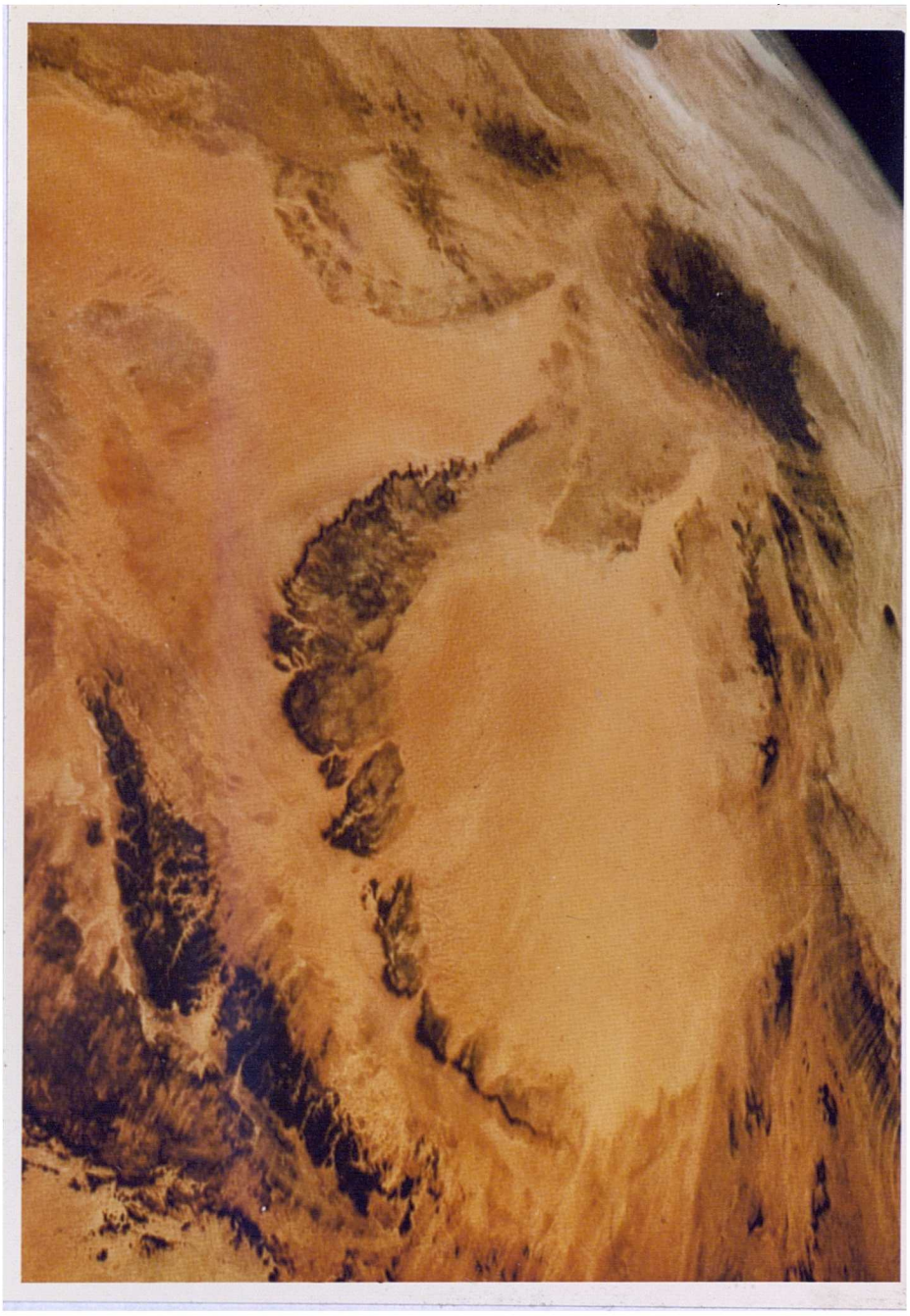
CHAPTER 1

INTRODUCTION

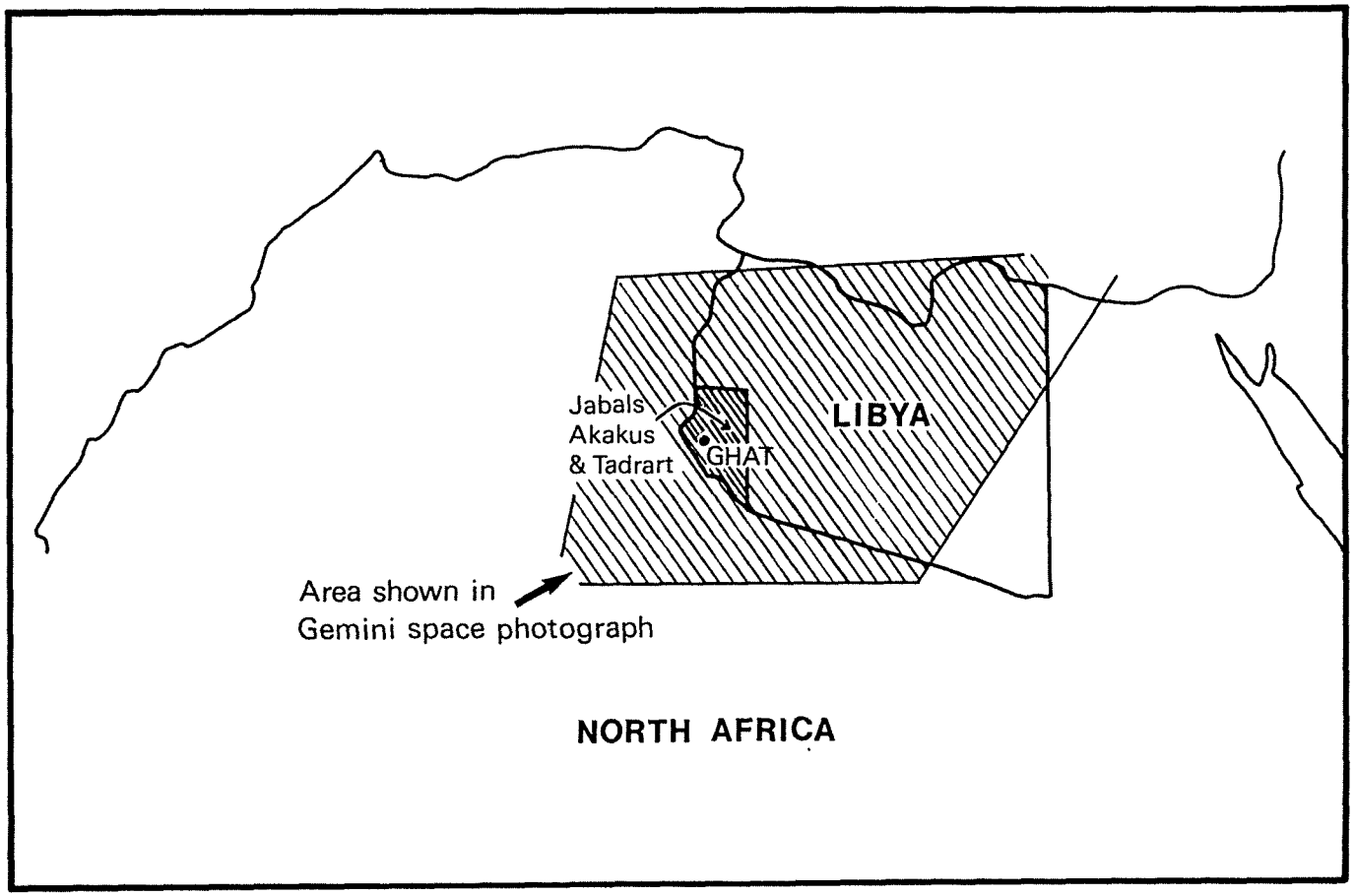
1.1 General

The opportunity to study the Cambro-Ordovician and Lower Devonian blanket sandstones of Southwest Libya and Saudi Arabia between 1977 and 1979 provided the material from which this thesis has been written. The field work undertaken in Saudi Arabia also included a study of the Ordovician and Silurian sediments from which some tentative evidence for the occurrence of Upper Ordovician glacial deposits, similar to those of the Central Sahara, was derived. The sediments studied belong to Palaeozoic successions which were laid down on the margins of a continental mass which was probably not very different from the Gondwanaland of the early Mesozoic (Du Toit, 1937; Shackleton, 1981). This continent extended beyond North Africa and Arabia and is thought to have included parts of the continental crust now forming France, Italy, Spain, Turkey and Iran (Fig. 1.1) (Scotese *et al.*, 1979; Smith *et al.*, 1973). Throughout Cambrian to Devonian times, this continental mass was surrounded by ocean crust which was subsequently subducted, and orogenic activity was restricted to the margin comprising parts of western South America, Antarctica and Australia (Du Toit's Samfrau Geosyncline (1937), see Figs. 1.1 & 1.2). The margin comprising North Africa and Arabia, therefore, may be described as a trailing-edge, passive margin. Continental drift of Gondwanaland during this time resulted in a southward movement of North Africa and Arabia relative to the poles, followed by a northerly movement after the Ordovician (Scotese *et al.*, 1979; McElhinny & Opdyke, 1968) (Figs. 1.1 & 1.2). In the Carboniferous conditions changed considerably. The North African/Arabian margin of Gondwanaland reached low latitudes and suffered continental rupture and collision. During the Early Carboniferous the Atlas Mountain chain first came into existence, during the Marsian orogenic events (Sander, 1968) and Spain and Italy were separated from the rest of Gondwanaland. This was followed in the Late Carboniferous by collision between Gondwanaland and the North American Craton, causing the development of a major orogenic belt in the Mauretanides (Asturian Orogeny, Sander, 1968; Scotese *et al.*, 1979). The sediments of the formations studied were laid down prior



FRONTISPIECE

Gemini XI spacecraft photograph of Libya viewed from the southwest (from Pesce, 1968). Field work was carried out in the Jabals Akakus and Tadrart which are visible in the lower left of the photograph. To the right, the symmetrical Murzuq Basin with its rim of Mesozoic Nubian-type sandstones and central sand sea is visible. In the upper right, Tertiary lava flows of the Jabal As Sawda and the Al Haruj Al Aswad appear as dark areas. To the west (left) of the lavas, the Qarqaf Region is visible, flanked by the Awbari sand sea.



FRONTISPIECE LOCATION MAP

to these Carboniferous events and may be considered, therefore, to provide evidence for sedimentation processes characteristic of middle to high latitudes on the stable cratonic margins of Gondwanaland during the Palaeozoic.

The study of the Palaeozoic blanket sandstones of Southwest Libya was undertaken on the western flank of the Murzuq Basin (Fig. 1.3). In this region Cambrian to Devonian strata crop out as a belt around the northeast of the Hoggar Massif and are exposed in a dissected plateau termed the Tassili N'Ajjer (Kilian, 1931). Since the Murzuq Basin has been cited as a classical example of an intracratonic basin (Selley, 1976), this study has been used to draw conclusions not only about sedimentation processes in cratonic regions, but also about the nature and behaviour of the basins in which the Palaeozoic cratonic sediments are preserved.

The Cambrian to Devonian strata of North Africa and Arabia display a remarkably similar stratigraphy from the successions of the Tindouf and Taoudeni basins in the west to the succession of the Arabian Basin in the east (Fig. 1.3) (Bellini & Massa, 1980; Bender, 1963, 1968, 1975; Beuf *et al.*, 1971; Freulon, 1964; Klitzsch, 1981; Powers *et al.*, 1966; Selley, 1972, 1978, p. 185; Turner, 1980; Whiteman, 1971; Wolfart, 1981). A basal Cambrian to Lower Ordovician blanket sandstone passes upwards into Ordovician sandstones, shales and diamictites. These are overlain, generally abruptly, by Llandoverian graptolitic shales. The Silurian strata are characterised by a gradational vertical transition from shales to sandstones. These are overlain by the second major blanket sandstone in the succession. This unit is of Lower Devonian age, and it passes upwards into shales and sandstones of the Middle and Upper Devonian.

A general indication of the degree of marine influence may be determined by the percentage of shales bearing a marine fauna in the successions. Marine influence can be shown by this approach to increase towards the northwest in North Africa, from the Kufrah Basin to the Mac-Mahon Basin (Fig. 1.3) (Read & Watson, 1975; Freulon, 1964), and towards the north and northeast in Arabia (Wolfart, 1981). This suggests that although Cambrian to Devonian strata are present at outcrop or in the subsurface across most of North Africa and Arabia they may never have been present over much of the Nubian and Arabian shields. In the Palaeozoic, prior to the Tertiary opening

Figure 1.3 OUTCROP AND SUBSURFACE OCCURRENCE OF CAMBRIAN TO DEVONIAN STRATA IN NORTH AFRICA AND ARABIA

Modified after Whiteman (1971)

Additional sources:

International Tectonic Map of Africa (1968),

International Geological Map of Africa (1963), and

Beuf *et al.* (1971).

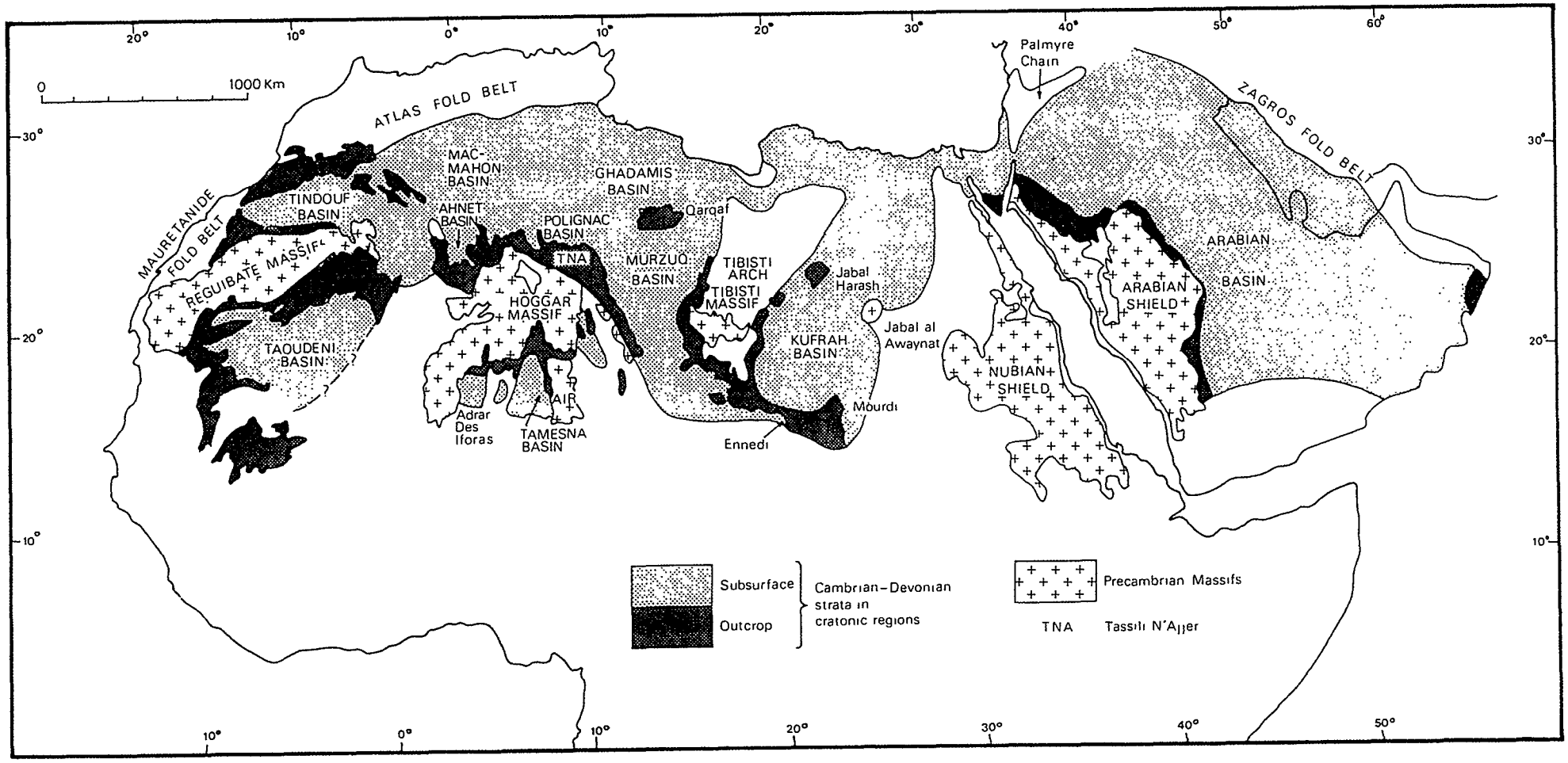


Figure 1.3 OUTCROP AND SUBSURFACE OCCURRENCE OF CAMBRIAN TO DEVONIAN STRATA IN NORTH AFRICA AND ARABIA

of the Red Sea, these two shields were united as one major shield area (Open. Univ., 1980). Palaeocurrent evidence from the eastern Kufrah Basin (Turner, 1980) indicated a source area situated in the southeast whilst the palaeocurrent data obtained in this study in Saudi Arabia suggested a source area in the southwest. Powers *et al.* (1966) and Selley (1972) provided additional evidence in Arabia which supported the location of a source area in the region of the Arabian and Nubian shields.

The palaeocurrent data from the blanket sandstones cropping out around the margins of the Tibisti Massif indicate that this massif also acted as a source area, although intermittently (Klitzsch, 1966a; Oasis, 1971), whereas palaeocurrent evidence exists to indicate that, at least for some of the time, the Hoggar Massif did not have any influence upon current flow direction and channel orientation, and that sediment was transported from southeast to northwest across the massif (Beuf *et al.*, 1971).

The Palaeozoic sediments of North Africa and Arabia (Fig. 1.3) are present in basins divided from each other by arches over which the Palaeozoic succession is severely attenuated. Similar arches are present between the cratonic basins of North America (Clark & Stearn, 1968, p. 126 *et seq.*). The Murzuq Basin is divided from the Ghadamis Basin by the Qarqaf Arch, the southwesterly extension of which marks the boundary with the Polignac Basin. Another arch extends in the subsurface from the Tassili N'Ajjer to the north, dividing the Polignac and Ghadamis basins from the Mac-Mahon Basin (Choubert & Faure-Muret, 1969). In Arabia an arch extends eastwards from the most easterly part of the Arabian Shield, the Central Arabian Arch, dividing the Arabian Basin into two (Powers *et al.*, 1966). Where these structural axes have been mapped in areas of outcropping Precambrian rocks, they have been shown to be complex fault zones activated both vertically and transcurrently during a phase of compressional tectonics which marked the final stages of the Pan-African Orogeny (Bertrand & Caby, 1978; Fabre & Moussine-Pouchkine, 1977). Kennedy (1965) coined the term "Pan-African Orogeny" for the stabilisation of the crust of Gondwanaland in late Precambrian times by the welding together of a number of Archaean cratons and Pan-African mobile belts (Clifford, 1970; Gass, 1977). The main structural axes of North Africa lie within or at the margins

of crust formed of "cratonised" (Greenwood *et al.*, 1976) Pan-African mobile belts, and it is thought that the influence of these crustal structural lineaments was one of the main controls on basin development during the Phanerozoic (Fabre & Moussine-Pouchkine, 1977).

Very little evidence exists, following the termination of the Pan-African Orogeny in late Precambrian to early Cambrian times (Clifford, 1970), for any major tectonic activity on the North African and Arabian craton during Cambrian to Devonian times. The sediments laid down at that time are present in strata which have experienced only minor folding and only local igneous intrusions. Dolerite dykes, probably of late Palaeozoic age and associated with the stabilisation of the Mauretanide Mobile Belt (Fig. 1.3), are present in the succession in Morocco and Algeria, but are absent further east. Devonian vulcanism is reported in the Hoggar Massif (Beuf *et al.*, 1971) and granite intrusions, in Cambrian, Lower Ordovician and Carboniferous sediments, are present in Egypt, Sudan and southeast Libya (on the eastern side of the Kufrah Basin at Jabal al Awaynat) (Burollet, 1963). The earlier granites may be associated with the closing stages of the Pan-African Orogeny, but those in Carboniferous rocks, and the evidence of Devonian vulcanism, indicate that some regions of the craton, particularly those along the major fault systems, experienced intermittent tectonic activity after the Pan-African Orogeny.

1.1.1 Nubian-Type Sandstones

The Cambro-Ordovician and Lower Devonian blanket sandstones of North Africa and Arabia have been termed "Nubian Sandstones" (McKee, 1962; Selley, 1968, 1972) on account of their similarity to the Upper Cretaceous blanket sandstones of Nubia on the southern borders of Egypt (Russegger, 1847). The term "Nubian Sandstone" has subsequently been used to refer to a large number of sandstone units of differing ages from North Africa and Arabia (Picard, 1943; McKee, 1962) which led to a debate about the validity of stratigraphic terminology employing the word "Nubian" (Pomeyrol, 1968; Whiteman, 1970). Despite the fact that the type area of Nubia has not been sufficiently accurately defined geographically, nor the sandstone succession sufficiently well studied for the establishment of a stratigraphic type section (Pomeyrol, 1969), the characteristics

and interpretation of the Cretaceous sandstones of Nubia are relevant as an analogy for the Palaeozoic Nubian Sandstones. The Upper Cretaceous blanket sandstone ranges in thickness from 70 m to 162 m and may be divided into three parts (Attia, 1955). The lower and upper units are composed predominantly of sandstone, whereas the middle unit is an iron-rich oolitic claystone containing a mollusca fauna with both fresh-water and marine types present. This unit is thought to be estuarine on this evidence (Newton, 1909; in McKee, 1962). The presence of land plants in the overlying cross-bedded sandstone unit suggests near-shore or terrestrial conditions. Planar tabular cross-stratification is abundant in this unit, characteristically in 30-100 cm sets. Recumbent-folded deformed cross-stratification is abundant and the cross-strata indicate palaeocurrents directed towards the north. McKee (1962) considered these sediments to have been laid down in back-barrier or other protected areas of mildly agitated waters where fluvial flood currents introduced the sediments that made up the tabular sets of cross strata. Ward and McDonald (1979) studied the same succession and proposed an interpretation involving predominantly alluvial plain sedimentation with subordinate marginal marine and coastal plain facies present as interbedded units.

The term "Nubian Sandstone" was adopted by Said (1962) for the Triassic, Jurassic and Lower Cretaceous successions of southern Egypt. These have many similarities with the rocks of the Upper Cretaceous succession described by Russegger (1847) and Attia (1955). The term "Nubian" has, since the 1960's, been most commonly used to describe the Lower Cretaceous arenaceous successions of North Africa, for example in Egypt ("Nubia Group" in Klitzsch *et al.*, 1979) and in Libya ("Nubian Sandstone" in Barr and Weegar, 1972; and Maghrabi & Cheshitev, 1977). The Lower Cretaceous succession of southwest Egypt has been studied recently (Klitzsch *et al.*, 1979) and it was proposed that the sequence of deposits was laid down on a northward-sloping alluvial plain in a warm and humid, or semi-humid, climate. The streams crossing this plain are thought to have been straight and slightly sinuous and to have contained sandwaves as the predominant bedform. The flood plain deposits contain plant remains, from which the climatic deductions were made, and rare lacustrine and marginal-marine facies are also present. The fauna of this succession was

studied by Barthel and Boettcher (1978) who proposed terrestrial and estuarine environments of deposition.

In a study of a large number of Nubian-type sandstones, mainly of Mesozoic age, McKee (1962) used the term "Nubian Sandstone" in a broad facies sense to refer to mineralogically mature and characteristically cross-bedded sandstones of North Africa which have widespread blanket geometries. In his study of the facies of these sandstones, McKee (1962) identified deposits of alluvial fan, fluvial, lacustrine, estuarine and lagoonal environments and pointed out that the characteristics of these facies have led geologists to propose a range of interpretations from aeolian to fluvial and open marine. The use of the term "Nubian" in a facies sense has gained many supporters. Such a usage was proposed in Libya by Desio (1936b) and has been maintained in many subsequent papers (Whiteman, 1970; Bonnefous, 1972; Van Houten, 1980). Selley (1968) argued for its use as a very general facies term and emphasised the importance of regional geology in its definition. He described the Nubian Sandstones as those which rim the Arabian and Nubian shields, regardless of age or facies, but which result from a remarkably similar geological history, common to the whole region. Van Houten (1980) used the term "Nubian Facies" to refer to a recurring non-marine to paralic facies, composed primarily of quartzose, cross-bedded sandstones, which accumulated on stable platforms and in intracratonic basins in many different places at different times. It is clear that the characteristically cross-bedded, quartz arenitic, blanket sandstone formations of North Africa and Arabia are sufficiently distinctive to deserve an all-embracing name. Very significant differences exist, however, between the formations, and within the formations a number of different facies may be defined which represent the deposits of a variety of non-marine and marine environments. To attach anything but the most general environmental label to the term "Nubian Facies" would be most dangerous, therefore. Selley (1968) pointed out what he considered to be one of the key problems in the interpretation of these facies. Planar tabular cross-stratified beds of considerable lateral dimensions, composed of quartz arenitic sandstones, and suggestive, therefore, of open marine environments and marine winnowing, are often associated with a continental fauna and flora of wood and terrestrial vertebrate remains. He suggested that these characteristics

imply the past existence of depositional settings not common at the present time.

The Palaeozoic blanket sandstones of North Africa and Arabia are composed of similar facies and are alike in that they lack body fossils and consist of cross-stratified sandstones and siltstones which contain channel and scour structures and abundant trace fossils. Plant remains are present in some of the Lower Devonian blanket sandstones and these provide some evidence for the environment of deposition. The Cambro-Ordovician sandstones, however, lack even these environmental indicators and are, consequently, very difficult to interpret. Some distinctive problems are involved in the interpretation of Cambro-Ordovician, cratonic sediments throughout the world. These problems arise in the interpretation of the sedimentary successions of North Africa, the sediments of which were deposited in middle to high palaeolatitudes (Bennacef *et al.*, 1971; Beuf *et al.*, 1971; Burollet & Byramjee, 1969), and in the interpretation of the sedimentary successions of North America, the ancient North Atlantic Continent, and Scandinavia, sediments deposited, by contrast, in low palaeolatitudes (Scotese *et al.*, 1979; Dott & Byers, 1980; Swett & Smit, 1972; Seilacher, 1967, p. 415). The sediments of these successions are notable for the compositional and textural maturity of the sandstones, and for a dearth of shale (Dott & Byers, 1980; Beuf *et al.*, 1971). The absence, or paucity, of fauna in the sandstones results in a reliance on petrography, sedimentary structures and biogenic traces for the purposes of environmental interpretation. In adopting a uniformitarian approach and using facies models based on modern depositional environments to aid interpretation of these successions, some caution is appropriate. It seems likely that the differences between Lower Palaeozoic and present-day faunas and the absence of terrestrial vegetation in the Cambro-Ordovician affected the spatial distribution and abundance of biogenic traces as well as the fluvial geomorphology of the sediments of these successions (Bambach, 1979; Schumm, 1968; Cotter, 1978). Whilst the validity of the study of modern processes to help in understanding ancient sequences is not doubted, it may be pertinent to question whether a modern analogue exists for the epeiric seas or extensive shallow continental shelves that may have been characteristic of some of the Palaeozoic Era (see Selley, 1968).

Beuf *et al.* (1971) interpreted both the Cambro-Ordovician and the Lower Devonian blanket sandstones of the Sahara as being broadly transgressive sequences consisting of fluviatile sediments which pass upwards into transitional-marine, estuarine deposits. They recognised, however, that the relationship of the fluvial facies to the transitional-marine facies is complex, especially in the Lower Devonian successions. They proposed a model (Beuf *et al.*, 1971, p. 213) in which rivers opened out into the sea through estuaries where no clear distinction existed between fluvial and shallow-marine sand bars and where no shoreline, delta or barrier-lagoon complexes were developed. Burollet and Byramjee (1969), on the other hand, considered that the enigmatic Cambro-Ordovician sandstones were probably largely marine, having been deposited in an environment of very low gradients and vast shallow seas, where frequent, extensive transgressions and regressions dominated the processes of deposition. On a field trip to the Tassili N'Ajjer to study the same Cambro-Ordovician sediments (Bennacef *et al.*, 1971), the interpretation of Beuf *et al.* (1971) of the lower part of the succession as fluvial, was the source of great controversy. P. E. Potter and J. J. Bigarella considered that the lithologies, the maturity of the sediments, the sedimentary structures and the presence in associated beds of the trace fossils *Skolithos* and *Cruziana*, suggested a shallow-marine origin. A. Seilacher supported this view on the basis of the trace fossils present (*ibid*). P. Allen and L. M. J. U. Van Straaten, on the other hand, supported Beuf *et al.* (1971) in their interpretation of these sandstones as fluvial on the basis of the sedimentary structures and the unidirectional palaeocurrent indicators. In a similar Cambro-Ordovician succession in Jordan, Selley (1970), faced with the same problem of interpreting a succession of sandstones with the characteristics of fluvial sediments, proposed that the presence of *Cruziana* trace fossils at certain horizons might be explained by the invasion of the lower reaches of a braided river system by trilobites during a dry spell or a quiescent period between seasonal floods. The location of the *Cruziana* traces at two distinct stratigraphic horizons within this Jordanian succession may, however, indicate that minor transgressions were responsible for their presence. In a recent study of the Cambrian sandstones of Southwest Libya, Cepek (1980) proposed a model which, to some extent, provided a compromise between

the two separate schools of thought. He proposed a model involving sub-tidal to inter-tidal deposition adjacent to a fluvially-dominated delta complex.

The interpretation of the facies described by Beuf *et al.* (1971) as transitional marine is just as contentious as their fluviatile facies. On the basis of work undertaken in Jordan on Ordovician Nubian-type sandstones, Selley (1978, pp. 185-189) argued that this facies represents the deposits of migrating shoals in an open-marine environment. He argued that this facies is also present in the Lower Ordovician sediments of Libya which were interpreted, by contrast, in a recent study by Vos (1981b) as the deposits of a fan-delta complex.

At a Penrose-type conference on cratonic sedimentation sponsored by the Society of Economic Palaeontologists and Mineralogists in Wisconsin, in the summer of 1980, it became clear that the problems of interpretation encountered in the cratonic Cambro-Ordovician succession of North America (Dott & Byers, 1980) were very similar to those associated with the Palaeozoic Nubian-type sandstones of North Africa and Arabia. It was interesting to note that the discussion which took place at the exposure of the Mount Simon Sandstone (the basal Cambrian sandstone in the Wisconsin area) between many of the leading authorities on sedimentary geology revealed that no consensus of opinion existed about the interpretation of these sediments. The succession comprised cross-stratified sandstones which displayed unimodal palaeocurrents and contained bedding-plane trace fossils including *Rusophycus* and *Cruziana*. Despite the fact that an interpretation was proposed for these sandstones by Driese (Driese *et al.*, 1981) based upon a tide-dominated sub-tidal model in which the dominant preserved bedding was that of tidal channels, it was evident that a consensus did not exist about how fluvial, estuarine and shallow-marine bars could be distinguished on the basis of physically formed sedimentary structures, biogenic traces and patterns of palaeocurrents, in Lower Palaeozoic sandstones.

Despite the inconclusive result of the discussion about the Mount Simon Sandstone, the debate over the interpretation of these sandstone successions reflected recent advances in knowledge of modern sedimentary environments and an understanding of the problems involved in determining ancient depositional settings. An improvement over the

last decade in understanding processes operating on the continental shelf (Swift *et al.*, 1972; Stanley & Swift, 1976) and in fluvial and subaerial sand transport and deposition (Hunter, 1977; Miall, 1978) has made a number of alternative, well-documented analogues and models available for application where before only simplistic and idealised models existed. The initial adoption of a number of alternative interpretations, however, (a multiple-hypothesis approach) is counselled, for instance, by Frey and Seilacher (1980, p. 188) concerning the interpretation of trace fossil data. They warn against

> "...(an) almost overwhelming obsession to find unique, infallible ethological and environmental indicators (and against a) fevor for exact, one-to-one analogs in uniformitarian comparisons."

They also argue that

> "...intent quests for (unique environmental indicators) have dimmed the view of alternate possibilities or interpretations. Both comprehensive models and individual generalities tend to be overworked and over-extended (and this failing includes the) inflexible interpretation of the ichnofacies concept (of Seilacher (1967) and Frey (1975))...".

Finally, they emphasise (pp. 203-204) that biogenic structures, like sedimentary structures, are subject to the strengths and weaknesses of uniformitarianism. They represent behaviour and activity at the time of sedimentation just as sedimentary structures indicate the processes operating at the time. The interpretation of these biogenic traces therefore is dependent upon the degree to which the behaviour and activity of modern organisms is similar and is well understood.

The improvement in knowledge of modern sedimentary environments has emphasised how much is still not known. In shelf environments the internal structures of offshore sand ridges is imperfectly known and in fluvial sedimentology, despite a profusion of facies models for both braided streams (Miall, 1978; Rust, 1978) and meandering streams (Jackson, 1978), a healthy regard for their limitations is necessary. These concern the interpretation of sediments laid down before the development of land plants (Long, 1978; Winston, 1978) and the need to consider three-dimensional geometry of sand bodies (Nami & Leeder, 1978). Further, it is necessary to

consider the potential for preservation of structures and facies in relation to their internal controls (e.g. exceptional flood events) and external controls (eustacy and tectonics) (Lucchi, 1979). All too frequently facies models are not based upon a good knowledge of stratigraphic vertical sequences in the recent.

Considering the advances that have taken place in sedimentology, however, since Beuf *et al.* (1971) published their exhaustive study of the Cambrian to Devonian succession of the Sahara, it seems appropriate to reconsider the interpretation of their fluvial and transitional-marine facies and to attempt to refine the interpretation of the depositional settings of the Palaeozoic Nubian-type sandstones of North Africa and Arabia.

1.2 The Present Study

The facies studies presented in this thesis were undertaken in two field areas (Fig. 1.4). The field area in Southwest Libya was visited in late 1977 and early 1978 whilst the Saudi Arabian field area, in the Qassim District, was visited in late 1979. No facies studies had been undertaken previously on any of the four formations described in this thesis, and Beuf *et al.* (1971, p. 456) specifically mentioned the need for a detailed study of the Lower Devonian of Southwest Libya as it represented an important gap in their palaeogeographic reconstructions of the strata flanking the Hoggar Massif in the Central Sahara. One reason so little work has been undertaken on these strata is their extremely remote location. The field work on these sediments reported here was carried out with the use of a landrover and by camping at the site of the measured sections. These were situated between 80 and 150 kms from the nearest oasis, Al Awaynat, a small settlement with one shop, across a barren desert of soft sand, dunes and rocky outcrops. Al Awaynat is, in turn, 120 km from Ghat, the largest oasis in the southwest of Libya.

A brief account of the history of this research project follows. In the summer of 1977 a research programme sponsored by the Libyan National Oil Corporation and administered through the geological consultancy MARENGO (Mediterranean Arabian Resource Evaluations for Natural Gas and Oil) was established to study the facies of the subsurface and outcropping Palaeozoic strata of the Murzuq Basin. It

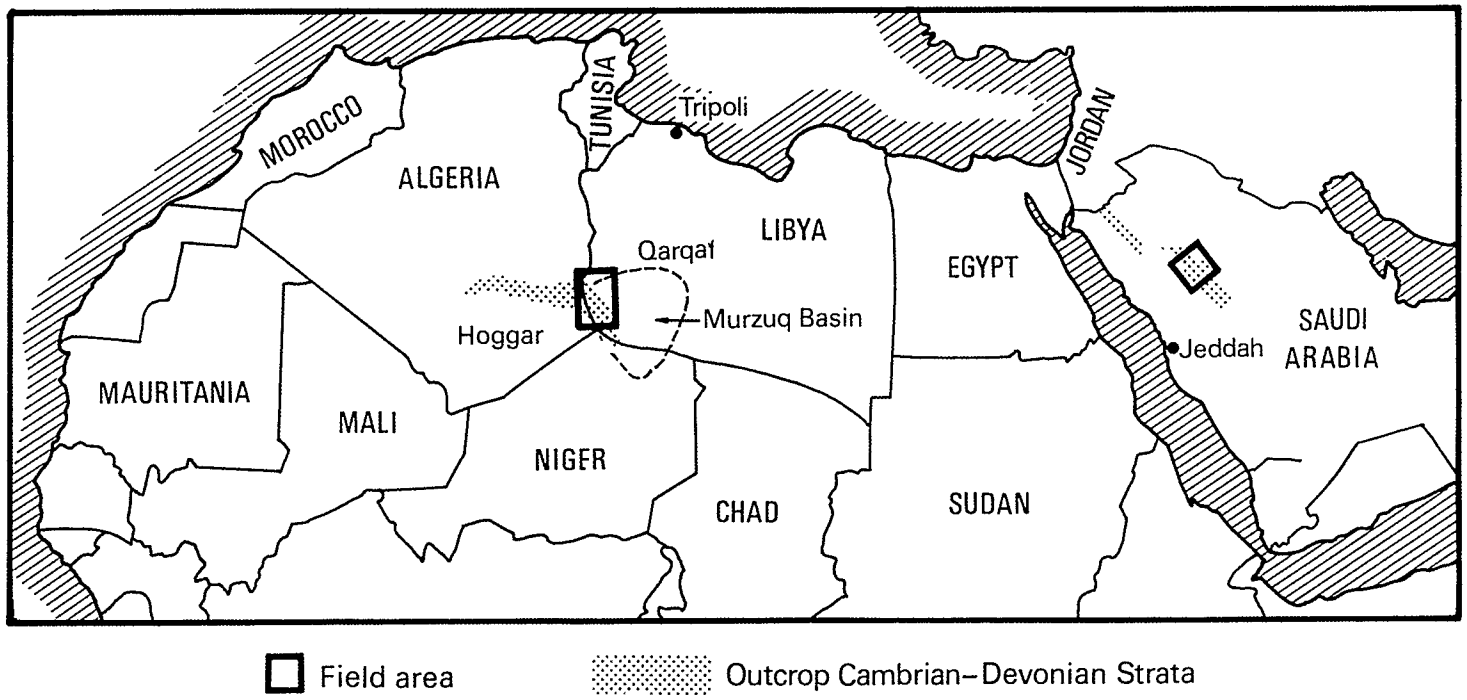


Figure 1.4 LOCATION OF TWO FIELD AREAS

was intended that each of the Palaeozoic formations which were well exposed in the field should be studied by a research student who would undertake outcrop facies analysis and subsurface studies from well data to obtain a regional understanding of the origin and distribution of the formation. These different studies were to have been brought together to generate a basin analysis for the Palaeozoic of the Murzuq Basin. The author was enrolled as a research student at Imperial College in August 1977 with the intention that a considerable amount of time would be spent in Libya on field work and subsurface studies. The Tadrart and Ouan Kasa formations of the Lower and Middle Devonian were assigned to the author and, following arrival in Tripoli, the first field trip was undertaken in late October. Two weeks were spent in the field looking at the Palaeozoic rocks of Southwest Libya. However, as a result of a restriction by the local authorities, owing to a border dispute with Algeria, on the use of the main access route (Takarkhouri Wadi) to the area where the Tadrart and Ouan Kasa formations are most well exposed, it proved possible only to spend one day studying these formations. On a second field trip of 18 days, during December 1977, a longer but politically acceptable access route was used and 12 days were spent on the exposures of the two Devonian formations.

At the end of October a decision was made, for internal political reasons, by the Libyan National Oil Corporation, to terminate the MARENGO contract at the end of the year and to cancel the study of the Murzuq Basin. An attempt was made by the author to continue with the project by taking an academic research post at the Al Fateh University in Tripoli. However, it proved possible only to make one more field trip to the study area, after which all further attempts to undertake field work in this area were frustrated, largely as a result of increasingly severe restrictions on travel by expatriates to the southern borders of Libya. During the one field trip that was successfully arranged, a period of 29 days in February and March, 1978, a total of 22 days were spent measuring sections on the exposures of the Tadrart and Ouan Kasa formations. During the other days some time was spent studying the Middle and Upper Devonian Aouinet Ouenine Formation and the Silurian and Ordovician rocks of the study area. Apart from a five day field excursion to the Qarqaf Region on the north of the Murzuq Basin (Fig. 1.4) with the Earth Science Society of Libya at the

end of April 1978, no further visits to Southwest Libya were made, although efforts to arrange field work continued actively until March 1979.

The author was, therefore, able to spend only five weeks in total, in the field, on the study of the Tadrart and Ouan Kasa formations, and this field work was undertaken within the first seven months of the project. For a long time it was considered that the results of this work, which initially appeared to leave the majority of the major interpretation problems unsolved, would be unsuitable material for a thesis. During 1979 and 1980, therefore, time was devoted to three different projects, all of which, it was felt, might provide material more suitable for a thesis. The first of these projects was a study of the Palaeozoic stratigraphy, sedimentology and oil prospects of the Murzuq Basin based upon the author's field work, and subsurface data from wells in the Murzuq Basin. This project, which was started in mid 1978, was sponsored by Occidental of Libya, and a report was completed in 1980 (Clark-Lowes, 1980a). Although this research provided the financial backing necessary to allow the author to continue with the thesis project, it did not, in the final analysis, provide material ideally suited for inclusion in a thesis because of the poor quality of the majority of the subsurface data. Also, the report concentrated upon applying what was known about the stratigraphy and sedimentology of the Palaeozoic to the search for hydrocarbons, rather than attempting to refine in any detail the existing interpretations.

The second project grew out of the opportunity to visit Saudi Arabia in late 1979 to study Nubian-type sandstones in the Palaeozoic succession of the Qassim District. The project was undertaken as a part of the Imperial College Saudi Arabian Cover Rock Project through which assistance in the field was provided by Riofinex (a subsidiary of Rio Tinto Zinc) and the Directorate General of Mineral Resources of Saudi Arabia. The objective of the Cover Rock Project was to identify potential strata-bound mineral deposits. The author was required to undertake a facies study of the complete Cambrian to Lower Devonian succession. Three weeks were spent studying this succession and for the majority of this time field work was undertaken, as in Libya, with the use of a four-wheel-drive desert vehicle. For a period of six days, however, a helicopter was available which helped

greatly in speeding up the field work. It also allowed some reconnaissance field mapping using 1:100,000 scale controlled photomosaics to be undertaken. This provided a refinement of the existing 1:500,000 scale geological map of the area (Bramkamp *et al.*, 1963) and proved important to the understanding of the facies relationships in the Ordovician succession. The results of this work were presented in a report completed in 1980 (Clark-Lowes, 1980b).

A third subsidiary project was undertaken, mainly in 1979, to investigate whether scanning electron microscope work on the collected samples of Nubian-type sandstones might provide evidence, from grain surface textures, for the transporting medium in which the grains of the sandstone were carried prior to being deposited. Whilst this study provided some valuable petrographic and diagenetic data on the sandstones studied, overgrowths drastically limited the number of grains displaying original surface textures, and those that did appeared not to be susceptible to the type of analysis proposed by Krinsley and Doornkamp (1973) and Krinsley (1978) for environmental discrimination. It was concluded that little useful information could be obtained about depositional settings from these techniques in the case of the Palaeozoic samples under study.

In writing up, as a thesis, the results of the research work undertaken between 1977 and 1980 a decision was made to concentrate on the outcrop facies analysis of the sediments studied in both Southwest Libya and Saudi Arabia, and to include only that part of the subsurface study of the Murzuq Basin which has bearing upon the nature of cratonic sedimentation and cratonic basins. Similarly, only that work on the scanning electron microscope that has relevance to sediment petrography and diagenesis is included. This decision was taken in the knowledge that, had it been possible to do more field work, some remaining interpretation problems may have been solved, but also in the belief that, however much time had been spent studying these sediments in the field, they would have remained enigmatic, just as the interpretation of the Cambro-Ordovician blanket sandstones of North America continues to be the source of considerable dispute (Dott & Byers, 1980). Since the reports on the subsurface study of the Murzuq Basin and the facies analysis of the Saudi Arabian succession were not completed until the end of the third year of this project, this text has been largely written whilst in

full time employment.

In summary, therefore, an attempt is made in this thesis, through the facies work undertaken, to provide detailed descriptions of the facies of the Palaeozoic blanket sandstones of Libya and Saudi Arabia, and to discuss their interpretation, and secondly, to discuss the facies analysis of the more marine Ordovician and Silurian succession of Saudi Arabia. In the remaining pages of this chapter the geology and geomorphology of the two study areas are described. In Chapter 2 the techniques and terminology employed in the outcrop facies work is discussed in order to clarify the definitions of terms which, especially in the field of ichnology, are frequently used in different ways by different authors and in which duplication of terms is common. In the succeeding chapter (Chapter 3) a description and interpretation is given of the facies of the Libyan Lower Devonian blanket sandstones (Tadrart Formation) and the overlying marine sediments of Ouan Kasa Formation. The facies of the Cambrian to Lower Ordovician blanket sandstones of Saudi Arabia (Saq Formation) are similar to those of the Lower Devonian of Libya and are treated next (Chapter 4). In the study of the facies of the Ordovician to Lower Devonian sediments of Saudi Arabia (Tabuk Formation) (Chapter 5) a different aspect of cratonic sedimentation is discussed. Evidence for deltaic sedimentation is provided by the Lower Ordovician sediments whilst the study of an extensive palaeotopography in which valleys were filled in the late Ordovician with probable fluvio-glacial and glacial deposits, provides some evidence for a late Ordovician glacial history in parts of Saudi Arabia similar to that of the Central Sahara. In the final chapter some conclusions are drawn about cratonic sedimentation and the development of cratonic basins.

1.3 The Geology and Geomorphology of the Study Area, Southwest Libya

1.3.1 General

The study area lies on the west of the Murzuq Basin in Southwest Libya in the region of Ghat, Jabal Akakus and Jabal Tadrart (Fig. 1.5). In this region the Cambrian to Devonian strata are progressively truncated northwards by Upper Devonian and Carboniferous strata which lie above a major basin-wide unconformity. The significance of this unconformity is illustrated by the successions encountered in ten oil exploration wells in the Murzuq Basin (Table 1.1). In this stratigraphic tabulation it can be seen that the Lower Devonian and Silurian successions are only partially preserved, the amount of preservation being a function of the degree of erosion beneath the "Chatti Formation". It has been necessary in the stratigraphic table (Table 1.1) to divide the Middle and Upper Devonian Aouinet Ouenine Formation into two informal formations: an upper "Chatti Formation" and a lower "Ouenine Formation" (Table 1.2) following the usage adopted by Collomb (1962) in his measured section in the Wadi ash Shati. This division is adopted because it recognises the importance of the unconformity at the base of the "Chatti Formation" (Table 1.2) and because the succession measured by Collomb (1962) provided a much better reference section for the wells of the Murzuq Basin than did the type section at Awaynat Wanin on the North of the Qarqaf Arch (Collomb, 1962; Massa & Moreau-Benoit, 1976). This subdivision has only been adopted in Table 1.1, however, since the use of the formation name "Aouinet Ouenine" has become well established for outcropping Devonian sediments (Banerjee, 1980). From the well data and the Wadi ash Shati succession (Tables 1.1 & 1.2) it is evident that the age of the uplift and erosion associated with the major unconformity is Frasnian and that the uplift and erosion was most pronounced in the northwest, the north and the east, where the "Chatti Formation" lies above Cambro-Ordovician sediments. In the west, southwest and south, however, the Cambrian to Devonian succession is fully preserved. The subsurface data also provides evidence to support the proposed existence of a Palaeozoic Brak-Ghanimah Arch separating two Palaeozoic sub-basins, the Murzuq-Jadu and the Dur al Qussah sub-basins (Klitzsch, 1963, 1966a). The present depth of burial of the Precambrian basement (see 2000 m basement

Figure 1.5 GENERALISED GEOLOGICAL MAP OF THE MURZUQ BASIN

Modified after

Geological Map of Libya (Second Edition) (Conant and Goudarzi, 1964; Maghrabi and Cheshitev, 1977),

Unpublished Oasis Oil Company map of Western Libya and Niger, and

Furst and Klitzsch (1963).

Tectonic Elements from Hoen (1968) and Goudarzi (1980).

Structure contour on Precambrian basement based on data from oil exploration wells in the basin (e.g. AINC 34, see Table 1.1).

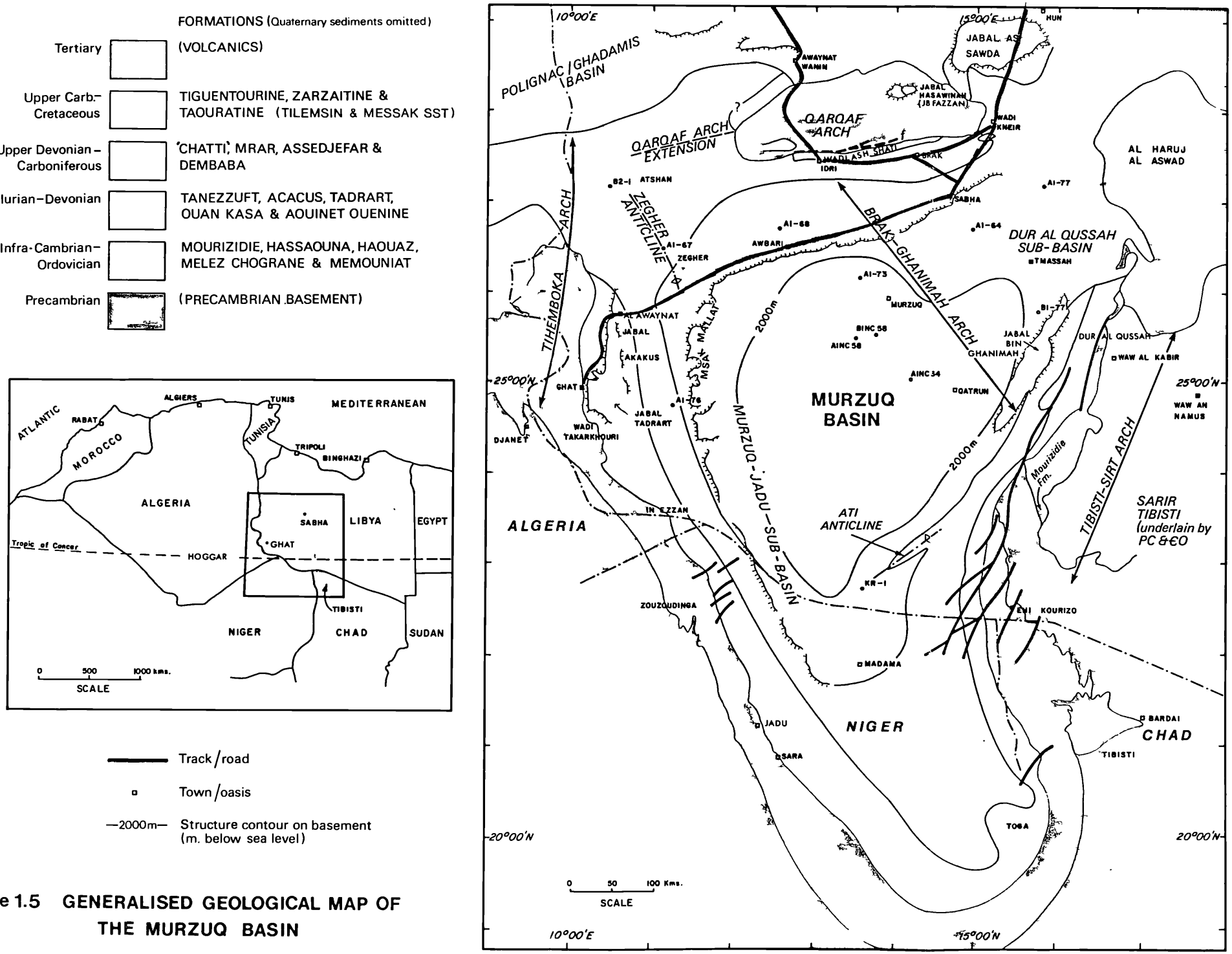


Figure 1.5 GENERALISED GEOLOGICAL MAP OF THE MURZUQ BASIN

	TOPS OF FORMATIONS	A1-64	A1-67	A1-68	A1-73	A1-76	A1-77	B1-77	AlNC34	AlNC58	KR-1
PERMO-CARBONIFEROUS	Tiguentourine	488		?172	1087		352	165	1476	1624	396
	Dembaba	701		535	1615		481	558	1919	2106	840
	Assedjefar	?		?	1637		?	725	1981	2150	?
	Mrar	?	Surf.	?	?	Surf.	?	910	2119	2277	?
DEVONIAN	"Chatti"	1120	?	1088	1977	?	886	1260	2332	2469	?
	"Ouenine"	(1137)	(297)	(1208)	(2021)	361	(941)	(1410)	(2432)	(2522)	1312
	Ouan Kasa	-	-	-	-	419	-	-	-	-	1362
	Tadrart	-	-	-	2021 TD 2063 in Tad.	445	-	-	2432	2522	?
SILURIAN	Acacus	-	297	-		698	-	1410	2563	2565	?
	Tanezzuft	-	331	1208		1065	-	1485	2637	2728	1528
CAMBRO-ORDOVICIAN	Memouniat & Melez Chograne	1137	448	1413		1481	941	1513	2710	2847 TD 3012 in Mem.	1636
	Haouaz	?	?	1524		1630 TD 1669 in Hao.	c.1100	TD 1714 in Cambro-Ord.	2886		TD 1994 in Cambro-Ord.
	Hassaouna	1280	?	TD 1755 in ?Has.			c.1300		(2954)		
Basement (Top)		1634	978		c.2679	c.1955	2369	c.2297	2954	c.3389	c.2600
Total Depth in basement		1662	1006				2377		3035		
Elevation (m. above m.s.l.)		486	620	451	479	705	491	497	543	539	c.800

KEY

488 = Depth below surface (metres) of top of formation.

(1137) = Depth below surface of unconformity where one or more formations are absent.

c.2679 = Estimated depth.

Surf. = Formation outcrops at surface.

- = Formation(s) absent in sub-surface.

? = Formation(s) present, boundaries uncertain.

TD = Total Depth.

LOCATION MAP

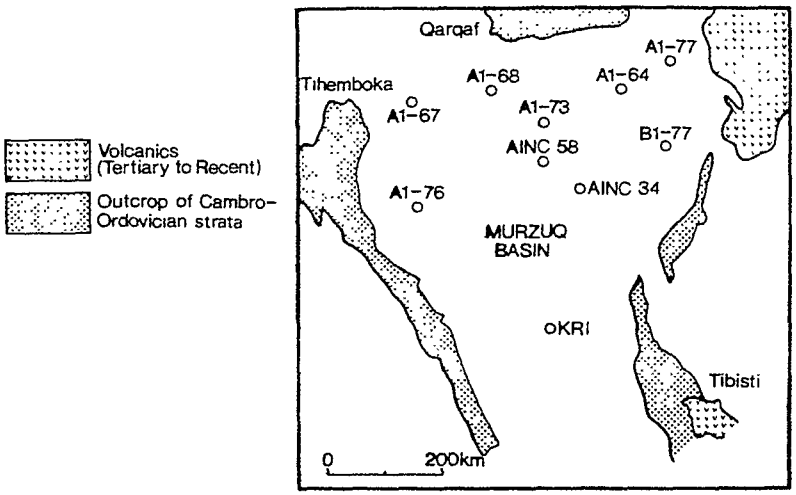
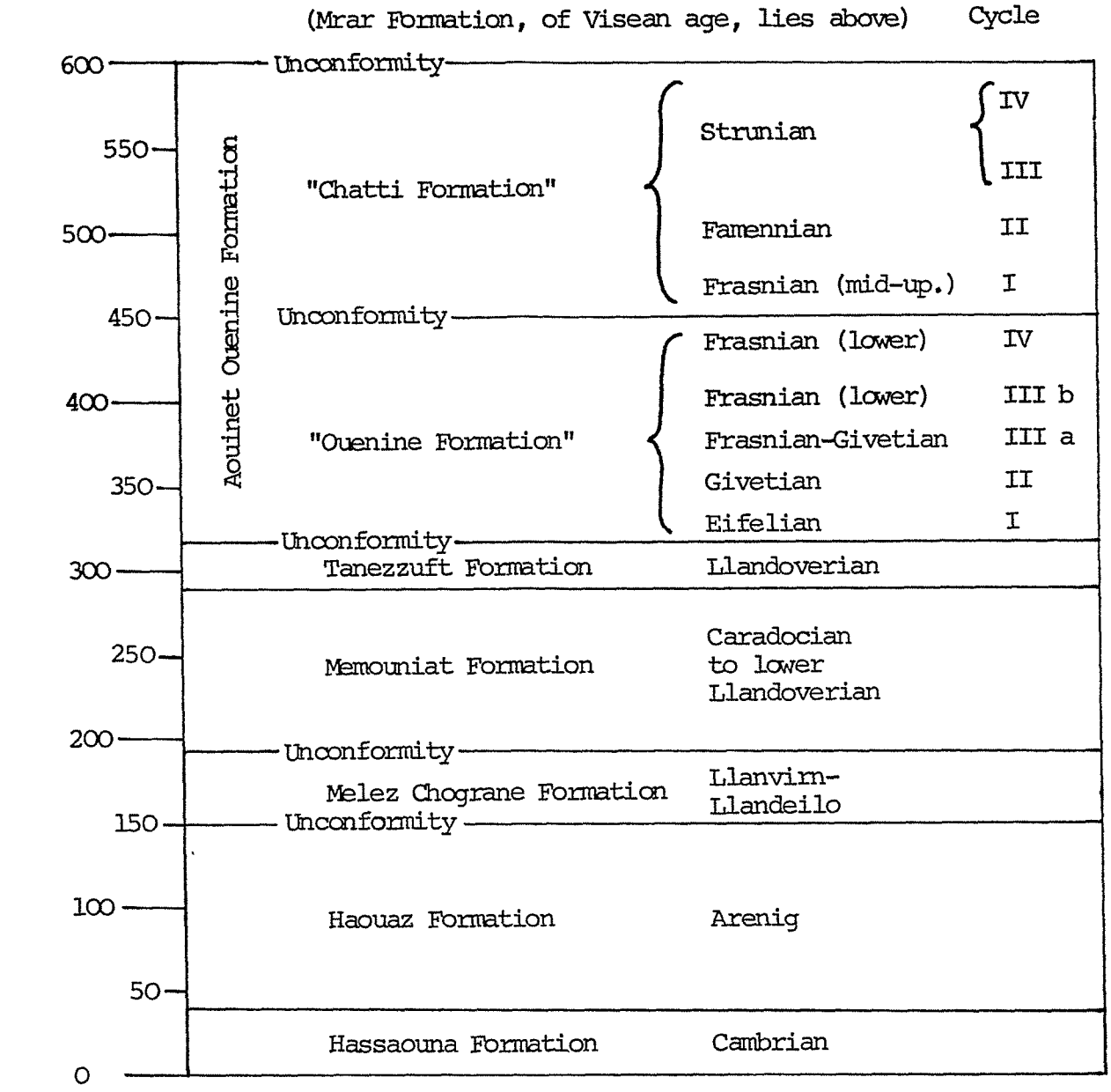


Table 1.1 COMPILATION OF STRATIGRAPHIC DATA FROM TEN OIL EXPLORATION WELLS IN THE MURZUQ BASIN.



Modified after Collomb (1962) and Turk (1978)

Table 1.2 STRATIGRAPHIC SUCCESSION IN WADI ASH SHATI

structure contour, Fig. 1.5) and the cuesta-like escarpment formed by the Mesozoic Messak Sandstone indicate that the Brak-Ghanimah Arch subsequently ceased to be a positive structural element and collapsed as part of the floor of the Mesozoic Murzuq Basin. The Cambrian to Devonian succession studied on the west of the Murzuq Basin, therefore, is preserved within a Palaeozoic sub-basin which is bounded in the west by the Tihemboka Arch and in the east by the Brak-Ghanimah Arch.

The sandstones of the Cambrian to Devonian succession of the Central Sahara have been assigned to three broad facies types by Beuf *et al.* (1971): fluviatile deposits, transitional-marine deposits and glacial/periglacial deposits (Fig. 1.6). They argued that the blanket sandstones of the Cambro-Ordovician succession throughout the Central Sahara are characterised by fluviatile deposits at the base, overlain by transitional-marine deposits overlain in turn by Upper Ordovician glacial and periglacial deposits. Flanking the northeast of the Hoggar Massif these sediments are exposed in the range of jabals and plateaux of the Tassili N'Ajjer (Tassili Interne, Kilian, 1931) (Fig. 1.6). Beuf *et al.* (1971) argued, similarly, that the blanket sandstones of the Tassili Externe are characterised by a generalised vertical succession of Silurian transitional-marine deposits overlain by Lower Devonian fluviatile deposits with transitional-marine deposits at the top (Fig. 1.6). The successions of the Tassili Externe were acknowledged, however, to be more complex and variable than those of the Tassili Interne. The Internal and External Tassili of the Tassili N'Ajjer are separated by a low-lying region of broad wadis in which marine graptolitic shales of Silurian (mainly Llandoverian) age crop out (Berry & Boucot, 1973a). This succession is termed the Intra-Tassilian (Beuf *et al.*, 1971).

1.3.2 Previous Geological Research

The earliest geographical records about southern Libya are those of Herodotus who visited the Fazzan during the Roman occupation of North Africa. Until the early 19th Century maps including the Fazzan were based largely on his observations. In the late 18th Century and during the 19th Century the natural history notes of the early North African explorers included the first scientific and geological work undertaken in the Fazzan. In 1797 two Germans, Frederick Horneman,

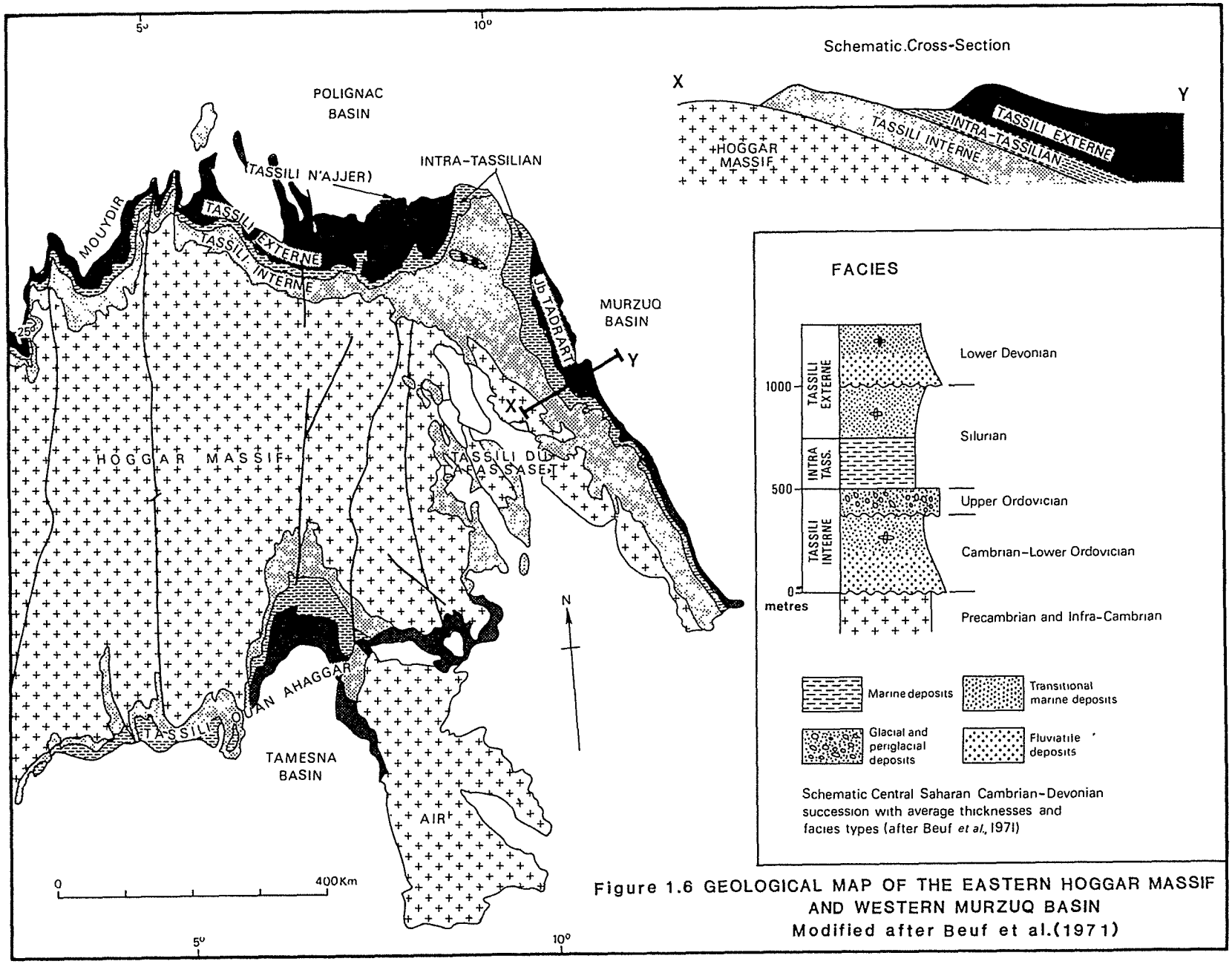


Figure 1.6 GEOLOGICAL MAP OF THE EASTERN HOGGAR MASSIF AND WESTERN MURZUQ BASIN
Modified after Beuf et al.(1971)

a student of natural sciences and an explorer, and Joseph Freudenberg, visited Murzuq on a scientific mission of the African Association of London, the aim of which was to reach the River Niger from the Mediterranean coast and to record natural historical and cultural observations. In 1799 Horneman succeeded in reaching the River Niger via Bornu, but died in 1800 at the age of 28. In 1822, members of Captain Dixon Denham's expedition, including H. Clapperton, were the first Europeans to visit and draw up maps of the Ghat-Al Awaynat district. In 1849 a scientific expedition led by a cleric, J. Richardson, and including a German naturalist and geologist, A. Overweg, and an archaeologist, H. Barth, made detailed reports on the oases of the Fazzan, including Ghat and Awaynat Wanin. Before Overweg died in 1851 he made detailed observations of many of the Palaeozoic rocks of the Murzuq Basin (Pesce, 1969).

During the Italian administration work was carried out by Desio, Borghi, Chiesa and others (Desio, Ed., 1939 - 1953) who provided the first fairly comprehensive topographic documentation of southern Libya, and the first geological framework of the Murzuq Basin. The establishment by Desio of the Natural History Museum in Tripoli in the late 1930's facilitated detailed recording of stratigraphic data. During the French administration of the Fazzan (1942-51) accurate geological mapping and surveying was continued on a systematic basis and much of it was published by the Institut de Recherches Sahariennes at the University of Algiers (e.g. Dalloni, 1948). This work resulted in the publication of a 1:2,000,000 scale map of the Central Sahara for the International Geological Congress of 1952. Geological mapping and field work was continued by the French after 1951, the key work on Southwest Libya being that of Freulon (1964) and that published by Beuf *et al.* (1971). Full reviews of the geological exploration of the Fazzan are provided by Lelubre (1952) and Furon (1955). Freulon (1964) undertook a study of the Palaeozoic succession of the Central Sahara based upon field mapping in the Tassili N'Ajjer of Algeria and in the Fazzan. During five field seasons, of six to eight months duration each, he surveyed these areas topographically and undertook primary geological mapping. The work was done between 1950 and 1955 on foot, on camel and, in the later years, with jeeps. The Jabals Akakus and Tadrart of Ghat and Al Awaynat lay at the centre of his field study area, and he was

responsible for its first geological surveying, the resulting map being published as the Fort Charlet map at a scale of 1:500,000. Despite the central position of the Ghat-Al Awaynat area the majority of Freulon's stratigraphic research was undertaken in southern Algeria and in the Qarqaf region. His work in the Ghat-Al Awaynat region was largely undertaken in the winters of 1950/51 and 1952/53.

In the publication of Beuf *et al.* (1971) there is compiled, arguably the most comprehensive account of siliciclastic cratonic sedimentation in the geological literature. The excellent exposure of these undeformed sediments makes the outcrops surrounding the Hoggar and Reguibat massifs (see Fig. 1.3) ideal areas for such a study. The research recorded in the compilation of Beuf *et al.* (1971) is largely that of French Saharan geologists who had been working in the area since 1958. The succession of Southwest Libya was only briefly studied, however, in 1961 and 1962.

In the late 1950's and early 1960's petroleum exploration programmes in Southwest Libya, in the Murzuq Basin, and on the shoulder between the Murzuq and Ghadamis basins, were undertaken by a number of companies after oil had been discovered in the Polignac Basin. 15 boreholes in total were drilled in 7 concessions in the Murzuq Basin (Fig. 1.5 & Table 1.1) and geological field parties studied the outcropping Palaeozoic strata on the margins of the basin (e.g. Collomb, 1962; Furst & Klitzsch, 1963; Collomb *et al.*, 1960, 1961; Craig *et al.*, 1958; Collomb & Heller, 1958). The need for an accepted stratigraphic nomenclature became clear and in 1960 the Lexique Stratigraphique International (Libye) was published through the efforts of A. Desio and the Names and Nomenclature Committee of the Petroleum Exploration Society of Libya. In 1965, E. Klitzsch, as part of a revision of the Palaeozoic part of the Lexicon under his chairmanship, published the details of sections which he measured in the Silurian to Lower Devonian successions on the west of the Murzuq Basin, and of associated palaeontological work, and proposed the adoption of these as type sections (1965a).

The most recent reviews of the geology of southwest Libya are those of Bellini and Massa (1980), who drew together a large amount of previously published and unpublished academic and oil company data in a major review of the Palaeozoic of the southern sedimentary basins of Libya, and Klitzsch (1981), who has provided a valuable review of

the extensive research he and his co-workers have undertaken in the Murzuq Basin.

1.3.3 Stratigraphy and Structure

In the study area (Fig. 1.7) blanket sandstones of both the Cambro-Ordovician and the Lower Devonian are exposed, on the west and in the east respectively of the Wadi Tannezzuft. In this area the sandstones of the Cambro-Ordovician, which are exposed in the Tassili Interne in Algeria, are weathered to a relatively low lying and featureless region (Fig. 1.8). By contrast the sandstones of the upper part of the Silurian and the Lower Devonian successions (the equivalent of the Tassili Externe) crop out in a large ravined plateau (a "Tassili", to use the Tamahak word) as the Jabal Akakas and Jabal Tadrart (Fig. 1.8). Figure 1.9 illustrates the geology of the area which is dominated by gently eastward dipping strata which form outcrop belts, aligned north-south, of well exposed sandstones and relatively poorly exposed shales. The Silurian shales of the Tanezzuft Formation crop out in the Wadi Tanezzuft to the east of which the sandstones of the Acacus and Tadrart formations (Silurian and Lower Devonian) are exposed. The whole succession dips eastwards marking the western boundary of the Murzuq Basin at an average dip of less than 1^{o}. Gentle flexuring of the strata has taken place (Fig. 1.9) and dips of up to 7^{o} are recorded with exceptional dips of up to 12^{o} (Collomb & Heller, 1958; author's measurements). Some north-south aligned faults are present, usually throwing towards the east, i.e. into the basin. A structural cross-section constructed from the topographic and geological maps (Figs. 1.8 & 1.9) and drawn from Ti'n Alkum northeastwards through the Wadi Teshuinat (Fig. 1.10) shows that the average dip must be considerably below 1^{o} and that the geomorphology of the Jabal Tadrart is related to a flexure or monocline aligned approximately north-south and running along the dip slope up the back of the dissected plateau.

The stratigraphic succession of the Ghat Region is illustrated in Figure 1.11 in which the Cambro-Ordovician and Silurian to Lower Devonian blanket sandstones are schematically illustrated as cross-stratified and bioturbated units (see Fig. 2.1 for symbols and abbreviations used in figures). Three major unconformities are recognised in the succession. At the base the Palaeozoic rocks lie

Figure 1.7 JABAL AKAKUS AND JABAL TADRART : LOCATION MAP

Scale of map : approximately 1 cm = 8 km.

Drawn from NASA satellite image at approximate scale of 1:250,000 (Fig. 1.15).

Additional sources:

Mori (1965), Collomb and Heller (1958), Freulon (1964), Salem and Busrewil (1980) and U.S. Army Map of Fort Charlet area (NG 32) (1955).

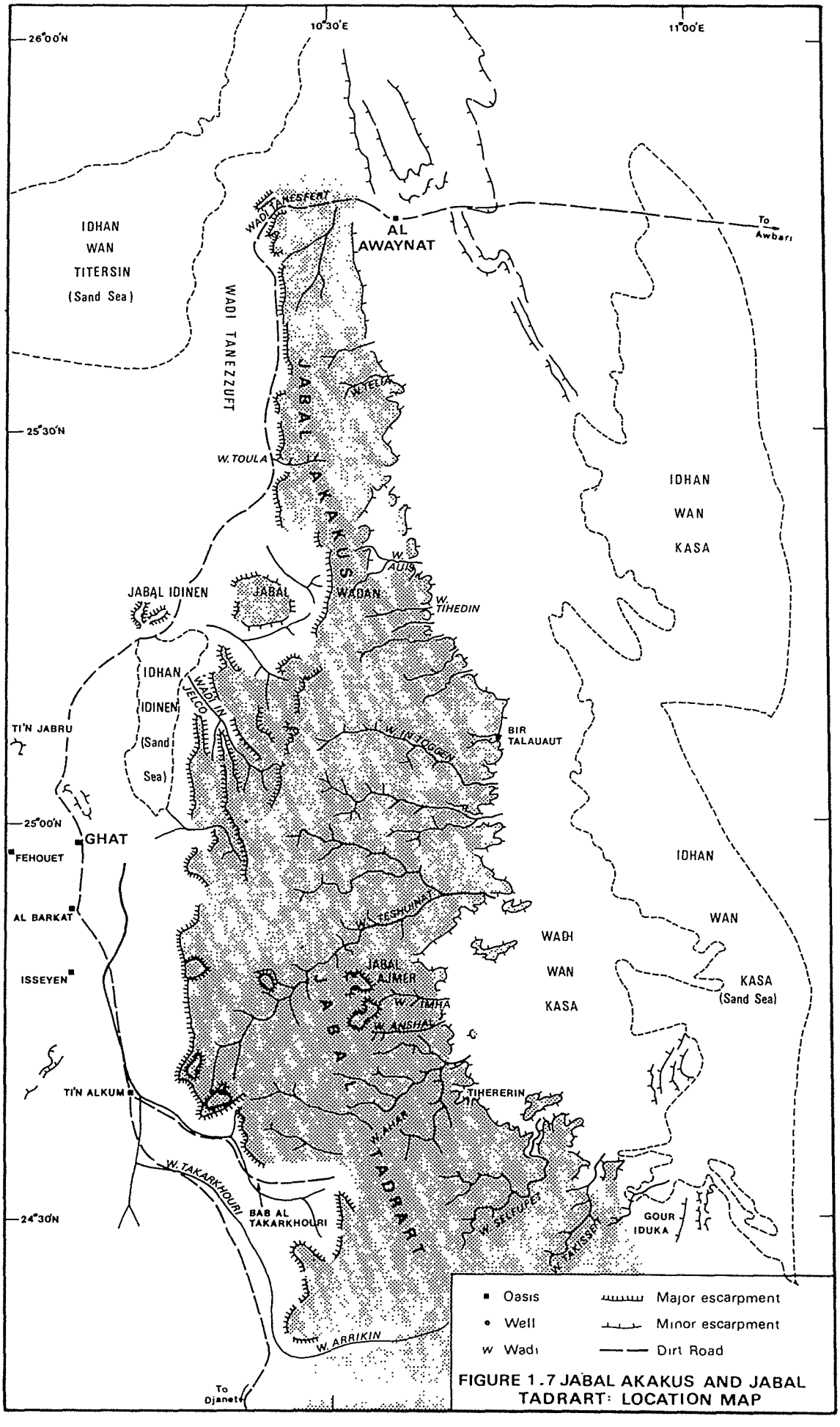


FIGURE 1.7 JABAL AKAKUS AND JABAL TADRART: LOCATION MAP

Figure 1.8 JABAL AKAKUS AND JABAL TADRART : TOPOGRAPHY

Map constructed from Figure 1.7. Spot heights (metres above mean sea level) are from Mori (1965), Collomb and Heller (1958), Collomb *et al.* (1961), Craig *et al.* (1958) and Freulon (1964).

Note that northwards, along the axis of the jabals, the elevation decreases, the highest ground lying southeast of Ghat and north of Wadi Takarkhouri. This has been attributed to northeastward tilting of a Tertiary peneplain. However, the situation is complicated by the presence of a number of Tertiary peneplains (Barth & Blume, 1975), the oldest of which may be responsible for the high plateau south of Ghat, while northwards the lower plateaux may be the remnants of younger and lower peneplains. The highest plateau is located where the topographic extent of the resistant sandstones is thought to have been greatest at the time of peneplanation, offering, therefore, greater resistance to weathering.

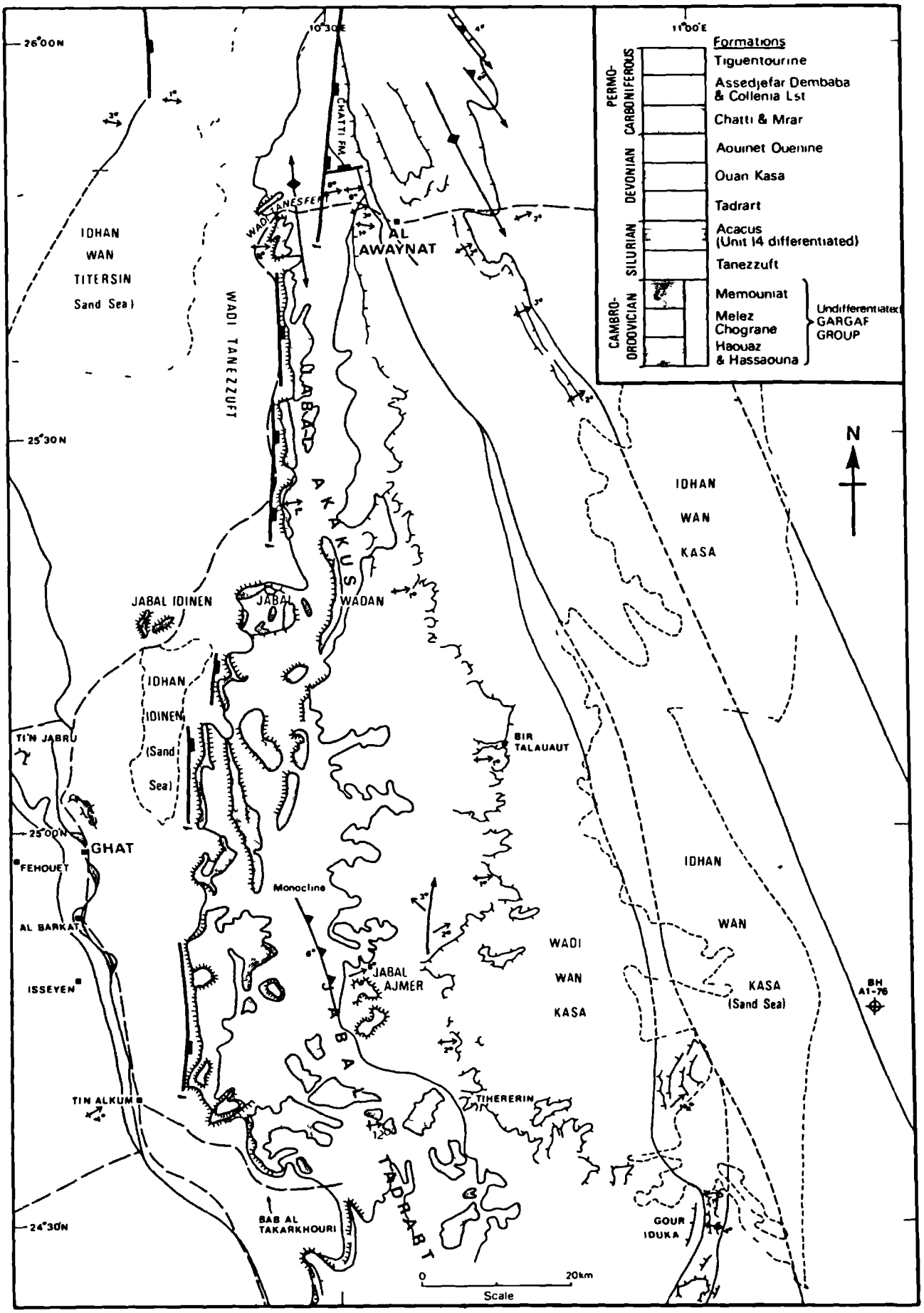

26°00N
25°30N
25°00N
24°30N
10°30E
11°00E
Formations
PERMO-CARBONIFEROUS
Tiguentourine
Assedjefar Dembaba & Collenia Lst
Chatti & Mrar
DEVONIAN
Aouinet Ouenine
Ouan Kasa
Tadrart
SILURIAN
Acacus (Unit 14 differentiated)
Tanezzuft
CAMBRO-ORDOVICIAN
Memouniat
Melez
Chograne
Haouaz & Hassaouna
Undifferentiated GARGAF GROUP
CHATTI FM.
WADI TANESFET
AL AWAYNAT
IDHAN WAN TITERSIN Sand Sea)
WADI TANEZZUFT
DJADO
AKAKUS
JABAL IDINEN
JABAL WADAN
IDHAN IDINEN (Sand Sea)
TI'N JABRU
GHAT
FEHOUET
AL BARKAT
ISSEYEN
TIN ALKUM
Monocline
JABAL AJMER
BIR TALAUAUT
IDHAN WAN KASA
WADI WAN KASA
KASA (Sand Sea)
BH A1-76
TIHERERIN
BAB AL TAKARKHOURI
DJADO
GOUR IDUKA
N
0
20km
Scale

Figure 1.9

Figure 1.9 JABAL AKAKUS AND JABAL TADRART : GEOLOGY

Map constructed from Figures 1.7 and 1.8. Field geological data, including structural dip measurements, from author's sections, Collomb and Heller (1958), Collomb *et al.* (1961), Craig *et al.* (1958), Klitzsch (1965a,1969), Martindale (1977) and BEICIP (1972).

Acacus and Tadrart formation boundaries revised after consideration of topography, regional and local variations in structural dip, and morphological boundaries mapped from the satellite image of the area (see Fig. 1.15). Extensive exposure of the massively bedded sandstone unit of the Acacus Formation (unit 14, Klitzsch, 1965a) is distinguished from that of the thinly bedded units where possible. The Collenia Limestone (uppermost beds of the Mrar Formation) has been lumped with the Assedjefar Formation for convenience of mapping.

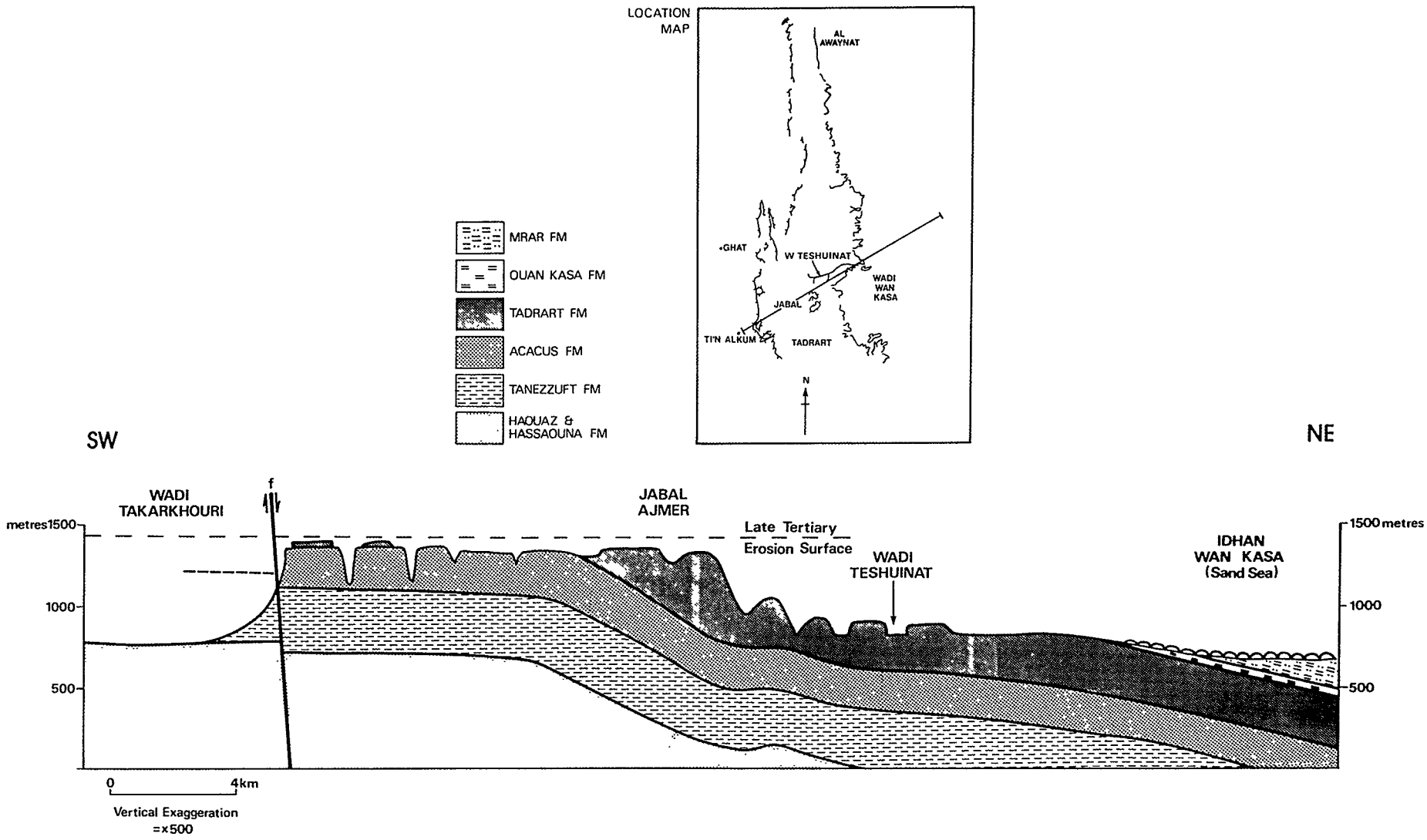


Figure 1.10 STRUCTURAL CROSS-SECTION THROUGH JABAL TADRART

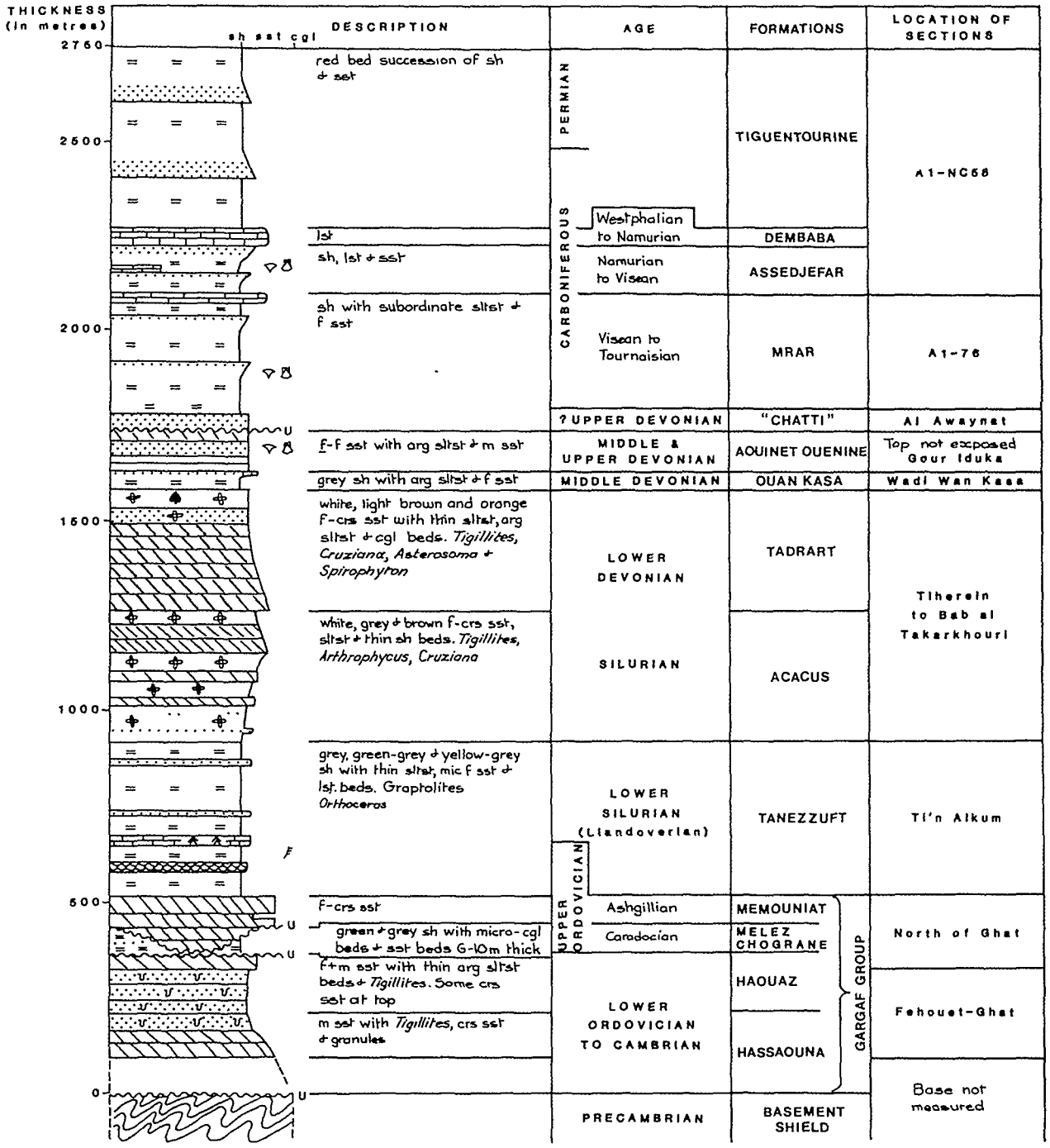


Figure 1.11 STRATIGRAPHIC SUCCESSION OF THE GHAT REGION

unconformably upon eroded Precambrian igneous and metamorphic rocks. At the base of the Upper Ordovician Ashgillian succession an unconformity surface is present which is characterised by an eroded topography of considerable relief (BEICIP, 1972) and which commonly combines with a subsidiary unconformity at the base of the Caradocian succession. Although shales are present in the Caradocian succession they are commonly truncated by the upper of these two unconformities with the result that the complete Cambro-Ordovician succession is generally composed of sandstones. The third unconformity is that already discussed (see section 1.3.1) at the base of the "Chatti Formation". The age of this unconformity in the Ghat Region has generally been considered to be earliest Carboniferous to latest Devonian (e.g. Bellini & Massa, 1980) but evidence from measured sections near Al Awaynat (Collomb & Heller, 1958; Collomb *et al.*, 1961) indicated that the oldest sediments lying above the unconformity surface were probably Upper Devonian. Their lithology, oolitic ironstones and sandstones, suggested a correlation with the "Chatti Formation" of the Qarqaf Region (Collomb, 1962; Turk, 1978; Van Houten & Karasek, 1981) which indicated that the age of the major unconformity in the Ghat Region may be the same as in the Qarqaf Region, i.e. Frasnian. Some support for this is provided by the presence of a hiatus of Frasnian to Famennian age in the Gour Iduka region (Fig. 1.9) (Collomb & Heller, 1958). Since the formation or group name of "Aouinet Ouenine" has gained general acceptance in the literature for the outcropping Middle and Upper Devonian rocks of the Ghat Region (e.g. Banerjee, 1980; Klitzsch, 1981; Bellini & Massa, 1980) this name has been retained for the outcrops of the study area. However, it is used here to embrace rocks of Eifelian to Frasnian age and not those of Famennian age.

The stratigraphy of the Cambrian to Lower Devonian successions of the Central Sahara is presented in Table 1.3 to provide a framework for the description of the Cambro-Ordovician and Lower Devonian blanket sandstones, and to show the lithostratigraphic equivalents of the formations of the Murzuq Basin in the basins of southern Algeria, Niger, and West Libya (the Ghadamis Basin). For convenience of reference a generalised description of the lithologies and the facies is provided from Beuf *et al.* (1971).

AGE	ALGERIA (NORTH HOGGAR)			LIBYA		ALGERIA & NIGER (SOUTH HOGGAR)	LITHOLOGY	FACIES	GENERAL SAHARAN TERMINOLOGY
	Ahnet	Mouydir	Tassili N'Ajjer	Murzuq Basin	Ghadamis Basin	Ouan Ahaggar			
LOWER DEVONIAN	Oued Samene Fm.	Oued Samene Fm.	Oued Samene Fm.	Tadrart Fm.	Tadrart Fm.	Edikel Fm.	Cross-stratified and channelled sandstones with subordinate siltstones	Fluviatile passing upwards to Transitional Marine	External Tassilian Sandstone Succession
SILURIAN	Assejrad Fm.	Tamelrik Fm.	Tamelrik Fm. / Tifernine Fm } Oued Karkai Fm.						
		Assejrad Fm.	Atafaitafa Fm.	Acacus Fm.	Acacus Fm.	Efeimazerta Fm.			
						?Zone de Passage	Sandstones, siltstones and shales	Transitional Marine	
	Imirhou Fm.	Imirhou Fm.	Imirhou Fm.	Tanezzuft Fm. / Iyadhar Fm.	Tanezzuft Fm.	Imirhou Fm.	Graptolitic shales, siltstones and subordinate sandstones	Open Marine	Intra-Tassilian Succession
CAMBRO-ORDOVICIAN	Tamadjert Fm.	Tamadjert Fm.	Tamadjert Fm.	Memouniat Fm. / Melez Chograne Fm.	Djeffara Fm.*	Tamadjert Fm.	Sandstones, conglomerates and shales	Glacial & Periglacial (Glacio-marine passing upwards to fluvio-glacial)	Internal Tassilian Sandstone Succession
	In Tahouite Fm.	In Tahouite Fm.	In Tahouite Fm.	Haouaz Fm.	Bir Ben Tartar Fm.	In Azaoua Fm.	Sandstones, siltstones & subordinate shales	Fluviatile passing upwards to Transitional Marine	
					Kasbah-Leguine Fm.				
	Ajjers Fm.	Ajjers Fm.	Ajjers Fm.	Hassaouna Fm.	Sanrhar Fm. / Sidi-Toui Fm.	Ajjers Fm.	Cross-stratified and channelled sandstones with subordinate siltstones		

Note: Lithostratigraphic Units are correlated in this chart and age equivalence is not implied.

*The Djeffara Formation awaits redesignation as it may be confused with a Quaternary formation of the same name.

Table modified after Beuf *et al.* (1971)
Bellini & Massa (1980)
and Banerjee (1980)

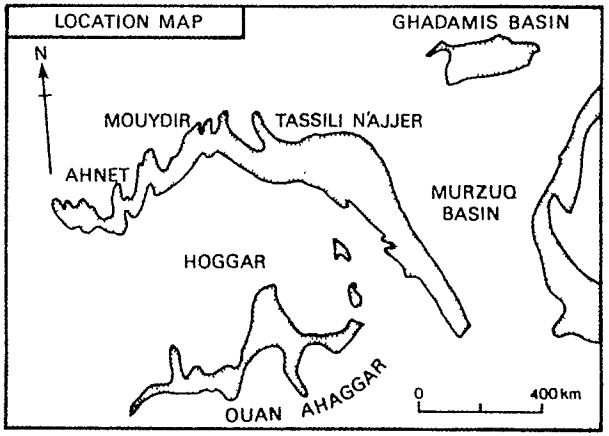


Table 1.3 STRATIGRAPHIC CORRELATION CHART FOR CAMBRIAN TO LOWER DEVONIAN STRATA OF THE CENTRAL SAHARA

The type sections for the four formations of the Cambro-Ordovician Gargaf Group lie in the Qarqaf region, after which the group is named. They have only been briefly studied in the Ghat region and the lower part of the Hassaouna Formation, lying across the Libyan border in Algeria has not been measured. The Hassaouna Formation comprises at least 120 m of medium and coarse-grained cross-stratified sandstones with *Tigillites* (vertical burrow trace fossils, see Chapter 2) in the uppermost 30 m of the formation. Some coarse beds of sandstone with granules are present (Bellini & Massa, 1980). The overlying Haouaz Formation is 130 to 160 m thick and is composed of well-bedded sandstones containing *Tigillites* interbedded with thin argillaceous siltstones (*ibid.*; BEICIP, 1972). These two formations are the lithostratigraphic equivalents of the Ajjers and the younger In Tahouite Formations of Algeria which were interpreted as being of fluvial and transitional-marine origin by Beuf *et al.* (1971). The Haouaz Formation is overlain by the Melez Chograne Formation which comprises up to 80 m of green shales, coarse-grained sandstone beds with granules, and calcareous lenses. A haematitic sandstone bed lies at the base of the formation. The Memouniat Formation is 60-100 m thick in the Ghat region and lies above a strongly erosive lower bounding surface which, in places, completely truncates the Melez Chograne Formation. It is composed primarily of fine-grained sandstones, but coarse-grained and conglomeratic sandstones are present (Bellini & Massa, 1980; BEICIP, 1972; personal observation). Ashgillian brachiopods (*Hirnantia* and *Plectothyrella*) have been recorded from the Memouinat Formation and a Caradocian fauna of orthids, trilobites, cystoids, bryozoans and brachiopods was recorded from the Melez Chograne Formation from samples in the Qarqaf and Ghat regions (Havlicek & Massa, 1973). The age of the underlying Hassaouna and Haouaz formations, however, has not been determined directly in this region. The presence of striations in the Memouniat Formation in the Ghat region has been used as evidence for a glacial origin for this formation (Bellini & Massa, 1980; Beuf *et al.*, 1971). BEICIP (1972) undertook further work on the Memouniat Formation north of Ghat and recognised a number of features typical of glacigene deposits.

The Silurian succession in the Ghat region is composed of graptolitic shales which pass upwards gradationally into a well-bedded

succession of bioturbated and rippled sandstones, siltstones and shales which are overlain by thick coarser-grained and cross-stratified sandstones. The shales crop out in the Wadi Tanezzuft and are impressively exposed at Jabal Idinen and at the base of the escarpment of the Jabal Akakus (Fig. 1.12). Desio (1936a, 1940a) measured a type section in the Silurian succession of the area lying to the east of Jabal Idinen and divided it into a lower shaly Tanezzuft Formation and an upper sandy Acacus Formation. Freulon (1964) carried out further work on the succession and Klitzsch (1965a, 1969) established new type sections, measured at Ti'n Alkum and Bab Al Takarkhouri, in which the base and top of the succession are exposed. These new type sections are now accepted as replacing the earlier incomplete type sections (Banerjee, 1980). Massa and Beltrandi (1975), and Massa and Jaeger (1971) showed, on the basis of graptolite faunas, that the Silurian succession of western Libya comprises a prograding sequence in which the sandstones lying above the shales of the Tanezzuft Formation are progressively younger northwards. In the Ghat Region the Tanezzuft Formation is lower Llandoverian in age, whereas to the north, in the Ghadamis Basin, it is Llandoverian to Ludlovian in age. During field work carried out with the author, Martindale (1977) interpreted the sediments of the Silurian succession as the deposits of a prograding delta system which passed upwards into braid-plain deposits. Palaeocurrent readings displayed a bipolar distribution aligned northwest to southeast. It was concluded, from somewhat limited data, that the predominant direction of delta progradation may have been to the northwest, as had been previously suggested by Berry and Boucot (1973a).

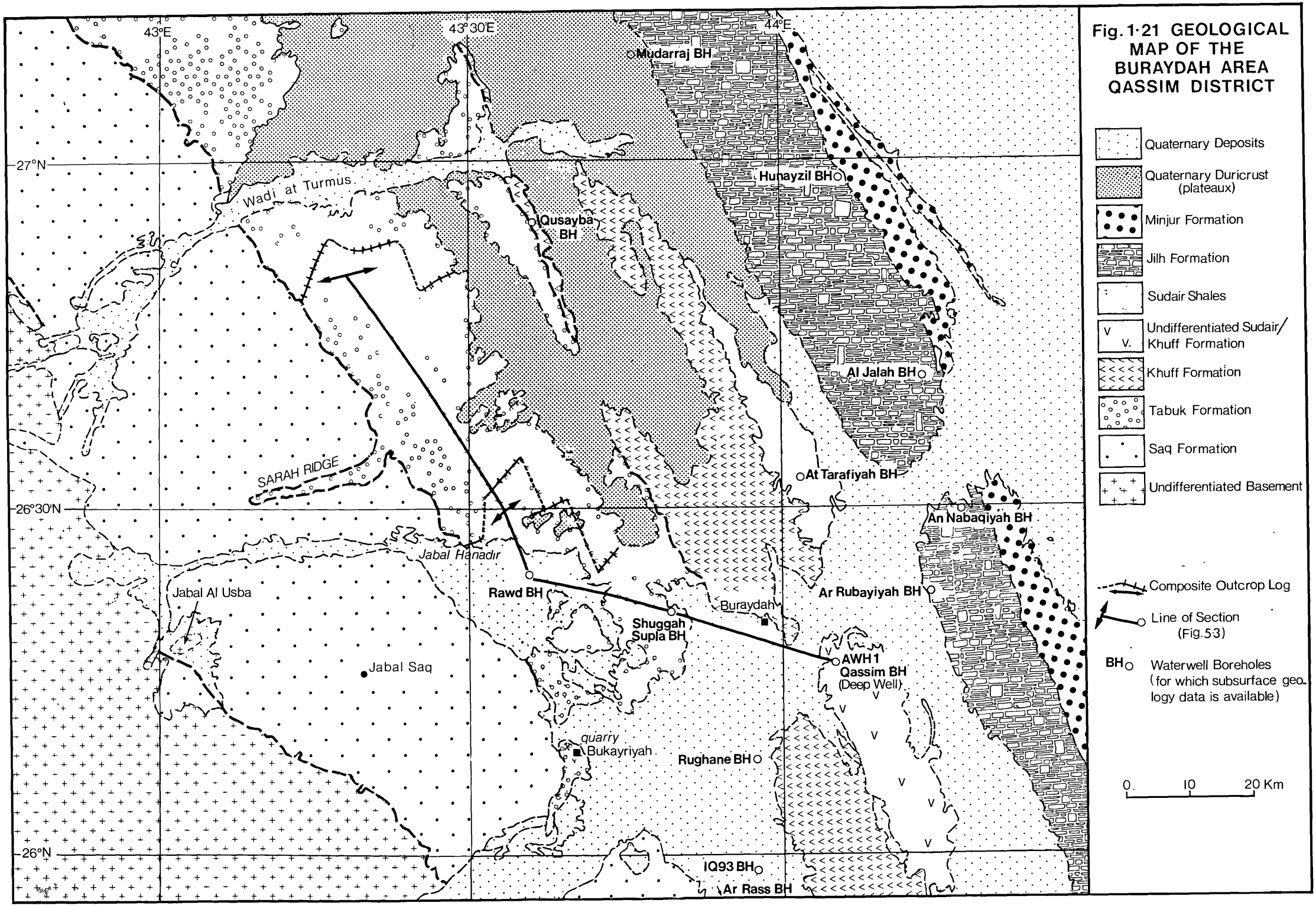
Towards the top of the succession of the Acacus Formation a thick sandstone horizon (135 m thick in the type section) is present which weathers to a major cliff at Jabal Idinen and in the Jabal Akakus escarpment (Fig. 1.12). This sandstone horizon has many of the same characteristics as the sandstones of the overlying Tadrart Formation. The base of this horizon, therefore, may be treated as the base of the Silurian to Lower Devonian blanket sandstone of the area.

The Devonian succession of the study area is exposed in the Jabals Akakus and Tadrart and in the low-lying region to the east (Fig. 1.9). The succession is divided into the Tadrart Formation

Jabal Akakus, viewed towards the north from Wadi Tanezzuft (Fig. 1.7). The straight escarpment, which is up to 300 m high, suggests fault control.

Jabal Idinen, viewed here from the southeast, is an outlier of sandstones of the Acacus Formation lying above shales of the Tanezzuft Formation in Wadi Tanezzuft (Fig. 1.7). It is also named Qaf Ajnoun (the Devil's Mountain).

Figure 1.12 JABAL AKAKUS AND JABAL IDINEN : GEOMORPHOLOGY


43°E
43°30'E
44°E
27°N
26°30'N
26°N
Mudarraj BH
Hunayzil BH
Qusayba BH
Wadi at Turmus
SARAH RIDGE
Al Jalah BH
At Tarafiyah BH
An Nabaqiyah BH
Jabal Hanadir
Jabal Al Usba
Rawd BH
Buraydah
Ar Rubayiyah BH
Shuggah Supla BH
AWH1 Qassim BH (Deep Well)
Jabal Saq
quarry
Bukayriyah
Rughane BH
IQ93 BH
Ar Rass BH
Fig. 1·21 GEOLOGICAL MAP OF THE BURAYDAH AREA QASSIM DISTRICT
Quaternary Deposits
Quaternary Duricrust (plateaux)
Minjur Formation
Jilh Formation
Sudair Shales
Undifferentiated Sudair/ Khuff Formation
Khuff Formation
Tabuk Formation
Saq Formation
Undifferentiated Basement
Composite Outcrop Log
Line of Section (Fig.5·3)
BH Waterwell Boreholes (for which subsurface geology data is available)
0
10
20 Km

(meaning "the formation of the mountain massif" in Tamahak, the language of the Touaregs), the Ouan Kasa Formation of Eifelian (Middle Devonian) age and the Aouinet Ouenine Formation of Givetian to Frasnian (Middle and Upper Devonian) age (Fig. 1.11). The sediments of the Tadrart Formation are exposed on the dip slope of the escarpment of the Jabals Akakus and Tadrart (Fig. 1.10) where they have been carved by the agents of desert weathering and erosion into impressive "ruiniform" stacks, rounded buttes, natural rock arches and rugged jabals (Figs. 1.13 & 1.14). The formation is composed of fine- to medium-grained sandstones with subordinate coarse-grained sandstones, conglomerates and siltstones and is 317 m thick in the type section (Klitzsch, 1965a, 1969). In the lower part it is thick-bedded and cross-stratified, and contains rare plant remains, whereas in the upper part it is thinly bedded and displays evidence of considerable biogenic activity. Klitzsch (1969) considered the lower part to be continental to subcontinental and the upper part marine, and Beuf *et al.* (1971), although only having detailed knowledge of the Jabal Tadrart in the far south, broadly supported this interpretation by assigning the lower part to their fluviatile facies and the upper part to their transitional-marine facies. Burollet (1967) argued, on the basis of the lateral extent of the sandstones and on textural considerations, however, that these sandstones should be interpreted as shallow marine, the coarse-grained sandstones having been shed from uplifted arches into adjacent subsiding basins. The unimodality of the palaeocurrents were interpreted as the result of the activity of strong waves in a shallow sea.

The earliest scientific papers on the Tadrart Formation were those of Desio (1936b, 1940b, 1942), Lelubre (1952) and Freulon (1953b). Later work was undertaken by Furst and Klitzsch (1963), Burollet (1967), Burollet and Manderscheid (1967) and BEICIP (1972). The first type section of the formation was measured in the Jabal Tadrart (25°0'N. - 10°30'E., Burollet, 1960) (Fig. 1.9). A more complete type section, measured further to the south, soon superseded this, however (Klitzsch, 1965a, 1969). The age of the formation at the type section has not been determined directly since studies of the very sparse fauna and flora of the sediments have yielded no evidence on dating. Indirect evidence for a Siegenian/Emsian age rests upon its stratigraphic position and the ages of lithostrati-

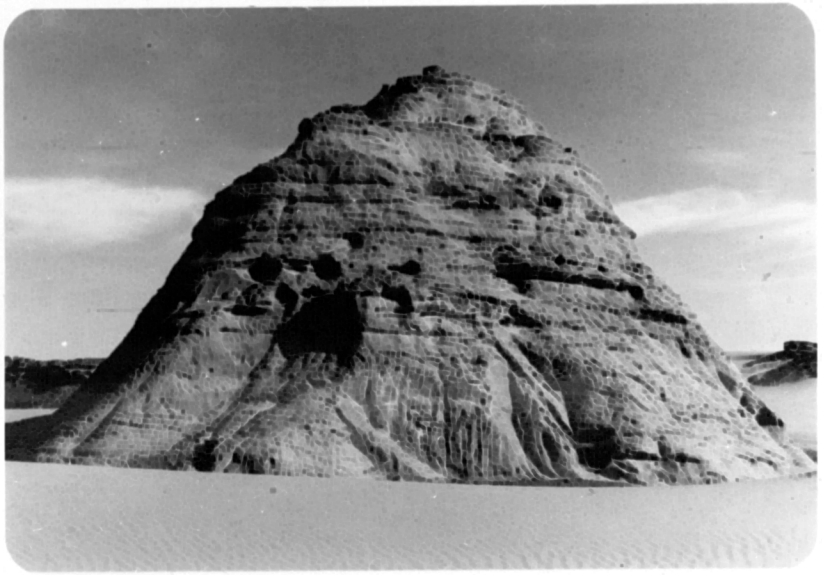
"Ruiniform" weathering of Tadrart Formation sandstones, 43-100 m stratigraphically above the base of the formation, Wadi Anshal (Fig. 1.7).

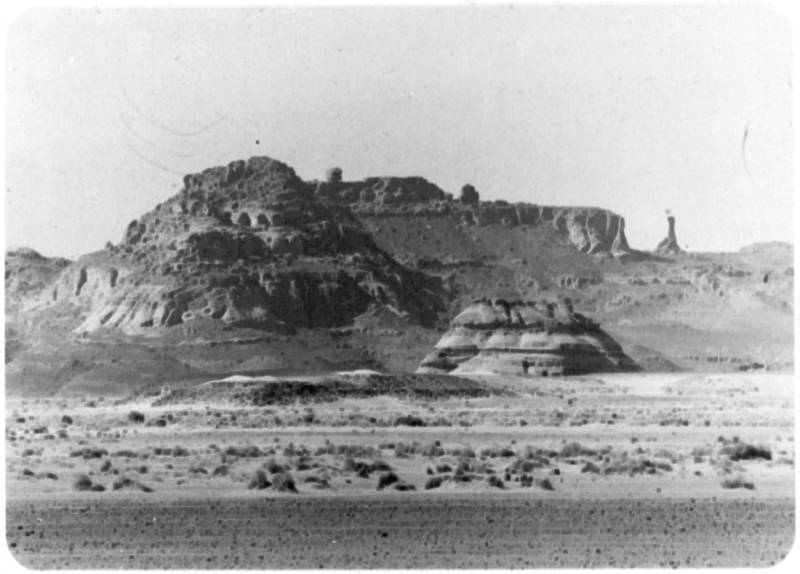
Rock stack developed in the sandstones of the Tadrart Formation, 90-125 m stratigraphically above the base of the formation, Wadi Talauaut (Fig. 1.7).

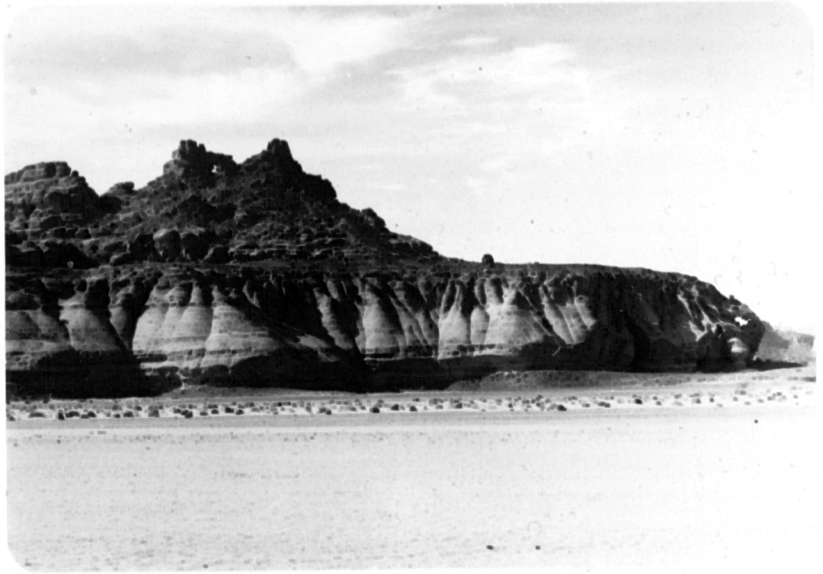
Figure 1.13 JABAL TADRART : "RUINIFORM" WEATHERING

Figure 1.14 JABAL TADRART : GEOMORPHOLOGY

Mesas, buttes, rock stacks and bluffs (slab walls); typical weathering forms of the Tadrart Formation







graphical equivalents to the north in the Ghadamis Basin and to the west and northwest in Algeria and Tunisia. These equivalents, however, are recognised by the coarse-grained character of their sandstones and there is evidence not only that the upper boundary to this lithostratigraphical unit is diachronous, but also that the first medium- and coarse-grained sandstones in the Saharan Silurian/Lower Devonian successions were deposited earlier in the southwest than in the northeast (Beuf *et al.*, 1971, p. 103). Nonetheless, the evidence of lateral interfingering of marine sediments with coarse-grained sandstones in Algeria and Tunisia provides some support for a Siegenian/Emsian age for the Tadrart Formation in Southwest Libya. Upper Siegenian/lower Emsian faunas are present in marine limestones which have been shown to be the lateral facies equivalents of the upper part of the Tadrart Formation at Dahar in the extreme south of Tunisia (Burollet & Manderscheid, 1967). Furthermore, a late Siegenian/early Emsian micro-flora was obtained from the sandstones of the Tadrart Formation in the subsurface of the Ghadamis Basin (Massa & Moreau-Benoit, 1976). In southeast Algeria the External Tassilian Sandstones are thought to be Siegenian in age on the basis of a "few scarce fossils" reported close to the top of the formation by Dubois *et al.* (1969) and also on the basis of a lateral facies change westwards to finer-grained marine sediments in which Siegenian faunas are present (Legrand, 1967). Klitzsch (1969) argued that these External Tassilian Sandstones are Siegenian and Emsian in age on the basis of work carried out on brachiopods collected from these sandstones (Massa & Manderscheid, pers. comm. to Klitzsch; Claracq *et al.*, 1963). The External Tassilian Sandstones pass laterally northwestwards in Algeria into successions with marine sediments of Gedinnian/Siegenian age (Beuf *et al.*, 1971, pp. 160-161).

The Palaeozoic succession above the Tadrart Formation, cropping out east and northeast of Jabal Tadrart in the Ghat region (Fig. 1.9) is less well exposed than the Cambrian to Lower Devonian succession and will be described here only briefly. East of the Jabal Tadrart, the Middle and Upper Devonian succession is exposed in only three localities: east of Wadi Auis, Wadi Wan Kasa and Gour Iduka (Fig. 1.9). The Ouan Kasa Formation is present at all three localities, whereas the Aouinet Ouenine Formation is only present at the southernmost locality, i.e. at Gour Iduka. The most well exposed section of

the Ouan Kasa Formation is at Wadi Wan Kasa where a 43 m thick succession of shales and subordinate sandstones and siltstones was first measured by Borghi and Chiesa (1940) and was later established as the type section by Klitzsch (1965a, 1969) who dated these sediments as Middle Devonian (Eifelian) on the basis of their abundant spirifers. The Aouinet Ouenine Formation, at Gour Iduka, is composed of a succession, at least 107 m thick, of very fine to medium-grained sandstones with some thin argillaceous siltstone horizons, which contains a rich fauna of brachiopods and bivalves in parts of the succession (Bellini & Massa, 1980). The Aouinet Ouenine Formation is overlain unconformably by oolitic ironstones, shales and sandstones of the "Chatti Formation" and by Carboniferous marine shales with subordinate siltstones, sandstones and limestones (the Mrar, Assedjefar and Dembaba Formations). A terrestrial red bed succession of shale and sandstone (the Tiguentourine Formation) lies at the top of the succession. Much of this Permo-Carboniferous succession crops out beneath the sandsea of the Idhan Wan Kasa, but where it is exposed it forms low-lying escarpments revealing some gentle folding in the area to the east of Al Awaynat (Collomb & Heller, 1958) (Fig. 1.9).

1.3.4 Geomorphology

In the Mesozoic of Southwest Libya, the Messak Sandstone (Upper Jurassic-Lower Cretaceous, Fig. 1.5) displays many of the same characteristics as the Nubian-type sandstones of the Tadrart Formation (Lorenz, 1980). These two Nubian-type blanket sandstones have weathered to form easterly-dipping escarpments which dominate the scenery of the region (Fig. 1.15, cf. Frontispiece) and between which less resistant beds have weathered to a lower topographic level and aeolian sands have accumulated as sand seas. The Landsat image (Fig. 1.15) reveals evidence of four different types of drainage pattern (cf. Fig. 1.7): a well developed dendritic pattern of wadis draining towards the east and northeast off both escarpments; a poorly developed irregular drainage pattern of wadis draining to the west, off the Jabal Akakus into the Wadi Tanezzuft; the northward draining Wadi Tanezzuft; and lastly, an ancient drainage pattern of large rivers draining from the Hoggar Massif towards the east and northeast which is thought to be responsible for the sand-filled wadis which

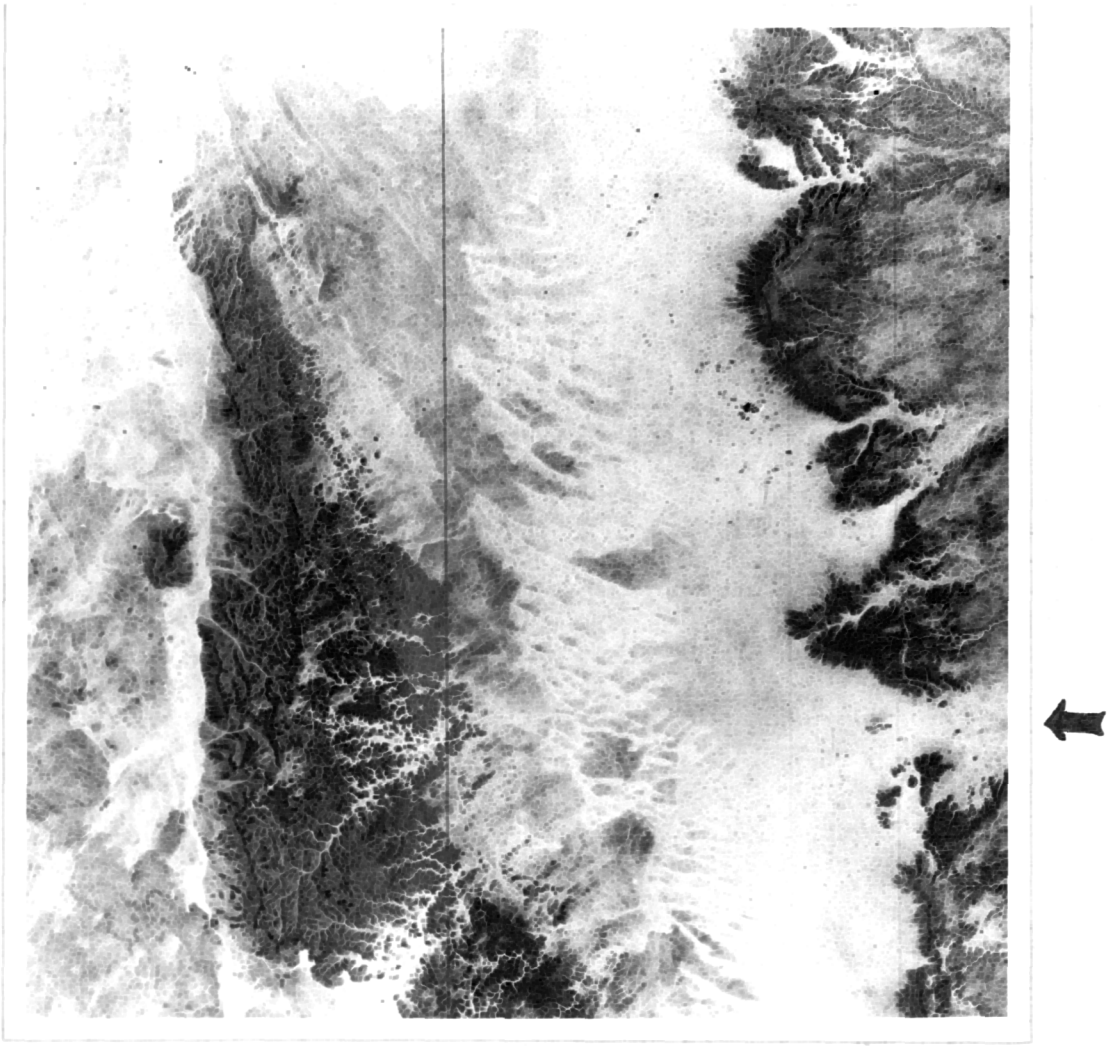


Figure 1.15 JABAL AKAKUS, JABAL TADRART AND MSAK MALLAT : LANDSAT IMAGE.

Wadi systems indicating eastward and westward drainage patterns are visible in the Jabals Akakus and Tadrart. The most pronounced northeastwards-draining wadis are Wadi Teshuinat and Wadi Takarkhouri. To the east, the Msak Mallat is also dissected by a number of major wadis. The one in the south, named Tehi-n-Tilemsin and marked by an arrow, is thought to have been part of the same drainage system as the Wadi Takarkhouri.

break the escarpments, e.g. Wadi Takarkhouri and Tehi-n-Tilemsin in the Msak Mallat (Busche, 1980; Barth & Blume, 1975). The very flat top to the Msak Mallat escarpment has been attributed to a Tertiary peneplain (at 1000 to 1250 m above mean sea level) which has subsequently been dissected. Busche (1980) argued that the Msak Mallat escarpment was developed through a process of differential vertical erosion under non-arid conditions rather than by slope retreat under conditions of arid pedimentation, as is usually assumed for this type of landform. The evidence for this is provided by the discovery of weathering features on the Tertiary peneplain surface which indicate that chemical weathering typical of hot humid climates occurred in the late Tertiary. The development of escarpments typical of the present arid weathering cycle can only have occurred in the relatively recent past therefore. The absence of a rectilinear drainage pattern on the dip slopes of the present escarpments suggests that the channels were not controlled by the underlying sandstone but rather by a soil covered slope (*ibid.*). It is considered probable, therefore, that the development of the present scenery took place as a result of differential weathering, in the late Tertiary or later, of a peneplain across which major rivers flowed from west to east under conditions which were initially moderately humid but which became increasingly arid. As the softer beds weathered more rapidly, valleys formed along the strike of the argillaceous Palaeozoic strata subsequent to the main drainage system (e.g. Wadi Wan Kasa and Wadi Tanezzuft). The general lack of truncated headwaters in the dendritic drainage systems of the escarpment dip slopes suggests that little slope retreat has occurred since their development.

Busche (1980) argued, therefore, that the wadis which cut the major escarpment of the Jabals Akakus and Tadrart and which drain to the east (Wadi Takarkhouri and Wadi Arrikin/Wadi Takisset) are not examples of drainage capture as suggested by Pesce (1968) but are evidence of a drainage pattern which existed on a late Tertiary peneplain prior to the development of the escarpment.

Whereas the Msak Mallat escarpment displays a rounded and irregular, cuesta-like character in plan, the trace of the scarp slope of the Jabals Tadrart and Akakus is much more regular, with remarkably straight cliffs which are progressively displaced to the west southwards along the escarpment (Fig. 1.9). It has been

suggested that this escarpment may be fault controlled (e.g. Collomb & Heller, 1958; Klitzsch, 1969). No direct evidence may be assembled to show that a fault or a series of faults are present parallel to the escarpment on the east of the Wadi Tanezzuft, but the character of the escarpment is clearly very different from that of the Msak Mallat. Like the Msak Mallat the remnants of a Tertiary erosion surface which dips gently to the north and east, in this case at 1400-1500 m above mean sea level (Fig. 1.8), can be seen between the deep wadis (Fig. 1.16). This surface dips eastwards towards the Msak Mallat at an angle that is very much less than the regional eastward dip of the strata. Consequently, it follows that during the Tertiary the resistant sandstone formations wedged out westwards beneath this surface to a feather edge . Later differential erosion of such a feather edge of resistant sandstone might be expected to produce a somewhat irregular topographic feature. However, if the sandstone did not wedge out to a feather edge because of displacement along a fault, then subsequent down-weathering would result in the exposure of a fault escarpment (Fig. 1.17). Collomb and Heller (1958) mapped a major fault aligned approximately north-south and throwing basinwards, i.e. towards the east, in the Al Awaynat area and it seems quite reasonable to propose that an en-echelon southwards extension of this fault system might have been responsible for the geomorphology of the Jabal Akakus (Figs. 1.9 & 1.15). The Jabal Idinen outlier may have resulted from the weathering and erosion of a part of the ancient fault line where displacement was absent or at a minimum. The absence of similar fault controls on the Msak Mallat escarpment may be an indication that the faults mapped in the Ghat Region are late Palaeozoic or early Mesozoic in age.

Barth and Blume (1975) showed that the character of the escarpments around the Murzuq Basin was determined to a large extent by the lithological contrast at the feather-edge of the resistant sandstone. Where the resistant sandstone lay above a relatively pure shale, differential erosion would be very effective in producing a steep escarpment. This situation applies to both the escarpments of the Jabal Akakus and the Msak Mallat, although the clean shales of the Tanezzuft Formation may be slightly less resistant than those of the Early Mesozoic Tilemsin Formation. Nevertheless, the two escarpments are not only different in shape but vary in the way that they have

View of Jabal Akakus escarpment.

View towards the northeast showing the Idhan Idinen and Jabal Idinen (far distance).

Both photographs illustrate a widely developed Tertiary peneplain surface.

Figure 1.16 AERIAL VIEWS OF JABALS TADRART AND AKAKUS

Figure 1.17 GEOMORPHOLOGICAL EVOLUTION OF THE JABAL AKAKUS ESCARPMENT

(a) Mesozoic tilting and normal faulting of the western margin of the Murzuq Basin.

(b) Late Tertiary uplift and peneplanation.

(c) Differential weathering, down-wearing and erosion during the Quaternary.

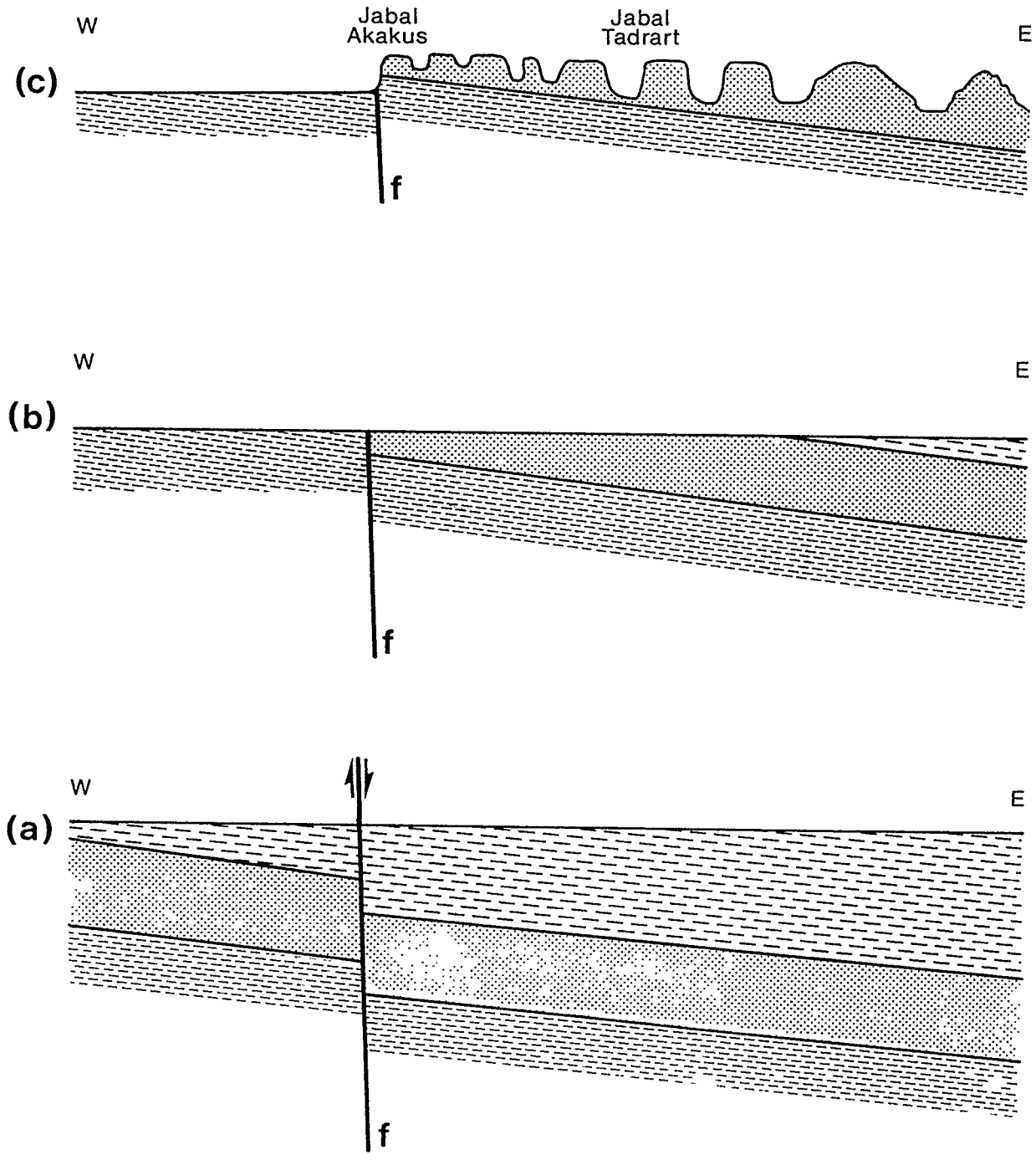

W
Jabal Akakus
Jabal Tadrart
E
(c)
f
W
E
(b)
f
W
E
(a)
f
External Tassilian Sandstones
Sltst, sh & (sst); Intra-Tassilian Succession
Pre-Tassilian Succession (post Lower Devonian)

Figure 1.17

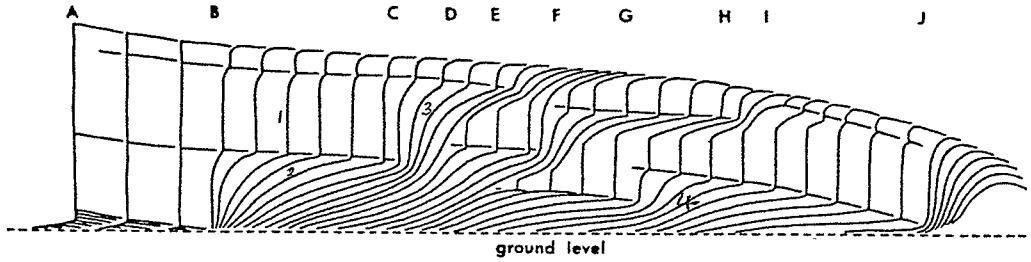
weathered since they became topographic features. Whereas the Msak Mallat escarpment is dominated by landslides (Grunert & Busche, 1980) the Jabal Akakus has been dissected by short wadis draining westwards into the Wadi Tanezzuft. This young drainage system may well have developed by the exploitation of joints and lines of weakness in an exposed cliff line of sandstone, probably under arid conditions, and the lack of a thin feather-edge of sandstone may explain the absence of landslides. Weathering and erosion occur by progressive back-wearing of an escarpment in arid conditions where thick, homogeneous and relatively poorly-cemented sandstone units lie above thinly-bedded horizons containing shale partings (e.g. southeastern Utah, Koons, 1955; Oberlander, 1980). This occurs by the undermining of the massive sandstone unit along the less resistant horizons in the thin-bedded strata, and the calving of sandstone flakes away from the cliff face which collapse down the cliff and generally break up into sand over a short period of time.

The sandstones of the Tadrart Formation which are dissected by major wadi systems display a variety of geomorphological forms which may be understood in terms of arid weathering processes which must have been initiated in latest Tertiary times. These processes are active at present in southeast Utah, and the scenery of Canyonlands, where the Jurassic Entrada Formation is exposed (Oberlander, 1980), bears a striking similarity to the dissected plateau of the Jabal Tadrart (Fig. 1.18). It seems likely therefore that, whereas the drainage pattern which had been inherited by downcutting was initiated in a more humid climate, the present wadi margin scarps, slopes, bluffs, rock arches and stacks were developed under fairly arid conditions. Following the terminology of Oberlander (*ibid.*) the two main geomorphological forms into which the massive sandstones of the Tadrart Formation are weathered may be classified as "slick-rock slopes" and "slab walls". Convex and concave-shaped slopes are termed "slick-rock slopes" after the Slick Rock Member of the Entrada Formation. The way in which the slopes and walls are related to thin-bedded (and less resistant) horizons and to intraformational partings is illustrated in Figure 1.19. In simple terms thinly-bedded intercalations commonly cause a break within the escarpment with slab walls developed above, and slick-rock slopes below. However, slick-rock walls may develop by the downward weathering of

Wadi Teshuinat, viewed towards the east (Fig. 1.7). The photograph shows the basal 40 m sandstone bluff of the Tadrart Formation (arrows), and overlying slopes and slab walls.

Basal 40 m unit of the Tadrart Formation, Wadi Auis (Fig. 1.7). Slick rock slopes are well developed in the foreground whilst slab walls predominate on the far side of the wadi. A well-developed plateau is present above the slab walls, stratigraphically 40 m above the base of the formation.

Figure 1.18 THE TADRART FORMATION IN WADIS AUIS AND TESHUINAT



Key: 1 - slab wall; 2 - slick-rock slope; 3 - slick-rock ramp, and 4 - concave slope break.

Development of escarpment forms in massive sandstone through time and space, after study of segmented cliffs in massive sandstones of Southeastern Utah.

For simplicity a constant ground level (dashed line) is assumed during escarpment retreat, along with equal thicknesses of removal from major slab walls in each unit of time. At A a through-going cliff is present due to sapping above a thin-bedded substrate. At B the thin-bedded substrate passes below ground level, escarpment retreat slows, and effective intraformational partings assume control of escarpment form. Effective partings close at C, E, F, G, H and J, leading to local slab wall stagnation and rounding into slick-rock slopes. Partings that open at D, E, F, G, and I initiate growth of new slab walls. Note that the effect of former partings in rocks that have been removed continues to be expressed in the form of slick-rock ramps and concave slope breaks.

Figure 1.19 THE EVOLUTION OF SLICK-ROCK SLOPES AND SLAB WALLS THROUGH ESCARPMENT RETREAT, AFTER OBERLANDER (1980)

a slick-rock slope, beneath an intraformational parting, by water runoff from higher levels. This has the effect of creating a slope-wall break which provides no indication of the presence of an argillaceous horizon. These horizontal partings may be as thin as 10 cm to be effective in causing an undermining of the overlying sandstone and its consequent collapse.

In the Wadi Auis (Fig. 1.18) slick-rock slopes are developed where the wadi floor lies above the thin-bedded sandstones and siltstones of the Acacus Formation. Where these less resistant beds are exposed water-sapping and erosion has caused collapse and slab walls are developed. Collapse alcoves are frequently present at the base of these slab walls, formed by small-scale detachment and collapse of a few cubic metres of rock above a thin-bedded horizon. These alcoves act as a precursor to large-scale slab wall stoping. The morphology of the lowest weathered sandstones of the Tadrart Formation in Wadi Auis and Wadi Teshuinat suggests that slick-rock slope forms dominated initially and that down-cutting of the wadi floor led to the exposure of the thin-bedded upper beds of the Acacus Formation which allowed water-sapping, the undercutting of the slick-rock slopes and the development of slab walls by collapse of material which was subsequently weathered to sand.

The Entrada Formation study (Oberlander, 1980) drew attention to the fact that a massive sandstone may weather to a slick-rock slope or a slab wall depending on geomorphological controls, and that the form developed may have little relationship to lithology or cementation. Used with caution, however, slope breaks and tableland plateaux were found to be useful in the study of the Tadrart Formation for identifying thin horizons of finer-grained sandstone and argillaceous siltstone within the succession and for tracing their lateral extent.

Plateaux are well developed at different levels within the wadis of the Jabal Tadrart (Fig. 1.18). They may have developed in a number of different ways and great care must be taken before assuming that they correlate with lithological boundaries or disconformity surfaces within the stratigraphic succession. Barth and Blume (1975) showed that not only is there a Tertiary peneplain developed at 1400-1500 m in the Jabals Akakus and Tadrart, but that a series of terraces at lower levels are related to pauses in down-wearing as the old peneplain was dissected and the initiation of a new period of lateral

weathering and erosion occurred. This process was never brought to completion, however, since a new period of uplift and dissection ensued before sufficient time had elapsed for complete peneplanation to have occurred. Strongly developed plateaux are present, therefore, which are subparallel to the bedding. This is not always the case, however, and it seems likely that, during the later periods of dissection of the Jabal Tadrart, peneplanation would have taken place by back-wearing of escarpments. Given the very gentle dip of the strata, the back-wearing process would have been controlled to some extent by lithological breaks within the succession, as discussed above.

The abundant rock paintings and engravings found in collapse alcoves and on overhanging slab walls, or shelters as they are usually termed in the Sahara, of Savannahland hunting scenes (Mori, 1965), are evidence that the climate was more humid only a few thousand years ago. Such a climate, however, is quite consistent with the arid weathering processes active at present in Utah and is more appropriate for water sapping of thinly-bedded horizons.

1.4 The Geology and Geomorphology of the Study Area, Northwest Saudi Arabia

1.4.1 General

A far greater literature exists on the Palaeozoic of Southwest Libya than on that of Northwest Saudi Arabia. Geological research in Southwest Libya was dominated by the studies of Italian and French geologists working prior to Libyan independence (1951) and, thereafter, by exploration related to the search for hydrocarbons. By contrast, the geological exploration of the Palaeozoic succession in Saudi Arabia was largely undertaken in the 1950's and 1960's by German geologists, the geologists of the United States Geological Survey (Powers *et al.*, 1966, 1968) and the Arabian American Oil Company (Aramco). Unlike the study of the Palaeozoic of the Murzuq Basin, the author did not have access to unpublished oil company reports in Saudi Arabia. Consequently the only major available source material on the Palaeozoic of the study area at the time the writer's field work was undertaken was that of Powers *et al.* (1966, 1968) who quoted

extensively from the work of Helal (1964a, 1965). Since then Wolfart (1981) has published a major review of the Lower Palaeozoic of the Middle East.

Despite the limited geological literature available on the Palaeozoic of Northwest Saudi Arabia, relatively recent geological and sedimentological studies on the Lower Palaeozoic of Southern Jordan and Southern Israel (Table 1.4) provided valuable insight into their lithostratigraphical equivalents in Saudi Arabia (Selley, 1970, 1972, 1978; Sneh & Weissbrod, 1978). A seminal paper on the nature of Nubian-type sandstones in Arabia is that of Picard (1943).

Geological field work was undertaken in Northwest Saudi Arabia, in the Qassim District, just northwest of Buraydah (Fig. 1.20). A short visit was also made to Al Mayyah, north of the main study area, to examine outcropping Lower Devonian Sandstones. The main study area (Fig. 1.21) is bounded to the north by the Wadi at Turmus (just south of $27^{o}00'$ latitude) and to the west by the contact of the Saq Formation with the basement rocks of the Arabian Shield. In the east, the boundary follows the basal unconformity of the Permian Khuff Formation which cuts across the Tabuk Formation; while to the south the area is bounded by a line which runs east-west through Buraydah (approximately $26^{o}20'$ latitude). This area provides good exposure and nearly complete sections through the Lower Palaeozoic succession. It includes two topographic features, Jabal Saq and the Sarah Ridge, which have provided important sedimentological data (Figs. 1.21 & 1.22). Some geological data are available from the logs of a number of water supply boreholes (Fig. 1.21).

1.4.2 Stratigraphy, Structure and Geomorphology

The Lower Palaeozoic succession of Northwest Saudi Arabia has been divided into two formations, the Saq and Tabuk formations (400 m and 610 m thick respectively in their type sections; Powers *et al.*, 1966) (Table 1.4), the sediments of which are well exposed in the study area in northwest-southeast striking escarpments (Figs. 1.21 & 1.22). The succession of siliciclastic sediments which comprises the Saq and Tabuk formations rests on the Arabian Shield with very strong angular discordance, and is itself unconformably overlain by the Permian Khuff Formation (Fig. 1.21). Stratum contours, based on the logs of 18 boreholes in the southeast part of the Qassim District

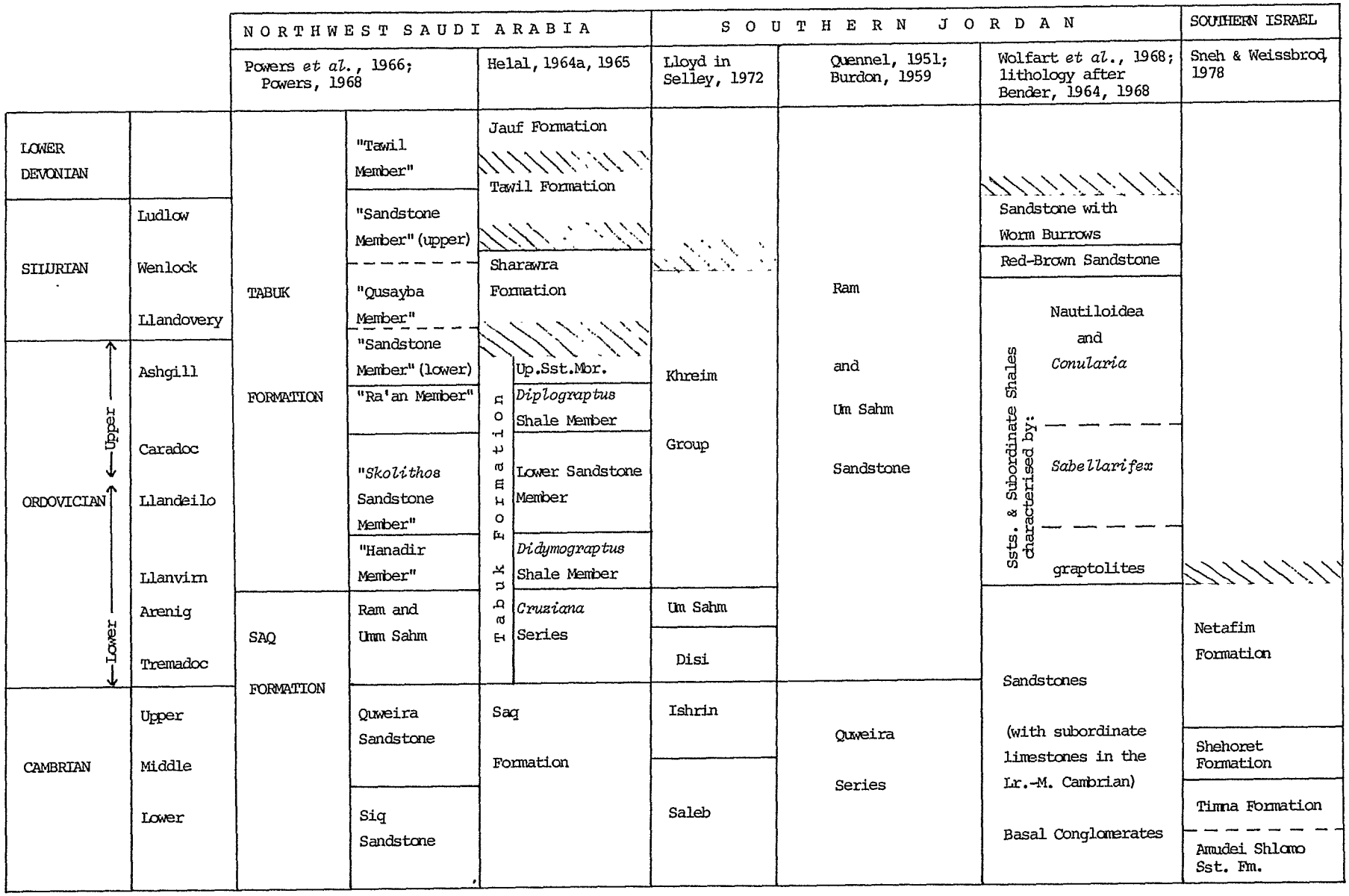


Table 1.4 STRATIGRAPHIC CORRELATION CHART FOR CAMBRIAN TO LOWER DEVONIAN STRATA OF NORTHWEST ARABIA

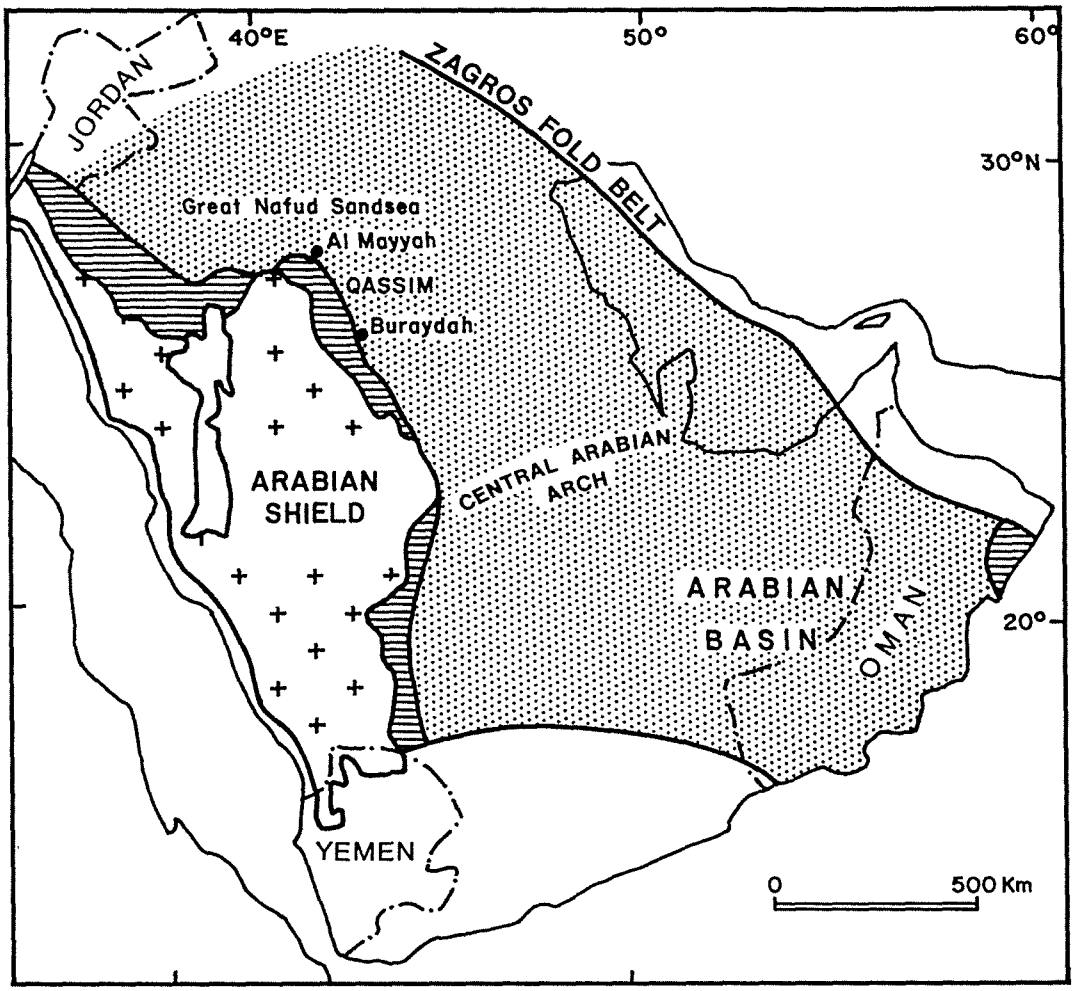


Figure 1·20 LOCATION MAP OF ARABIA

Figure 1.22 LANDSAT FALSE-COLOUR IMAGES OF NORTH CENTRAL SAUDI ARABIA AND OF STUDY AREA

The vertical margins of the photographs are orientated towards the north northeast.

Upper : in the north of the image Palaeozoic strata are hidden beneath the sands of the Great Nafud Sandsea. In the southwest, rocks of the Precambrian Shield are exposed. The area studied lies in the southeast where the east-west trending Sarah Ridge may be seen.

Lower : detail of study area. The western boundary of the Tabuk Formation is clearly visible as a northwest-southeast aligned escarpment. The boundary between the Saq Formation and the Precambrian Shield is discernible as a marked colour change. The Sarah Ridge and the outcrop of the"Sarah Member"to the east of the Tabuk escarpment may also be seen.

A fossil dendritic pattern visible in the south, on the Precambrian Shield, provides evidence of easterly and northeasterly drainage associated with the Wadi Ar Rumah lineament.

(Parsons-Basil, 1968), drawn on the Precambrian basement and on the base of the Tabuk Formation, indicate a regional dip to the northwest of less than one degree. The Tabuk Formation decreases in thickness southwards from Qusayba, as a result of erosion prior to deposition of the Khuff Formation (Fig. 1.21).

The Lower Palaeozoic succession has many similarities with that of the Central Sahara comprising Nubian-type fluvial, shallow-marine and probable fluvio-glacial facies as well as deposits of deltaic origin (see Figs. 1.6 & 1.11). The Cambrian and Lower Ordovician Saq Formation is composed predominantly of cross-stratified sandstones, similar to those of the Hassaouna and Tadrart formations of Libya (Fig. 1.11), whilst the succeeding Lower Ordovician to Lower Devonian Tabuk Formation, composed of sandstones, siltstones and subordinate shales and conglomerates, provides evidence of somewhat more marine depositional settings than those of their lithostratigraphical equivalents in Libya. Nevertheless both areas appear to be characterised by late Ordovician glacial or periglacial deposits and by early Silurian graptolitic shales.

The sediments of the Tabuk Formation have been assigned to seven informal members by Powers *et al.* (1966) (Table 1.4) largely on the basis of the recognition of three graptolitic shale members, the "Hanadir" (Llanvirnian), "Ra'an" (probably late Caradocian) and "Qusayba" (Llandoverian) members. The remaining members of this Formation are composed predominantly of sandstones, the "*Skolithos* Sandstone Member" being distinctive by the abundance of vertical burrows. The "Tawil Member" lies at the top of the succession and is composed of cross-stratified sandstones that are very similar to those of the thickly-bedded intervals of the Tadrart Formation (Fig. 1.11).

CHAPTER 2

TECHNIQUES AND TERMINOLOGY EMPLOYED IN FACIES STUDIES

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CHAPTER 2

TECHNIQUES AND TERMINOLOGY EMPLOYED IN FACIES STUDIES

2.1 Introduction

The methods of facies analysis adopted in this thesis have been described by Reading (1978, Chap. 2) and Selley (1978). In the study of Nubian-type sandstones the subjects of palaeocurrent analysis and ichnology have received considerable attention (e.g. Beuf *et al.*, 1971; Selley, 1970, 1972) because of the general lack of variation in the other parameters relevant to facies analysis: sedimentary structures, bed geometry and lithology. The problems associated with the interpretation of palaeocurrent data and ichnology in Palaeozoic Nubian-type sandstones are central themes in any study of these sandstones. Consequently a review of these subjects as they apply to this research is presented below. First, however, some aspects of bed geometry, sedimentary structures and lithology are briefly discussed.

Despite the presence of some steep rock faces in the Jabal Tadrart which provided excellent exposure in two dimensions of bed geometry, in the majority of the sections measured the identification of the type of boundary surface at the base of each bed proved to be impossible to ascertain with confidence. Observations in some well-exposed three-dimensional sections of broad shallow scour horizons encouraged a scepticism about whether these could be distinguished from planar erosion surfaces except in unusually well-exposed sections normal to scour/channel axes. Detailed work on these broad shallow scours suggested that they were far more abundant in the cosets of cross-strata typical of Nubian-type sandstones, and generally described as having a planar geometry, than was previously appreciated.

The terminology employed in describing sedimentary structures is based largely upon the classic descriptions of stratification and cross-stratification of McKee and Weir (1953), Allen (1963), Southard (1975) and Harms (1975a, 1975b). The conventions of symbols and abbreviations used in diagrams largely follow those of Selley (1978) and Shell's Standard Legend (Fig. 2.1)*. Since lithological and grainsize variations within many of the sections measured were limited it was considered reasonable to combine data on lithology, grainsize and sedimentary structures in one column of varying width.

* See copy in pocket at end of Volume II

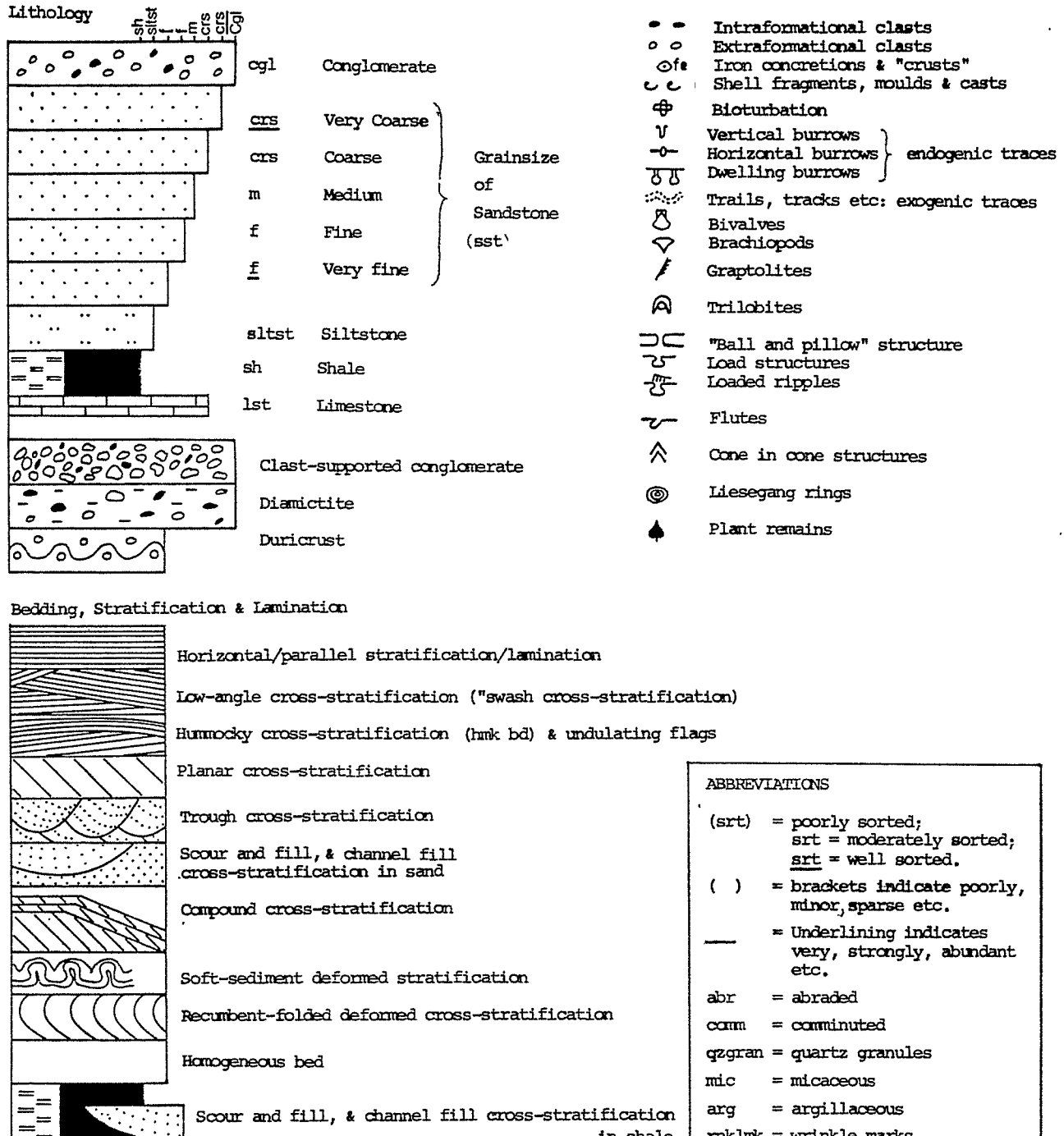


Figure 2.1 KEY TO SYMBOLS AND ABBREVIATIONS USED IN FIGURES

It should be noted that the terms "planar" and "trough" cross-stratification have been used to distinguish sets with flat bounding surfaces from those with curved trough-shaped bounding surfaces. "Tabular" and "wedge" cross-strata are varieties of planar cross-strata. The lower bounding surfaces of sets and beds may be "erosional", if discordance is demonstrable, or "abrupt" if no discordance is visible. Additionally, foresets are described as angular, tangential or concave following Jopling (1965). Lastly scour and fill, and channel fill cross-stratification are distinguished by their lower bounding surface, the former being trough-shaped, the latter cylindrical (Allen, 1963).

Thin section and scanning electron microscope petrographic work was undertaken on samples collected in the field. Although sandstones predominated, other lithologies present in the successions studied were conglomerates, diamictites, siltstones and shales (Fig. 2.1). The classification of these siliciclastic rocks adopted here, is based upon grainsize (Wentworth, 1922, in Blatt *et al.*, 1972) and the percentages of matrix and quartz detrital components (Pettijohn, Potter and Siever, 1973, p. 158, after Dott, 1964). The sandstones studied were all quartz arenites or subarkoses and were remarkable for their uniform and mature mineralogy. The texture and diagenetic components of the samples studied varied considerably however. The term "quartz arenite" is considered preferable to Krynine's term "orthoquartzite" (1941) which came to be used synonymously with "quartz arenite" (Pettijohn, 1957, p. 295), because it is felt that any use of the term "quartzite" should be avoided for the frequently very friable Nubian-type sandstones.

Petrographic studies of Palaeozoic Nubian-type sandstones have been undertaken largely by French geologists (Beuf *et al.*, 1971; Mennig & Vittimberga, 1962; Chauvel & Massa, 1981) who have concentrated upon the graphic presentation of petrographic data in vertical logs as a means of characterising and correlating apparently barren and monotonous siliciclastic successions. The petrographic data recorded for this purpose included the following factors: abundance of mica and heavy minerals, degree of rounding of grains, proportion of grains to cement, grainsize and sorting, and grain composition. These studies illustrated the petrographic variation that exists in Nubian-type sandstones, but nonetheless confirmed the predominance of mineralogically-mature sandstones. Additionally, emphasis has been laid upon the study of iron-rich oolitic sediments which are present at various stratigraphic horizons, but not within the blanket sandstones of the Cambro-Ordovician and the Lower Devonian (Chauvel & Massa, 1981).

2.2 Palaeocurrent Analysis

The use of palaeocurrent indicators in facies analysis is a very important aspect of determining the depositional setting of any suite of rocks under study (Selley, 1978). The construction of a number of palaeocurrent rose diagrams may be used to distinguish environments of broadly unidirectional flow from those of scattered flow azimuths, and those of bimodal current azimuths. Furthermore, the calculation of vector mean azimuths from broadly unidirectional palaeocurrent data provides valuable evidence for the orientation of the palaeoslope (Potter & Pettijohn, 1977). Once a general environment of deposition of the sedimentary rocks under study has been established the palaeocurrent data available should be analysed more critically. Rather than lumping readings together to create rose diagrams, each palaeocurrent reading should be interpreted in the light of the type of stratification and the type of bed from which it was measured, and in the light of its relationship to readings obtained from vertically and laterally adjacent beds. Comparison of the readings obtained from similar types of stratification may also provide evidence which can be used to establish the nature of the subenvironments within the established depositional setting. Attempts have been made to use palaeocurrent analysis to refine established broadly-defined environmental interpretations, invariably with mixed success. For instance Ore (1963) and Smith (1972) both reached different conclusions about how to distinguish braided from meandering fluvial deposits. Banks and Collinson (1974) showed how the palaeocurrent data from preserved braided stream deposits should display a lower variance than those from a modern braided river. Throughout the debate which centred on these papers, no consensus was established about the validity of any particular approach to distinguishing the two environments. On the one hand there exists the need to obtain as many readings as possible to obtain statistically valid results. On the other hand, any resulting rose diagram suffers from comparing unlike with unlike. However, if the palaeocurrent data are used initially to provide an indication of the orientation of the palaeoslope and then to provide additional evidence for the interpretation of each bed in a succession, the palaeocurrent data can play a valuable part in determining the subenvironments present and the processes operating in a depositional

setting. This approach is considered more valuable in refining broad environmental interpretations than a statistical manipulation of a large data set of palaeocurrent readings.

2.3 Ichnology

In a detailed, meticulous and well illustrated piece of work Desio (1940b) described the "vestigia problematica" of Libya and first drew attention to the abundance of trace fossils in the Jabal Tadrart. In his work, many of the structures recorded within the beds and on bed surfaces were recognised as being fossilised "structures within the sediment left by living organisms", i.e. they were trace fossils or fossil lebensspuren (Abel, 1912; Haas, 1954). Desio (1940b), therefore, undertook the earliest study in Libya of palaeoichnology (Buckland, about 1830, in Hantzschel, 1975). In attempting to define which trace fossils occurred together he also initiated a study of trace fossil associations or ichnocoenoses (Davitashvili, 1945; Lessertisseur, 1955) a subject later expanded by Seilacher (1969) after work in the same area. Table 2.1 shows the genera and species Desio (1940b) identified in the Silurian and Devonian successions of southern Libya together with their modern generic names (after Hantzschel, 1975). It will be noted that some of Desio's "vestigia problematica" are of inorganic origin and will not be further discussed here (although it is interesting to note that he saw mudcracks - his *Polygonolites* - in the Lower Devonian succession). Further it should be noted that two of the Middle and Upper Devonian species are no longer considered valid names (Hantzschel, 1975): *Myrianites libycus* is a horizontal trail composed of a rounded furrow, marked transversely with equidistant grooves, which may be referable either to *Gyrichnites* WHITEAVES, 1883, or to *Palaeophycus* HALL, 1847; *Terebellolites fezzanensis* is a pencil-like tube which branches in an irregular fashion and which may be the cast of a burrow, the orientation of which is not known; it is possible, however, since it is usually enriched in iron oxides, that it is of inorganic, recent origin related either to the development of vertical tubes of precipitated minerals caused by capillary action at the rock surface, or by iron enrichment on the surface of the rock. It should be noted that

AGE	ICHNOSPECIES & - GENERA (DESIO, 1940b)	FULL GENERIC NAME (HANTZSCHEL, 1975) † = inorganic origin * = name no longer in use	LOCATION
Middle & Upper Devonian	*Myrianites libycus* n. f. *Bifungites fezzanensis* n. f. *Terebellolites fezzanensis* n. f.	* *Bifungites* Desio, 1940 *	Near Al Mahruqah (Wadi Ash Shati) & Dur al Beddadah (Wadi Ash Shati)
Lower Devonian	*Eoclathrus Balboi* n. f. *Harlania Harlani* Conr. *Cruziana* cfr. *furcifera* (Orb.) Delgado *Cruziana* cfr. *Prevosti* Rou. *Arenicolites* sp. *Kinneyia dubia* n. f. *Kinneyia labyrinthica* n. f. *Nereites Castroi* Delg. *Sabellarifex parvus* n. f. *Climactichnites Chiesai* n. f. *Polygonolites* *Interconulites*	† *Arthrophycus* Hall, 1852 *Cruziana* D'Orbigny, 1842 *Cruziana* D'Orbigny, 1842 *Arenicolites* Salter, 1857 † (ripple-like structures) † (ripple-like structures) *Nereites* Macleay, 1839 *Sabellarifex* Richter, 1921 *Climactichnites* Logan, 1860 † (Mudcracks) † (Cone-in-Cone structures)	Jabal Akakus, Jabal Tadrart & Jabal Idinen
Silurian-Devonian	*Harlania Harlani* Conr. *Cruziana furcifera* (d'Orb.) Delgado *Cruziana Saportai* Delgado *Fraena* cfr. *Rouaulti* (Lebesc.)	*Arthrophycus* Hall, 1852 *Cruziana* d'Orbigny, 1842 *Cruziana* d'Orbigny, 1842 *Fraena* Rouault, 1850	Jabal Eghei
Upper Silurian	*Nereites Castroi* Delgado *Sabellarifex parvus* n. f. *Interconulites*	*Nereites* Macleay, 1839 *Sabellarifex* Richter, 1921 † (Cone-in-Cone structures)	Jabal Akakus
Silurian	*Rouaultia Lyelli* (Rouault)	type species of *Rouaultia*, now *Fraena* Rouault, 1850	Gilf Kabir
Lower Silurian	*Sabellarifex tassiliensis* n. f.	*Sabellarifex* Richter, 1921	Ti'n Alkum

SAMPLE LOCATION MAP

A = Near Al Mahruqah & Dur al Beddadah, Wadi Shati.
B = Jabal Akakus
C = Jabal Idinen
D = Jabal Tadrart
E = Ti'n Alkum
F = Jabal Eghei
G = Gilf Kabir

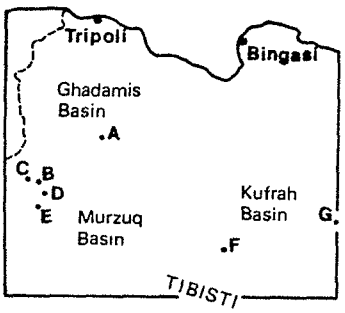


Table 2.1 TRACE FOSSILS OF THE SILURIAN AND DEVONIAN STRATA OF SOUTHERN LIBYA (MODIFIED AFTER DESIO, 1940b).

Hantzschel (1975) recommended that the generic terms *Harlania* and *Rouaultia* be dropped in favour of *Arthrophycus* and *Fraena* respectively. In the former case Goeppert (1852) failed to designate a type species for *Harlania* and consequently Hall's term, *Arthrophycus* (1852), is favoured. It is adopted here despite the fact that the term *Harlania* is in more common use in Saharan geology.

Beuf *et al.* (1971) reviewed the trace fossils present in the Lower Palaeozoic and in the Lower Devonian of the Sahara (Tables 2.2 & 2.3). The nomenclature listed as being proper, or correct, as opposed to that popularly and widely used in Saharan geology, includes only one name not approved in the treatise on the subject (Hantzschel, 1975): *Pelecypodichnus* SEILACHER, 1953, should more properly be *Locheia* JAMES, 1879. However, the popular terms commonly used are an unhappy mixture of generic names (many of which are now obsolete (Hantzschel, 1975) or have been used in a broader sense; *Tigillites*, for instance, is widely used to describe all vertical burrows), and descriptive terms derived from the morphological classification of Lessertisseur (1955) (e.g. Bilobite, from bilobate tracks). It is considered desirable to discontinue the use of these terms replacing them by generic (and specific) names where possible, and where not possible, by morphological or ethological terms.

The Palaeozoic trace fossils of Arabia have attracted less attention than those of North Africa. However, Powers *et al.* (1966) and Wolfart (1981) recorded the presence of a number of ichnogenera in the Palaeozoic strata of Arabia, and Selley (1970) undertook detailed ichnofacies work on the Cambro-Ordovician succession of Jordan (Table 2.4).

The morphological description of trace fossils favoured by Gubler (1966), and based to a large extent on samples collected and observed in the Palaeozoic succession of the Sahara, was a modified version of the largely morphological classification of Lessertisseur (1955). In Lessertisseur's classification (1955) the traces of invertebrate activity were divided into those on the bedding surface (exogenic traces; e.g. simple bilobate and trilobate crawling trails, spiral and starlike feeding traces and resting traces), and those created within the sediment (endogenic traces; e.g. feeding traces, vertical and horizontal burrows and tunnels). Although Lessertisseur (1955) divided his two main categories largely on the basis of their

CHARACTER	TERMINOLOGY	PRINCIPAL SYNONYMS	POPULAR TERMS	AGE RANGE
Simple vertical burrows	?*Skolithos* Haldeman, 1840 *Monocraterion* (has enlarged funnel shaped mouth)	*Scolecolithus* Roemer, 1840 *Tigillites* Torell, 1870 *Monocraterion* Torell, 1870 *Scolithus* "Hall", James, 1881 *Sabellarites* Richter, 1920 *Sabellarifex* Richter, 1921 *Lepocraterion* Stehmann, 1934	*Tigillites* *Scolite*	Precambrian-present
Vertical, spiral and helicoidal burrows	*Daedalus* Rouault, 1850	*Vexillum* Rouault, 1850 *Humilis* Rouault, 1850	*Vexillum*	Cambrian?-Silurian-Devonian
Walking and crawling traces, prints of repose, and horizontal burrows.	*Pholeus* Fiege, 1944 *Palaeophycus* Hall, 1847 *Arthrophycus* Hall, 1852	*Fucoides* Brongniart, 1823 *Harlania* Goeppert, 1852 *Phycodes* Richter, 1850	*Harlania*	Cambrian?-Silurian-Oligocene?
	Cruziana d'Orbigny, 1842	*Bilobites* d'Orbigny, 1839 *Bilobichnium* Krejci-Graf, 1932	*Bilobite*, *Cruziana*	Cambrian-Devonian
	C. furcifera *C goldfussi*			Arenigian
	Rusophycus	*Rhysophycus* Hall, 1852		
	Fraena Rouault, 1850	*Rouaultia* Tromelin and Lebesconte, 1875 *Fraena* Peneau, 1946	*Fraena*	
Traces of the embedded position of live lamellibranches in sediment	*Pelecypodichnus* Seilacher, 1953 (= *Lockeia* James, 1879 (Hantzschel, 1975))	*Sagittichnus* Seilacher, 1953		Ordovician-Silurian-present

Table 2.2 TRACE FOSSILS OF THE LOWER PALAEOZOIC STRATA OF THE SAHARA (MODIFIED AFTER BEUF *et al.*, 1971).

CHARACTER	TERMINOLOGY	PRINCIPAL SYNONYMS	POPULAR TERMS	AGE RANGE
Spiral and helicoidal burrows	*Zoophycus* Massalongo, 1885 *Daedalus* Rouault, 1850	*Taonurus* Fischer-Ooster, 1858 *Spirophyton* Hall, 1863 *Alectorurus* Schimper, 1869 *Cancellophycus* Saporta, 1873	*Spirophyton*	Silurian-Cretaceous
Simple vertical burrows	*? Skolithos* *Monocraterion*		*Tigillites* *Tigillites*	
Crawling traces	*Cruziana*		*Bilobite, Cruziana*	

Table 2.3 TRACE FOSSILS OF THE LOWER DEVONIAN STRATA OF THE SAHARA (MODIFIED AFTER BEUF *et al.*, 1971).

AGE	ICHNOGENERA & SPECIES
Devonian	*Fraena* Rouault *Skolithos*
Silurian	*Rusophycus* *Skolithos* *Harlania*
Cambro- Ordovician	*Harlania* *Rusophycus* *Sabellarifex dufrenoyi* (Rouault) *Cruziana furcifera* d'Orbigny *Cruziana goldfussi* (Rouault) *Cruziana huberi* (Meunier) *Diplichnites* *Ichnyspica* *Incifex* *Tasmanadia* *Merestomichnites* *Skolithos* *Fraena* Rouault

Modified after Powers *et al.* (1966), Selley (1970) and Wolfart (1981)

Table 2.4 TRACE FOSSILS OF THE CAMBRIAN TO DEVONIAN STRATA OF ARABIA

three-dimensional geometry, he also divided them according to an ethological classification (i.e. one based on interpretation of the type of activity represented by the trace, whether it be resting, feeding or locomotion). Gubler (1966) dropped the ethological component of Lessertisseur's classification (1955) and adopted a purely morphological classification which is in substance very similar to that adopted by Horowitz (in Frey, 1971). Horowitz (*idem*) also used a bipartate division between intrastratal structures and bedding-surface structures and divided up the trace fossils within these headings according to their shape. His classification also took account of the filling or internal pattern, the size and the orientation of the trace.

A modification of Gubler's classification provided the basis for an initial description of the trace fossils commonly found in the Palaeozoic successions studied in this thesis (Table 2.5). It should be noted that the disadvantage of this classification is that some traces may be considered both exogenic and endogenic. For example, rectilinear, regular surface traces such as *Arenicolites* may also be classified as simple non-rectilinear internal traces. This can have the result that the same trace fossil may have two different names. For instance, *Bifungites* is thought to be the surface expression of *Diplocraterion* or a similar U-shaped internal trace (Seilacher, 1969). Additional problems arise in attempting to define which traces are surface traces, which are burrows in relief, and which are traces at bed interfaces. *Arthrophycus*, and the other endogenic burrows in relief (e.g. *Scolicia*) have appearances characteristic both of bedding plane and internal trace fossils. They might therefore be considered both endogenic and exogenic, but the presence of internal structures within the trace fossil indicates that they are best considered endogenic traces which have modified the bedding interface as well as the internal structure of the bed. In the case of *Arthrophycus* (see Fig. 2.6), which has a complex internal structure of spreiten which creates an arc-like webb of disturbed sediment, it is fairly easy to make the distinction. However, in many cases, no internal structure is visible. A further classification problem is created by traces which vary laterally and consequently belong to different subdivisions of the classification. *Fraena*, for instance, can be unilobate or bilobate.

EXOGENIC TRACES (tracks & traces on the bed surface)	Irregular	Unilobate	*Agrichnium* *Fraena* *Gyrichnites*
		Bilobate	*Climactichnites* *Halopoa* *Cruziana* *Isopodichnus* *Diplichnites* *Rusophycus* *Fraena* *Gyrochorte* *Sagittichnus*
		Trilobate	*Kouphichnium* *Merestomichnites* *Scolicia* *Protichnites*
	Regular	Meandering	*Cochlichnus* *Nereites*
		Cuneiform	*Paleodictyon*
		Rectilinear	*Arenicolites* *Bifungites* *Corophioides* *Diplocraterion*
ENDOGENIC TRACES (tubes & burrows within the bed)	Burrows in relief		*Arthrophycus* *Planolites* *Fucusopsis* *Phycodes* *Palaeophycus* *Scolicia* *(Scoyenia)**
	Traces at bed interfaces		*Neonereites*
	Internal traces	Simple rectilinear	*Cylindricum* *Skolithos* *Monocraterion* *Tigillites* *Sabellarifex*
		Simple non-rectilinear	*Arenicolites* *Corophioides* *Diplocraterion*
		Complex helicoidal	*Daedalus* *Spirophyton* *Zoophycus*
		Other complex arrangements	*Asterosoma* *Rosselia*
		Others	*Lockeia* (= *Pelecypodichnus*)

*Not present in rocks older than Permian but used by Seilacher (1964) to define a non-marine ichnofacies

Modified after Gubler (1966) and Lessertisseur (1955)

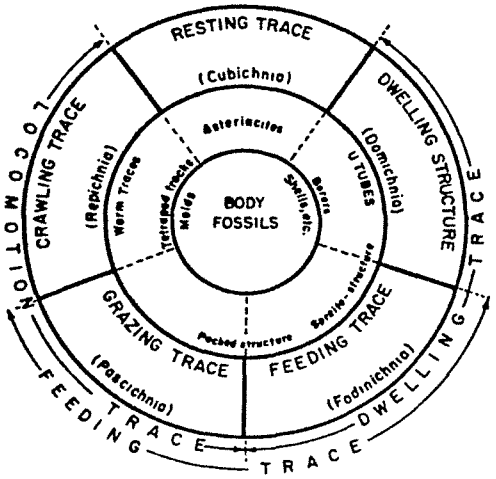
Table 2.5

MORPHOLOGICAL CLASSIFICATION OF TRACE FOSSILS WITH EXAMPLES FROM CAMBRIAN-DEVONIAN STRATA, NORTH AFRICA AND ARABIA.

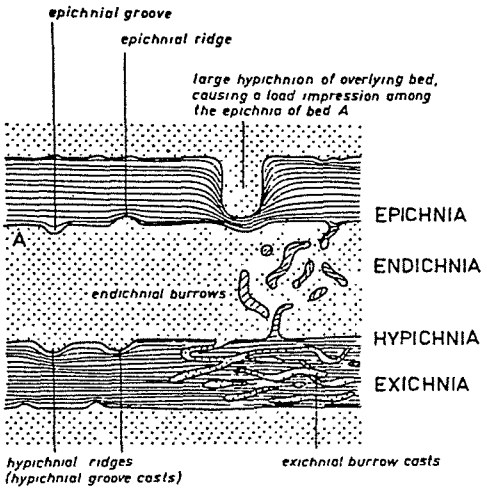
Martinsson (1970) devised a descriptive system for trace fossils to distinguish traces on the upper and lower surfaces of a bed (the casting medium), traces within a bed, and other unusual types of traces (Fig. 2.2). The main distinction in Martinsson's system (1970) is between traces which are present on the upper surface of a bed (Epichnia), on the lower surface of a bed (Hypichnia) and traces present within the bed (Endichnia). In Table 2.5 all the endogenic traces are Endichnia, whereas the exogenic traces may be preserved as indentations and grooves on the upper surface of a bed (Epichnia) or as protuberances and ridges on the sole of a bed which was laid down on a surface with tracks and traces on it (Hypichnia).

Trace fossils lend themselves to an ethological, genetic interpretation because they generally reflect the behaviour and activity of the organisms which produced them rather than the physiology of that organism. Indeed different organisms with similar behaviour may create identical traces. Their traces, therefore, are of great value in interpreting the conditions which existed when they were alive, but are of limited value in dating the rocks in which they left their mark. Seilacher (1953) recognised that organisms generated traces not only through the activities of resting, feeding and locomotion (as in the classification of Lessertisseur, 1955) but also as a result of their occupation of dwelling structures. His ethological classification (1953) (Fig. 2.2) also took into account the fact that a locomotion trace created while feeding (a grazing trace) should be distinguished from one where the trace was produced purely by locomotion (a crawling trace) and, similarly, that feeding structures within the sediment should be distinguished from structures created purely for dwelling. The Latin terms for resting, crawling, grazing, feeding, and dwelling traces used by Seilacher are not generally adopted here, but in using the English terms it should be noted that a feeding trace is one created within the sediment and not on it, whilst a grazing trace is a surface trace which may or may not form a complex surface pattern.

The fact that organisms may burrow along buried bed interfaces creates a difficulty in relating morphological to ethological classifications. Whereas all endogenic traces were created within the sediment body not all surface traces were created at the sediment-water interface. As already noted, most surface traces are



Ethological classification of trace fossils proposed by Seilacher (1953) (from Osgood, 1970).



Diagrammatic representation of toponomic terminology suggested by Martinsson (1970) and shown in cross section (Martinsson, 1970, p. 327). (Stippled areas are siltstones and ruled areas, shales. For descriptive terms at right refer to bed A.)

Both diagrams from Hantzschel (1975).

Figure 2.2 ETHOLOGICAL AND TOPONOMIC TERMINOLOGIES FOR TRACE FOSSILS

resting, crawling or grazing traces, but some were created by horizontal burrowing along a bed interface and by the subsequent collapse of the overlying bed into the burrow (e.g. some varieties of *Cruziana*, Webby, 1969). Such a trace fossil, of burrow origin, is difficult to distinguish from a hypichnial cast of a resting, crawling or grazing trace, because no internal structure of the burrow has been preserved.

Trace fossils and their associations have been shown to be particularly useful in palaeogeographic investigations because of their autochthonous nature. Since they cannot be displaced, they provide unequivocal evidence of life on or within the sediment. Since they also provide evidence of life-style or behaviour, which was determined to a large extent by the local environment, they are good facies indicators. Seilacher (1967), after more than ten years' work on the subject, showed that characteristic trace fossil assemblages occur in the same facies in rocks of different ages and locations. Each such assemblage, he argued, is characteristic of a particular environment and is composed of associations of trace fossils which constitute ichnofacies. The identification of different ichnofacies represents one part of the process of defining a facies. As a rough generalisation, shallow water ichnofacies are characterised by vertical tubes, burrowing structures, dwelling burrows, and resting impressions whereas, in rocks of deeper water origin, trace fossils with complicated spreiten structures and many varied and patterned grazing trails are characteristic. Shallow-water environments, therefore, are characterised by high-energy currents and consequently by suspension feeders. These organisms are gradually replaced in deeper water, lower energy environments, by increasingly elaborate sediment feeding organisms, living both within the sediment and at the sediment-water interface. It was also shown (Seilacher, 1967) that the trace fossil assemblages of flysch deposits are dominated by grazing trails, those of epicontinental and paralic molasse deposits are characterised by resting traces, and those of predominantly fluviatile and continental deposits, with only periodic marine inundations, are characterised by all the ethological associations except complex grazing trails.

The trace fossils observed in this study in the Palaeozoic successions of Libya and Saudi Arabia are illustrated in Figures 2.3 to 2.9. The formations and facies in which the trace fossils were

Figure 2.3 EXOGENIC TRACES : SURFACE TRAILS AND TRACKS

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively; hammer head, 15 cm long.

For locality numbers see Fig. 3.5

Upper far left : network of trails on a rippled surface of fine-grained sandstone. From facies RBSdSt, Tadrart Formation, locality 13. (A)

Upper middle left : arthropod track consisting of narrow median furrow, flanked on both sides by regularly-spaced leaf-shaped lobes. Similar to *Nereites* but not meandering. From facies RBSdSt, Tadrart Formation, Tihererin. (B)

Upper middle and far right : *?Diplichnites*. Locomotion traces of arthropods, ?trilobites. Note tricuspate scratch marks in right hand specimen (Seilacher, 1970). This specimen is also similar to *Merestomichnites*, a eurypterid track. Left hand sample from facies 2, Saq Formation, near Anz. Right hand sample from the top of the Tanezzuft Formation, Jabal Idinen. (C,D)

Lower far left : *Fraena*. Bilobate and unilobate simple surface tracks that may be striated longitudinally or transversely. Also named *Rouaultia*. This example is a bilobate track on a rippled sandstone surface. From facies 2, Saq Formation, near Anz. (E)

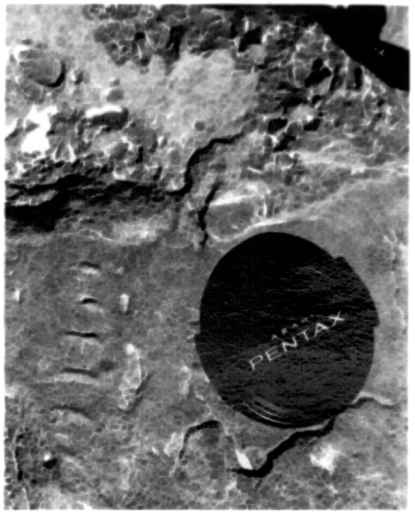
Lower middle left : *Fraena* or *Agrichnium*. Simple unilobate sinuous surface track, 2-4 mm wide. Developed on the surface of a rippled sandstone of facies RBSdSt, Tadrart Formation, locality 2. (F)

Lower middle and far right : arthropod tracks. Similar tracks to those shown in the upper middle left photograph. From facies RBSdSt, Tadrart Formation, Tihererin. (G,H)

A

B

C



D

E

F

G

H

Figure 2.4 *CRUZIANA, RUSOPHYCUS* AND *SCOLICIA*

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively; penknife, 2 cm wide.

For locality numbers see Fig. 3.5.

Upper far left : *Rusophycus*. Bilobate resting trace of an (A) arthropod, ?trilobite. From facies RBSdSt at the top of the Acacus Formation, locality 9.

Upper middle left : *Cruziana acacensis*. Bilobate deep resting (B) burrow or nest of a trilobite that is characteristic of the Acacus Formation (Seilacher, 1970). From facies RBSdSt at the top of the Acacus Formation, locality 9. Sample is present on a slab of rock used as a beduin tombstone.

Upper middle right : *Cruziana*. Bilobate crawling trace of an (C) arthropod, ?trilobite. From facies RBSdSt, near the top of the Tadrart Formation, Tihererin. This is an example of the smallest of the *Cruziana* species seen in this study.

Upper far right : *Cruziana*. From facies RBSdSt at the top of (D) the Acacus Formation, locality 9. Smaller species than *C. acacensis*.

Lower left : *Scolicia* (and *Cruziana acacensis* at bottom). (E) Weakly trilobate and bilobate band-like, gastropod trails. Occur both as surface and internal trails. From facies RBSdSt at the top of the Acacus Formation, locality 9. In Lower Palaeozoic and Devonian sediments, *Scolicia* is thought to be a marine indicator since gastropods first occupied non-marine habitats in the Carboniferous (Clarkson, 1979).

Lower middle and right : *Cruziana*. From fine-grained inter- (F, G) calations within the sandstones of the Saq Formation, facies 1, in the region of Jabal Saq. Generally present on the sole of fine-grained cross-laminated sandstones lying above thin micaceous siltstone beds.

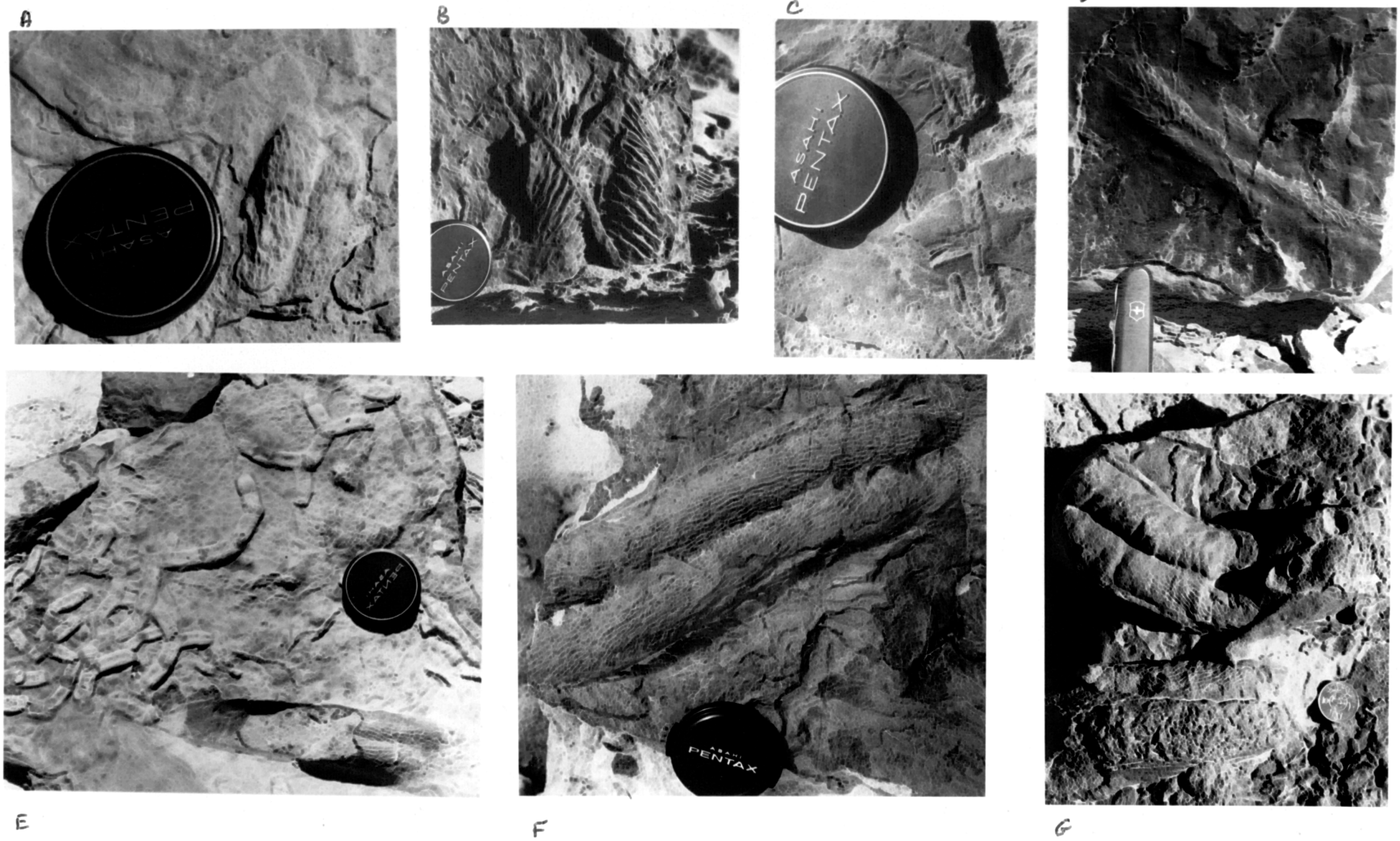

A
B
C
D
E
F
G
ASAHI
PENTAX

Figure 2.5 FEEDING BURROWS AND SIMPLE BUT IRREGULAR SURFACE TRAILS

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively.

Left upper and lower : small subparallel sinuous traces. May (A,D) be either internal burrows (e.g. *Palaeophycus* or *Fucusopsis*) or simple, but irregular unilobate grazing trails (e.g. *Fraena* or possibly *Agrichnium*). Upper sample from the top of the Tanezzuft Formation, Jabal Idinen; bottom sample from facies 3, Tabuk Formation, Jabal Hanadir.

Middle upper : *Palaeophycus*. Cylindrical or subcylindrical (B) burrow that is commonly unbranched and has smooth wall surfaces. Burrows may be orientated more or less obliquely to bedding. From facies RBSdSt, Tadrart Formation, Bir Talauaut.

Middle lower and right upper : *Palaeophycus* and/or *Fucusopsis*. (E,C) Cylindrical or subcylindrical burrows. Sinuous burrows with smooth walls or faint longitudinal striae may be attributed to *Palaeophycus*, whilst straighter burrows with thread-like sculpture may be assigned to *Fucusopsis*. Lower sample from the Tawil Member of the Tabuk Formation (Lower Devonian), near Al Mayyah; upper sample from a fine-grained sandstone in the Ouan Kasa Formation (Middle Devonian).

Right lower : *Gyrichnites*. Unilobate trail comprising a rounded (F) furrow marked transversely by nearly straight grooves. From Mrar Formation (Carboniferous), 30 km southeast of Al Awaynat.

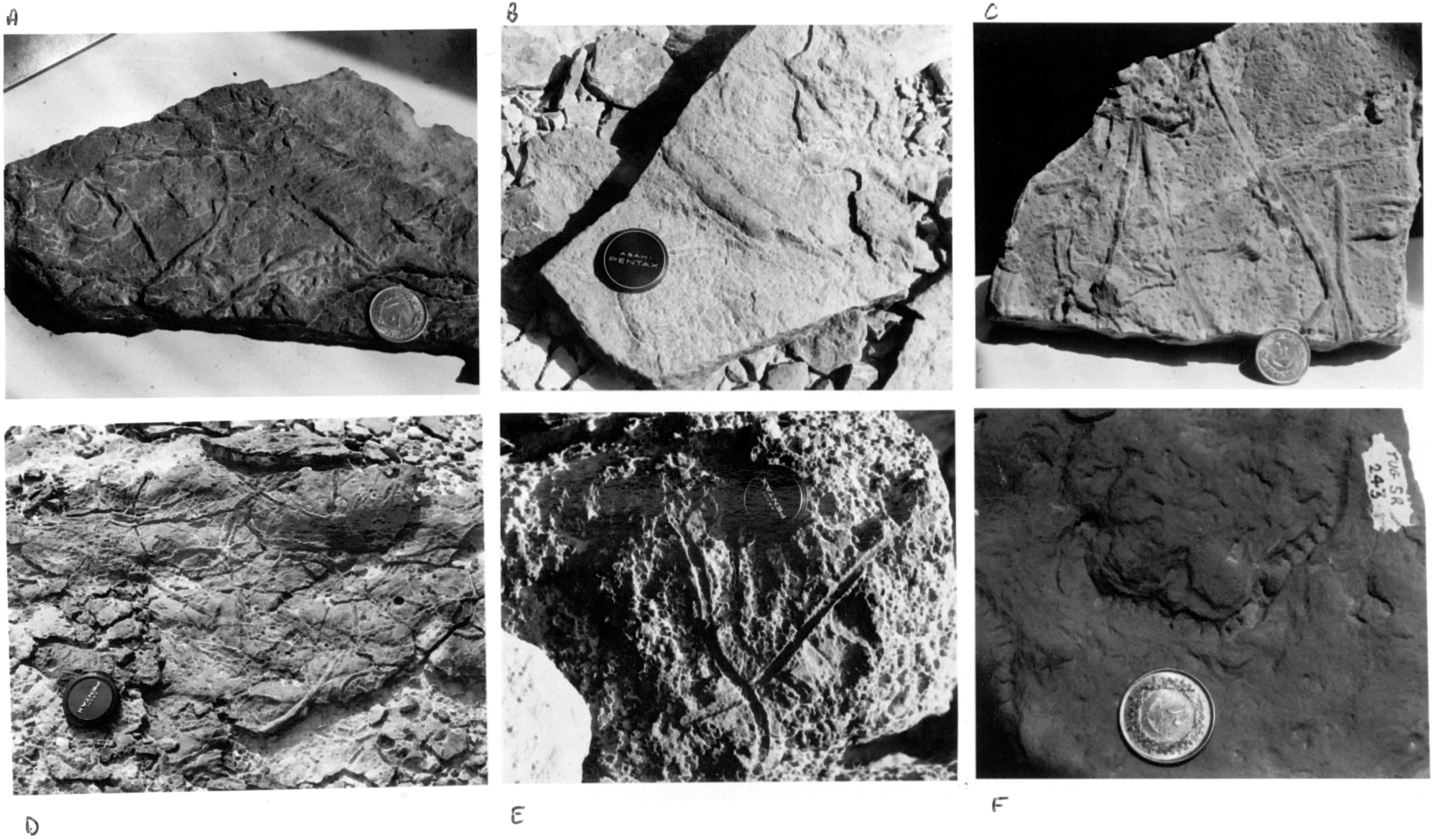

A
B
ASAHI
PENTAX
C
TUG SR
243
D
E
F

Figure 2.6 *ARTHROPHYCUS* AND *ASTEROSOMA*

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively; penknife, 9 cm long.

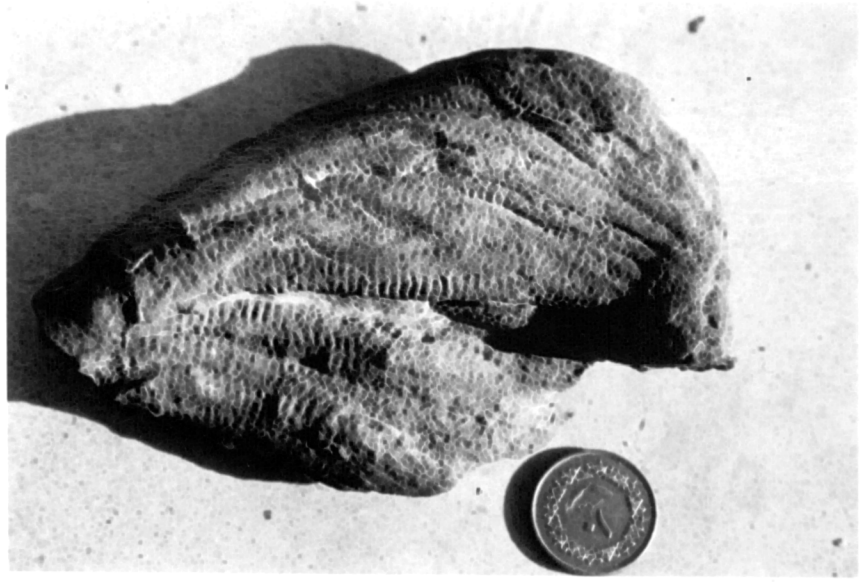
For locality numbers see Fig. 3.5

Left and upper right : *Arthrophycus*. Bundles of backfilled, annulated sediment-feeding burrows, with an internal chevron pattern of spreiten; characteristically arranged into a palm-frond appearance, but also present as nearly straight burrows (upper left). Generally present on the sole of fine-grained sandstone beds lying above argillaceous siltstone or shale beds in facies CSdSt. The palm-frond variety are characteristic of the Acacus Formation whereas both varieties are present in the Tadrart Formation. Upper left, from Tadrart Formation, locality 4. Lower left, from Tadrart Formation, locality 7. Upper right, from Acacus Formation, locality 10.
(A, B, C)

Lower right : *Asterosoma*. Radially backfilled burrow system of a sediment-feeding organism. From facies RBSdSt, Tadrart Formation, locality 2.
(D)

A

B



C

D

Figure 2.7 *SKOLITHOS* VERTICAL BURROWS

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively.

For locality numbers see Fig. 3.5

Upper left : characteristic vertical burrows of facies VBSdSt.
(A) Tadrart Formation, locality 13.

Lower left : characteristic vertical burrows of the Aouinet
(D) Ouenine Formation with annular ornament on the inside of the burrow. This specimen is from Gour Iduka.

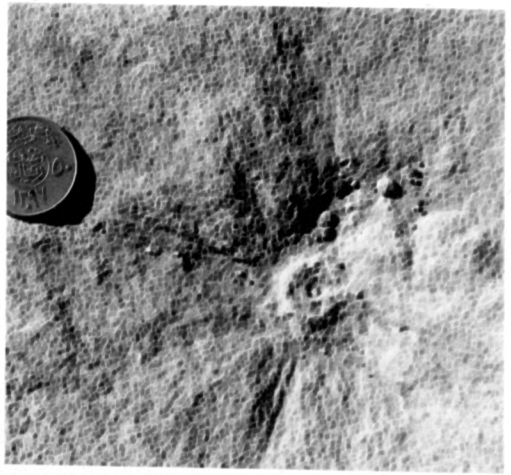
Middle and lower right : two varieties of vertical burrow
(B,E,F) characterise the Haouaz Formation both of which are smaller than those of the Tadrart Formation. Simple tubes (upper middle) contrast with poorly defined vertical tubes which are slightly larger and which produce an irregular nobbly appearance on the upper bedding plane (lower middle and right). The distinction may be in part due to differing degrees of cementation, however. All three examples from the Qarqaf region.

Upper right : vertical burrow displaying a radial stellate
(C) pattern on the upper bedding plane of a sandstone of facies 3, Tabuk Formation, Al Qara.

A

B

C



D

E

F

Figure 2.8 ENDOGENIC TRACES

Scales: lens cap and Libyan coin, 5 cm and 2 cm diameter respectively.

For locality numbers see Fig. 3.5

Upper left (A) : *Planolites*. Cylindrical or subcylindrical sediment-feeding infilled burrows, horizontal and oblique to bedding planes and penetrating the sediment in an irregular fashion. Similar to *Palaeophycus*. From the top of the Tanezzuft Formation, Jabal Idinen.

Lower left (D) : *Bifungites*. Dumb-bell-like structure, present on sandstone bedding planes, that may be the surface expression of *Diplocraterion* or similar U-shaped burrows (Seilacher, 1969). This specimen is from the Lower Bifungites Zone of the Aouinet Ouenine Formation in the Qarqaf region.

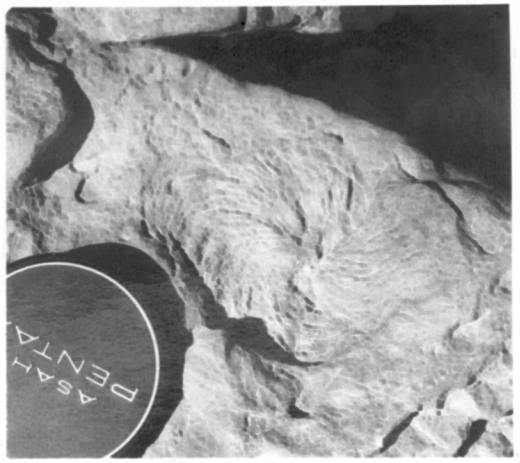
Middle upper and lower (B,E) : *Corophioides*. Rod-shaped structure on sandstone bedding planes. Thought to be the surface expression of U-shaped spreiten burrows that were perpendicular to the bedding plane. Both specimens from facies RBSdSt, Tadrart Formation. Upper sample from locality 2, lower sample from locality 16.

Right upper and lower (C,F) : *Spirophyton*. Complex internal sediment-feeding trace, similar to *Zoophycus*, but smaller, and displaying a circular outline around the spreiten. Spirally-coiled conical and planar forms exist. The conical-circular trace is attributed to lateral migration of the organism, and backfilling of the burrow (Seilacher, 1969). Both samples from facies RBSdSt, Tadrart Formation. Upper sample from locality 17, lower sample from locality 2.

A

B

C



D

E

F

Figure 2.9 VARIETY OF ENDOGENIC AND EXOGENIC TRACE FOSSILS, SOME OF UNCERTAIN AFFINITIES

Scale: lens cap is 5 cm diameter.

For locality numbers see Fig. 3.5.

Upper far left : *Rosselia*. Cylindrical burrows, commonly oblique to bedding, present within a chamber filled with concentric layers of matrix that are generally strongly weathered. Thought to be either the dwelling burrow of a semisessile suspension-feeding organism, or the burrow system of a sediment-injesting organism similar to that of *Asterosoma* (Seilacher, 1969). The central burrow on the right of the picture is 4 cm long. From facies CSdSt, Tadrart Formation, locality 17. (A)

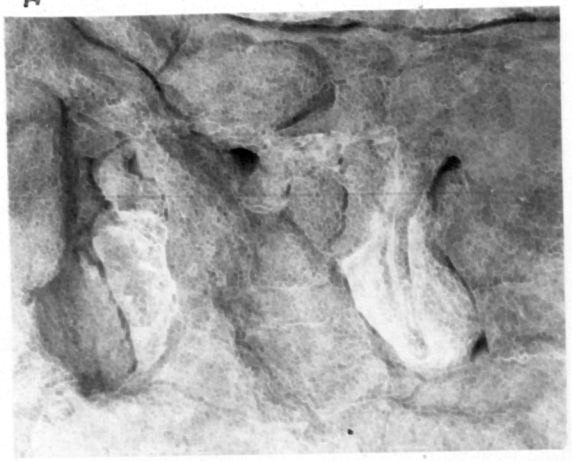
Upper middle left : crescentic bedding plane trace fossil with longitudinal ribbing. Probably attributable to *Arenicoloides*, and possibly the preserved lowest part of a *Corophioides* U-tube (Hantzschel, 1975). The trace fossil in this photograph is 2.5 cm from horn to horn of the crescent. From facies RBSdSt, Tadrart Formation, locality 5. (B)

Upper middle and far right : two varieties of horizontal feeding burrows of uncertain affinities. The one on the right may be a variety of *Phycodes*. From facies RBSdSt, Tadrart Formation. (C, D)

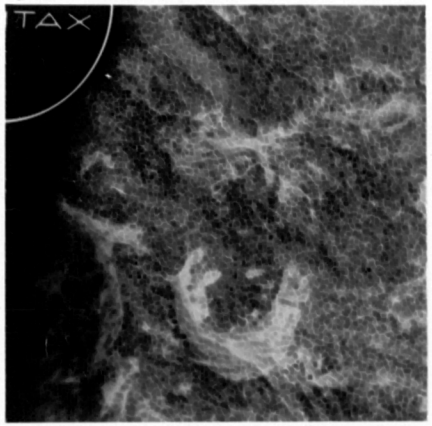
Lower left : cylindrical, "spirally" annulated trace fossil of uncertain affinities. Similar to *Diplichnites*. From facies RBSdSt at the top of the Acacus Formation, locality 8. (E)

Lower middle and right : trails and resting traces. Possibly attributable to *Sagittichnus*. From facies RBSdSt, Tadrart Formation, localities 1 and 13. (F, G)

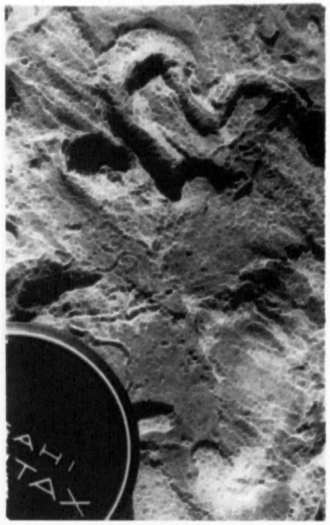
A



B



C



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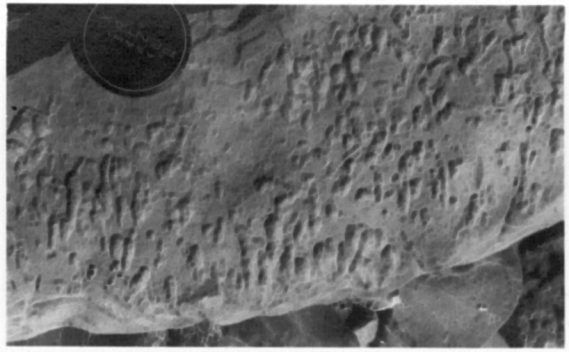

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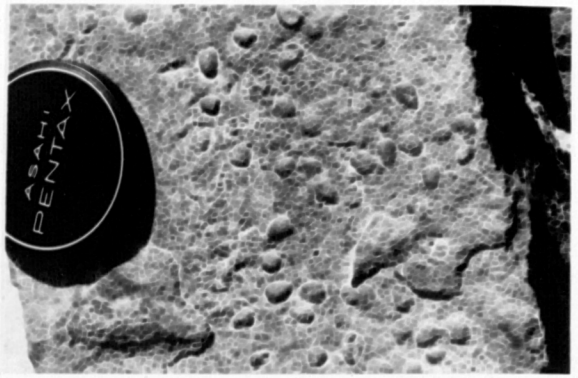

G

found, and their localities, are recorded in the descriptions of these figures since they act, in this thesis, as a reference set of photographs for the description of the facies given in Chapters 3, 4 and 5. Exogenic traces are illustrated in Figures 2.3 and 2.4, the majority of which are irregular (cf. Table 2.5) and are thought to be arthropod tracks and trails (i.e. crawling and possibly grazing traces), and resting traces. In most cases the type of arthropod responsible for these traces is uncertain but, in the case of *Merestomichnites*, eurypterids are thought to have been responsible (Hantzschel, 1975), and *Cruziana* and *Rusophycus* are thought to be the crawling and resting traces of trilobites (Crimes, 1970; Seilacher, 1970). *Scolicia* (Fig. 2.4) may be both an exogenic trace and an endogenic trace (Table 2.5), and is thought to be the grazing and feeding trace of a gastropod (Seilacher, 1978, pers. comm.). In Figure 2.5 feeding burrows and simple but irregular surface trails are illustrated. In some cases it is very difficult to establish whether the traces are exogenic or endogenic (e.g. *Palaeophycus*, *Fucusopsis* and *Fraena*). In other cases, however, a burrow-origin may clearly be established, and in the case of *Gyrichnites* a surface crawling-origin may be confidently proposed.

Endogenic traces are illustrated in Figures 2.6, 2.7 and 2.8. *Arthrophycus*, a burrow trace preserved in relief, and *Asterosoma*, a complex internal trace (cf. Table 2.5), are both sediment-feeding organisms (Fig. 2.6). *Skolithos* (Fig. 2.7) is a simple rectilinear trace (cf. Table 2.5) and is perhaps the most common trace fossil in the sediments studied. This ichnogenus is badly in need of monographic treatment (Osgood, 1970). Although *Skolithos* and *Tigillites* are distinct but overlapping genera (the description of the former includes mention of a funnel-shaped top to the tube whereas no mention is made of this in the description of the latter; Hantzschel, 1975) the two terms have often been used synonymously. Furthermore, the term used for vertical burrows tends to be determined by the nationality of the user rather than by any characteristics of the trace: the French and Germans use the term *Tigillites* (Gubler, 1966; Beuf *et al.*, 1971; Klitzsch, 1969), the Italians use the name *Sabellarifex* (Desio, 1940b) and the Americans use *Scolithus* (Powers *et al.*, 1966). In the text of this thesis the name used by the author quoted has been employed unless the present author has

personally studied the succession in question, as to do otherwise would lead to an even greater uncertainty about the character of the trace fossil being described than already exists. *Monocraterion* is distinguished from *Skolithos* by the presence of depressed laminae adjacent to the burrows, annulation marks on the burrow, and "trumpet pipe" funnel structures. The difference between these two trace fossils may be a result of the different activity of one burrowing and filter-feeding organism, the former being created by upward movement of the organism in the burrow, the latter from continuous occupation of the burrow (Goldring, 1962; Hantzschel, 1975). A variety of endogenic trace fossils is illustrated in Figure 2.8, two of which have exogenic, regular and rectilinear surface expressions: *Bifungites* and *Corophioides* (see Table 2.5). These two trace fossils have a similar appearance to *Planolites* but are distinct in being the traces of U-shaped burrows whereas *Planolites* is the trace of a burrowing organism which created an undulating burrow system through the sediment. *Spirophyton* is a complex helicoidal, internal trace fossil formed by a sediment-feeding organism. It is similar to *Asterosoma* (Fig. 2.6) and *Rosselia* (Fig. 2.9) both of which share a complex internal structure. Finally three examples of endogenic, and three of exogenic traces of uncertain affinities, all from the Jabals Akakus and Tadrart, are illustrated in Figure 2.9. The endogenic traces (top right) were probably produced by sediment-feeding burrowing organisms whilst the exogenic traces (bottom) may be the crawling and resting traces of arthropods.

The majority of the trace fossils recorded in the Cambrian to Devonian sediments of North Africa and Arabia (Tables 2.1-2.4; Figs. 2.3-2.9) belong to the following ichnofacies of Seilacher's bathymetric zonation (1967): *Scoyenia* (non-marine), *Skolithos* (littoral) and *Cruziana* (shallow-marine, sub-littoral) ichnofacies (Table 2.6). Only a few ichnogenera are present which belong to the ichnofacies characteristic of deeper water depositional environments (*Zoophycus* and *Nereites* ichnofacies). These are *Daedalus*, *Spirophyton*, *Zoophycus* (complex helicoidal, internal, sediment-feeding traces), and *Nereites* (meandering, surface grazing traces) (Table 2.5). When Seilacher visited the Jabal Tadrart and Ghat regions he recognised trace fossils typical of the *Cruziana* and *Skolithos* ichnofacies in the Lower Palaeozoic and Devonian successions. He considered, however, that the

ICHNOFACIES (named after type ichnogenus)	DESCRIPTION	GENERA
Scoyenia	Non-marine assemblage characterised by tracks and vertical burrows	*Arenicolites* *Cochlichnus* *Clyindricum* *Fuersichnus (Phycodes* type) *Isopodichnus* *Kouphichnium* *Lockeia* *Merestomichnites* *Palaeophycus* *Phycodes* *Planolites* *Protichnites*
Skolithos	Assemblage of vertical and U-shaped burrows created by suspension feeders adjusting position in littoral sands in response to sedimentation	*Arenicolites* *Bifungites* *Diplocraterion* *Monocraterion* *Skolithos* *Tigillites*
Cruziana	Assemblages of feeding traces and tracks created in well-sorted shallow-marine sands and silts	*Arthrophycus* *Palaeophycus* *Phycodes* *Planolites*
Asterosoma		*Asterosoma* *Cruziana* *Neonereites*
Spirophyton		*Corophioides* *Scolicia* *Spirophyton*

Modified after Seilacher (1964, 1967, 1969) and Pollard (1980).

Table 2.6 ICHNOFACIES OF CAMBRIAN TO DEVONIAN STRATA, NORTH AFRICA AND ARABIA

Scoyenia ichnofacies was absent in the successions he studied (1969). He further argued that the *Spirophyton* in the Devonian successions should be assigned to the *Cruziana* ichnofacies and that two separate trace fossil assemblages existed within this ichnofacies: the *Spirophyton* and *Asterosoma* assemblages (Table 2.5). He considered these assemblages, by analogy with Ordovician Nubian-type sandstones in southern Jordan (Bender, 1963; cf. Selley, 1970), to be related to different subenvironments of a channel system characterised by different current strengths. This is an example of the way in which Seilacher's initial bathymetry model (1967) has been modified over the years as it has been realised that marine trace fossil assemblages are not solely depth-dependent but, as Osgood (1970) and Frey (1971) have pointed out, such factors as nutrient supply and substrate type probably influenced the composition of the ichnocoenoses to an equal degree. Indeed Osgood and Szmuc (1972) questioned the value of *Zoophycus* (similar to *Spirophyton*, see Fig. 2.8) as a bathymetry indicator, as they observed that it can occur both in shallow and deep water sediments. In the light of this evidence, the presence of *Daedalus* and *Spirophyton* cannot be considered to provide any strong support for deeper-water ichnofacies than the *Cruziana* ichnofacies. *Nereites* was recorded in the Acacus Formation by Desio (1940b, Plate XI and p. 70) (Table 2.1) and a trace fossil similar to *Nereites* from the Tadrart Formation is illustrated in Figure 2.3. In both specimens the track is fairly straight with occasional minor changes in direction. They are distinct, therefore, from the classical meandering grazing traces of *Nereites* figured in Hantzschel (1975) and may, perhaps, be more reasonably assigned to *Gyrochorte* or *Halopoa* (Hantzschel, 1975). In conclusion, little evidence exists for the presence of trace fossils of the *Zoophycus* and *Nereites* ichnofacies, which provides some evidence for shallow marine (sub-littoral and littoral) depositional environments, notwithstanding the arguments of Osgood (1970) and Crimes (1980) that Seilacher's ichnofacies bathymetry concept (1967) was a serious oversimplification.

It is very difficult to establish from the trace fossil assemblages present in the Palaeozoic successions studied whether any evidence exists for non-marine depositional environments. Studies of terrestrial trace fossil assemblages have concentrated on Mesozoic sediments, particularly those of the Triassic (Seilacher, 1967;

Pollard, 1980), since the development of a full complement of terrestrial organisms did not occur until much later in geological history than the development of a full marine fauna. However, in the Devonian, insects and terrestrial arthropods occupied non-marine habitats (Clarkson, 1979; Pollard, 1980) and, therefore, it should be possible to use the traces of these organisms in ichnofacies studies. The trace fossils characteristic of the *Scoyenia* ichnofacies are listed in Table 2.6. They comprise exogenic crawling and resting traces and endogenic vertical and horizontal burrows. The vertical burrows may have been occupied by insects (dwelling traces) whilst the horizontal burrows were probably formed by terrestrial soft-bodied, sediment-feeding organisms similar to their marine counterparts. Distinguishing these trace fossils from those formed in the shallow-marine realm is dependent, therefore, upon the recognition of insect-formed burrows, and traces formed by terrestrial arthropods. *Scoyenia*, the type trace fossil of the non-marine ichnofacies, is not found in earlier sediments than those of the Permian but most of the other trace fossils of this ichnofacies are recorded in Devonian rocks and some of them in Lower Palaeozoic rocks (Pollard, 1980; Hantzschel, 1975). Unfortunately, the trace fossil considered by Seilacher (1970) to be a facies indicator for a non-marine environment, *Isopodichnus*, is considered by others to belong to the genus *Cruziana* and/or *Rusophycus* (Hantzschel, 1975; Bromley & Asgaard, 1979) implying that, in this case, the traces produced by marine and non-marine arthropods may not be distinguished at generic level. Similarly *Cylindricum*, a genus typical of the *Scoyenia* ichnofacies and thought by Pollard (1980) to be the dwelling burrow of insects, is considered by Bromley and Asgaard (1979) to be synonymous with *Skolithos*. Arthropod tracks and resting traces other than *Cruziana* and *Rusophycus* were only rarely observed in this study. Some examples of the foot-like impressions, scratch marks and resting traces observed are illustrated in Figures 2.3 and 2.9. Apart from a tentative recognition of *Merestomichnites* in two specimens (Fig. 2.3, upper middle and far right) none of the trace fossils of arthropods characteristic of the *Scoyenia* ichnofacies was observed. This suggests that subaerially-formed exogenic traces were only rarely preserved, if at all, in the sediments studied. Of the other trace fossils observed which are characteristic of the *Scoyenia* ichnofacies

none were exclusive to that ichnofacies, being also characteristic of the *Skolithos* or *Cruziana* ichnofacies. Clearly ichnofacies defined at generic level and considered alone do not provide sufficient evidence to adequately distinguish the ichnocoenoses of sub-littoral and littoral from fresh-water and subaerial environments in this study. It is necessary to define the trace fossils more precisely, to relate these assemblages to observations of similar assemblages elsewhere, to make ethological interpretations of the trace fossils and to relate them to all the sedimentary evidence on depositional environment of the beds in which the ichnofossils were found.

The most common trace fossils in the sediments studied are all characteristic ichnogenera of the *Skolithos* and *Cruziana* ichnofacies. Little evidence exists, therefore, for a well-developed *Scoyenia* ichnofacies. It was found, however, that the following assemblage of trace fossils was characteristic of thin bioturbated, argillaceous intercalations within the coarser-grained beds of the Nubian-type sandstones: *Arthrophycus*, *Skolithos*, *Cruziana* and *Palaeophycus*. These coarser-grained beds have been interpreted as fluviatile and estuarine deposits (see Chaps. 3 & 4) and, consequently, this "*Arthrophycus* assemblage" is considered, contrary to Seilacher (1969), to be evidence for restricted marine or non-marine conditions, contrasting with the open-marine conditions inferred from the much more diverse trace fossil assemblages characteristic of the finer-grained beds of the Nubian-type sandstones.

Finally, the trace fossils illustrated in Figures 2.3-2.9 draw attention to age-related scale changes in three ichnogenera: *Cruziana*, *Skolithos* and *Arthrophycus*. Cambro-Ordovician and Silurian *Cruziana* species are larger than their Lower Devonian counterparts (Fig. 2.4), whilst the *Skolithos* species characteristic of the Ordovician Haouaz Formation is significantly larger than its equivalent in the Devonian Tadrart and Aouinet Ouenine formations (Fig. 2.7). Comparison of *Skolithos* species must be treated with care, however, owing to the poor definition of the genus. Lastly Silurian examples of *Arthrophycus* from the Acacus Formation are considerably larger than those from the Lower Devonian Tadrart Formation, although a range of scales and types is present in the Tadrart Formation (Fig. 2.6).

CHAPTER 3

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CHAPTER 3

THE FACIES OF THE TADRART AND OUAN KASA FORMATIONS (DEVONIAN), SOUTHWEST LIBYA

3.1 Introduction

3.1.1 Stratigraphy

In this chapter a facies study of the Devonian Tadrart and Ouan Kasa formations is presented. The Tadrart Formation comprises a thick succession of sandstones and very subordinate siltstones which forms the upper part of the blanket sandstone known as the External Tassilian Sandstones (Kilian, 1931) (Table 3.1). The lower part of this blanket sandstone is composed of the massive sandstones of the upper part of the Acacus Formation. The sandstones of the Tadrart Formation have not been dated on direct evidence in Southwest Libya but they are thought to be Siegenian/Emsian on indirect evidence (see Chapter 1, section 1.3.3). Equally the upper part of the Acacus Formation (units 14 & 15, Table 3.1) has not been dated on direct evidence (Klitzsch, 1965a, 1969) although it has been argued, again on indirect evidence, that it is not younger than Llandoverian (Klitzsch, 1969), of Ludlovian age (Freulon, 1964), or of Wenlockian/Ludlovian age (Bellini & Massa, 1980). The shales and subordinate sandstones and siltstones of the Ouan Kasa Formation, on the other hand, have been dated on the basis of a marine fauna of brachiopods as Middle Devonian (Eifelian) (see Chapter 1, section 1.3.3).

Klitzsch (1965a, 1969) measured sections through the Silurian to Middle Devonian succession of Southwest Libya which he established as the type sections for the Tanezzuft, Acacus, Tadrart and Ouan Kasa formations (Fig. 3.1) (Banerjee, 1980). He followed the divisions of the succession proposed by Desio (1936a, 1940a) for the Silurian and Burollet (1960) for the Devonian. Burollet (1960) based his type sections on the work of oil company geologists who had measured sections in the Jabal Tadrart (Collomb & Heller, 1958; Collomb *et al.*, 1961). They recognised a regional unconformity/disconformity surface at the base of a succession of "ruiniform" sandstones which they used to define the base of the Tadrart Formation. Freulon (1964) published the results of sections he measured in the Jabals Akakus and Tadrart at Wadi Takarkhouri and Takisset (Fig 3.2) and further south at In

J A B A L S A K A K U S A N D T A D R A R T						IN EZZAN	GENERAL SAHARAN TERMINOLOGY Beuf *et al.* (1971) Freulon (1964)
AGE (Jabals Akakus & Tadrart succession)		SEQUENCES T=Transgressive R=Regressive	LITHOLOGY	S T R A T I G R A P H Y			
				Klitzsch (1965a, 1969) & Banerjee (1980) (Takarkhouri Section)	Freulon (1964) (Takisset Section)	Freulon (1964), Beuf *et al.* (1971), Furst & Klitzsch (1963)	
MIDDLE DEVONIAN Eifelian (Ouan Kasa Fm.)		(OPEN MARINE)	shales, siltstones & sandstones	OUAN KASA FORMATION (43 m)	SERIE D'OHEBRAN	OUAN KASA FORMATION	
LOWER DEVONIAN TO UPPER SILURIAN		T4	Cross-stratified and channelled sandstones with subordinate siltstones	TADRART FORMATION (317 m) Units 16-24	SERIE DE TAKISSET (also, Série des Grès Rouge) (300 m)	GRES SUPERIEURS	EXTERNAL TASSILIAN SANDSTONE SUCCESSION (also GRES SUPERIEURS)
		T3					
		T2			IKNIOUEN HORIZON (1-3 m)		
		T1		ACACUS FORMATION (345 m) Units 14 & 15 (154 m)*	GRES A *HARLANIA* (150-250 m)	SERIE LITEE A SCOLITES	
LOWER	?Upper-Middle Llandoverian	R	sandstones siltstones & shales	Units 8-13 (191 m)*	ZONE DE PASSAGE	SERIE DE REMPLISSAGE GRES INFERIEURS PLISSES ET DECOLLE SERIE PSAMMITIQUE	INTRA-TASSILIAN SUCCESSION
SILURIAN	Lower Llandoverian	(OPEN MARINE)	shales, siltstones & subordinate sandstones	TANEZZUFT FORMATION (400 m) Units 2-7	ARGILES A GRAPTOLITES	ARGILES A GRAPTOLITES	

*Division of Acacus Formation after Klitzsch (1969)

Table 3.1 STRATIGRAPHIC CORRELATION CHART FOR SILURIAN TO MIDDLE DEVONIAN STRATA, JABAL AKAKUS, JABAL TADRART AND IN EZZAN

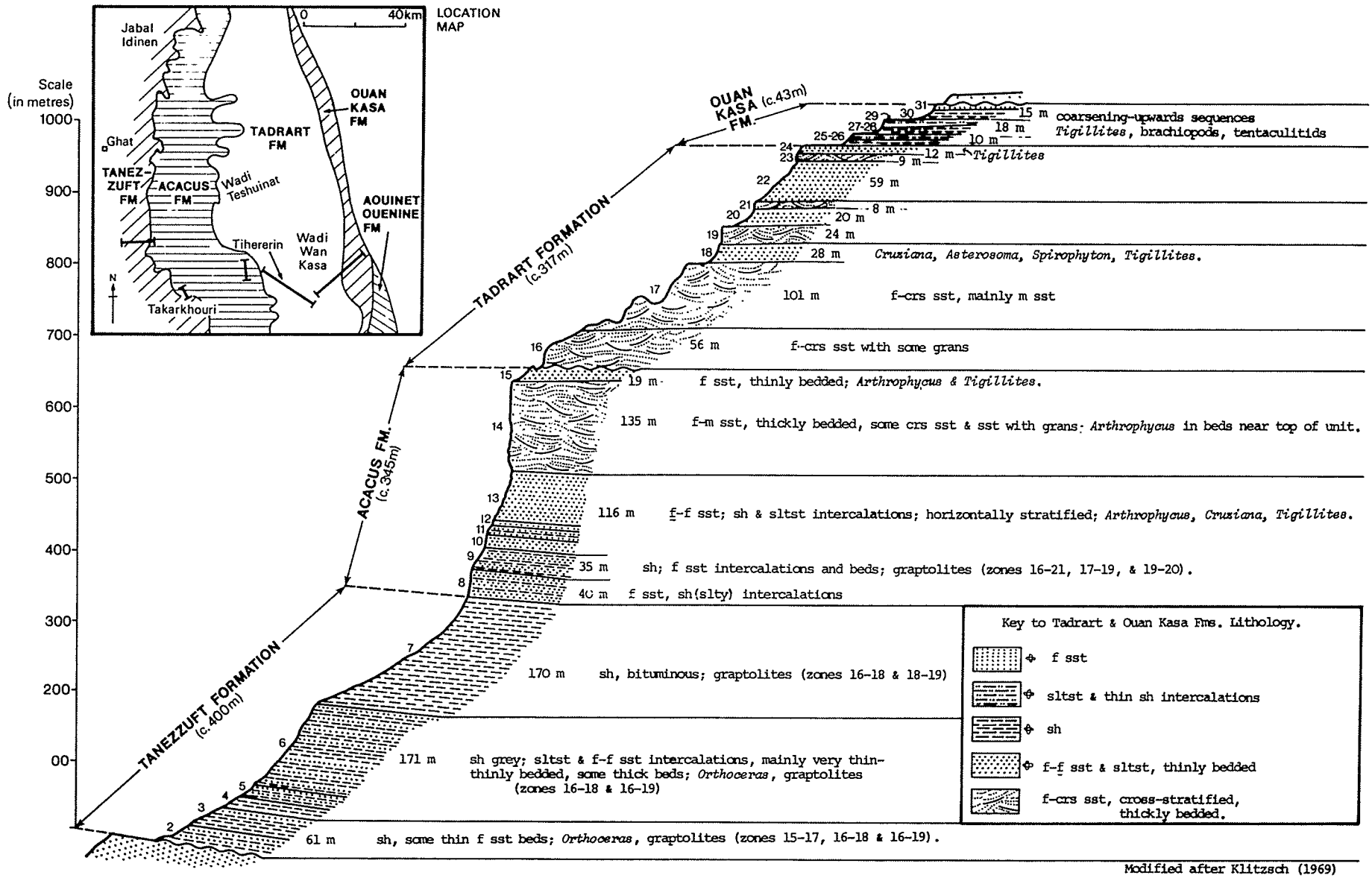


Figure 3.1 TYPE-SECTION OF SILURIAN TO MIDDLE DEVONIAN STRATA, WADI TANEZZUFT TO WADI WAN KASA.

Figure 3.2 LOCATION OF MEASURED SECTIONS, JABALS AKAKUS AND TADRART

Map constructed from Figures 1.7 and 1.9.

The author's measured sections (marked in red) and those reported by Bellini and Massa (1980), Collomb *et al.* (1961), Freulon (1964) and Klitzsch (1965a) give an indication of the availability of stratigraphic data in the area. The dashed lines show the correlations established between individual sections to make up five composite sections, the names of which are underlined in red.

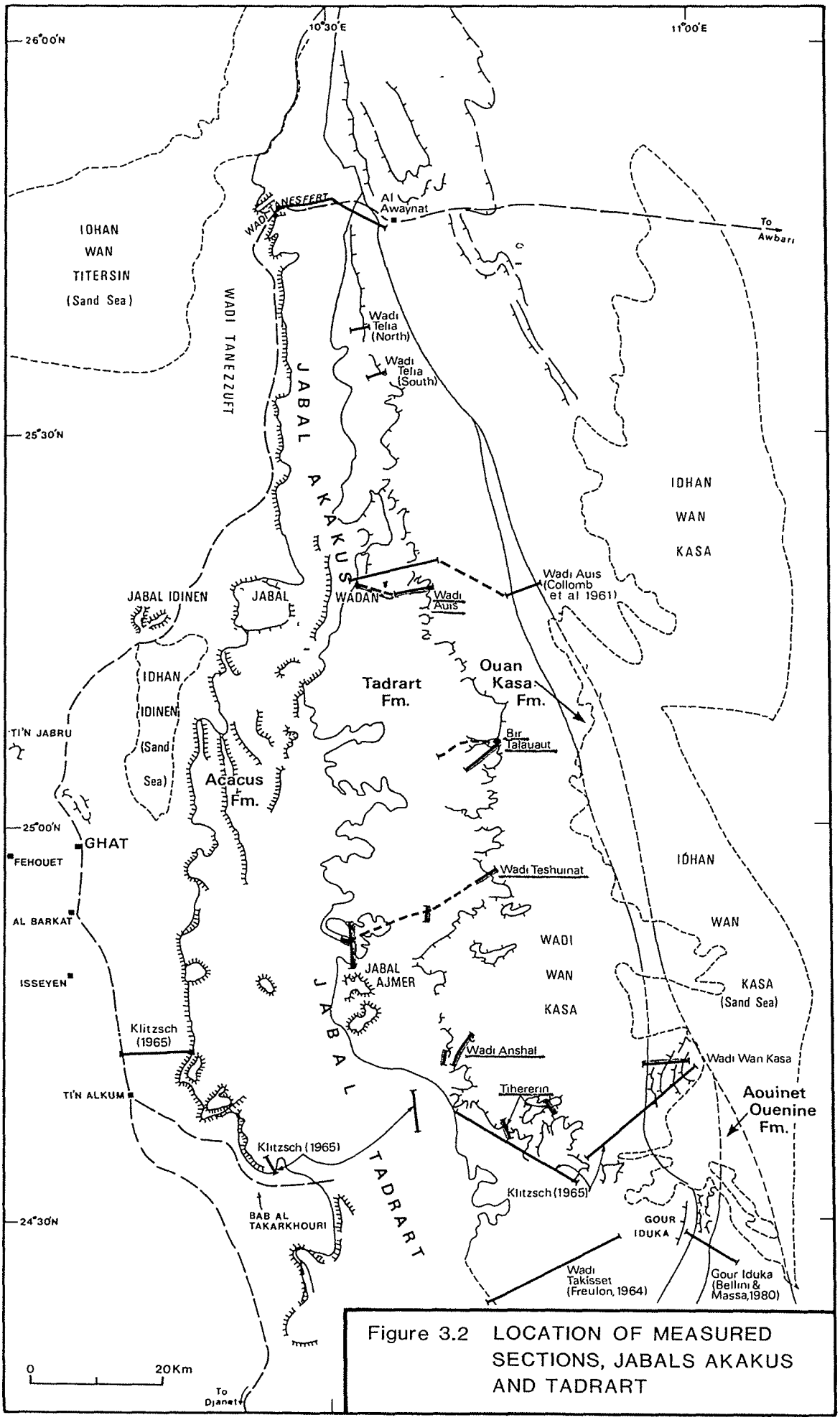


Figure 3.2 LOCATION OF MEASURED SECTIONS, JABALS AKAKUS AND TADRART

Ezzan, in which he divided the External Tassilian Sandstones into a lower unit (Grès à *Harlania*) and an upper unit (Série de Takisset, also Série des Grès Rouge), separated by an iron-oxide enriched bed, the Iknioven Horizon (Table 3.1). Freulon (1964), therefore, recognised two major stratigraphic breaks within the External Tassilian Sandstones, that at the base of the Grès à *Harlania*, and that of the Iknioven Horizon. In the author's field work two further important stratigraphic breaks were recognised, both within the Tadrart Formation, on the basis of which the External Tassilian Sandstones have been divided into four sequences (Table 3.1), each of which is characterised by a fining-upwards grainsize pattern and each of which lies above an unconformity/disconformity surface. The interpretation of the facies of these sequences indicated an upward transition from continental to marine deposition (see Fig. 3.6). Consequently they have been termed transgressive sequences (Table 3.1).

The recognition of a number of transgressive sequences and unconformity/disconformity surfaces in the External Tassilian Sandstones of the Jabal Tadrart suggests that this blanket sandstone has similar characteristics both in the Jabal Tadrart and in the Tassili N'Ajjer (Dubois *et al.*, 1969). Klitzsch (1965a, 1969, pers. comm. 1978) and Bellini and Massa (1980), however, claimed that only one major regional unconformity exists, that at the base of the Tadrart Formation. Additionally Bellini and Massa (1980, p. 29) argued that this unconformity is marked by a sharp grainsize break which clearly distinguishes the Tadrart Formation from the Acacus Formation and which represents the initiation of a cycle of sedimentation through to the Middle Devonian. Furthermore it has been argued that the unconformity at the base of the Tadrart Formation was caused by movements associated with the Caledonian Orogeny (Ardennian phase) (Furst & Klitzsch, 1963; Bellini & Massa, 1980), and that the Tadrart Formation, therefore, may be interpreted as a post-orogenic deposit (Bellini & Massa, 1980, p. 16). It was found that little evidence existed to indicate that the surface of unconformity and its correlative disconformity surface at the base of the Tadrart Formation was any more or any less significant than the other three similar surfaces recognised within the succession. It was neither clearly associated with any greater degree of angular truncation nor with any sharper grainsize transition than

any of the other unconformity/disconformity surfaces. Consequently, no evidence appears to exist to support the contention that the time break at the base of the Tadrart Formation was greater than at any of the other three unconformities. The evidence of four fining-upwards sequences lying above regional unconformity/disconformity surfaces is support for an episodic, pulse-like sedimentation pattern of epeirogenic or eustatic origin (or a combination of both) rather than support for a concept of pre- and post-Caledonian Orogeny sedimentation. Indeed, it is arguable whether the term "Caledonian Orogeny" is applicable to the tectonic history of North Africa and Arabia since the continent of which they were a part was involved in no continental collision or crustal rupture at the time of the closing of the Iapetus Ocean (Scotese *et al.*, 1979) (see Chapter 1). The similarity of the facies of the Tadrart Formation with those of the Internal and External Tassilian Sandstones of the Tassili N'Ajjer (Beuf *et al.*, 1971) suggests that the stable cratonic tectonic setting thought to be applicable to the interpretation of the Cambro-Ordovician blanket sandstones (Beuf *et al.*, 1971) is probably also applicable to these blanket sandstones.

When the first field sections were measured by the author, the base of a major sandstone bluff, which correlated with a massive sandstone unit at the base of the type section (Fig. 3.1, see Fig. 1.18), was taken as the base of the Tadrart Formation. It was observed, however, that the base of this sandstone unit lay conformably above thinly bedded sandstones and siltstones in some localities, whereas at others the boundary was a low-angle angular unconformity. It was also observed that similar unconformity/disconformity surfaces existed both stratigraphically higher, within the Tadrart Formation, and lower, within the Acacus Formation. Consequently, although the majority of the sections measured were measured on the exposures of the Tadrart Formation, some sections were also measured on the exposures of the Acacus Formation in order that the facies of the upper beds of this formation could be compared and contrasted with those of the Tadrart Formation. It was concluded that the thick, massive cross-stratified sandstone at the top of the Acacus Formation, and the thin overlying well-bedded succession (units 14 & 15, Fig. 3.1) display facies and vertical sequence characteristics which are

very similar to those of the Tadrart Formation, and quite distinct from those of the underlying well-bedded succession of sandstones, siltstones and shales of the lower part of the Acacus Formation (units 8-13, Fig. 3.1). The conviction grew, therefore, that the subdivision adopted by Freulon (1964) (Table 3.1) was a more valuable stratigraphic framework than that of Klitzsch (1965a, 1969). The reason that more than one regional unconformity/disconformity surface was not previously recognised is thought to be related to the nature of these surfaces. They are easily recognised where a well-bedded, fine-grained sandstone succession is truncated by massive sandstone of an overlying transgressive sequence. However, where a massive sandstone unit completely truncates a well-bedded sandstone succession down to an underlying massive sandstone unit, the boundary is very difficult to discern because of the similarity of the facies of the beds above and below the boundary. It is a characteristic of these massive sandstones that they display abundant scour and channel erosion surfaces. To recognise a regional unconformity/disconformity surface within a succession of the massive cross-stratified sandstones, therefore, it is necessary to distinguish this type of surface from penecontemporaneous channel and scour erosion surfaces. The only way this can be done is by lateral correlation of detailed measured sections. Ideally these would be spaced much more closely than the five composite measured sections of this study (Fig. 3.2) but, in the time available, it was not possible to obtain a denser grid of stratigraphic data. Nevertheless, the correlation of these composite sections provided strong evidence for the presence of four fining-upwards sequences each lying above a regional unconformity/disconformity surface. The geomorphological study of the Jabal Tadrart and Jabal Akakus (see Chapter 1, section 1.3.4) also provided support for the presence of four sequences in the External Tassilian Sandstones. The weathering of these sandstones in the study area into bluffs, slab walls and slick-rock slopes revealed the presence of an alternation of massive, laterally-persistent sandstone units and relatively fine-grained, thinly-bedded sandstone units. The massive sandstones have sharp basal boundaries, often erosive, and are persistent as stratigraphic horizons throughout the study area. They merge together in the north of the study area and are difficult to distinguish. This is

thought to be a reflection of greater truncation at each unconformity surface in this area caused by contemporaneous uplift of the Tihemboka Arch (see section 3.7).

Before describing the sections measured, and the facies recognised, a brief description of the Silurian to Middle Devonian succession exposed in the Jabals Akakus and Tadrart is given. The lowermost unit, the Tanezzuft Formation (Fig. 3.1), comprises a lower 60 m unit of grey and yellow shales with some thin fine-grained sandstone beds. This is overlain by a middle unit, 170 m thick, of grey shale with very fine to fine-grained sandstone and siltstone intercalations and thin to thick beds. The upper unit is 170 m thick and is composed of green-grey to black, bituminous and graptolitic shales. The occurrence of a 40 m thick fine-grained sandstone unit above the graptolitic shales is used to mark the conformable boundary of the Tanezzuft Formation with the Acacus Formation. The Tanezzuft Formation is of Lower Llandoverian age (Klitzsch, 1965a, 1969) based upon its graptolite assemblages, and is interpreted as being of open marine origin (Berry & Boucot, 1973a) (Table 3.1). The succeeding Acacus Formation is composed of:
a basal 40 m thick, white to grey sandstone, in which some silty shale intercalations are present, overlain by a 35 m thick grey, graptolitic shale. The shale is silty in parts of the succession and fine-grained sandstone intercalations are present which display flute marks, load casts and small-scale fining-upwards grainsize profiles. Above the shale lies a succession of brown to grey, very fine to fine-grained sandstones passing upwards to white sandstones. The sandstones and intercalations of shale and siltstone are well-bedded in thin to very thick bioturbated and ripple-laminated beds. Some interference rippled surfaces are present. The next unit upwards (unit 14, Fig. 3.1) comprises white to dark brown, fine to medium-grained and cross-stratified sandstones, 135 m thick, which generally weather to a massive bluff and are responsible for the vertical cliffs of the Jabal Akakus (see Fig. 1.12). The sandstones are thickly bedded and friable except where a strong ferruginous matrix is present. Unlike the succession beneath, biogenic structures are absent except in the uppermost beds where *Arthrophycus* is found. A 19 m thick grey, fine-grained and poorly-sorted, argillaceous sandstone lies above in thin beds (unit 15, Fig. 3.1). The sandstones are bioturbated and strongly

ferruginous, even "quartzitic" at the top, this being characteristic of the boundary between the Acacus and Tadrart formations (Klitzsch, 1965a, 1969). The succession beneath unit 14 comprises a regressive coarsening-upwards sequence of beds (Martindale, 1977) (see Chapter 1, section 1.3.3). It has been assigned to the middle and possibly upper Llandoverian by Klitzsch (1965a, 1969) on the basis of the graptolites in unit 9 (Fig. 3.1). No fossils were found in units 10 to 13 to provide any evidence of their age. Freulon (1964) described the equivalent beds at Takisset (Fig. 3.2) as the "Zone de Passage" (Table 3.1). He collected graptolites from a shale bed near the top of this succession which were dated as lower Ludlovian. This is in such stark contrast to the evidence of Klitzsch (1965a, 1969) and Massa and Jaeger (1971) that some doubt must be cast upon its validity. The Grès à *Harlania* beds lie above the Zone de Passage beds in Wadi Takisset and are the equivalent of units 14 and 15 (Fig. 3.1 & Table 3.1). These cross-stratified massive sandstones which pass upwards into well-bedded fine-grained sandstones and siltstones are the beds of the lowest transgressive sequence of the External Tassilian Sandstones. The only fossils found in these beds were reported by Freulon (1953a, 1955). He found "Gigantostraces" (the traces of eurypterids) and *Pterygotus* in the Grès à *Harlania*, on the basis of which he argued that these beds are Silurian in age (Freulon, 1964). The value of these fossils, however, is very limited for precise age determination.

The base of the Tadrart Formation, and the Iknioun Horizon at the base of the Série de Takisset mark the base of the second transgressive sequence (Table 3.1). Klitzsch (1965a, 1969) described this boundary as an angular unconformity, dividing a succession which weathered to rounded jabals lying to the west (and stratigraphically below) from one which weathered to "ruiniform" jabals to the east (Figs. 3.1 & 3.2). He noted that the bed lying immediately above the boundary is commonly an iron-oxide cemented and silicified argillaceous sandstone, although a white silty shale up to 1.2 m thick may also be present. Freulon (1964) described the Iknioun Horizon as a one to three metre thick ferruginous crust of red and brown micaceous sandstone which commonly displays cylindrical ferruginous Liesegang bands. He noted that the Iknioun Horizon is one of a number of ferruginous crusts developed within the External Tassilian Sandstones,

associated with minor surfaces of erosion which display some relief, but are not associated with any angular unconformity. He interpreted these horizons as palaeosoils resulting from hiatuses and minor localised uplift caused by Ardennian phase epeirogenic movements (Freulon, 1964). Adjacent to structurally positive arches, however, such as the Tihemboka Arch, Freulon (1955, 1964) noted that these ferruginous crusts may be locally associated with angular unconformities of up to 7°. In contrast to the view held by Klitzsch (1965a, 1969) that the boundary is recognisable on the basis of a geomorphological change, Freulon (1964) observed that, without the particularly strongly developed "Iknioven Horizon" ferruginous crust, the sandstone succession of the Grès à *Harlania* would be difficult to distinguish from that of the "Série de Takisset". It was found in the author's field work that geomorphological boundaries could provide very misleading impressions about the nature of the stratigraphic boundaries within these weathered strata (Chapter 1, section 1.3.4).

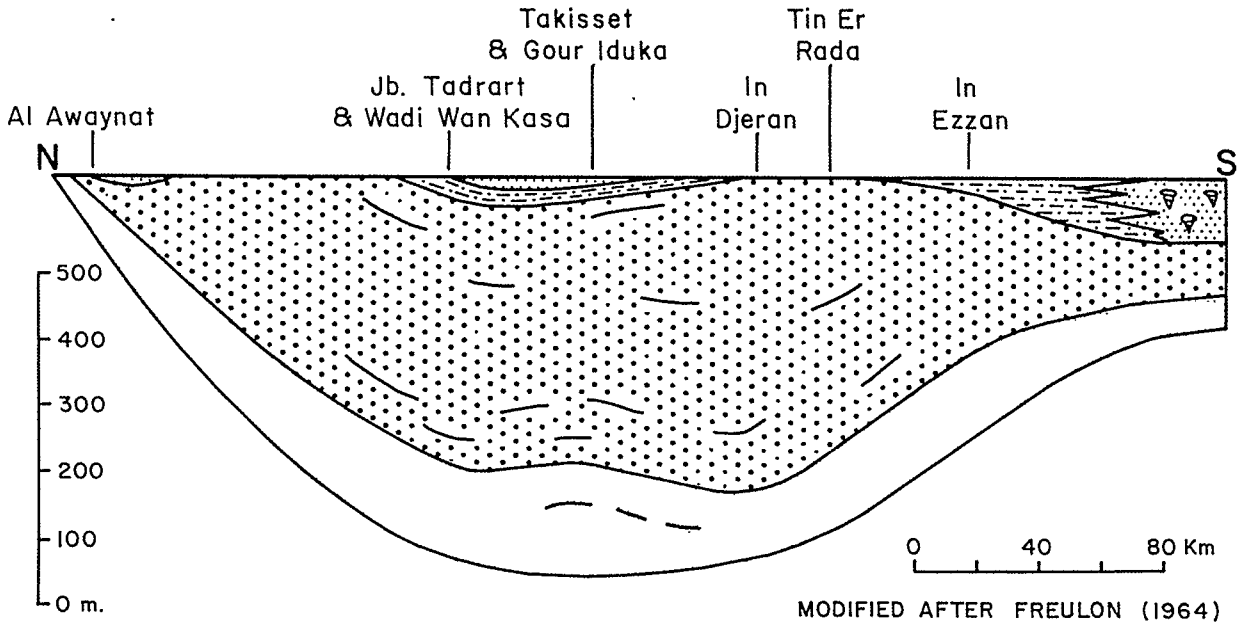
In the type section, the Tadrart Formation comprises approximately 317 m of white to light brown and orange cross-stratified fine to coarse-grained sandstones in the lower part, and well-bedded fine-grained sandstones in the upper part (Klitzsch, 1965a, 1969) (Fig. 3.1). Plant remains have been reported from the lower beds, and the upper beds are bioturbated and contain abundant trace fossils (*ibid.*). A review of the plant remains found in the Lower Devonian of Libya was given by Lejal-Nicol and Massa (1980). The Tadrart Formation comprises an alternation of two main lithological assemblages already encountered lower in the section in units 14 and 15 of the Acacus Formation: a thickly-bedded, cross-stratified sandstone succession which weathers to bluffs, slab walls, mesas, buttes and rock stacks, and a thinly-bedded and bioturbated succession of fine-grained sandstones and siltstones which weathers to slopes, commonly covered with scree (Figs. 3.1 & 3.3). The sandstones are white to light-brown, dark brown and grey. The first unit of the formation comprises fine- to coarse-grained sandstones (with granules occasionally present in the matrix) which are generally friable except where a strong ferruginous matrix is present. This unit is capped by a 50 cm thick, red, silicified siltstone. Another thickly-bedded unit lies above, comprising 101 m of cross-stratified sandstone, mainly medium-grained and containing

Figure 3.3 JABAL TADRART : BLUFFS AND SLOPES

Major bluff developed in the basal sandstones of transgressive sequence 3, Middle Wadi Teshuinat (97-173 m stratigraphically above the base of the Tadrart Formation). The left hand arrow marks the base of the third transgressive sequence. It also indicates the location of the measured sections illustrated in Figures 3.19 and 3.20. The right hand arrow indicates the position of a major erosion scour surface illustrated in Figure 3.62. Scree-strewn slopes and slick rock slopes are developed in the upper beds of the second transgressive sequence. The plateau at the top of the mesa is thought to be a Tertiary peneplain surface.

rare ferruginous beds. There follows an alternation of the thinly-bedded and thickly-bedded lithological assemblages. Unit 18 is characterised by an abundance of trace fossils including *Cruziana*, *Asterosoma*, *Spirophyton* and *Tigillites*, and units 19 and 21 are characterised by alternating fine to coarse-grained sandstones with granules, and fine to medium-grained sandstones. The fine-grained sandstones and siltstones of unit 22 are characterised by poor sorting and an argillaceous matrix. Unit 23 is the last thick-bedded unit in the succession, comprising fine to coarse-grained sandstones, before the succession of sandstones and siltstones passes up into the predominantly argillaceous Ouan Kasa Formation. It is noteworthy that the thin-bedded assemblage of fine-grained sandstones and siltstones gains predominance over the coarser-grained, thick-bedded units upwards in the succession and, therefore, the formation may be considered to represent, in gross terms, a fining-upwards sequence, possibly of transgressive origins (Fig. 3.1, see Fig. 1.11). However, the presence of fine to coarse-grained and granular sandstones in the thick-bedded units towards the top of the succession indicates that a number of grainsize cycles are present. Freulon (1964) described the equivalent beds in Wadi Takisset as comprising a 300 m thick succession of coarse-grained, cross-stratified sandstones in the lower part which pass upwards into a finer-grained succession of sandstones displaying some rippled surfaces. This thickness for the Série de Takisset corresponds well with the thickness of 317 m for the sandstones of the Tadrart Formation measured in the same area by Klitzsch (Fig. 3.2).

Freulon (1964) measured sections through the Upper Silurian and Devonian successions at at least six different localities between Al Awaynat and south of In Ezzan (Fig. 3.4, see Fig. 1.5) and showed that major lateral thickness variations exist, largely in the Série de Takisset, but also in the Grès à *Harlania*. These lateral thickness variations were also noted by Collomb and Heller (1958) and Collomb *et al.* (1961). The outcrop pattern of the Acacus and Tadrart formations (Figs. 3.3 & 3.4) and the topography of the Jabals Akakus and Tadrart reflect the lateral variation in thickness of the two formations, the width of outcrop being greatest (up to 50 km) and the most rugged jabals being located where the formations are thickest just



Grès a bois
Grès à *Spirophyton* } MIDDLE & UPPER DEVONIAN
Série d'Ohébran — MIDDLE DEVONIAN
Série de Takisset
Grès à *Harlania* } LOWER DEVONIAN to UPPER SILURIAN : External Tassilian Sandstones

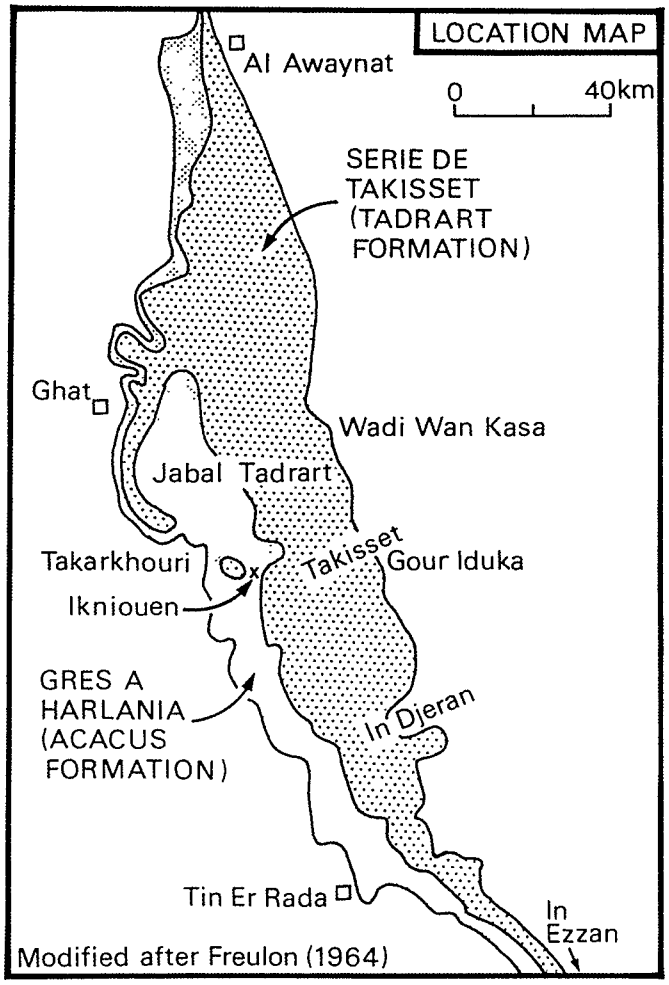


Figure 3·4 LATERAL THICKNESS VARIATION IN EXTERNAL TASSILIAN SANDSTONES, SOUTHWEST LIBYA.

north of the Bab al Takarkhouri. The outcrop pattern is also, however, a reflection of both the truncation of the Tadrart and Acacus formations in the north by the predominantly argillaceous sediments of the "Chatti Formation" and the Carboniferous succession, and the presence of a structural flexure southeast of Ghat (cf. Figs. 1.9 & 1.10).

The Ouan Kasa Formation is thought to lie conformably above the Tadrart Formation and, in the type section, comprises 43 m of shale with argillaceous siltstone and fine-grained sandstone intercalations (Klitzsch, 1965a,1969). The contact between the Tadrart and Ouan Kasa formations is poorly exposed, however, and it has been suggested that a conglomeratic bed lying above an erosion surface patchily exposed at Gour Iduka, south of Wadi Wan Kasa (Fig. 3.2), may indicate an unconformity at this boundary (Massa, pers. comm. 1978). Bellini and Massa (1980, pp. 29-30) nonetheless argued that the Tadrart and Ouan Kasa formations represent a fining-upwards "evolutionary cycle". Three lithologies are present in the Ouan Kasa Formation all of which are bioturbated: grey shales with gypsum beds and some silty-horizons, light grey to grey siltstones with thin shale intercalations, and thinly-bedded, fine-grained sandstones (Fig. 3.1). These are arranged in coarsening-upwards sequences, in the lower two of which brachiopods, styliolines and tentaculitids were collected by Klitzsch (1965a,1969). In siltstones of the lower sequence he recorded the presence of *Strophomena carianata* CONRAD, and in the middle sequence, *Chonetes orientalis* var. *admixta* COWPERREID and *Mucrospirifer mucronatus* CONRAD. Some dispute exists about the age of the marine fauna present in the Ouan Kasa Formation. Borghi and Chiesa (1940) first described the fauna and assigned it to the Middle Devonian (Eifelian). This was later supported by Burollet and Manderscheid (1967) and Hinsch (in Klitzsch, 1969). Burollet and Manderscheid (1967) reviewed the palaeontological data on the Ouan Kasa Formation and concluded that it was diachronous, being Emsian in the Ghadamis Basin and Eifelian in the Ghat region. Dubois *et al.* (1969) argued that it is of Siegenian/Emsian age in the Polignac Basin. Nevertheless Bellini and Massa (1980) argued that because the succession at Awaynat Wanin, on the southern border of the Ghadamis Basin (see Fig. 1.5) is well studied and because the

Ouan Kasa Formation is Emsian in that locality, the Eifelian age determined for the Ghat region exposures must be erroneous. Until their proposed new type section through the Devonian succession at Gour Iduka is published with direct evidence that the fauna of the Ouan Kasa Formation is Emsian at that locality, it cannot be accepted that they have disproved the Eifelian age determination. If Burollet and Manderscheid (1967) were correct in their claim that the Ouan Kasa Formation is diachronous then it is likely that the Tadrart Formation is also diachronous. Evidence also exists that the Silurian Tanezzuft and Acacus formations are diachronous across western Libya, the arenaceous beds of the Acacus Formation lying above progressively younger beds towards the north, the whole succession representing, therefore, a northwards prograding system (Massa & Jaeger, 1971). The arguments of Bellini and Massa (1980) and others (see Chapter 1, section 1.3.3) about the age of the External Tassilian Sandstones depended upon age determinations of lithostratigraphical equivalents to the north and northwest in the Ghadamis and Polignac basins (see Fig. 1.5) and upon age determinations of interfingering shales with these lithostratigraphical equivalents. Additionally, these arguments depended upon the ages of the youngest known underlying Silurian beds and the oldest known overlying Devonian beds, as well as evidence for the age of the Ardennian phase of the Caledonian Orogeny which, it is argued, demonstrably changed the nature of sedimentation. The evidence available suggests that the Zone de Passage beds (units 8 to 13, Table 3.1) are upper to middle Llandoverian, and the Ouan Kasa Formation is Eifelian. Given the very nature of lithostratigraphical correlation and the uncertainty about the influence of Caledonian events in this area, the Siegenian/Emsian age proposed for the Tadrart Formation and the Wenlockian to Ludlovian age for the Grès à *Harlania* (Bellini & Massa, 1980) must be treated with some caution. It would seem prudent to consider the External Tassilian Sandstones in the Ghat region as upper Silurian (Wenlockian to Ludlovian) to Lower Devonian (Table 3.1), until further work is undertaken on the age of these sediments.

3.1.2 Measured Sections

The Tadrart and Ouan Kasa formations strike north-south in the study area (Fig. 3.2) and are truncated in the north by a Late Devonian unconformity. The northernmost outcrop of the Ouan Kasa Formation is 50 km south of Al Awaynat, whilst the Tadrart Formation extends to a point just north of Al Awaynat before it is entirely truncated by the "Chatti" and M'rar formations. In the south both formations extend into Algeria and Niger (Furst & Klitzsch, 1963; Plauchut & Faure, 1959). In this study, however, it was not possible to visit the outcrops of the two formations south of the latitude 24°30'.

The Akakus escarpment in the study area may only be penetrated from the west through to the exposures of the Tadrart Formation by two routes: the Wadi Tannesfert (to the west of Al Awaynat, Fig. 3.2), and the connected wadis of Wadi Takarkhouri and Wadi Ahar (Fig. 3.5). It was not possible to use the latter approach route as a means of access to the eastern exposures, from Ghat, since part of the Wadi Takarkhouri lies within the politically sensitive "no man's land" of the border between Libya and Algeria. Access to the exposures of the Tadrart Formation from Ghat was, therefore, achieved via Wadi Tanezzuft, Wadi Tannesfert and Al Awaynat and then across sand and gravel plains (sarir) and along shallow, sand-clogged dry wadis to Bir Talauaut (80 kms to the south) where a small waterwell exists.

The best topographic and geological maps of the Jabals Akakus and Tadrart available to the author when the field work was undertaken were those of the United States Army Map Service ("Fort Charlet" area, 1955, 1:1,000,000 scale) and of Collomb and Heller (1958) respectively. The geological map of Collomb and Heller (1958), at a scale of 1:200,000 was based upon previous maps of the area, mainly at 1:1,000,000 scale, and upon photogeological interpretation of aerial photographs backed up by field observations. Detailed structural sections could not be produced, however, as data on topographic elevations were sparse. Consequently, although dip measurements were taken, the mapping of formation boundaries depended more on photo-geological character than upon the calculation of bedding inter-sections with topographic relief. No sufficiently accurate contour map existed, therefore, to make estimates of stratigraphic positions within the Tadrart Formation from values of elevation. Experience

Figure 3.5 JABAL TADRART LOCATION MAP

Detailed map of Jabal Tadrart to show measured section localities (numbers 1 - 18). Where sections were measured over a considerable distance these are shown with a solid bar. The section measured in the Ouan Kasa Formation (locality 18) could not be accurately located on the map and this is shown as a broken bar. Similarly the type sections measured by Klitzsch (1965a, 1969), which are also shown on this map, are only approximately located. These sections are distinguished from those measured by the author by the letter "K".

Localities with reasonably well established names are listed below:

2 : Tadrart Rock Arch
9 : Wan Muhuggiag
10 : Wan Amil
12 : Ruin Mountain (see Fig. 1.13, upper)
14 : Arc de Triomphe (see Fig. 3.37)

Map constructed from NASA satellite image of the area (approximate scale 1:250,000) (see Fig. 1.15).

Additional sources: Collomb and Heller (1958), Collomb *et al.* (1961), Craig *et al.* (1958), Freulon (1964), Klitzsch (1965a, 1969) and Mori (1965).

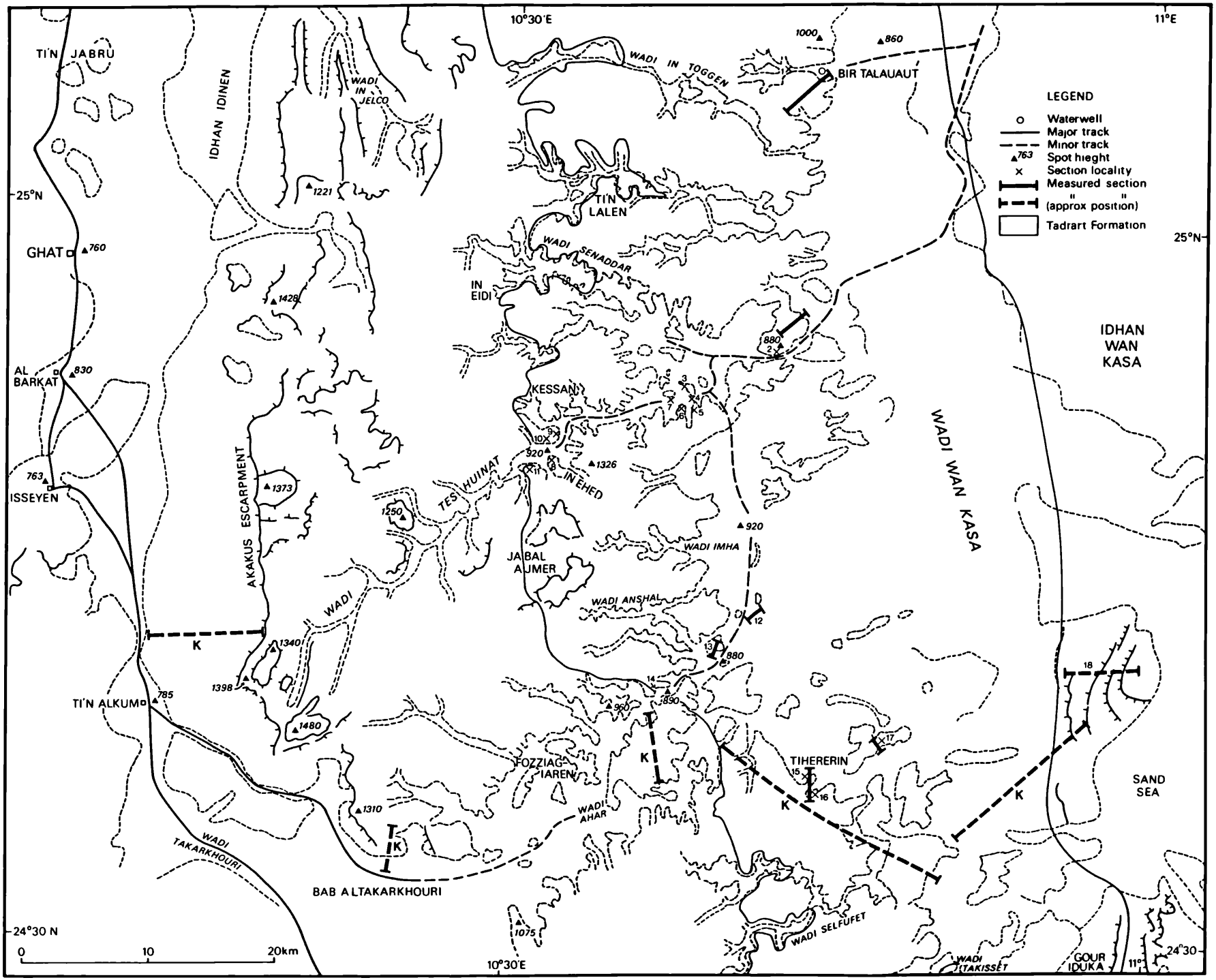


Figure 3.5 JABAL TADRART LOCATION MAP

with barometric altimetres in deserts, where daily pressure fluctuations are considerable, led the author to the belief that the results of the use of an altimeter would be ambiguous and of little value. The use of an Abney level in measuring sections is thought to have provided reasonably accurate section thicknesses, however. The Landsat Satellite image of the study area (see Fig. 1.15) at an approximate scale of 1:250,000, although not available at the time of the field work, was subsequently of great use as a "base map", since wadi systems stood out more clearly on this than on the photogeological map, and it served as the basis for the construction of Figures 1.7, 1.8, 1.9, 3.2 and 3.5.

Locating measured sections on the Collomb and Heller (1958) base map in the field was initially based upon compass bearings and kilometer readings from the odometer of the landrover. After some time had been spent exploring the different wadi systems three major known reference points were established, Bir Talauaut, Wadi Teshuinat and the eastern extension of Wadi Takarkhouri, where it was possible to recognise some of the features described in the type section of the Acacus and Tadrart formations (Fig. 3.1). Since the wadi systems were filled with sand and existing landrover tracks followed circuitous courses in almost all directions, estimates of directions and distances travelled proved to be highly inaccurate. It was only by establishing a network of known major topographic features, recognised on the photogeological map and later on the satellite image, outwards from known reference points, that it proved possible to locate the measured sections to a fair degree of accuracy (say to within 1 km). A local Tuareg beduin acted as a guide and helped with route finding and naming wadis and massifs.

A total of at least 1200 m of section in the Tadrart Formation was measured in the Jabal Tadrart and Jabal Akakus in five separate composite sections (Fig. 3.2). In the northern two of these, Wadi Auis and Bir Talauaut, the basal part of the Tadrart Formation was not measured. At Bir Talauaut the base was thought to be 90 m or less below the base of the section measured. At Wadi Auis, although not measured in detail, the basal 90 m was visible (see Fig. 1.18). The next composite section southwards lies approximately 24 kms to the south of Bir Talauaut and was measured in three separate sections

in Wadi Teshuinat, along 21 km of the wadi (Fig. 3.5). The stratigraphically highest part was measured at locality 2. The middle part of the Tadrart Formation was measured at localities 3-7, and the lowest part of the formation at localities 8-11. The Wadi Anshal composite section lies 20 kms to the southeast of Wadi Teshuinat (localities 12 & 13) and the southernmost composite section lies a further 15 to 22 kms to the southeast at Tihererin (localities 15-17). South of the Bir Talauaut section all the composite sections were measured from the base of the formation. The top of the formation, however, is only poorly exposed, in the Wadi Wan Kasa and at Gour Iduka, and the contact between the Tadrart Formation and the overlying Ouan Kasa Formation was not observed. In addition to the major composite sections measured in the Tadrart Formation, short sections were also measured in the Tadrart Formation at Al Awaynat, at a point 12 kms south of Al Awaynat (Wadi Telia North), and at Wadi Telia South (Fig. 3.2). Additional sections were measured at Wadi Wan Kasa (in the Ouan Kasa Formation, locality 18), and in Wadi Ahar, Upper Wadi Teshuinat and Wadi Tanezzuft (all in the Acacus Formation) (Figs. 3.2 & 3.5). In the composite sections the practise adopted to achieve a full vertical section was to allow considerable overlap of stratigraphic intervals in order to allow correlation of separate sections into a composite. A number of separate vertical sections were measured in the Wadi Teshuinat, Wadi Anshal and Tihererin composites over intervals of particular interest. In establishing composite sections it was usually possible to trace stratigraphic horizons down-dip eastwards along approximately east-west orientated wadis so that up to three different measured vertical sections could be combined to produce a composite section.

3.1.3 Schematic Vertical Succession

Based largely on the composite sections of Wadi Anshal and Tihererin (Fig. 3.5) a schematic vertical section has been drawn up to show the succession from the lowest major thick-bedded and cross-stratified unit (unit 14 of the Acacus Formation) up through the alternations of thinly- and thickly-bedded units of the Tadrart Formation to the marine succession of the Ouan Kasa Formation (Fig. 3.6). This section, therefore, illustrates the main characteristics of the four transgressive sequences of the External Tassilian

Figure 3.6 SCHEMATIC VERTICAL SECTION THROUGH THE ACACUS, TADRART AND OUAN KASA FORMATIONS, JABAL TADRART.

NOTE:

Apart from transgressive sequence 1, this vertical section is based upon the author's measured sections at Wadi Wan Kasa (locality 18, Fig. 3.5), where a section through the sediments of the Ouan Kasa Formation was measured, and sections at Wadi Anshal (localities 12, 13 & 14, Fig. 3.5) and Tihererin (localities 15, 16 & 17, Fig. 3.5). The composite sections established for Wadi Anshal and Tihererin were correlated with each other (see Fig. 3.54) to provide a generalised and schematic, yet characteristic, vertical succession through transgressive sequences 2, 3 and all but the uppermost part of sequence 4. The section of transgressive sequence 1 is taken from the type section of the Acacus Formation (Klitzsch, 1965a, 1969) and is assigned facies on the basis of the author's observations in Wadi Ahar (Fig. 3.5) and on work undertaken jointly with Martindale (1977).

Detailed actual measured sections from Tihererin and Wadi Anshal are presented in sections 3.2-3.5 (this Chapter) to illustrate the characteristics of the different facies.

Sandstones. Each sequence displays a fining-upwards grainsize pattern, and is bounded at the base by an unconformity/disconformity surface which is commonly marked by ferruginous sandstones immediately above and below the surface.

The massive sandstones of sequence 1 generate the largest bluffs and cliffs of the succession and are characterised by major subhorizontal internal scour horizons. It should be noted that this sequence has been illustrated with a break in the vertical scale on Figure 3.6 for reasons of space. It appears thinner than it really is, therefore. The equivalent unit in sequence 2 is broken by a remarkably laterally persistent plateau at 40 to 45 m above the base of the Tadrart Formation (see Fig. 1.18) beneath which slab walls and slick-rock slopes are present in a homogeneous succession of sandstone. Above this plateau, which is developed along a thin fine-grained ferruginous sandstone horizon containing, in places, *Arthrophycus* and *Skolithos*, the succession varies somewhat. Where fine-grained partings are present, an irregularly-weathered steep slope is developed composed of small slab walls, slick-rock slopes and rubble-strewn slopes. These are developed on weathered surfaces of well-bedded fine-grained sandstones with thin argillaceous siltstone intercalations. Where the succession above the 40 m plateau is more homogeneous, slick-rock slopes and ramps are developed (see Fig. 1.19). At a stratigraphic position 90-127 m above the base of the Tadrart Formation the massive sandstones of sequence 3 form a major bluff or slab wall lying, usually erosively, above a thinly-bedded succession (Fig. 3.3). The massive sandstones of sequence 3 pass upwards into a thinly-bedded succession of up to 80 m thickness which is generally weathered to a series of rubble-strewn slopes. A final sandstone bluff (termed the "Sandstone Marker Bluff"), lying above this thinly-bedded succession at 200-240 m up in the Tadrart Formation, is thinner than the lower two, but is persistent throughout the south of the study area. In the north the thinly-bedded horizons are thinner or absent and the whole succession is less well differentiated geomorphologically. However, the sandstones of the "Sandstone Marker Bluff" may be tentatively identified on the basis of their sedimentological characteristics in these northern sections.

In this study of the four sequences and the relationships between

them it is shown that the lower three have similar palaeocurrent orientations whereas the uppermost sequence displays an enigmatic palaeocurrent pattern indicating, perhaps, some change in structural and sedimentological environment (section 3.7).

A tentative correlation between the schematic vertical section and the type section (Fig. 3.1) is proposed in the right hand column of Figure 3.6. It seems likely that the base of sequence 3 was not recognised in the type section because of truncation of the thin beds at the top of sequence 2 in that section.

The lateral continuity of major cross-stratified sandstone bodies and of the "40 m plateau" is a striking characteristic of the succession which allows correlation and makes the recognition of sequences possible. However, not all the cross-stratified sandstone bodies are similarly laterally continuous and many can be seen to thin and pinch out over a distance of 50 m or so, being enveloped in the bioturbated sandstones and siltstones of the thinly-bedded units. Correlation with the type section suggests that units 19 and 23 may be of this type. The cross-stratified sandstone succession at the base of sequence 4 (unit 21, Fig. 3.1), although thinner than those of sequences lower in the succession, is remarkably continuous, however, creating a recognisable "Sandstone Marker Bluff" in the rubbly, scree-strewn slopes characteristic of the exposure of the upper beds of the Tadrart Formation.

The thick-bedded and cross-stratified sandstones have been divided into two broad facies types, the multi-storey cross-stratified sandstone facies (facies MSd) and the channelled sandstone and siltstone facies (facies CSdSt) interpreted as the sandstones of fluvial and estuarine complex channel systems. The thin-bedded and bioturbated sandstones have also been divided into two broad facies types, the rippled and bioturbated sandstone and siltstone facies (facies RBSdSt) and the vertically-burrowed sandstone facies (facies VBSd). The latter has been tentatively interpreted as the deposits of tidal flats whilst the former is thought to have been deposited in a shallow-marine environment. Two facies have been recognised in the Ouan Kasa Formation: massive shales (facies MSh) and interbedded shales, sandstones and siltstones (facies IShSdSt), the deposits of an open marine environment. The four sequences recognised in the succession

represent, therefore, transgressive sequences of fluvial and estuarine channel deposits passing upwards in each case to sediments showing evidence of stronger marine influence and weaker detrital supply to the depositional setting. In the fourth sequence the transgression is complete with the deposition of open marine sediments. In the second sequence a transition upwards from sediments showing the characteristics of fluvial deposits (facies MSd) to those showing predominantly estuarine channel characteristics (facies CSdSt) occurs at the "40 m plateau" break. In sequences 1 and 3, facies MSd and CSdSt are more complexly related, whilst in sequence 4 facies MSd is virtually absent.

A description and interpretation of these facies follows in the next sections of this chapter. The finer-grained facies of the Tadrart Formation (facies RBSdSt and VBSd), and the facies of the Ouan Kasa Formation (Fig. 3.6) are given a less detailed treatment than facies MSd and CSdSt since these coarser-grained sediments comprise the characteristic facies of Nubian-type sandstones. Finally an account of the relationships of the facies to each other and to the transgressive sequences is given in section 3.7.

3.2 Multi-Storey Cross-Stratified Sandstone Facies (MSd)

3.2.1 Description

3.2.1.1 General

The sandstones of the multi-storey cross-stratified sandstone facies are quartz arenites which are mature mineralogically but are usually immature texturally. Sorting is generally poor although some well sorted sandstones are present and they are generally fine to medium-grained. Some very fine-grained sandstones and some coarser-grained sandstones are present and very coarse quartz sand grains and granules as well as argillaceous siltstone intraformational clasts are commonly present as lags to foresets and sets. In a few cases intraformational sandstone clasts of up to 30 cm diameter were observed. The sandstone grains are usually subangular to subrounded, the larger grains commonly being better rounded than the smaller ones. The sandstones are fawn, orange-buff and brown in colour, are commonly extremely friable (disintegrating in sample bags) and poorly cemented.

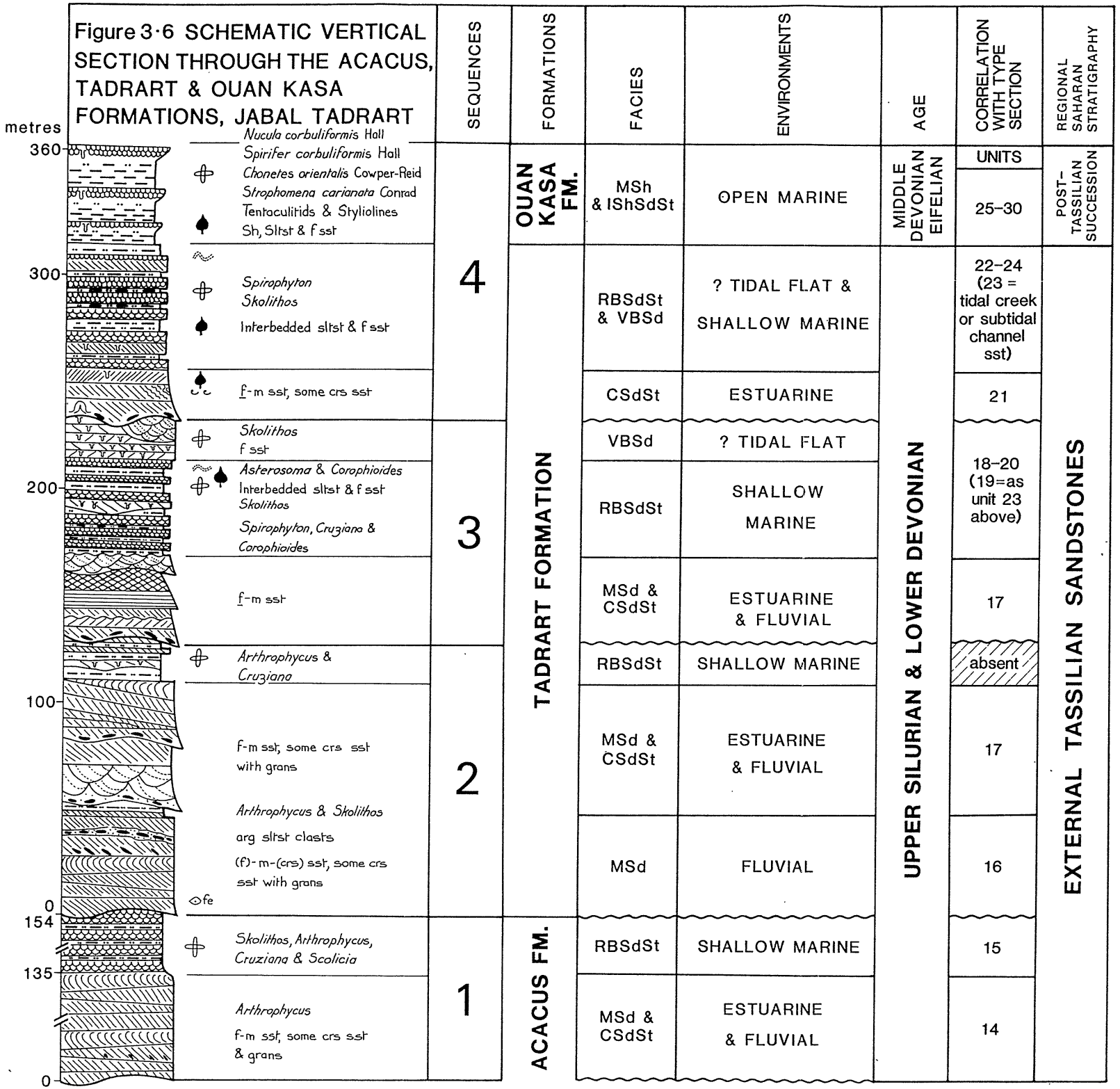


Figure 3·6 SCHEMATIC VERTICAL SECTION THROUGH THE ACACUS, TADRART & OUAN KASA FORMATIONS, JABAL TADRART
metres
360
300
200
100
0
154
135
0
Nucula corbuliformis Hall
Spirifer corbuliformis Hall
Chonetes orientalis Cowper-Reid
Strophomena carinata Conrad
Tentaculitids & Styliolines
Sh, Sltst & f sst
Spirophyton
Skolithos
Interbedded sltst & f sst
f-m sst, some crs sst
Skolithos
f sst
Asterosoma & Corophioides
Interbedded sltst & f sst
Skolithos
Spirophyton, Cruziana & Corophioides
f-m sst
Arthrophycus & Cruziana
f-m sst, some crs sst with grans
Arthrophycus & Skolithos
arg sltst clasts
(f)-m-(crs) sst, some crs sst with grans
fe
Skolithos, Arthrophycus, Cruziana & Scolicia
Arthrophycus
f-m sst, some crs sst & grans
SEQUENCES
4
3
2
1
FORMATIONS
OUAN KASA FM.
TADRART FORMATION
ACACUS FM.
FACIES
MSh & IShSdSt
RBSdSt & VBSd
CSdSt
VBSd
RBSdSt
MSd & CSdSt
RBSdSt
MSd & CSdSt
MSd
RBSdSt
MSd & CSdSt
ENVIRONMENTS
OPEN MARINE
? TIDAL FLAT & SHALLOW MARINE
ESTUARINE
? TIDAL FLAT
SHALLOW MARINE
ESTUARINE & FLUVIAL
SHALLOW MARINE
ESTUARINE & FLUVIAL
FLUVIAL
SHALLOW MARINE
ESTUARINE & FLUVIAL
AGE
MIDDLE DEVONIAN EIFELIAN
UPPER SILURIAN & LOWER DEVONIAN
CORRELATION WITH TYPE SECTION
UNITS
25–30
22–24 (23 = tidal creek or subtidal channel sst)
21
18–20 (19=as unit 23 above)
17
absent
17
16
15
14
REGIONAL SAHARAN STRATIGRAPHY
POST-TASSILIAN SUCCESSION
EXTERNAL TASSILIAN SANDSTONES

3.2.1.2 Sedimentary structures, bedding and palaeocurrents

Sets of planar tabular and wedge cross-strata are commonly stacked in multi-storey cosets of up to 8 m thickness. Sets are generally approximately 50-70 cm thick and show a slight tendency to be thinner upwards in the coset. They are commonly up to 1 m thick and exceptionally up to 2.8 m thick. Their basal boundaries may be clearly erosive or merely planar and abrupt. Grainsize grading within cosets is generally more marked within the foresets of each set than between sets, although it is common to find finer-grained beds at the top of a coset of cross-strata. Reactivation surfaces and recumbent-folded deformed cross-strata are commonly present punctuating the succession of planar cross-strata (Figs. 3.7, 3.8 & 3.9). The axis of the fold of the deformed cross-strata commonly occurs high up in the set and gets progressively lower in the set in a "down-current" direction as the deformed foresets pass laterally into soft-sediment deformed stratification. This is not always the case, however (Fig. 3.8), and it was found that a great variety of types of soft-sediment deformation characterises these structures. "Diapiric" structures are another common form of soft-sediment deformation (Fig. 3.8). In some cases deformed strata may be the dominant stratification type in both this facies, and in facies CSdSt, for up to 10 m of vertical section. It is usually associated with the coarser-grained sandstones of the succession. Beds which appear to be lacking obvious internal structure, such as that marked with an arrow in Figure 3.9 (upper picture), often prove on investigation laterally to be deformed beds "down-current" from recumbent-folded cross-strata.

Planar cross-stratified sets commonly have no top set or bottom set and are composed of angular foresets inclined to the horizontal at generally between 20^{o} and 27^{o}, although angles of repose of up to 31^{o} were measured. In sets over 50 cm thick foresets are commonly thicker than 1 cm (they may be up to 10 cm thick) and show fining-upwards grainsize grading normal to the foreset surface (Fig. 3.10). It is also common for foresets to thicken downwards to the toe of the foreset. An effect of the grainsize grading in the foresets is that, whilst the sandstone of the set may be said to be poorly sorted, small samples of sandstone from the upper part of a foreset are generally moderately to well sorted. Coarser-grade material is also

OVERLAY ON PHOTOGRAPH ABOVE

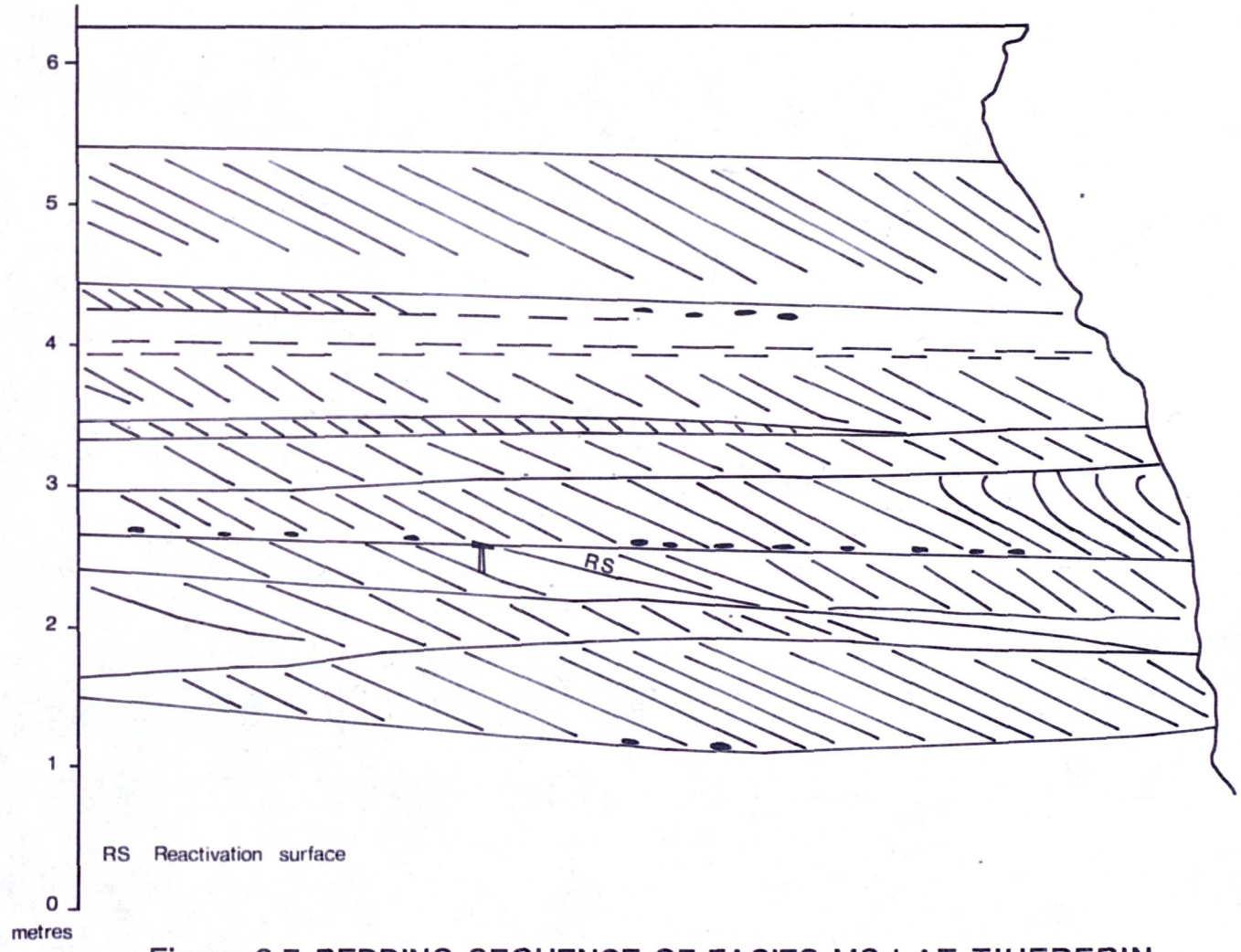
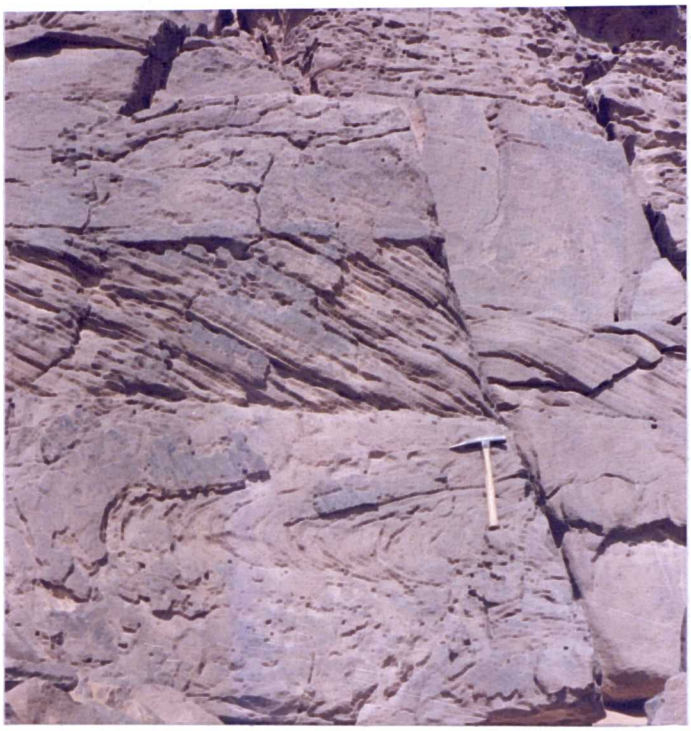


Figure 3.7 BEDDING SEQUENCE OF FACIES MSd AT TIHERERIN

"Diapiric" soft-sediment deformed strata in medium-grained sandstone with coarse sand grains and granules. The granules are concentrated along the deformed strata. Penknife is 8 cm long.

Perfectly-formed, recumbent-folded foresets (3-8 cm thick). The axis of the fold climbs towards the left ("down-current") and shows a marked thickening of the foresets across the crest of the fold. Deformed foresets are present to the right of the 10 cm scale bar.

Figure 3.8 SOFT-SEDIMENT DEFORMED STRATA AT BIR TALAUAUT



Bedding sequence at Wadi Anshal, 90 m stratigraphically above the base of the Tadrart Formation. The planar cross-stratified bed above the hammer is 70 cm thick and slightly wedge-shaped, thickening "down-current". The pitted appearance of the base of the bed of recumbent-folded cross-strata is due to the former presence of shale clasts which have been preferentially weathered.

Bedding sequence at Bir Talauaut, 144 m stratigraphically above the base of the Tadrart Formation in transgressive sequence 4, displaying reactivation surfaces and a wedge-shaped planar cross-stratified bed which thickens "down-current".

Figure 3.9 BEDDING SEQUENCES OF FACIES MSd AT WADI ANSHAL AND BIR TALAUAUT

Grainsize graded foresets of 1-3 cm thickness, composed of medium- to coarse-grained sandstone, with the coarse-grade sand lying at the base of the foreset. (The foreset dip azimuth of these cross-strata is 007^0).

Depressions, at the base of a set of cross-strata, marking the former presence of tabular-shaped clasts of shale/argillaceous siltstone (5-12 cm in length), the margins of which are iron-oxide enriched. These clasts were deposited together with quartz granules and very coarse-grade sand.

Figure 3.10 GRAINSIZE CHARACTERISTICS OF CROSS-STRATA AT BIR TALAUAUT

present in accumulations at the toe of the foreset (i.e. at the base of the set, where no bottom set is present) which may include coarse sand grains, granules and intraformational shale and argillaceous siltstone clasts commonly of pebble size (a few centimetres diameter). These weather out leaving holes in exposed sets of cross-strata giving the base of the set a homogeneous, pitted appearance (Fig. 3.9). Sets may be planar and wedge-shaped and it is common for sets to thicken in a "down-current" direction (Fig. 3.9).

A large variety of subsidiary stratification types is present in the sediments of this facies. Trough cross-stratification, scour and fill, and channel fill cross-stratification, parallel stratification, compound cross-stratification and cross-lamination are all present. Planar cross-stratification is also commonly present at a smaller scale than the 50-70 cm sets. This smaller-scale stratification type is usually present in sets averaging 15 cm in thickness and is often associated with compound cross-strata. The variety of stratification types in the facies is contrasted by the homogeneity of lithologies. Some argillaceous siltstones and silty shales, in thin bioturbated intercalations and beds generally only a few centimetres thick, are the only subsidiary lithologies present.

Within the stacked sets of planar cross-strata it is sometimes possible to find angular and tabular intraformational sandstone clasts of large cobble size (25 cm diameter). The sandstone of the clast, however, is barely distinguishable from that surrounding it, and it is usually very friable, suggesting little transport before deposition.

The measured section at Tihererin (Fig. 3.11) shows a characteristic succession of bedding and stratification types present in this facies as well as a number of features only infrequently observed. The succession comprises poorly-sorted medium to coarse-grained sandstones at the base of the Tadrart Formation. These lie above ferruginous beds of facies RBSdSt at the top of the Acacus Formation. Four major scour surfaces are present in the succession and other scouring events may be inferred by the sedimentary structures. The major scour surfaces are characterised in this succession by coarser-grade sediment above the scours compared with that below, but scoured surfaces are also commonly present in this facies where no grainsize variation exists across the boundary. In some cases, as at the base of unit 5 for instance, the sediment above the scour surface is

Figure 3.11 MEASURED SECTION OF FACIES MSd AT TIHERERIN

Notes:

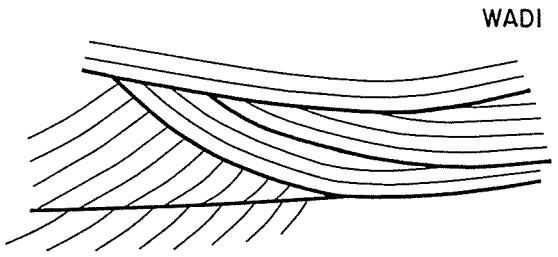
1) The section was measured from the base of the Tadrart Formation to 10.5 m stratigraphically from the base, at locality 15 (Fig. 3.5).

2) The succession has been divided into units for reference purposes.

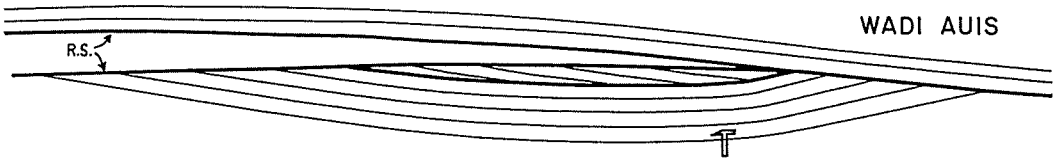
3) All the palaeocurrent readings are from foreset dip azimuths.

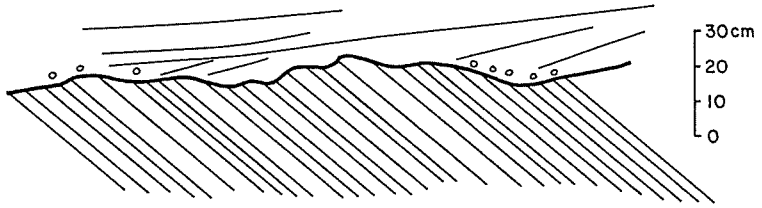
4) The sketches are from photographs of scour and fill/ channel fill cross-stratification which display concordant strata within scours, and an irregular, scoured surface. These structures were seen at Wadi Telia (North) and Wadi Auis (Fig. 3.2). In the scoured-surface feature a poorly-sorted, coarse-grained sandstone with granules lies above an irregular surface that has positive relief over the thicker foresets of the underlying cross-strata.

5) Abbreviations: UFR = upper flow-regime,
LFR = lower flow-regime.

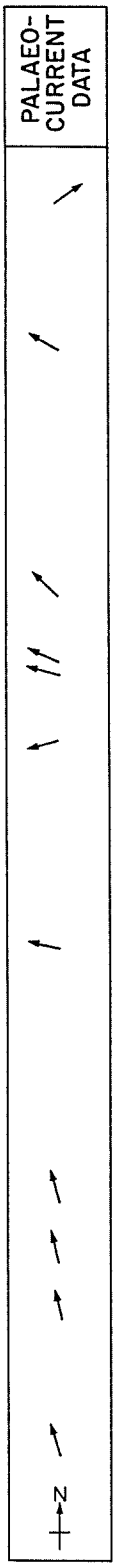


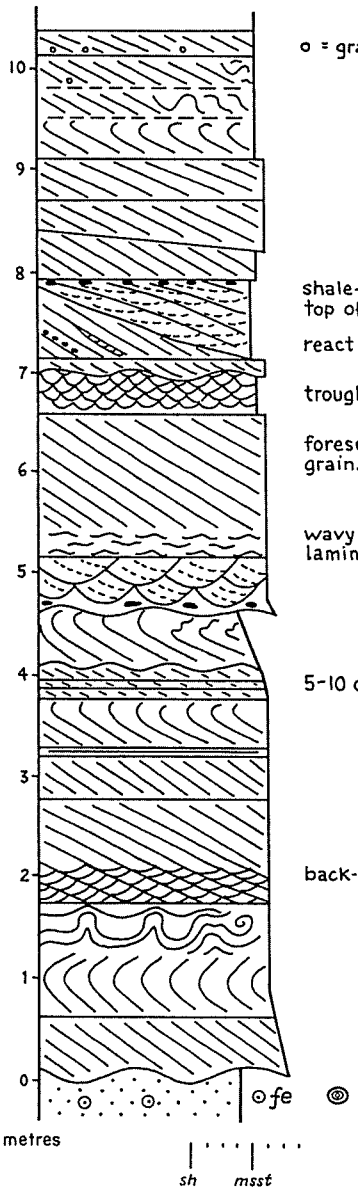
Scour and fill / channel fill cross-stratification in f-m sst.





Irregular scoured surface (20 cm relief) above cross-stratified m sst.





	INTERPRETATION	
10	—? scour event High-stage, LFR bar construction with UFR scouring events. Low stage strata not preserved	TRANSVERSE BAR(S)
9	Low-stage modification of bar	DISSECTED TRANSVERSE BAR
8	Bar top	TRANSVERSE BAR
7	Bar construction across channel Effective lee-side eddy	
6	Dune migration in channel	CHANNEL
5	? LFR bar construction	BAR
4	Low-stage bar top construction	TRANSVERSE BAR(S)
3	? UFR parallel strata & LFR bar construction	
2	Straight-crested dune with effective lee-side eddy	
1	High-stage LFR bar construction	

Figure 3·11

finer-grained than that below. The recumbent-folded deformed cross-strata in the succession are commonly overlain by visibly erosively-based beds, as at the top of unit 5, but they may also be overlain by soft-sediment deformed strata, as in unit 1, or by abruptly based sets of planar cross-strata.

Three bedding sequences are recognised within the succession at Tihererin (Fig. 3.11) which show an upwards trend towards beds of decreasing inferred, depositional flow strength (units 1-4, 6-8, and unit 9). This trend is discernible in upwards-decreasing set thickness, a generally poorly-developed, fining-upwards grainsize gradation and the occurrence, in the upper part of the sequence, of cross-laminated sandstones. The palaeocurrent data obtained from foreset dip azimuths from within each bedding sequence are generally unidirectional and very tightly arranged about the vector mean orientation. A much wider dispersal of orientations, although still unimodal, is obtained when all the palaeocurrent data from a succession are grouped together especially when low-stage structures are included. The bedding sequences described are only rarely recognisable, however. Successions are more common in which vertically stacked sets of planar cross-strata, some trough cross-strata and deformed strata are present, with recurrent scoured surfaces and reactivation surfaces. The succession above that of Figure 3.11 at Tihererin, for instance, comprises 22 m of such a succession in which no argillaceous siltstone or cross-laminated sandstone beds are present and no similar bedding sequences were identified.

Some noteworthy small-scale structures within the succession at Tihererin (Fig. 3.11) were observed. Angular foresets of a set of cross-strata (unit 9), showing grainsize grading from coarse to medium-grade sand, are truncated by a reactivation surface which is overlain by tangential foresets of slightly finer-grade sandstone. Multiple reactivation surfaces overlain by tangential foresets are developed above the lowest of these. Another structure of interest lies between the angular cross-strata of the truncated set (unit 9). A flaggy, cross-laminated, argillaceous siltstone drape of up to 1 cm thickness is present covering the lower half of the foreset indicating slack-water deposition from suspension. At the top of the set, and overlying both the angular and the tangential foresets, a pitted surface with granules is present indicating the former existence of a pebble

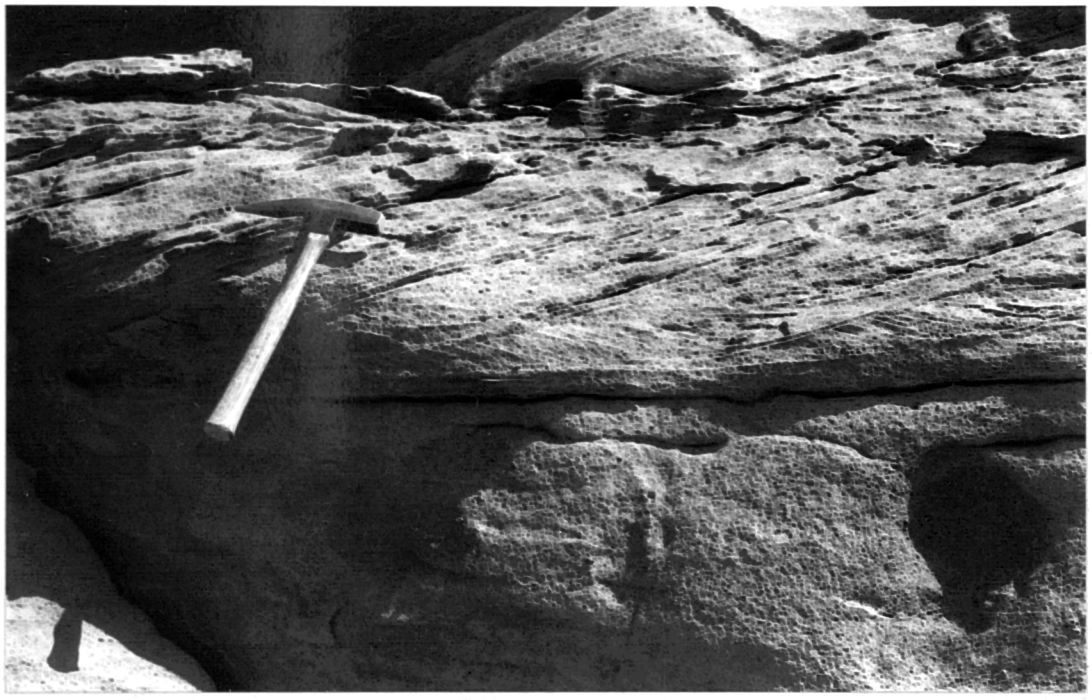
lag of shale intraformational clasts which have weathered out leaving only their impressions (Fig. 3.12). At the toe of the foresets of the cross-strata of unit 2, back-flow cross-strata are clearly developed (Boersma *et al.*, 1968) (Fig. 3.12).

Visibly erosive basal boundaries to beds and sets are generally planar or scour/channel-shaped surfaces although irregular scoured surfaces are also rarely present. In Wadi Auis an example of an irregular scoured surface was observed (Fig. 3.11, sketch) where the underlying bed comprised cross-strata of medium-grained sandstone in foresets of variable thickness and in which the irregular relief of the erosive surface appears to have developed in response to the variation in thickness between the foresets. The palaeocurrent data from the beds beneath and above the scour show that the scouring current was divergent by 120° from that inferred to have formed the planar cross-strata. This relationship may account for the unusual nature of the scoured surface. The poorly-sorted, coarse-grained sandstones lying above the scour display large-scale trough and planar cross-stratification.

The measured section at Tihererin did not show an example of the commonly-present scour and fill or channel fill cross-stratification. Small and large-scale examples of this structure were seen, however, at Wadi Telia and Wadi Auis respectively (Fig. 3.11, sketches). In neither case, however, was it possible to determine the three-dimensional character of the basal scour surfaces to distinguish channel fill from scour and fill cross-stratification.

In the measured section of the basal 43.5 m of the Tadrart Formation in Wadi Teshuinat (Figs. 3.13 & 3.14), where finer-grained sediments than those of the Tihererin measured section are present, some further characteristics of facies MSd are displayed. The succession is dominated by fine and medium-grained sandstones with ten thin and very thin maroon and pale yellow/brown argillaceous siltstone intercalations which have been used to divide the succession into reference units. The thin argillaceous intercalations are laterally impersistent in some cases but in others they were traced along exposed faces of slick-rock slopes and broken cliffed sections for more than 60 m. The eleven reference units have thicknesses varying from less than 1 m up to 10 m. Although marked vertical grainsize grading within these units is absent, an upwards decrease

Coarse-grade sand and granules concentrated in a lag on the top surface of a bed of cross-stratified sandstone which also displays the impressions of clay clasts. This phenomenon is thought to be created by winnowing of the surface of a fluvial bar during falling stage. From unit 9, Fig. 3.11.



Back flow cross-strata developed in medium to coarse-grained, poorly-sorted sandstone. The foresets are 3-8 cm thick. From unit 2, Fig. 3.11.

Figure 3.12 SHALE-CLAST LAG, AND BACK-FLOW CROSS-STRATA, FROM MEASURED SECTION AT TIHERERIN

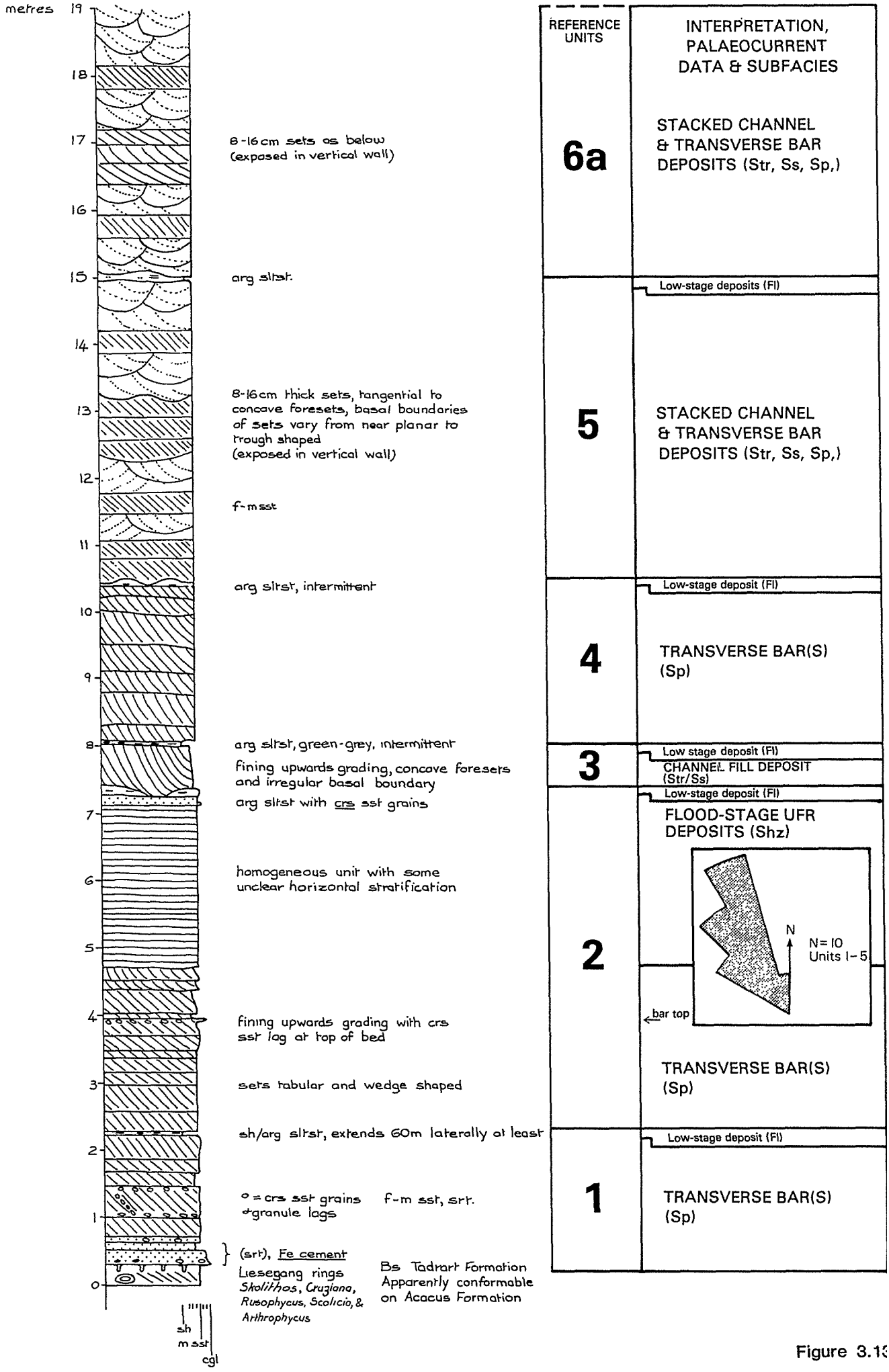

metres
19
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
8-16cm sets as below (exposed in vertical wall)
arg sltst.
8-16cm thick sets, tangential to concave foresets, basal boundaries of sets vary from near planar to trough shaped (exposed in vertical wall)
f-m sst
arg sltst, intermittent
arg sltst, green-grey, intermittent
fining upwards grading, concave foresets and irregular basal boundary
arg sltst with crs sst grains
homogeneous unit with some unclear horizontal stratification
fining upwards grading with crs sst lag at top of bed
sets tabular and wedge shaped
sh/arg sltst, extends 60m laterally at least
° = crs sst grains
*granule lags
f-m sst, srt.
(srt), Fe cement
Liesegang rings
Skolithos, Cruziana, Rusophycus, Scolicia, & Arthrophycus
Bs Tadrart Formation
Apparently conformable on Acacus Formation
sh
m sst
cgl
REFERENCE UNITS
INTERPRETATION, PALAEOCURRENT DATA & SUBFACIES
6a
STACKED CHANNEL & TRANSVERSE BAR DEPOSITS (Str, Ss, Sp,)
Low-stage deposits (Fl)
5
STACKED CHANNEL & TRANSVERSE BAR DEPOSITS (Str, Ss, Sp,)
Low-stage deposit (Fl)
4
TRANSVERSE BAR(S) (Sp)
Low stage deposit (Fl)
3
CHANNEL FILL DEPOSIT (Str/Ss)
Low-stage deposit (Fl)
2
FLOOD-STAGE UFR DEPOSITS (Shz)
N
N=10
Units 1-5
bar top
TRANSVERSE BAR(S) (Sp)
Low-stage deposit (Fl)
1
TRANSVERSE BAR(S) (Sp)

Figure 3.13

Figures 3.13 and 3.14 MEASURED SECTIONS OF FACIES MSd AT WADI TESHUINAT

Notes:

1) The sections were measured from the base of the Tadrart Formation to 43.5 m stratigraphically from the base, at locality 10 (Fig. 3.13) and at locality 7 (Fig. 3.14) in Wadi Teshuinat (Fig. 3.5).

2) The succession has been divided into units for reference purposes.

3) All the palaeocurrent readings are from cross-stratification and trough axes.

4) Abbreviations:
   a) Sp, Str, Ss, etc. = subfacies, modified after Miall (1977; see Fig. 3.16)
   b) UFR = upper flow-regime.

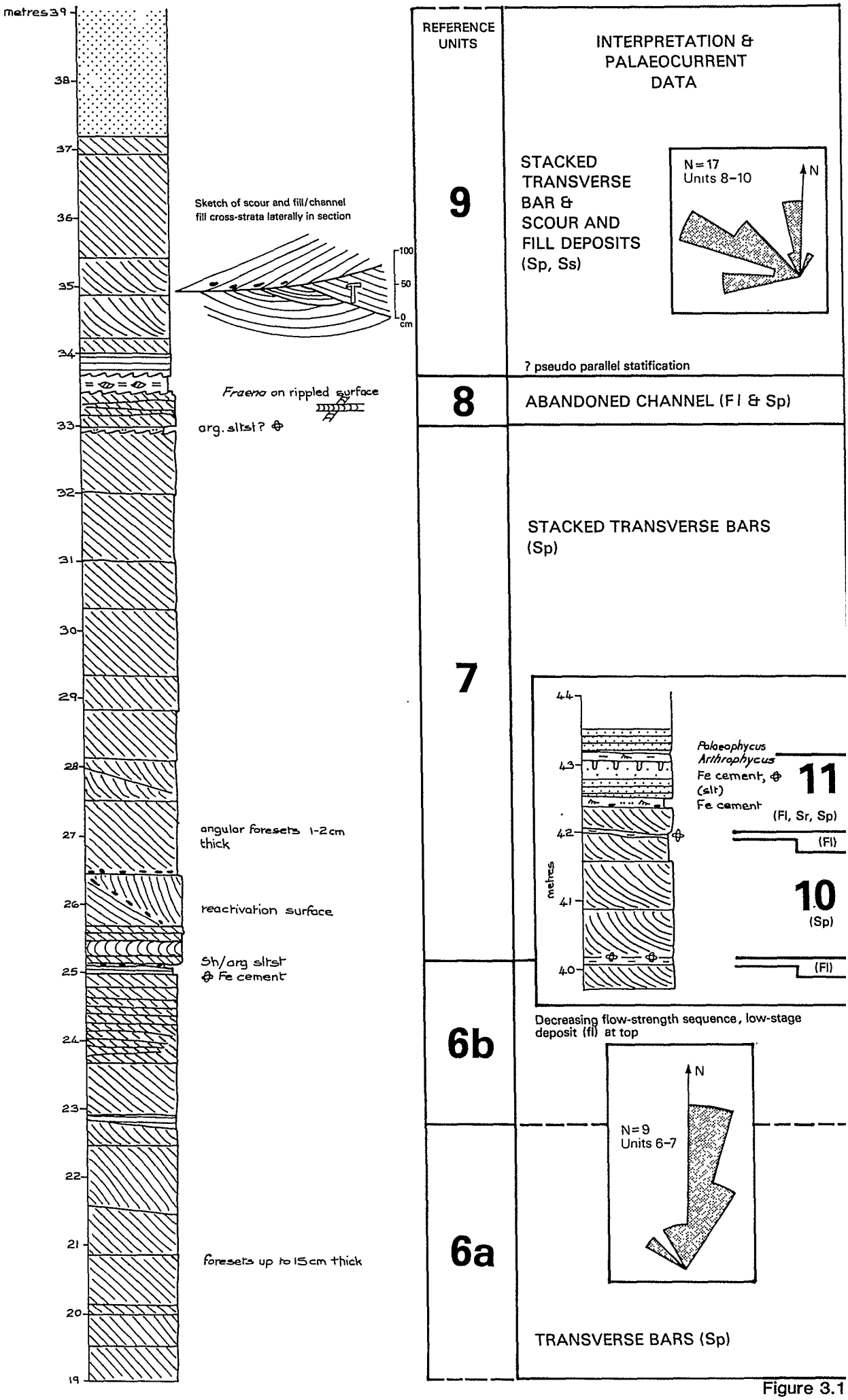

metres
39
38
37
36
35
34
33
32
31
30
29
28
27
26
25
24
23
22
21
20
19
Sketch of scour and fill/channel fill cross-strata laterally in section
100
50
0
cm
Fraena on rippled surface
arg. sltst? ⊕
angular foresets 1-2cm thick
reactivation surface
Sh/arg sltst ⊕ Fe cement
foresets up to 15cm thick
REFERENCE UNITS
INTERPRETATION & PALAEOCURRENT DATA
9
STACKED TRANSVERSE BAR & SCOUR AND FILL DEPOSITS (Sp, Ss)
N=17
Units 8-10
N
? pseudo parallel statification
8
ABANDONED CHANNEL (Fl & Sp)
7
STACKED TRANSVERSE BARS (Sp)
44
43
42
41
40
metres
Palaeophycus
Arthrophycus
Fe cement, ⊕
(slt)
Fe cement
11
(Fl, Sr, Sp)
(Fl)
10
(Sp)
(Fl)
Decreasing flow-strength sequence, low-stage deposit (fl) at top
6b
N
N=9
Units 6-7
6a
TRANSVERSE BARS (Sp)

Figure 3.1

in set thickness is quite pronounced in some cases (e.g. units 4 & 6, particularly 6b). The bedding sequence of unit 6b also shows evidence of decreasing flow-strength in the upwards transition from horizontally-stratified to planar cross-stratified sandstones to thinly-bedded and bioturbated argillaceous siltstones. The base of the units may be characterised by visibly erosive scoured horizons or by planar abrupt bed boundaries above the thin argillaceous siltstone intercalations. Trough-shaped scours were not observed, however.

Throughout the 43.5 m section all the cross-strata appeared to dip in approximately the same northwesterly direction. No apparent foreset dip azimuths were observed which were at variance with this direction. The steep slick-rock slopes and walls up which sections were measured did not provide much opportunity, however, for the measurement of true foreset dip azimuths. It was decided not to measure apparent azimuths as these would only have provided poor support for the general observation of unidirectional palaeocurrents and they would have provided no valuable information on the range of azimuths. The palaeocurrent roses plotted from true foreset dip azimuths (Figs. 3.13 & 3.14), therefore, are based on a rather small number of readings. Nevertheless, the evidence they provide for palaeocurrent directions towards the north, northwest and west is thought to be good. Although statistically somewhat unreliable, the impression is given that units 1-5 were characterised by northwesterly currents, units 6 and 7 by north-northeasterly currents and units 8-10 by a wider range of currents. Within a unit, or stacked succession of units, the measured azimuths generally create a fairly tight palaeocurrent rose pattern.

The sandstones of this succession are similar to those of the Tihererin section in being characterised by stacked planar cross-strata displaying grainsize graded foresets and lag deposits of coarse-grade sand, granules, and shale clasts which lie at the base of sets and, in some cases, on the top surfaces of beds. The lag deposits at the base of sets give individual sets a vertical fining-upwards grainsize character. Bottom sets are usually absent but in some sets a homogeneous to irregularly-stratified bottom set is present composed of poorly-sorted sandstone including a component of coarse sand. Cross-lamination is present in these bottom sets in some cases. Trough cross-strata, scour and fill strata (see sketch,

Fig. 3.14), reactivation surfaces and some recumbent-folded deformed cross-strata are also present, although this latter stratification type is not as abundant as in the coarser-grained sandstones of the Tihererin section.

Over the whole 43.5 m succession a change from predominantly medium-grained sandstones in the lower units to predominantly fine-grained sandstones in the upper units is discernible. There is also a gradual increase upwards in the section of trough cross-stratification, cross-lamination, and channel-like structures of 5 to 25 m width in which intraformational conglomerate channel lags are present.

3.2.1.3 Biogenic structures

The presence of diagenetic iron-oxide enrichment of argillaceous siltstone horizons in facies MSd, and the development of concretionary artifacts sometimes resembling biogenic structures, created difficulties in confidently identifying bioturbation structures when recognisable trace fossils were not present. Nevertheless, of the 12 argillaceous siltstone horizons within the succession studied at Wadi Teshuinat (Figs. 3.13 & 3.14), five contained biogenic structures, two contained possible bioturbation structures and the five without recognisable bioturbation lay at the base of the succession. The trace fossils recognised include *Fraena* (see Fig. 2.3), *Skolithos* (see Fig. 2.7), *Palaeophycus* (see Fig. 2.5) and *Arthrophycus* (see Fig. 2.6). *Fraena* is an exogenic trace which, in these sediments, is represented by unilobate, transversely-ribbed trails which are straight to moderately sinuous, commonly crossing each other, and may be present on current-rippled top surfaces of beds of planar cross-strata (unit 8, Fig. 3.14). They are commonly overlain by argillaceous siltstone beds which display lenticular cross-lamination. Where more than one argillaceous horizon is present within an interval of up to 1 m in thickness, these horizons are interbedded with thin sets of planar cross-stratified sandstone, and ripple cross-laminated sandstone beds which commonly display hard, haematite-enriched, bioturbated and/or rippled top surfaces (e.g. unit 8, Fig. 3.14). At the top of unit 11 (Fig. 3.14) *Skolithos* is present in a poorly-sorted sandstone bed of 50 cm thickness which is overlain by a maroon cross-laminated argillaceous siltstone of the "40 m plateau" in which

the feeding burrows *Arthrophycus* and *Palaeophycus* are present. As well as biogenic structures, the argillaceous siltstone horizons often contain abundant intraformational clay clasts.

3.2.1.4 Petrography

The sandstones of this facies are quartz arenites and subarkoses in which, in addition to quartz, a small percentage of detrital grains of feldspar, mica, and mature heavy minerals such as zircon, tourmaline and apatite are present. The quartz grains, some of which contain inclusions of tourmaline and rutile needles, are generally composed of monocrystalline quartz but some polycrystalline quartz grains are also present (less than 10% of the quartz grains). Grains composed of microcrystalline quartz are rare. Point counting was undertaken to establish the degree of sorting in the thin sections studied, and it was shown that a range of degrees of sorting existed. The coarser-grained sandstones are generally poorly-sorted, or moderately to poorly-sorted, whilst the finer-grained sandstones are generally moderately-sorted, except for those associated with argillaceous intercalations which have a range of textures from poor to well-sorted. Although a range of degrees of rounding of quartz grains also exists, roundness was not found to be directly related to grainsize. The occasional very well rounded grains are generally the larger monocrystalline quartz grains. In general the grains are subrounded to subangular and grains displaying high sphericity are rare.

Silica precipitation and dissolution features are common in these sandstones, and authigenic quartz overgrowths have developed on the majority of the quartz grains (Fig. 3.15). Surface erosion features were observed on some of the grains with overgrowths, suggesting that the overgrowths may have been inherited from an eroded sandstone, but the erosion features are difficult to distinguish from weathering phenomena developed recently at exposure. The common presence of sharply developed, sometimes interlocking, overgrowth crystal faces suggests that the majority were formed in place.

Both early and late haematite and subordinate limonite cement is present, commonly replacing any feldspar grains present, and corroding and replacing polycrystalline quartz grains. Rhombic crystals of haematite are present on the surface of some grains whilst on others radially fibrous haematite rims are present (Fig. 3.15). In some of

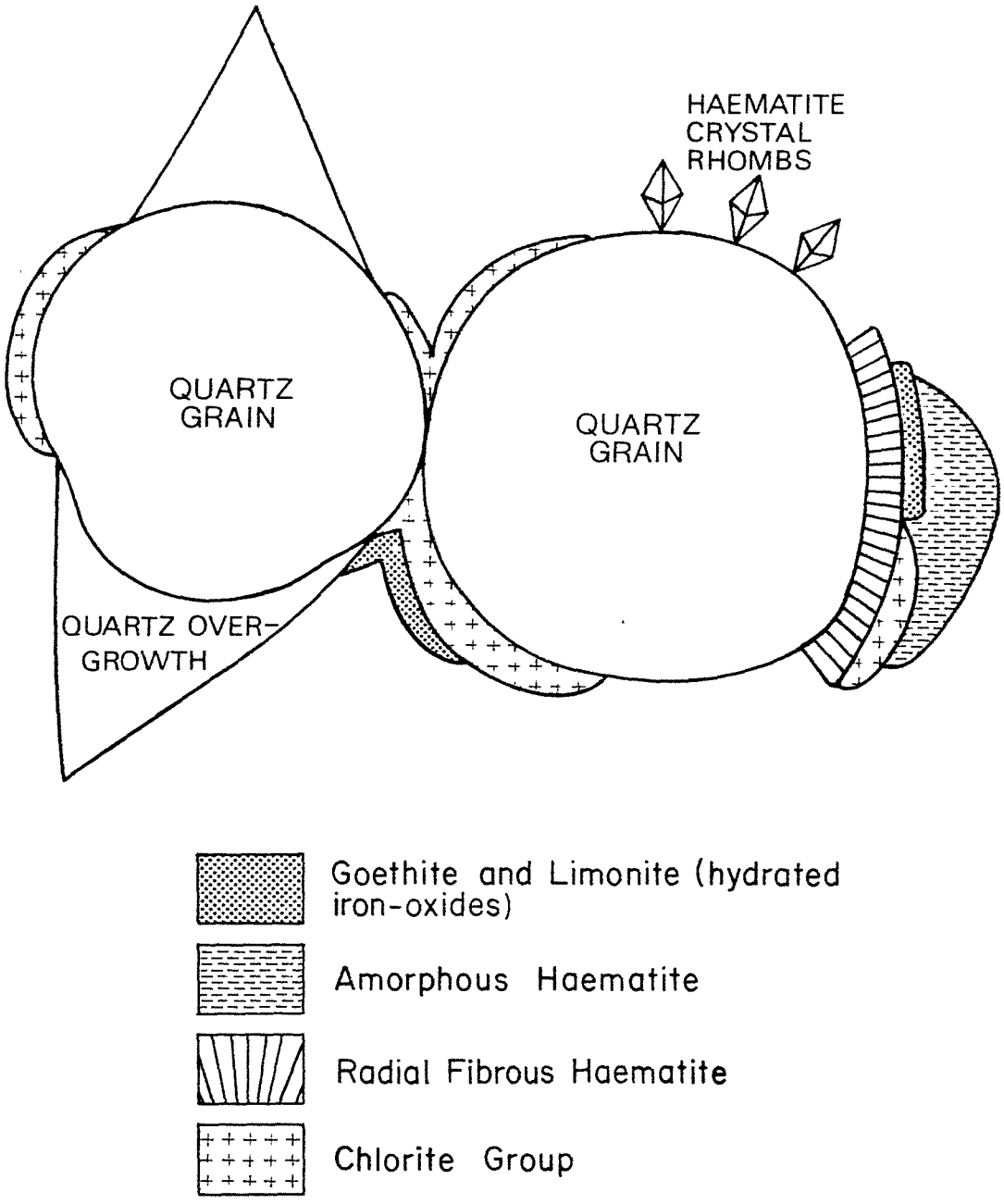


Figure 3.15 SCHEMATIC PETROGRAPHY OF SANDSTONES OF FACIES MSd

the sandstones studied, especially common in the fine-grained examples, a rim of green, yellow and brown, moderate birefringence minerals are present. Use of the electron probe on the scanning electron microscope allowed these rims to be identified as clay pellicles of Chlorite Group minerals (possibly chamosite). In a number of cases the chloritic pellicle was present over a fibrous haematite rim on the grain surface. This indicates that the haematite rim was developed first and is evidence that some of the haematite present is of early diagenetic origin. The chloritic clay pellicles are also thought to have an early diagenetic origin. Those coating quartz grains may well have formed by the infiltration of clay minerals into shallowly-buried sands during periods of high water-table level, and the subsequent coating of sand grains during periods of low water-table level (Walker, 1967). Diagenetic alteration of the clays present in these early pellicles is thought responsible for their present chloritic composition. The presence of chloritic rims over haematite rims suggests that an additional process was operating whereby authigenic chlorite was developed, possibly by the alteration of unstable detrital grains, and coated on the surface of quartz grains some time after burial. Haematitic clay pellicles comprise finely divided clay with a small proportion of included haematite and are responsible for the red pigmentation of red bed sandstones (Hatch *et al.*, 1971). These coatings have been ascribed to both crystallisation of original amorphous coatings and to diagenetic redistribution of iron derived from the decomposition of iron-bearing detrital minerals.

A third authigenic component is present in these sandstones. Titaniferous iron-oxides, including geothite, also coat quartz grains. The development of these rims is thought to have occurred later than the other authigenic processes described since they were observed to be present outside rims of haematite and chlorite. The presence of both chlorite and haematite rims was shown to inhibit the development of quartz overgrowths. This further supports the early diagenetic origin proposed for the texture of these sandstones, and provides an explanation, in part, for their moderately good porosities.

The mature heavy mineral assemblage of these quartz arenites and subarkoses and the contrast of considerable mineralogical maturity with only moderate to poor textural maturity may be evidence for a polycyclic origin for these sandstones, derived originally, perhaps, from

a granitic source area and deposited by processes, and in an environment, not associated with textural maturity. This evidence, and the presence of early stage haematite and chlorite authigenic components, supports a continental environmental interpretation.

The scanning electron microscope work undertaken on these sandstones is discussed at the end of section 3.3.1.7 (this Chapter) together with a discussion of the petrography of facies CSdSt (see Fig. 3.38).

3.2.2 Interpretation

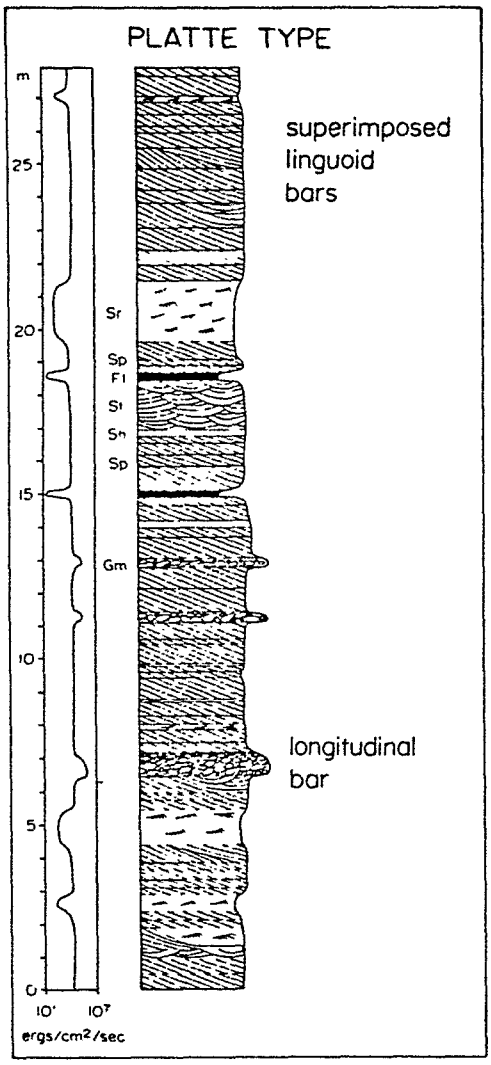
3.2.2.1 Sedimentary structures, models and vertical sequence analysis

The following characteristic sedimentary structures of facies MSd are all commonly present in transverse bars of braided river deposits (Ore, 1963): abundant planar, tabular and wedge-shaped sets of cross-strata; angular foresets with grainsize grading, and sets with poorer-sorted and coarser-grade sediment at the toe of the foresets; soft-sediment deformed stratification including recumbent-folded deformed cross-strata; and minor current-ripple cross-lamination and trough cross-stratification. The unidirectional palaeocurrent data are also characteristic of braided river deposits although similarly unidirectional palaeocurrent roses may be plotted from cross-strata of other fluvial environments and even tidally dominated marine environments (Ore, 1963; Reineck & Singh, 1973). Transverse bars are initiated in braided rivers as a result of an inability of the river to transport the coarsest fraction of its load and the consequent deposition of this fraction centrally in a channel. Seasonal fluctuations in discharge, load and gradient are responsible for this incompetence. The deposited material builds into a bar by the addition of more sediment to the top and downstream end of the incipient bar. As the bar is constructed, lateral channels take progressively more water and become unstable. They widen and, as downcutting occurs, the bar may become emergent. Emergence of transverse bars by this process, and during low-stage flow conditions between floods, allows minor channels, often orientated obliquely to the major channels of the braided system, to dissect the bar. Small-scale dunes and ripples are formed in the sand bedload of these channels. While emergence and dissection of transverse bars occurs, new transverse bars are initiated and constructed in the channels lateral to the emergent bar. By this process the braided-river channel

pattern is developed (Leopold & Wolman, 1957). The textural immaturity of some of the sandstones of this facies is further evidence for a fluvial interpretation (Leopold & Wolman, 1957), although this characteristic and those already mentioned are not inconsistent with a tidal/estuarine interpretation (Klein, 1977; Collinson, 1978b).

The presence of reactivation surfaces, scoured surfaces, argillaceous siltstone/shale intraformational clasts lying at the base of sets and of foresets, and the absence of major argillaceous beds, suggest a depositional setting of considerable fluctuation in flow conditions. The environment appears to have been one where high-velocity flows of considerable strength scoured away fine-grained sediments accumulated on bar tops, in abandoned channels or in overbank areas, and redeposited them in channels and in the construction of transverse bars. The presence of clasts of siltstone and shale indicates that fine-grained detritus was available in the sediment source area from which the river load was derived whilst the roundness of some sandstone grains suggests a polycyclic sedimentation history in which the sands of eroded Cambro-Ordovician Internal Tassilian Sandstones were redeposited during the early Devonian. The probable virtual absence of major terrestrial vegetation in the late Silurian/early Devonian may be a partial explanation for the absence of recognised overbank deposits and the inferred braiding of the rivers, since river bank margins would not have been stabilised by vegetation (Schumm, 1968; see Cotter, 1978 & Long, 1978).

Miall (1977) defined a number of facies types (subfacies Sp, Str, etc., Figs. 3.13 & 3.14), and four principal facies assemblages characteristic of braided river deposits. The Platte-type braided river model (Fig. 3.16), based on the work of Smith (1970, 1971a, 1971b, 1972), is the most appropriate analogue for facies MSd although the sediments of the Platte River are of slightly coarser grade (medium to coarse-grained and pebbly sandstones) than those of facies MSd. The major subfacies (Sp and Str) of the Platte model are both present although trough cross-stratified sandstones are not very abundant. Of the subsidiary subfacies of the Platte model only the conglomeratic subfacies (Gm) is not present in facies MSd. Cross-laminated sandstones (subfacies Sr) are present but are not very abundant. On the other hand scour and fill/channel fill cross-



↑ Generalised indication of changes in stream power

Note:

1) Absence of recognised cyclic sequences
2) Linguoid (transverse, lobate) bars are the dominant morphological features preserved
3) The river is shallow and lacks marked topographic features.

LEGEND

Major subfacies

Sp. Planar cross-stratified, medium to coarse-grained and occasionally pebbly sandstone. Set thickness 5-100 cm, occasionally up to 5 m. Sets traced laterally up to 20 m. Foresets are steep avalanche slopes. Deposits of lower flow-regime. Straight crested or slightly sinuous linguoid (transverse, lobate) bars and sandwaves (scour hollows absent).

Str. Trough cross-stratified, medium to coarse-grained and occasionally pebbly sandstone. Units solitary or in groups. Set thicknesses 5-60 cm. May contain shale intraformational clasts. Deposits of lower flow-regime dunes (sinuous crested ripple-forms with scour hollows).

Subsidiary subfacies

Gm. Massive or crudely-stratified, imbricated conglomerate with minor sandstone, siltstone or shale lenses. Deposits of longitudinal bars and channel lags.

Sr. Lower flow-regime, cross-laminated sandstone.

Shz. Lower and upper flow-regime, horizontally stratified/laminated sandstone. Primary current lineation may be present.

Fl. Interbedded very fine-grained sandstone, siltstone and shale. Cross-laminated and bioturbated with rootlets and caliche. Deposits of waning floods, and overbank areas.

Ss Scour and fill/channel fill cross-stratified sandstone.

Largely from work on the Platte River, Colorado-Nebraska (Smith, 1970, 1971a, 1972), modified after Miall (1977).

Figure 3.16 PLATTE-TYPE BRAIDED RIVER MODEL

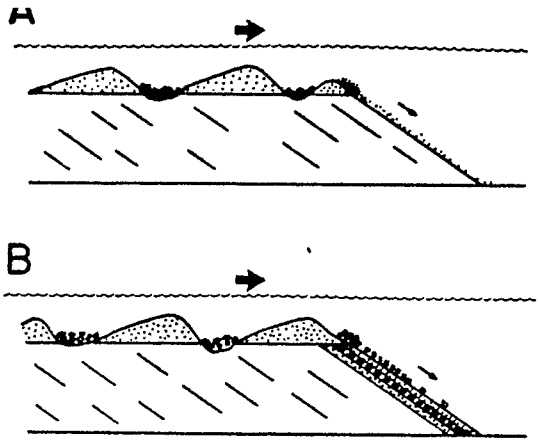
stratified sandstones (subfacies Ss) are more abundant in facies MSd than in the Platte model. The differences between facies MSd and the Platte model do not suggest, however, that one of Miall's three alternative facies assemblages (1977) would be a more suitable model. Since facies MSd differs from the Platte model in having less coarse-grade material, the finer-grained deposits of the Bijou-Creek model were considered as a possible analogue. These deposits, however, are characterised by a predominance of upper flow-regime, parallel-stratified sandstone with primary current lineation (McKee *et al.*, 1967; see Williams, 1971, for another example) and are inappropriate therefore as an analogue. In addition to the deposits of the Platte River, those of the Tana River, Norway (Collinson, 1970) and the South Saskatchewan River, Canada (Cant & Walker, 1978) also provide suitable analogues, the former being an example of the Platte-type facies assemblage (Miall, 1977).

Following the generalised interpretation of Miall's subfacies in the Platte model (1977), the sediments of facies MSd are inferred to have been deposited under the following conditions (see Figs. 3.11, 3.13 & 3.14). During floods, upper flow-regime conditions were responsible for scouring away low-stage, fine-grained deposits and ripping away the upper surfaces of transverse bars. Flood waters caused channel bank collapse and the incorporation of large friable sandstone clasts in channel deposits. Upper flow-regime parallel stratified sandstone (subfacies Shz) was deposited over the scoured horizons in places, sometimes incorporating shale and argillaceous-siltstone intraformational clasts that were transported by the flood waters. During the following high-stage flow, sinuous-crested dunes migrated down-stream in braided channels, constructing large-scale sets of trough cross-strata (subfacies Str). Straight and slightly sinuous-crested dunes, sandwaves and transverse bars were more abundant, however, and these migrated down-stream during high-stage flow, depositing large-scale sets of planar cross-strata (subfacies Sp). Sheet run-off processes on the tops of bars generated small bar-top channels at the downstream end of which delta-like lobate sediment accumulations developed beyond the bar margin (Collinson, 1970).

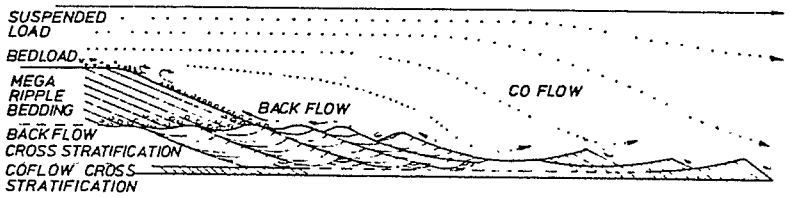
The horizontally stratified sandstone and the scour and fill/channel fill cross-strata (termed scour and fill hereafter) (Fig.

3.11) provide some evidence for upper flow-regime conditions. Primary current lineation was only occasionally observed on the horizontal strata, however, and it is probable that much of the horizontal stratification was formed by the process described by Smith (1971b), whereby migrating very-shallow, rhomboidal-shaped bars generate pseudo-planar stratification under lower flow-regime conditions. In the case where horizontal stratification was observed lying above a preserved rippled surface of beds of subfacies Fl (top of unit 8, Fig. 3.14) this interpretation seems the most likely. The concordant fill of the broad shallow scours of the scour and fill cross-strata was most probably deposited under upper flow-regime conditions (Miall, 1977). This sedimentary structure is thought to form by separate cutting and filling events (Miall, 1977). Scour hollows reminiscent of the scours at the base of scour and fill structures and caused, in some cases, by local vortices developed around obstructions have been seen in the Rio Grande and Brahmaputra rivers (Harms & Fahnestock, 1965; Coleman, 1969). Larger scours than these may represent major or minor channels, formed by avulsion or bar-top channelling during high-stage, or by bar dissection during low-stage conditions (Miall, 1977). The association of scour and fill cross-strata (subfacies Ss) with subfacies Sp, Str, and SS in facies MSd suggests that they were formed under high flow-velocity conditions by in-channel vortices and by high-stage avulsion processes.

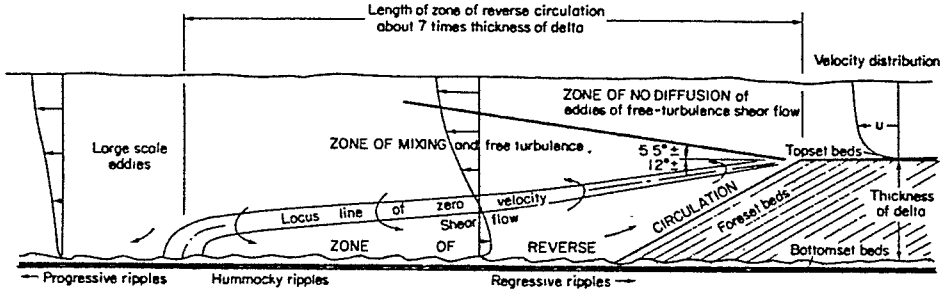
During falling-stage flow conditions, processes characteristic of high-stage flow conditions continued but smaller-scale bedforms were present in channels and on the top surface of bars which were responsible for smaller-scale sets (subfacies Str and Sp). The bedforms migrating over the surface of transverse bars, on reaching the bar crest, avalanched down the lee slope creating angular fining-upwards, grainsize graded foresets by the process described by Smith (1972) (Fig. 3.17). Segregation of bedload material in the migrating bar-top bedforms, by the concentration of coarse-grade sediment in lee-side scour pockets, is responsible for this phenomenon. The process ensures that large-scale sets of avalanche-foresets continue to be constructed on the lee slopes of bars during falling-stage flow conditions so long as bedforms migrate over the bar surface. These large-scale sets have a greater preservation potential than the small-scale sets formed on the bar top by migrating bedforms because of the



Origin of coarse and fine-grained foreset laminae by continuous avalanching. A - Finer sediment composing the small-scale asymmetrical bedform spills over the bar edge followed by B - avalanching of coarser sediment previously segregated into lee-side scour pocket. Thickness of laminae depends on length of foreset and size of surface bedforms. (After Smith, 1972).



Schematic illustration of flow pattern and resulting deposition on the lee-side of a megaripple with a well-developed backflow system. Various sedimentation units are laid-down simultaneously. Backflow beds are developed in the form of climbing ripples. (Modified after Boersma *et al.*, 1968; in Reineck & Singh, 1973.)



Development of lee-side eddy over large-scale ripple-form (microdelta). (After flume tank experiments by Jopling, 1961; in Blatt *et al.*, 1972).

Figure 3.17 CHARACTERISTICS OF THE LEE-SIDE OF RIPPLE-FORMS AND BARS

topographically high position of the small-scale sets. The angular characteristic of these foresets laid down on the bar-front avalanche slope may be attributed to the weakened separation eddy in the lee of the bar (Collinson, 1970).

The presence of lags of coarse-grade material, including abundant clay clasts, on the top of beds of cross-strata (Fig. 3.12) may have resulted from low-stage sheet run off events scouring sediment from part of a bar surface and redepositing it elsewhere on the top of the bar. The decreasing flow-strength of such sheet run-off flows is responsible for the deposition of coarse-grade material initially and for some winnowing of finer-grained sediment. The presence of this feature above low-stage bar-modification strata (unit 9, Fig. 3.11) supports this interpretation (see also unit 2, Fig. 3.13). The feature also provides evidence that in some cases the original top surface of the bed is preserved, whereas the top of many sets of cross-strata probably do not represent the original depositional sediment-water interface, the top of the set having been removed by a subsequent scouring event, preceding the deposition of the next set of planar cross-strata.

During low-stage flow, the following conditions of sediment accumulation may have applied. In general the deposits of these periods are poorly preserved, but a number of diagnostic sedimentary structures were produced and preserved. During periods of low-stage flow, low-velocity shallow flows characterised the braided channels, on the floors of which migrating ripples created beds of cross-laminated fine-grained sandstone (subfacies Sr). Similar deposits were also laid down on the tops of bars. Silt and clay was deposited from suspension in the quiet water conditions of abandoned channels and pools in the braid plain, and in scour hollows created during flood stage flows. Wind-driven currents and waves rippled the beds of silt and sand in these environments and created cross-laminated intercalations interbedded with argillaceous siltstones (subfacies Fl). Organisms bioturbated these quiet-water sediments creating the biogenic structures characteristic of subfacies Fl. The sediments of this subfacies in facies MSd differ from those of the Platte River in lacking any evidence for rootlet horizons, or for caliche beds formed as overbank soil horizons. This suggests that overbank deposits were rarely preserved, and it provides support for the

inference that, because of the probable absence of late Silurian/ early Devonian major terrestrial vegetation, channel banks were relatively unstable.

The division of facies MSd into subfacies (Fig. 3.16) allowed an analysis of vertical subfacies transition data. A clearly-visible scoured surface was considered as a subfacies for these purposes (subfacies SS). A most common vertical sequence was established which comprised stacked sets of planar cross-strata lying above a scoured surface and passing up eventually to an argillaceous siltstone bed (subfacies Fl):

A) SS → Sp → Sp - → SS → Sp → Sp - → Fl → SS.

Considering, in addition, the second and third most common vertical transitions of subfacies (second and third rank) it proved possible to establish the following sequences, given in order of rank. (It was necessary to go up to the fourth rank transition for subfacies SS whilst for subfacies Sp all the possible transitions were included).

B) SS → Str → Sp → Sp → Sr → Fl → SS

C) SS → Shz → Sp → Str → Sp → Fl → Sp etc.

D) SS → Fl → Ss → Str → Sp → Fl → Shz etc.

E) SS → Sp → Shz → Fl

F) SS → Sp → Ss → Sp → Sr → SS.

This analysis showed that, in some cases, high and low flow-strength deposits may be arranged vertically in a falling flow-strength sequence (e.g. sequence B, see units 6-8, Fig. 3.11, and unit 6b, Fig. 3.14). However, in other cases, no falling flow-strength sequence was discernible (e.g. sequences A, C and D). In the cases where "flow-strength sequences" were recognised, these were only rarely clearly associated with true vertical fining-upwards sequences. Apart from facies Fl which is defined as being composed of finer-grained sediment, it was common to find that the "low flow-strength" deposits were of the same grainsize as those of high flow-strengths. The value of this type of analysis is to establish trends not otherwise discernible in sequences showing some gradational or transitional variation. The evidence of soft-sediment deformed cross-strata, scour and fill cross-strata, abrupt planar basal set boundaries and intra-formational clasts, suggests that in facies MSd many more erosion surfaces, resulting from marked increases in flow strength, are present than are indicated by the visibly erosional scour surfaces

(subfacies SS). As a result it is common for beds in a succession to be virtually hydrodynamically independent of their neighbours in the measured section. Consequently the study of individual beds, or actual bedding sequences showing internal genetic relationships, and the lateral characteristics of these beds and bedding sequences, is likely to be the more rewarding approach to the interpretation of these sediments (Collinson, 1978a).

Some doubt has been cast upon the validity of vertical sequence analysis in generating models of fluvial alluvium and caution is required in relating idealised ancient sequences to the inferred vertical sequences of well-documented modern rivers (Collinson, 1981). Nevertheless, it is worth noting that the idealised vertical sequence determined for the Battery Point Formation (Devonian, Quebec), and interpreted as being analogous to the deposits of the South Saskatchewan River (Cant & Walker, 1976; Cant, 1978), has many similarities with sequence A, determined from this facies, and with some of the bedding sequences described in the measured sections at Tihererin and Wadi Teshuinat (Figs. 3.11, 3.13 & 3.14). Cant and Walker (1976) described a 10 m thick vertical sequence showing a fining-upwards grainsize trend and an upwards decrease in set thickness. In their model, a scoured basal surface is overlain by alternating sets of planar and trough cross-strata (with some scour and fill cross-strata) which are interpreted as channel deposits. These are overlain by thin sets of planar cross-strata and cross-laminated sandstones and interbedded shales. This fine-grained succession capping the sequence is thought to have been deposited on bar tops and by quiet water vertical accretion. Consequently, these deposits have a low preservation potential. Nevertheless, units 4 and 8 (Fig. 3.11) may be examples of these bar top deposits. Low stage deposits of fine-grained sediment lying in depressions in the braid plain are likely to have a higher preservation potential, however. The fine-grade sediments of the succession at Wadi Teshuinat (Figs. 3.13 & 3.14) are thought to have been deposited in this subenvironment of the braid plain.

The coarse-grained beds of facies MSd may be interpreted similarly to the coarse member of the Battery Point sequence as being in-channel deposits of mid-channel bars and migrating dunes. One of the vertical bedding sequences at Tihererin (units 6-8, Fig.

3.11) is similar to the Battery Point sequence, but is much thinner, whilst the thickest unit seen in the Wadi Teshuinat measured section (unit 6, Figs. 3.13 & 3.14) compares both in scale and sedimentary structures with the Battery Point sequence. The generally thinner nature of individual (rather than stacked) coarse members in facies MSd sequences suggests that this facies may have been deposited in a braid plain of shallower, less stable channels than that of the South Saskatchewan River. However, it is most probably an oversimplification to interpret all the cross-strata of facies MSd as channel deposits since the sediments deposited in intra-channel platform areas may also be preserved (see below).

In the following six sections consideration is given to the evidence provided by the sedimentary structures of facies MSd on the processes, and particularly the flow depths, which characterised the depositional setting of facies MSd. This approach was inspired to some extent by the similar approach adopted by Hodgson (1978) in his interpretation of the Lower Carboniferous Fell Sandstone braided river deposits of Northern England.

3.2.2.2 Cross-stratification

The abundant large-scale planar, and subsidiary trough cross-strata of facies MSd may have been created by a number of different processes. Under aggradational conditions migrating trains of straight and sinuous-crested ripple-forms (sandwaves, dunes, mega-ripples, etc.) create cosets of planar and trough cross-strata respectively (Allen, 1963; Dalrymple *et al.*, 1978; Harms, 1975a; Southard, 1975). Under stable flow conditions the dimensions of some of these ripple-forms are determined by flow strength. Large-scale cross-strata may also be created, however, during the construction of bars by a process of sediment avalanching on bar lee slopes from migrating bar-top ripples and ripple-forms (Smith, 1972) (Fig. 3.17). In this case set thickness is a response to the geometry and scale of the bar rather than to flow strength. The heterogeneous geometry of the sets of planar cross-strata of facies MSd, and the relatively rare occurrence of trough cross-strata, suggest that the process by which these sediments were laid down was more complex than that of the migration of a train of ripple-forms in stable flow conditions. Furthermore, the presence of grainsize graded foresets is evidence

for relatively small-scale ripple-forms superimposed on larger ripple-forms (see section 3.2.2.5). Both of these conclusions indicate that the model proposed by Jopling (1966) for the creation of planar cross-strata under unstable flow conditions may be more applicable to the sediments of this facies than models based upon the assumption of stable flow conditions. Jopling (1966) showed that planar cross-strata may be formed by a process of aggradation under conditions of changing flow strength and sediment discharge. Changes in base level may result in the development of an enlarged ripple which grows into a microdelta and which subsequently grows downstream by a process of foreset accretion. A set of cross-strata is created by this process, the thickness of which is related to attempts by the system to achieve an equilibrium profile in response to changes in flow depth.

Once a microdelta is produced it is only necessary to have continued aggradational conditions and some sediment transport over the top set of the microdelta for persistant downstream accretion of the avalanche slope. Sediment may be transported along the topset under any flow regime and still contribute to the foreset aggradation. Hence the structure has an inherent stability. Any change in flow conditions, which may include changes in flow velocity, flow depth, sediment discharge, sediment grainsize and flow direction, will, under conditions favouring aggradation, cause an enlarged ripple to form and grow into a microdelta, over which the consequent change in flow depth will cause downstream foreset accretion for as long as sediment is transported over the top of the microdelta (Jopling, 1966). Similarly any constriction existing in unidirectional or semi-unidirectional flow (resulting, for instance, from the development of dunes, sandwaves or bars during stable flow conditions) will be the site of flow expansion on the lee side of the bedform, bedload deposition and the creation of planar cross-stratification. Jopling (1966) argued, therefore, for an approach to the subject of the genesis of cross-strata that is based upon an understanding of the effects of the many changes in flow conditions which characterise natural depositional systems. Despite the emphasis that has existed in the literature for many years upon the study of bedforms generated under stable flow conditions, the earlier approach of Jopling (1966) has recently been supported (Leeder, 1981).

3.2.2.3 Controls on the dimensions and geometry of ripples, sandwaves and dunes

In attempting to establish what flow depths were characteristic of the depositional setting of facies MSd it is necessary to consider whether the thickness of sets of cross-strata may be used as evidence for depositional flow depth. Although it has been argued in the previous section that flow depths were undoubtedly constantly changing, it is probable that some of the cosets of trough and planar cross-strata of facies MSd were deposited under relatively stable flow conditions in which the bedforms that developed were in hydrodynamic equilibrium with flow strength. These conditions probably existed at certain times in the major channels of the braid plain. Leeder (1981) argued that bedform scale, an indication of which may be determined from set thickness, is only influenced by flow strength (and therefore flow depth) where outer flow zone processes are active in moulding the ripple-forms that are developed in migrating bedload material. The relationship of small-scale ripples and dunes to outer flow zone processes has been thoroughly documented in the literature but the position of sandwaves in the flow regime/bedform heirarchy is imperfectly understood. Allen (1968), Yalin (1964, 1977) and Guy *et al.* (1966) discussed the relationship between flow depth and the dimensions of ripples and dunes. In their work the boundary scale values between the two naturally-occurring distinct populations of ripple forms (ripples and dunes) are an amplitude of 4 cms and a wavelength of 60 cms (Allen, 1968). They concluded that ripples are formed in predominantly hydraulically smooth or laminar flows and that their dimensions are dependent not on flow depth but on grainsize and flow separation dynamics (Leeder, 1981), whereas dunes are limited to hydraulically rough, turbulent flows, and their dimensions are a response to grainsize and depositional flow strength, i.e. outer flow zone processes. Allen (1968) showed that a relationship exists between dune chord and flow depth, dune chord being approximately five times flow depth, whilst Yalin (1964) showed that dune height cannot exceed about one sixth of flow depth (*not* "about six times the flow depth" as stated in Allen, 1968, p. 138). From field evidence it has been shown that the ratio of height (H) to wavelength (λ_x) for ripples is different from that for dunes (Allen, 1970). The ratio

for dunes may be approximated by the equation $H = 0.074\ \lambda_x^{0.77}$ where $0.60 < \lambda_x < 2000$ metres. It should be noted that these relationships are established from lines of regression drawn through a scatter of data points and that the foregoing relationships are not consistent with each other. For instance, a dune of chord length 1 m should have a height of 7.4 cms and should have been generated in a flow depth of approximately one fifth of a metre on the basis of Allen's relationships (1968, 1970). According to Yalin (1964), however, a dune of 7.4 cms height should have formed in a flow of not less than 6 x 7.4 cms, i.e. approximately 44 cms. The flow depth calculation shows a discrepancy of just over a factor of two, therefore. It might be reasonable to assume from these calculations that flow depth must be greater than three times dune height and yet should generally be less than seven times dune height. This relationship is used a number of times in the following pages.

Sandwaves are difficult to interpret hydrodynamically because they only become clearly distinct from dunes in flow depths of greater than 50 cm (Southard, 1975). It is necessary, therefore, to include field evidence of flow depths unobtainable in flume tanks. Southard (1975) used evidence from estuarine medium-grained sandwaves to create a flow velocity/flow depth diagram. This leaves some doubt about whether the sandwaves studied were truly equilibrium bedforms. A further complication is provided by the fact that flume tank data on sandwaves in coarse-grade sands are well documented but data on fine to medium-grained sand (typical of facies MSd) is sparse. Despite these qualifications, Southard (1975), basing his interpretation on the straight-crested bars of Costello (1974), the intermediate flattened dunes of Pratt (1973) and on his own research, concluded that sandwaves belong in the stable-flow bedform heirarchy as a lower flow regime bedform type that is distinct from ripples and dunes. Sandwaves are characterised by a lower height to wavelength ratio than dunes, and a weaker dynamic coupling than exists with dunes between successive bedforms (the correlation of length from sandwave trough to crest with the height of the bedform immediately upstream is not as strong for sandwaves as for dunes (Costello, 1974)).

The distinction between sandwaves and dunes is complicated by the presence of bedforms that have characteristics lying between these two ripple-form types. Southard (1975) described dunes as bedforms

with sinuous (to irregular) crestlines and tangential to concave foresets, and sandwaves as bedforms with straight to slightly sinuous crestlines and generally angular foresets. The former generates trough cross-stratification whilst the latter generates planar cross-stratification. Dalrymple *et al.* (1978), however, recognised large-scale ripple forms (his megaripples type 1) in the tidal environment of the Bay of Fundy which had straight to slightly sinuous crests, which lacked scour pits and which had vertical form indices typical of dunes not sandwaves. These do not fit easily into the elegant but over-simplified bipartate classification. A migrating train of such megaripples would probably generate large-scale planar cross-stratification not dissimilar to that generated by migrating sandwaves.

Large-scale ripple forms vary, therefore, both in the degree to which the vertical form index complies with the equation established by Allen (1970) for dunes, and the degree to which their crests are continuous and straight, or curved and broken, the latter types generally, but not always, being associated with scour hollows. Allen (1968, after Conolly, 1965) suggested that a morphological gradation may exist between tabular sets of cross-stratification, broad sheets of sensibly constant thickness that were generated from straight-crested bedforms, and trough sets of cross-stratification, sets only four or five times wider than thick, generated from broken and sinuous-crested bedforms.

Leeder (1981) discussed the conditions of turbulent flow, sediment transport and bedform mechanics under which dunes and sandwaves are created and provided evidence to support the view that there is a considerable difference in the hydrodynamic conditions of generation of sandwaves and dunes. Jackson (1976, in Leeder, 1981) identified an inner and outer turbulent boundary layer. The inner layer (inner flow zone) corresponds approximately to Allen's laminar sublayer (1970, p. 37-39) and comprises spanwise streaks of low and high speed fluid, which are subject to periodic disruption and lift-up (bursts). The outer layer of the turbulent boundary layer (outer flow zone) comprises large-burst, "break up" vortices, and accelerated, downward-moving sweep fluid. A periodic "burst-sweep" cycle is identified which is responsible for most of the turbulence in a flow and for the vast majority of sediment transport.

Outer-layer flow variables were shown to determine mean burst

periodicity, even within the inner layer (Leeder, 1981). A coherent outer-layer flow structure exists (termed macroturbulence) which extends up to the free water surface and of which "surface boils" are an example. A poorly understood and complex relationship exists between macroturbulence and form-roughness of the bed, bursts being related to adverse pressure gradients in the lee of dunes, and burst spacial interval downstream, at a given point in time, being approximately five times water depth. This spacial burst interval may be related to bedform spacing, since, as noted earlier, dune chord length is frequently found to be approximately five times flow depth. This is supported by Yalin (1977) who developed the idea that bed discontinuities disturb the structure of turbulence at repeat distances of 4-7 flow depths. This work, therefore, supports the findings of earlier workers that dune spacing and geometry are determined by outer flow zone macroturbulence.

Bursts of the burst-sweep cycle are related to lee-side macro-separation eddies and may be responsible for the scour hollows frequently present at the base of lee slopes of dunes (Leeder, 1981). No evidence exists for a similar relationship, however, in sandwaves between flow depth, chord length and ripple-form height. Costello's flume tank studies of sandwaves (1974) indicated that these long, low bedforms with irregular wavelengths arise from chance bed defects produced by greater than average grain mounds, or bed irregularities. Leeder (1981) considered, on this evidence, that these bedforms may not be directly related to outer flow zone processes, but are probably related to grain to grain hindrance effects and are more similar therefore to ripples than they are to dunes. Ripples are unaffected by macroturbulence except to the extent that the burst-sweep cycle produces defects in the inner flow zone which, together with flow separation and reattachment, determines the ripple geometry for a given grainsize. Just as Jopling (1966) showed that micro-deltas may grow from enlarged ripples in unstable flow conditions, so Leeder (1981) argued that with increased but stable flow conditions small-scale ripples may be "cannibalized" by larger ripples which travel downstream at a greater rate assimilating smaller bedforms in the process. These "rogue" ripples may then grow into sandwaves. The transition from ripples to sandwaves with increasing flow strength would appear to be a gradational change, therefore.

As the slip face of a sandwave increases in size, effective lee-side macroturbulent bursts ("megabursts") may cause sufficient scour to transform the bed configuration from that of a bedform dominated by micro-flow separation dynamics to one controlled both by macro-turbulence and flow separation, i.e. sandwaves of long wavelength may be converted to dunes with shorter wavelengths. The trigger for this transformation may be the build-up of a critical excess pressure in the lee of a sandwave, which is sufficient to create an unfavourable pressure gradient and a separation eddy strong enough to interfere with the burst-sweep cycle of the outer flow zone. Such an analysis would suggest that the transition from sandwaves to dunes, with increasing but stable flow conditions, would be abrupt. Southard (1975, p. 24), however, argued that, whilst the ripple to sandwave transition is abrupt, the sandwave to dune transition is gradational. A great deal must depend upon the rate at which flow strength increases and the extent to which the bed has sufficient time to regain equilibrium with new flow conditions.

Although sandwaves are still poorly understood the work of Leeder (1981) and Jackson (1976) suggests that sandwaves and dunes are distinguished hydrodynamically by the presence or absence of the control of outer flow zone processes. Where macroturbulence is effective in moulding the bed, dunes that have wavelength to height ratios that conform approximately to the relationship established by Allen (1970) are created. The dimensions of these dunes are determined by burst/sweep cycle periodicity which in turn is related to flow depth. Where outer flow zone processes do not directly affect the bed geometry, sandwaves of variable dimensions are generated in which wavelength to height ratios are irregular but are always greater than those for dunes. The characteristics of sandwaves and the processes operating in their generation appear to be closer to those of ripples, "rogue" ripples and microdeltas, therefore, than they are to those of dunes. Since it is the effectiveness of flow macroturbulence in moulding bed geometry which determines whether sandwaves or dunes are created, and since a strong lee-side separation eddy may act as a trigger for effective macroturbulent control, the presence of distinct scour hollows in the lee of dunes may be considered their most intrinsic characteristic. The absence of scour hollows suggests that microflow separation dynamics (i.e. inner flow

zone processes) were dominant in their creation, in which case wavelength to height ratios are less likely to be constant and may be the same or larger than those for dunes. Dalrymple *et al.*'s megaripples type 1 (1978) are an example of the former, and sandwaves an example of the latter. Since migrating trains of dunes that have distinct scour hollows generate trough cross-strata, it may be concluded that flow depth calculations from set thicknesses are only valid in the case of the relatively rare cosets of trough cross-strata in facies MSd. Using the relationship established earlier in this section, set thicknesses of 50-100 cm which are characteristic of sets of trough cross-strata of facies MSd, suggest that depositional flow depth was between 1.50 m and 7 m. Although this statement is still very general it has relevance in considering whether a modern large river, such as the Brahmaputra River, would be a suitable alternative analogue to the Platte River. Furthermore it will be shown that the relationship of depositional flow depth to dune dimensions may be used to establish the depth of flow that was responsible for the creation of the larger of the grainsize graded foresets (section 3.2.2.5).

3.2.2.4 Lee-side flow structure

The flow structure on the lee-side of ripple-forms (ripples, sandwaves, dunes) is fundamentally the same as that of microdeltas (Jopling, 1961; Boersma *et al.*, 1968) (Fig. 3.17). This pattern of flow comprises a separation eddy of reverse circulation over which there is a zone of mixing and free turbulence and above which there is a zone of no diffusion. At the crest of the structure, flow separation or boundary layer separation (Allen, 1968) occurs which allows the separation eddy to form and which divides the transported sediment into two fractions: a) the bedload that accumulates at the brink of the lee slope until this slope exceeds the angle of repose of the sediment and avalanching takes place; and b) the suspended load that is transported further downstream. Downstream from the point of separation, at a distance commonly six to eight times the height of the set, the zone of diffusion expands to the extent that the flow becomes reattached to the bottom (Jopling, 1961). In ripple-forms, the point of reattachment is approximately two thirds of the distance up the next ripple-form stoss side (Reineck & Singh,

1973). In the zone of reverse circulation, flow velocities may be large enough to sweep fine-grained sediment (which has been deposited out of suspension through the zone of mixing) up to the bottom of the lee slope where it may be deposited either as fine ripple-laminae between coarser laminae formed by avalanching (zone of interweaving) or as a ripple-laminated bottom set layer over which avalanche foresets prograde (Boersma *et al.*, 1968) (Fig. 3.17). The strengths of the flow separation eddy, and the reverse circulation flow, vary between different large-scale ripple-forms and determine whether scour hollows are created, which, as noted earlier, is related to the degree to which outer flow zone processes control the geometry of the ripple-form.

The processes involved in the transition of sandwaves to dunes have been studied and summarised by Leeder (1981). In order to generate effective megabursts in the lee of sandwaves and to transform the whole bed geometry, flow strength must be increased but flow depth must not decrease to much below three times sandwave height. This is because the transformation from sandwaves to dunes must involve the creation of dunes of not less than the height of the superseded sandwaves and these must form in flows in equilibrium with dune height. Following the argument presented earlier it is also a requirement that flow over the sandwave must remain sufficiently deep for the generation of a critical excess pressure in the lee of the sandwave, a strong separation eddy, and interference with the burst-sweep cycle of the outer flow zone (Leeder, 1981). Since sets of planar cross-strata are commonly 70 cms thick it would seem that flow depths of 2.10 m to 4.90 m were not achieved (using the relationship flow depth equals 3 to 7 times dune height). In flow depths of less than 2.10 m the spacial periodicity of megabursts would be less than that of sandwave crests (Southard, 1975) with the result that bursting would break up flow on the stoss side of the sandwave causing superimposed dunes to form.

Further evidence that flow conditions suitable for the transformation of sandwaves/microdeltas to dunes were not achieved is provided by the characteristics of the foresets. Flow strength and the ratio of the water depth at the crest and at the base of the lee slope of a ripple-form are the essential controls on foreset type and shape (Jopling, 1966; Allen, 1970). Flow separation causes sediment

transported to the crest of the bedform to be continuously discharged over the ripple-form crest and convected and diffused in a turbulent-free shear layer before finally settling on the upper part of the lee slope and causing an increase in the slope of the avalanche face. Every time the angle of initial yield of the sediment is reached an avalanche of sediment leaves the crest. If sediment discharge is high, intermittent avalanching does not take place, but there is continuous general sliding down the lee slope. This latter process is not accompanied by any sorting of the sediment whereas grains carried down the lee slope by avalanches become size-sorted during the process, the coarser grains travelling furthest. Brush (1965, in Reineck & Singh, 1973) proposed that in avalanches down lee slopes coarser material should accumulate towards the outer side of the avalanching material and should travel furthest. This conclusion was reached by considering the fact that larger particles tend to move in the zone of lesser shear, and that, under the influence of gravity and by virtue of their size and higher settling velocities, they are transported the greatest distance. Inversely graded foreset laminae should result from these processes. Since these were not seen in facies MSd, nor were they reported by Smith (1971a, 1972) or Collinson (1970, 1978b), it may be assumed that an alternative process dominated.

Avalanching and lee-side rolling are characteristic of low flow strengths. As tractive force increases more and more bedload grains are catapulted into the turbulent eddy and are deposited on either the lower slopes of the avalanche slope or in front of it so that angular foresets become tangential or concave upwards. Similarly a decrease in sediment grainsize favours tangential and concave foresets over angular ones, and a shallowing of the basin in front of the dune has the same effect because, for a given tractive force, more of the suspended material is landing beyond the avalanche material and this emphasises the curved lower part of the foreset. The presence of angular foresets in the planar cross-strata of facies MSd indicates, therefore, that relatively low flow strengths and/or relatively deep lee-side basins prevailed.

The general absence of bottom sets, with the exception of the rare presence of wavy bottom set laminae, homogeneous bottom sets and back flow cross-stratification (Figs. 3.11 & 3.12) and the generally angular nature of foresets of cross-strata, is evidence, therefore,

that ripple-form, lee-side eddies were generally ineffective in the creation of "regressive ripples" (Jopling, 1961) or back-flow cross-strata (Boersma *et al.*, 1968) (Fig. 3.17). These characteristics also provide evidence for minimal suspension transport and, by implication, the absence of strong crestal flow separation and strong lee-side eddies. It also implies that flow depths over the crest of the bedform were relatively low and that disproportionately long foreset slopes existed. These flow strength implications indicate that conditions in the lee of the sandwaves were generally in disequilibrium with flow, and were considerably less turbulent than was necessary to achieve the transition to dunes, conditions characteristic of the downstream side of bars but generally not pertaining to an actively migrating train of dunes or megaripples. Further, it appears that, despite the existence of high flow strengths during flood-stage flows (evidence for which is provided by the presence of upper flow-regime horizontally-stratified sandstone beds), flow depths were only sufficiently great to create large-scale dunes or to transform microdeltas/sandwaves into dunes, in the deeper channels.

3.2.2.5 Grainsize graded foresets

It has already been noted (section 3.2.2.1) that alternating coarse and fine laminations in foresets of modern planar tabular cross-strata in the Platte River result from avalanching of sediment previously sorted by small bedforms on the surface of transverse bars, upstream from their slip faces (Smith, 1972). Lunate and linguoid dunes and ripples are usually preceded by scour pockets which contain the coarsest sediment in transport. These scour pockets, formed by lee-side eddies, move along in front of the migrating ripples and dunes, carrying the coarsest grains in them. As the bedform and its scour pocket approach the crest of the bar, the coarse material spills over first and is followed by continuous avalanching of the finer material in the ripple or dune. The result is a coarse and fine-grained lamina couplet, or grainsize graded foreset, the thickness of which is determined by the length of the foreset and the size of the parent bedform plus scour pocket (Fig. 3.17).

Dunes have been observed migrating over the tops of bars and over the stoss side of sandwaves, towards the crest, in a number of studies (Collinson, 1970; Smith, 1971a, 1972; Harms *et al.*, 1974;

Coleman, 1969; Boothroyd & Hubbard, 1976). Harms *et al.* (1974) considered that the sandwaves and superimposed dunes were in equilibrium with the flow, whereas Allen and Collinson (1974) argued that the phenomenon resulted from unstable flows, the sandwaves forming under different flow conditions than the superimposed dunes. McCabe and Jones (1977), after conducting flume experiments, cast doubt on the interpretation of graded foresets given by Smith (1972) and supported, in part, by McBride *et al.* (1975). They observed a concentration of coarse-grade material at the base of the slip face, resulting from avalanching of mixed-grade sediment, which produced a base-of-set lag deposit that was similar to those seen in facies MSd. They did not, however, observe any processes which created grainsize segregation of the whole foreset. Jopling (1964) noted, from flume microdelta studies, that probably the most important sorting process in the development of size-differentiated foreset laminae is not sorting as a result of the avalanching process itself, but rather segregation of grainsizes along the stream bed upstream from the avalanche face (see Boersma, 1967). He did not, however, consider that this segregation might occur by the process described by Smith (1972) despite his observation (Jopling, 1965) that sediment is often transported along the topset of microdeltas as ripples. Instead he considered that fine and coarse-grade, foreset-laminae alternations result largely from the alternation of traction bedload and suspension load deposition on the avalanche face (Jopling, 1966).

Collinson (1970) observed two foreset types in trenches through linguoid bars in the Tana River in which tabular sets of cross-strata with angular foresets were exposed. The finer-grained foresets decrease in thickness downwards and were thought to be deposited from suspension and to be composed of sediment transported over the stoss slope by saltation. The coarser-grained foresets, on the other hand, thicken downwards and were thought to have been deposited by avalanching of material transported to the crest of the bar as ripples. The observations in the Tana River suggest that Jopling's division of foresets into suspension and avalanche types is valid and that Smith's foreset couplets (1972) are to be found within the downward-thickening avalanche-type foresets. Although the hypothesis that variations in the heavy-fluid layer may be responsible for grainsize segregation of avalanched material (Jopling, 1966) cannot be ignored, it is considered

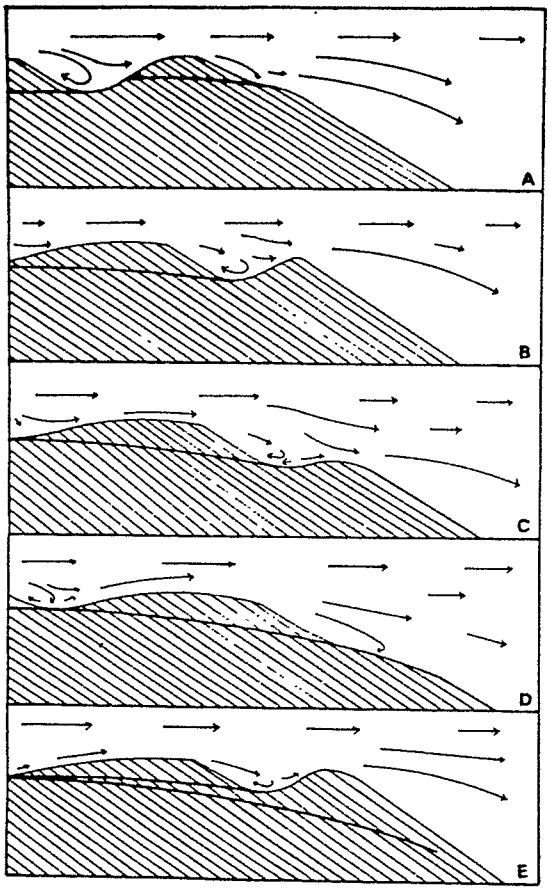
that the process described by Smith (1972) for the genesis of grain-size graded foresets offers a convincing explanation for this phenomenon in facies MSd. From the evidence that graded foresets provide about the dimensions of bar-top dunes, conclusions may be drawn about the outer flow zone conditions applying at the time of the migration of these dunes. This approach, therefore, provides evidence about the flow depths over the top of the bar or sandwave at the time of active foreset accretion.

It is a simple matter to calculate that, for a set thickness of 70 cms and a foreset angle of 30° from the horizontal (= foreset length of 140 cms) a foreset thickness greater than approximately 1.5 cms indicates that the parent bedform and its scour pocket must have been a dune rather than a ripple. Since many of the foresets of the planar cross-strata of facies MSd are thicker than 1.5 cms and lie in sets of at least 70 cms thickness, it may be assumed that superimposed dunes were responsible, at least in part, for the creation of the planar cross-strata. Graded foresets of up to 10 cm thickness must have been formed by dunes of up to 10 cm in height and 145 cm wavelength since the cross sectional area of a 10 cm thick foreset, in a set 70 cm thick, is the same as that of a dune and accompanying scour pocket of 10 cm height and 145 cm wavelength, assuming an angle of inclination of 30°. If lower angles of inclination existed the dimensions of the superimposed dune would have been slightly larger. These dunes provide evidence, therefore, using the relationship discussed in section 3.2.2.3, for flow depths over the stoss side of the order of 30-70 cm during active dune migration. These flow depths are not necessarily an indication of the maximum flow depth over the sandwave or bar in which the grainsize graded foresets were deposited. However, it seems reasonable to assume that deeper flows over the sandwave or bar were either of insufficient strength to cause a transformation to dunes, or of such strength that sediment was continuously transported to the crest of the bedform in upper flow regime conditions. The homogeneously-sorted foresets of the sandwaves may well have been formed by this process. Although flow depths may have exceeded 70 cm over the sandwave during high-stage flow conditions, therefore, flow depths appear to have been insufficient in the areas away from the main channels of the braid plain (i.e. generally less than 2.10 m water depth, see section 3.2.2.4) to transform the sandwaves or bars to dunes.

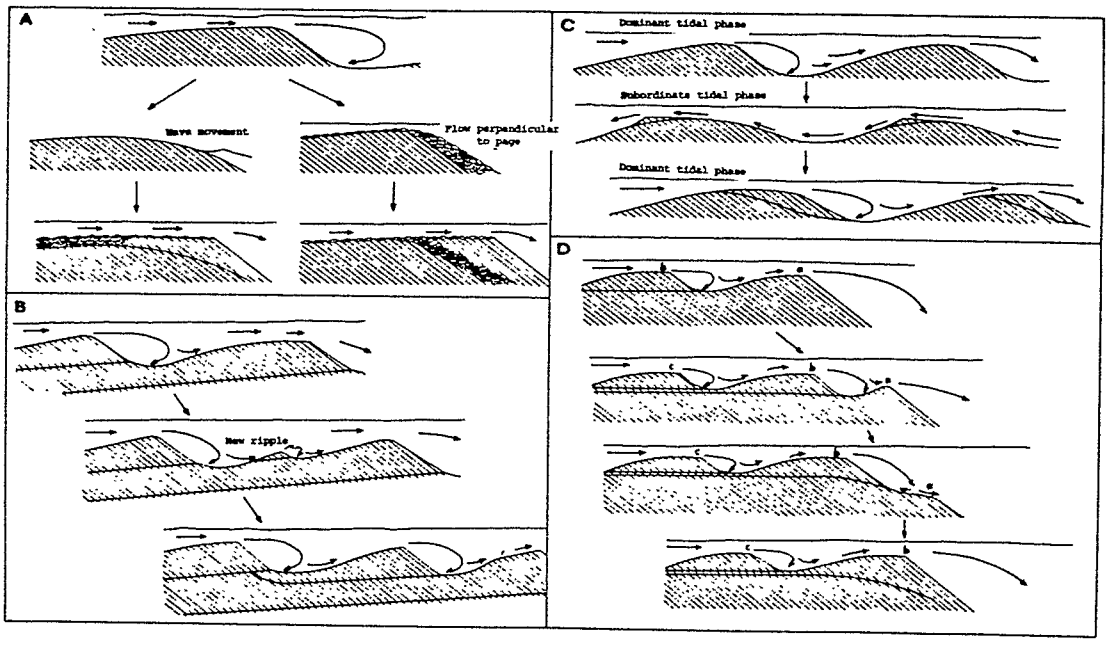
3.2.2.6 Reactivation surfaces

Evidence for modification of bars and ripple-forms during falling-stage flow conditions, and of dissection of bars during the low-stage, is provided by the presence of reactivation surfaces and small-scale cross-stratified sandstones lying above reactivation surfaces that are present within cosets of predominantly large-scale sets of cross-strata. McCabe and Jones (1977) showed how convex-upwards reactivation surfaces are produced by washing out of the top of large-scale ripple-forms by superimposed ripples. They also reviewed the processes by which reactivation surfaces are formed, concluding that only certain varieties of reactivation surfaces provide evidence of falling and low-stage flow conditions (Fig. 3.18). Reactivation surfaces may be formed by random interaction of bedforms (under constant flow stage) (Allen, 1973b) or by changes in flow direction in tidal environments (Klein, 1970) (Fig. 3.18). Collinson (1970) argued that reactivation surfaces in linguoid bars of the Tana River resulted from low-stage modification by wave action or flow around the bedform, particularly parallel to the slip face. Unit 9 (Fig. 3.11) is thought to have formed by this process. The bar which was dissected, in this case, shows evidence of an earlier hiatus when cross-laminated argillaceous siltstone was built up the lee face of the bar by low-stage currents. The presence of reactivation surfaces of the type described by Collinson (1970) and by McCabe and Jones (1977) in facies MSd is thought to provide evidence, therefore, for fairly rapid changes in flow stage in the former case and for fairly constant or slowly-changing flow conditions in the latter case.

A third type of "reactivation surface" is seen in facies MSd which should not strictly be termed a reactivation surface according to the definition of McCabe and Jones (1977, p. 713) since it is not present within a "cross-bed". Instead, it is an irregular surface over one or a number of planar sets of cross-strata. These sets are not continuous but terminate laterally "down-current" and are overlain by a planar wedge set of cross-strata which thickens down-current by overstepping and engulfing the underlying sets. The set above the "reactivation surface" is characterised by a planar top surface and an irregular basal surface, therefore (see reactivation surface in Fig. 3.9, lower). This type of reactivation surface is thought to



Modification of the crest of a large-scale ripple-form during washing out of a superimposed ripple. Convex-upwards reactivation surfaces that truncate the upper part of the foresets of the ripple-form are produced.



Methods of forming reactivation surfaces, modified from (A) Collinson (1970), (B) Allen (1973b), (C) Klein (1970), and (D) McCabe & Jones (1977).

Figure 3.18 FORMATION OF REACTIVATION SURFACES IN LARGE-SCALE RIPPLE-FORMS (AFTER McCABE & JONES, 1977)

result from bar construction during rising or high-stage flow conditions when slip-face foreset accretion causes the migration of a bar over earlier low-stage bedforms into a depression or pool in the braid platform down-current from the bar. It may also have been produced by the development of a delta lobe at the downstream end of a bar-top or braid-platform channel (Bluck, 1974).

Although the reactivation surfaces of the type described by Collinson (1970) provide evidence of low-stage flow around bars which may have been emergent, the other types of reactivation surfaces seen in facies MSd, whilst providing some evidence for stage fluctuations, do not provide any evidence for low-stage emergence of bars. No evidence for subaerial exposure of bars, or other parts of the braid plain, in the form of desiccation cracks or caliche horizons, was observed, and the biogenic traces of subfacies Fl, suggest that the abandoned channels and pools of the braid plain rarely dried up completely and, perhaps, that the tops of bars were rarely emergent either. The persistence of the "pools" and "abandoned channels" through time suggests that the braid plain may have been the site of some extensive lakes. The River Tana provides a good example of this type of despositional setting (Collinson, 1970). Bars within the channel system are commonly always submergent, and perennial lakes and pools in the braid plain contain sediment-feeding organisms (Collinson, 1978b).

3.2.2.7 Soft-sediment deformation structures

Recumbent-folded deformed cross-stratification is abundant in this facies and provides evidence for pore water pressure fluctuations in recently deposited sediment resulting from changes in superincumbent sediment load (McKee *et al.*, 1962). Alternatively this phenomenon may be attributed to changing water levels in the lee of a bar (Bluck, 1974). Reduced overburden and/or depth of water allows water to escape from previously slightly over-pressured sands causing fluidisation of the sediment. The fact that the foresets of this structure are invariably deformed "downstream" would seem to suggest that the deformation is to some extent current-induced. In a number of cases (e.g. unit 7, Fig. 3.14, just out of section) a reactivation surface was seen that truncated a succession of beds not thicker than 1 m and including a bed with recumbent-folded foresets. If the

truncation plane does not indicate erosion of a considerable thickness of sediment, an assumption that is difficult to verify, then the folding of the foresets would have occurred after only minor burial. In the case of unit 7 (Fig. 3.14) a probable explanation is that current-induced horizontal stresses, perhaps produced during flood stage when overburden sediment was removed, or during falling stage when water depths decreased, were responsible for deforming sediment that was undergoing fluidisation at the time. It is also possible that earthquake tremors were responsible for the fluidisation and sediment deformation (Allen & Banks, 1972) but this explanation alone does not account for the "down-current" character of the deformation.

3.2.2.8 Palaeocurrent data

The low-variance palaeocurrent rose diagrams plotted from foreset dip azimuths of stacked sets of planar cross-strata suggest that these represent the deposits of straight to slightly sinuous-crested, similarly-orientated transverse bars. Ore (1963) and Smith (1972) emphasised that the sinuosity of bar crests in straight braided channels determines the character of the palaeocurrent rose diagram and, whilst differing on certain aspects of palaeocurrent interpretation, they both concluded that a wide dispersion would be expected. Banks and Collinson (1974) argued, however, that the cross-strata deposited from a migrating transverse bar, with a convex-downstream crest, would show a much lower variance of azimuth orientations than a simple consideration of the geometry of the bar crest would suggest. They concluded that braided-stream deposits, as suggested by Kelling (1968), Allen (1970), Blatt *et al.* (1972) and others, are indeed characterised by low palaeocurrent variance. Miall (1977) argued that braided streams are characterised by lower directional variance than other more or less unimodal flow systems including meandering streams and many delta distributaries. He also noted that planar cross-strata in fluvial systems commonly show higher directional variance than trough cross-strata since bar avalanche slopes may be orientated at high angles to the channel direction. It follows that low directional variance from planar cross-strata suggests deposition from fairly straight-crested transverse bars migrating down channel, rather than from sinuous and horn bars, such as those observed in the South Saskatchewan River (Cant & Walker, 1978).

The degree of low-stage modification of high-stage deposits may also determine the degree of palaeocurrent variance (Smith, 1974). The low-variance palaeocurrent data from facies MSd would suggest that low-stage modification was not extensive. This is evidence that changes in discharge in the rivers responsible for facies MSd may generally have been very rapid, unlike those of the Platte River, not allowing time for significant bar modification (Jones, 1977) although the presence of convex-up reactivation surfaces is evidence, on the contrary, for gentle flow stage changes. Unfortunately insufficient palaeocurrent data were collected from low-stage bar dissection structures to establish the degree of divergence that these show from the prevailing palaeocurrent orientation, but divergences of up to 90^{o} were recorded. The divergence of orientations between groups of units showing low-variance palaeocurrents (Figs. 3.13 & 3.14) may indicate major changes in style of bar migration, accretion and preservation, or changes in the orientation of the channel in which the bars were situated. The palaeocurrent data from facies MSd in the lowest 40 m of the Tadrart Formation (second transgressive sequence, see Fig. 3.6) were plotted together for each of the successions measured in the study area and are discussed in section 3.7 (see Fig. 3.55). The three southern rose diagrams, with 20 readings each, or more, show an internal variance of 150^{o} to 180^{o}, a bimodal or trimodal distribution, and very similar vector mean directions. They are thought, therefore, to provide support for a fluvial interpretation although they represent such a grouping of data from thick successions of facies MSd that these rose diagrams are considered to provide little evidence about the type of fluvial system (Collinson, 1978b, pp. 57-58).

3.2.2.9 Biogenic structures

The trace fossil evidence of facies MSd, and the biogenic traces in subfacies Fl, cannot be used directly to distinguish continental and marine sediments although similar trace fossil assemblages have frequently been quoted in support of arguments for a marine depositional setting (see Chapters 1 & 2, sections 1.1.1 & 2.3). However, the presence of *Fraena*, *Arthrophycus* and *Palaeophycus* in the argillaceous siltstone intercalations of subfacies Fl, and of *Skolithos* in adjacent sandstone beds, indicates the former existence of organisms

which built complex internal sediment-feeding burrow systems (*Arthrophycus* and *Palaeophycus*), organisms which occupied vertical burrows and lived by filter feeding of suspended sediment (*Skolithos*) and organisms which crawled along the sediment surface (*Fraena*) (see Table 2.5). These trace fossils belong to the *Skolithos* and *Cruziana* ichnofacies of Seilacher (Table 2.6) but are thought, from their sedimentological context, to be evidence for the occupation by soft-bodied organisms of perennial lakes in the braid plain, abandoned channels and pools, and of restricted marine environments (lagoons and estuarine channels) created during transgressions of the sea over the braid plain. *Arthrophycus* is present in two distinct forms, a palm-frond shaped bundle of curved burrows and a straight to sinuous isolated burrow variety (Fig. 2.6). Facies MSd is dominated by the straight-burrow variety although small-scale examples of the bundle variety are also rarely present. The presence of *Arthrophycus* has frequently been quoted as evidence for a marine environment of deposition (e.g. Potter & Pettijohn, 1977, in the Silurian Tuscarora Sandstone Formation of the Appalachians) but Cotter, also working on the Tuscarora Formation (1978, & pers. comm. 1979), has cast doubt on this, suggesting that both marine and lacustrine varieties are present, the latter being observed in thin shale intercalations within a thick succession of cross-stratified sandstones interpreted as being of braided fluvial origin. It is possible that further work would show that the different varieties of *Arthrophycus* present may be related to different environments/water salinities (see section 3.3.2.8).

The sedimentary structures of the trace-fossil bearing beds, which include planar cross-strata, the upper surfaces of which commonly display current ripples with superimposed trails (subfacies Sp), lenticular cross-lamination and intraformational clay clasts (subfacies Fl), and ripple cross-lamination (subfacies Sr), provide support for an interpretation of these beds as the deposits of lakes, lagoons and abandoned channels since they are the product of sudden changes in flow strength and of gentle current winnowing in quiet shallow-water conditions (Collinson, 1978b). The sudden changes in flow strength occur when a channel is abandoned, or when flow into a lake/lagoon is abruptly terminated by channel avulsion. The difference between the trace fossils present in the sandstone beds and those found within the argillaceous beds is probably largely a function of

substrate type, nutrient availability and current conditions at the sediment/water interface.

The trace fossils and biogenic traces of this facies provide evidence, therefore, that the sediments of subfacies Fl and adjacent sandstone subfacies (Sr and Sp) were capable of providing a habitat for sediment and suspension-feeding organisms, which must have been one of quiet water and only rare subaerial emergence. Lagoons and extensive permanent lakes, abandoned channels and pools in fluvial floodplains provide this habitat. It seems unlikely that each of the trace-fossil bearing beds represents a transgression resulting in the development of estuarine/marine environments, since the fine-grained intercalations are laterally impersistent and are present at a large number of different stratigraphic horizons. The presence of facies Fl at the "40 m marker plateau", however, is thought to indicate the transition from fluvial deposits of facies MSd to estuarine/marine deposits of facies CSdSt.

The interpretation presented here for the trace fossils of facies MSd is similar to that proposed by Selley (1970) for Jordanian Nubian-type sandstones, but is perhaps more strongly dependent upon the sedimentological setting of the biogenic structures. This emphasis has been applied because of the somewhat inconclusive results of earlier attempts to discriminate marine and non-marine depositional environments using trace fossils (see Chapters 1 & 2, sections 1.1.1 & 2.3).

3.2.2.10 Conclusions

Cross-stratification is so abundant in this facies that the conditions under which it was formed have been discussed at some length. Cosets of trough cross-stratification are interpreted as the product of migrating sinuous and lunate-crested dunes which were present in the channels of a braid plain and from the dimensions of which, inferred from set thicknesses, flow depths of 1.5 m to 7 m have been calculated. The interpretation of planar cross-stratification is more complicated, since it may have formed by migration of sandwaves, straight-crested dunes or microdeltas/bars. Since it is not clear how the cross-stratification produced by straight-crested dunes may be distinguished from that produced by sandwaves, and since the dimensions of microdeltas/bars and sandwaves are not controlled

by outer flow zone processes, set thicknesses of planar cross-strata may not be used to infer depositional flow depths. However, the evidence provided by grainsize graded foresets for superimposed small-scale dunes indicates that flow depths, during active migration of the bedforms responsible for the planar cross-strata, was of the order of 30-70 cms over the crest of the bedform. The discrepancy between the flow depths calculated for the crest of these bedforms and the flow depths calculated from the trough cross-strata is thought to provide evidence that the trough cross-strata were formed during flood-stage deposition in the deeper of the braid-plain channels, whilst the down-stream foreset accretion of microdeltas/bars and sandwaves took place during all flow stages both within channels and in platform areas of the braid plain.

The evidence for upper flow regime conditions provided by the fine to medium-grained, horizontally-stratified sandstones (subfacies Shz) (Guy *et al.*, 1966), and the relatively rare occurrence of cosets of trough cross-strata, are thought to indicate that although flow velocities were occasionally sufficient to transform sandwaves into dunes this rarely occurred away from the major channels because water depth was generally too shallow, i.e. generally less than 2.10 m. The presence of abundant grainsize graded foresets, however, indicates that flow depth/flow strength conditions were frequently such that outer flow zone macroturbulence was capable of moulding the bedload material into small-scale dunes which migrated over a bed "landscape" the irregularities of which were too great to be eliminated or transformed into large-scale dunes.

Taking into account the evidence provided by the reactivation surfaces, low-stage modification structures and upper flow regime horizontal stratification of facies MSd, for fluctuating flow conditions, and allowing for the absence of evidence for subaerial exposure of the bedforms, the following interpretation of the planar cross-strata and associated beds of this facies has been established, based upon a refinement of the earlier interpretation based upon Miall's subfacies (1977) (see section 3.2.2.1).

The depositional setting was a braided river system in which the major channels were typically a few metres deep (characteristically perhaps 2-4 m, but possibly up to 7 m deep) and lay in a braid plain of low topographic relief and poorly developed channel banks, in parts

of which extensive lakes were developed. Transverse bars, typically 70-100 cm high, but in places up to 2.8 m high, and with straight to sinuous crests of up to 40 m in length, were present in wide channels, many of which were shallower than the major channels of the braid plain. During flood-stage flows, some scouring and erosion of previously-deposited sediment took place. Net aggradation of sediment occurred in the depressions of the fluvial plain on the lee-side of transverse bars by the avalanching of sediment carried as bedload downstream to the crest of the bar. Aggradation also took place by the initiation of new bars as a result of the growth of rogue ripples and sandwaves/microdeltas from pre-existing ripples in response to changes in flow conditions. At the height of the flood stage, upper flow regime conditions existed depositing horizontally-stratified sandstones on scoured surfaces and bar tops, and transporting sediment to bar crests. As the flood-stage waned, lower flow regime conditions predominated and dunes were created which migrated downstream on bar stoss slopes in flows of 30-70 cm depth. Flow depths between bars were typically of the order of 1-2 m and were insufficient even in flood-stage to transform the sandwaves/microdeltas into dunes. As flow stage continued to wane ripples instead of dunes migrated over the bar stoss slope, and bar dissection and modification processes predominated. Continued bar migration downstream resulted in the development of internal reactivation surfaces. Flow appears not to have waned to the extent that bars were subaerially exposed for long periods of time. During the following flood stage the uneven topography resulting from the previously described processes provided the catalyst for the development of lee-slope avalanche faces which migrated downstream.

Although evidence for low-stage emergence is not strong, the abundant reactivation surfaces and stratification types indicating fluctuations of flow strength and minor upper flow-regime conditions, and the strength of evidence in favour of abundant transverse bar deposits and very subordinate dune and megaripple-train deposits, suggests an interpretation for facies MSd based upon the Platte, Tana and South Saskatchewan rivers as modern analogues. The thick stacks of multi-storey planar cross-strata may have been deposited as successive sand flats (Cant & Walker, 1978). These have been recorded as growing from mid-channel transverse bars, a small part of which becomes

emergent during falling stage and acts as a nucleus for down-stream and lateral growth into sand flats 50 m to 2000 m long. They are composed of thick (1-2 m) sets of planar cross-strata (the original transverse bar) and smaller sets of planar tabular cross-strata, with subordinate parallel-stratified, trough cross-stratified and cross-laminated sandstones.

Although the sand flats are emergent during low flow stage, features developed on their surface providing evidence of their emergence are removed during the next high flow stage. The abundant clay clasts incorporated in cross-strata may have originally been desiccated clay drapes. Similar sand flats are developed in the Brahmaputra (Coleman, 1969) and the possibility that this big river might provide a modern analogue for facies MSd is considered in the interpretation of facies CSdSt (section 3.3.2.10). Bluck (1979) gave two examples of fine-grained braided alluvium, the characteristics of which are controlled by flow depth and channel floor relief. In the deeper of these two examples (flood depth less than 3 m) three stratification types are developed in the following vertical sequence; planar stratified sands overlie an erosion surface and planar cross-strata and pseudo-planar stratified sands lie above. Where the low flow stage channel has greater relief the rhomboidal-shaped bars responsible for the pseudo-planar stratification are replaced by planar cross-stratified sand flats. This suggests that the rivers responsible for facies MSd were typically characterised by channels in which flood water depths were not much shallower than 3 m, and it provides support for the South Saskatchewan River analogue, in which channels are typically 3-5 m deep. This conclusion is consistent with earlier conclusions that the depositional setting of facies MSd was one of slightly shallower channels than characterise the South Saskatchewan River, and that channels were perhaps typically 2-4 m deep.

An alternative model to that of the South Saskatchewan River was proposed by Bluck (1974) for the construction of preserved sheet-sandstone complexes (lithosomes) from his work on sandur deposits in southern Iceland. Transverse (or lateral) bars were observed to develop on the inside banks of sinuous reaches of major braided channels, where they were preserved as the channel migrated undercutting the outside bank of the bend in the river. As the channel

continued to undercut this bank a succession of transverse bars were stacked together in a platform area which was topographically higher than the main channel but lay within the channel complex. Transverse channels cut across this platform area and opened out into deeper slough channels, at the downstream margin of the platform, building small deltas as a result of flow expansion. Continued undercutting of the outside bank of the main channel left the early-formed transverse bars progressively further removed from the main channel and these were ultimately accreted to form a supra-platform area composed of sandstones bearing the sedimentary structures of transverse bars, transverse channels and small deltas. It was inferred that the bedforms of the main channel would not be preserved by this process (although those of slough and transverse channels would be) with the result that trough cross-stratification and major channelled sandstone bodies would not be a very significant component of the lithosome. This type of lateral accretion of transverse bar deposits may well explain the sheet-like character of facies MSd, and the relatively minor importance of the deposits of migrating dune-trains. The preserved deposits, studied by Bluck (1974), however, were characterised by very variable cross-strata foreset dip azimuths, as the cross-strata were formed mainly during low-stage flow when flow orientations were widely dispersed. The cross-strata of facies MSd, on the other hand, have a low variance of palaeocurrent orientations, suggesting that deposition took place predominantly from fairly straight-crested transverse bars during high flow stage and that the resultant cross-strata, once accreted into a lithosome, were not extensively modified.

Harms (1980) encountered the same low palaeocurrent variance in the late Jurassic-early Cretaceous Nubian-type sandstones of Egypt, in that part of the succession of cross-stratified sandstones that had very similar characteristics to those of facies MSd. The presence of rootlet horizons within beds similar in appearance to subfacies Fl, and the presence of abundant convex-upwards reactivation surfaces (see Fig. 3.18, upper), led to a fluvial interpretation for these beds based upon a model of low-relief, perennial rivers in an extensive alluvial plain in which falling-stage structures were created but in which low-stage bar-dissection events, such as characterise ephemeral, flashy braided-river systems, were not very significant. Harms (1980) proposed that the evidence suggested, therefore, that the channels of

this alluvial plain must have been straight to only slightly sinuous. The processes by which the transverse-bar, or sandwave, deposits were accreted into a lithosome may well have been similar to those described by Bluck (1974) and Cant and Walker (1978) in which case the palaeocurrent data should be interpreted as providing evidence for very wide channels and the former existence of straight-crested transverse bars rather than for straight channels, although, of course, no evidence exists in either the Mesozoic or the Palaeozoic Nubian-type sandstones under discussion for well-developed meandering-channel systems.

Harms (1980) described a vertical sequence of beds which is commonly present in the Nubian-type sandstones he studied, in which a coset of planar tabular cross-strata lies above a sharply-erosional base and is overlain by a haematitic, kaolin-rich root horizon of fine-grained sandstone in which plant fragments are commonly present. The cosets of cross-strata are characterised by upwards-thinning sets, grainsize graded foresets inclined at 25^{o}-26^{o} and displaying very low dip azimuth variances, and by soft-sediment deformation including recumbent-folded foresets. This vertical sequence of beds is very similar to those described from facies MSd (see Figs. 3.11, 3.13 & 3.14) and to those of the Upper Cambrian Mount Simon Sandstone (Driese *et al.*, 1981), and was interpreted by Harms (1980) to be channel sandwaves overlain by vegetated flood-plain deposits, the latter formed by the process described by Walker (1974) whereby clay deposited from suspension is infiltrated into the interstices of the previously-deposited sand in very shallow water in flood-plain pools, lakes and abandoned channels. The root horizons, plant remains and unimodal palaeocurrent data were sufficient evidence for Harms (1980) to propose a non-marine interpretation for his beds, whereas the presence in the clay-rich beds of trace fossils similar to those encountered in facies MSd persuaded Driese *et al.* (1981) to propose a marine estuarine interpretation for the Mount Simon Sandstone beds. It is considered that, despite the trace fossil evidence of facies MSd, the evidence of the petrography, the sedimentary structures and the unimodal low-variance palaeocurrent data (see Fig. 3.55) supports a fluvial interpretation. Evidence has been provided by Ashley (1978) and Middleton and Hereford (1981) that tidal influence in fluvial systems generally results in some bipolarity

of foreset azimuths, even if this is only very rare.

It has been noted earlier that the absence of preserved channel-shaped erosional scour surfaces in facies MSd may be attributed to the process by which transverse bars are accreted into a lithosome (Bluck, 1974) whereby the major channel sandstone bodies are only rarely preserved. An alternative explanation for this phenomenon is that the fluvial deposits were laid down on large-scale humid alluvial fans or inland deltas (Boothroyd & Nummedal, 1978; Bull, 1972; Friend, 1978; Gole & Chitale, 1966; Heward, 1978) where channel incision was not widespread and alluvium was deposited on convex-upwards lobes of varying scales from streams and rivers which decreased in size down-stream. Selley (1978) proposed that the cyclicity of coarser-grained, thickly-bedded sandstones with finer-grained bioturbated sandstones, which characterises the Jordanian Cambro-Ordovician Nubian-type sand-stones, and the Central Saharan External Tassilian Sandstones, may be a result of periodic uplift, or sea level fall, initiating, on each occasion, a cycle of alluvial fan development at the foot of a landward-retreating escarpment. He interpreted the coarser-grained cross-stratified sandstones, therefore, as the deposits of braided-streams which flowed over a coalescing complex of broad, imbricated alluvial fans on a piedmont plain, similar in some respects to the interpretation of some of the sandstones of the Torridon Group of Scotland (Stewart, 1975). Assuming that the escarpments were composed of earlier Nubian-type sandstone sequences, then the absence in facies MSd of conglomerates, which are normally characteristic of alluvial fans (Heward, 1978), and the commonly texturally-immature yet miner-alogically-mature petrography of the sandstones of facies MSd are not inconsistent with this hypothesis. However, all the evidence suggests that, although the rivers are thought to have been perennial and relatively shallow, they were not characterised by abundant low-stage reworking and subaerial emergence. This is generally inconsistent with the characteristics of alluvial fan deposits although Boothroyd and Nummedal (1978) noted from studies of a large-scale humid alluvial fan that sedimentary structures formed during subaerial exposure on the tops of bars are rarely preserved in the sediments deposited at the centre of braided channel complexes because of later scouring events. One line of evidence against the retreating-escarpment hypo-thesis is that, in some cases, the escarpment undergoing denudation

would not have been composed of sufficiently coarse-grade sediment to supply detritus to the alluvial fan supposedly deposited by this process (see section 3.7).

3.3 Channelled Sandstone and Siltstone Facies (CSdSt)

3.3.1 Description

3.3.1.1 General

Sandstone successions with more abundant argillaceous siltstone intercalations than are present in facies MSd, and characterised by channelling on a variety of scales, have been assigned to the channelled sandstone and siltstone facies. Although scour and fill and channel fill cross-stratification is present in facies MSd (subfacies Ss), this facies may be distinguished from facies MSd by the presence of recognisable channel-fill sequences and a variety of types of channelled scour surfaces that are clearly distinct from those at the base of scour and fill/channel fill cross-stratification. Nevertheless, facies MSd and facies CSdSt are very similar in a large number of respects and are difficult to distinguish without good exposure.

The channelled sandstones of facies CSdSt are present, usually associated with facies MSd, in all four transgressive sequences (Fig. 3.6) as one of the facies of the coarse-grained units of the Tadrart Formation identified by Klitzsch (1965a) (Fig. 3.1). This facies commonly lies above facies MSd, but it has a complex, inter-fingering relationship with both facies MSd and RBSdSt. The facies is characterised by multi-storey planar cross-stratified sandstones and bedding sequences of channel-fill sandstones lying above and within scoured channel-shaped basal boundaries. Grainsize variation and palaeocurrent directional variance are more marked within the stacked cross-strata of this facies than in those of facies MSd.

As with facies MSd, this facies has an abundance of reactivation surfaces, intraformational clay clasts, soft-sediment deformed strata (including recumbent-folded foresets), and grainsize graded foresets in thick planar sets of cross-strata of, exceptionally, up to 2.5-3 m thickness, but typically of 70 cm thickness. Compound cross-stratification is commonly present in both facies CSdSt and MSd, but this facies is less lithologically homogeneous than facies MSd.

Argillaceous siltstone horizons are more abundant, although generally laterally impersistent, in thin and very thin beds, and they show more evidence of bioturbation than those of facies MSd. The trace fossils present, however, are broadly the same, the argillaceous horizons yielding abundant *Arthrophycus* and examples of *Palaeophycus* and *Fraena*, whilst adjacent fine-grained relatively well-sorted sandstone beds commonly contain *Skolithos*. In addition, however, some examples of *Cruziana* were found in the argillaceous beds, whilst some of the channel sandstones also contain *Rosselia*. Intraformational conglomerates are present above major scour surfaces and are generally overlain by thick sets of tabular or trough cross-stratified sandstone. The conglomerates may be up to 60 cm thick and are generally monomict and clast supported, containing clasts of fine-grained sandstone or argillaceous siltstone of up to 40 cm in diameter, but commonly of 5-10 cm diameter. The clasts lie within a very poorly-sorted, sandstone matrix and may be slightly rounded or tabular in shape. It proved impossible to correlate intraformational conglomerates over any distance and it seems likely that they are not laterally persistent for more than a few hundred metres.

The sediments of facies CSdSt have been divided into four subfacies, the first three being characterised by different varieties of channel and scour structures, and the fourth being characterised by an abundance of clay drapes and intercalations. The sediments of subfacies 1, and their sedimentary structures, are similar to those of subfacies Sp of facies MSd, but are distinguished by the presence of abundant broad shallow scours that are concave upwards and channel shaped. Channel-fill vertical sequences, comprising a large variety of bedding types including beds of trough, planar and scour and fill cross-stratified, and cross-laminated sandstone, characterise subfacies 2 and 3, the latter being distinguished by the presence of abundant biogenic structures and by sparse shell moulds and casts. The sediments of subfacies 4 display no channelling but are characterised by an alternating sequence of beds of sandstone and argillaceous siltstone/shale.

3.3.1.2 Subfacies 1

Multiple shallow scour surfaces, up to 12 m wide and generally less than 1 m deep, overlain by thin argillaceous siltstone horizons which are overlain, in turn, by angular foresets of cross-strata filling the concave-up, channel-shaped scour, characterise this subfacies. Recognition of these scour surfaces is dependent upon good exposure without which the erosion horizons might be thought to be planar and the succession mistaken for facies MSd. Subfacies 1 is well displayed at locality 4 (Fig. 3.5) in Wadi Teshuinat where it lies both above and beneath beds of facies MSd (Figs. 3.19 & 3.20). The multiple scour horizons may appear to be concordantly filled with channel-fill cross-strata when viewed down the axis of the scoured surface (Figs. 3.21 & 3.22 lower), but the cross-strata lying within the scours are, in fact, composed of angular foresets up to a few centimetres thick which display grainsize grading and thinning-upwards from the toe of the foresets. They form sets of cross-strata which may be up to 175 cm thick and in which reactivation surfaces are present. No clear fining-upwards grainsize grading within the set was observed. A section orientated at 90° to that of the channelled sandstones illustrated, would have the appearance of a vertical coset of thick sets of planar cross-strata, since the axis of the scour is characteristically virtually horizontal and "planar" in two dimensions over distances of up to 20 m. The scours, therefore, have a shallow-channel geometry rather than a scour-and-fill geometry in which the axis of the scour plunges at both ends.

Foreset dip azimuths measured from the same set vary laterally by as much as 84°. The trace of foresets, corresponding to the line of the bar crest, may be followed on horizontal exposures giving evidence that bar crests were up to 40 m in length, and were sinuous. Straight crests of up to 10 m in length were also observed. These horizontal exposures displayed a complex interfingering of foresets, with foresets dying out laterally and interfingering with new foresets. The argillaceous siltstone drapes on the basal scour surfaces of the channel scours are impersistent laterally and, where they pinch out, clay clasts are commonly incorporated in the toeset of the cross-strata.

Compound cross-stratification (Harms, 1975a) is commonly present

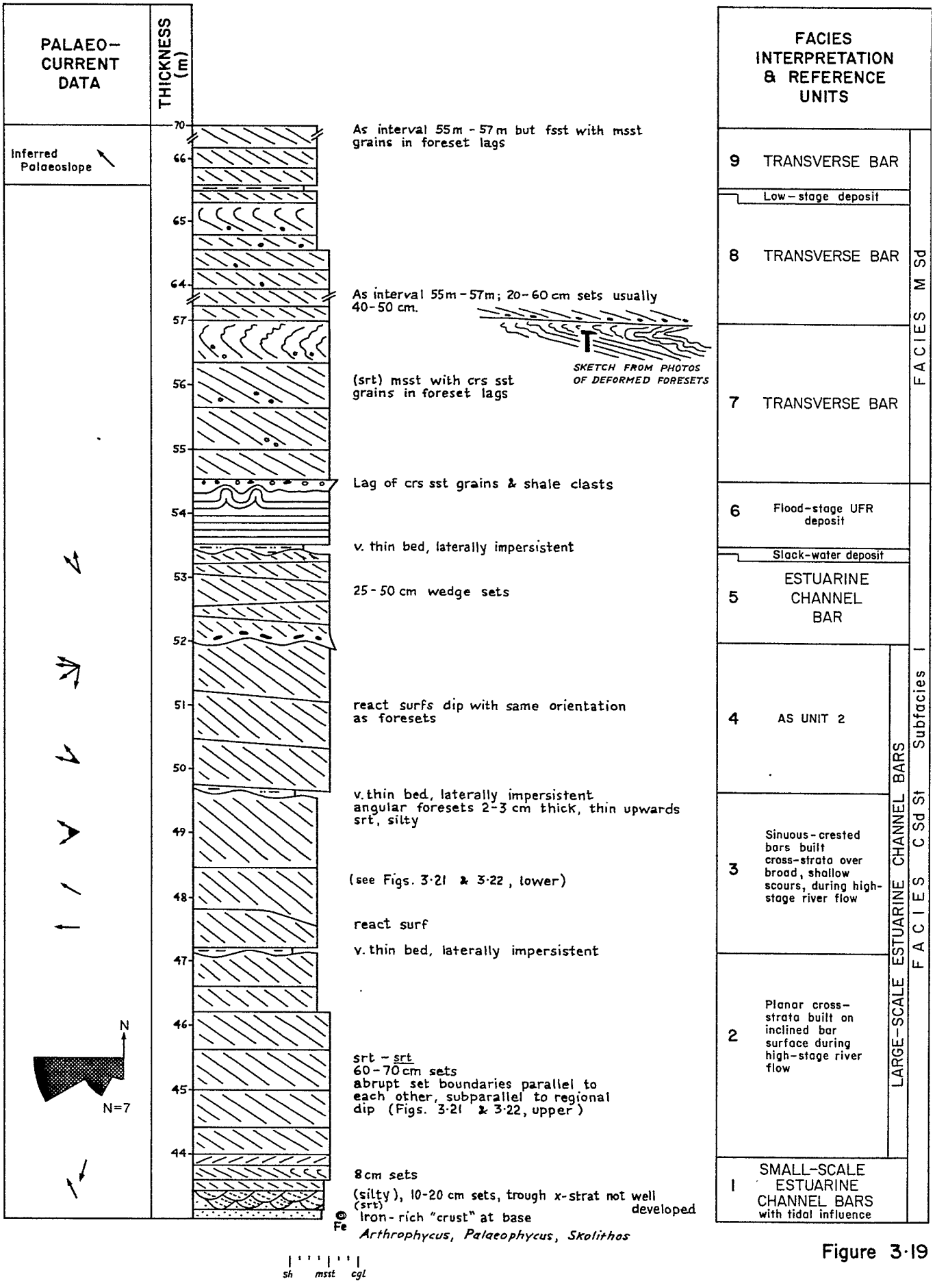

PALAEO-CURRENT DATA
THICKNESS (m)
FACIES INTERPRETATION & REFERENCE UNITS
Inferred Palaeoslope
70
66
65
64
57
56
55
54
53
52
51
50
49
48
47
46
45
44
As interval 55m - 57m but fsst with msst grains in foreset lags
As interval 55m - 57m; 20-60 cm sets usually 40-50 cm.
SKETCH FROM PHOTOS OF DEFORMED FORESETS
(srt) msst with crs sst grains in foreset lags
Lag of crs sst grains & shale clasts
v. thin bed, laterally impersistent
25 - 50 cm wedge sets
react surfs dip with same orientation as foresets
v.thin bed, laterally impersistent angular foresets 2-3 cm thick, thin upwards srt, silty
(see Figs. 3·21 & 3·22, lower)
react surf
v. thin bed, laterally impersistent
srt - srt
60 - 70 cm sets
abrupt set boundaries parallel to each other, subparallel to regional dip (Figs. 3·21 & 3·22, upper)
8cm sets
(silty), 10-20 cm sets, trough x-strat not well developed
(srt)
Iron-rich "crust" at base
Fe
Arthrophycus, Palaeophycus, Skolithos
sh msst cgl
N
N=7
9 TRANSVERSE BAR
Low-stage deposit
8 TRANSVERSE BAR
7 TRANSVERSE BAR
FACIES M Sd
6 Flood-stage UFR deposit
Slack-water deposit
5 ESTUARINE CHANNEL BAR
4 AS UNIT 2
3 Sinuous-crested bars built cross-strata over broad, shallow scours, during high-stage river flow
2 Planar cross-strata built on inclined bar surface during high-stage river flow
LARGE-SCALE ESTUARINE CHANNEL BARS
FACIES C Sd St Subfacies I
1 SMALL-SCALE ESTUARINE CHANNEL BARS with tidal influence

Figure 3·19

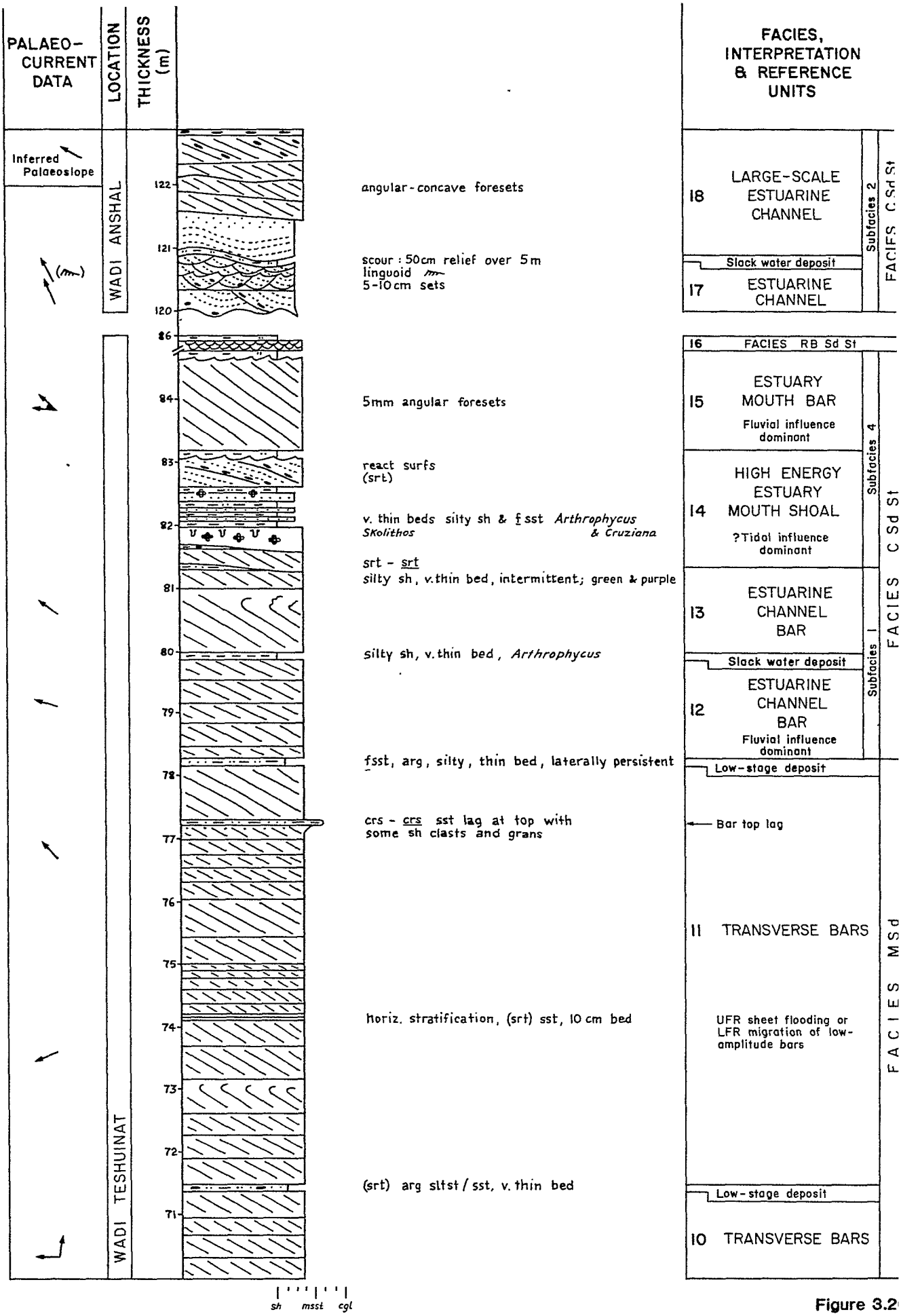

PALAEO-CURRENT DATA
LOCATION
THICKNESS (m)
FACIES, INTERPRETATION & REFERENCE UNITS
Inferred Palaeoslope
WADI ANSHAL
WADI TESHUINAT
122
121
120
86
84
83
82
81
80
79
78
77
76
75
74
73
72
71
angular-concave foresets
scour : 50cm relief over 5m
lingoid
5-10cm sets
5mm angular foresets
react surfs
(srt)
v. thin beds silty sh & f sst Arthrophycus
Skolithos & Cruziana
srt - srt
silty sh, v.thin bed, intermittent; green & purple
silty sh, v.thin bed, Arthrophycus
fsst, arg, silty, thin bed, laterally persistent
crs - crs sst lag at top with some sh clasts and grans
horiz. stratification, (srt) sst, 10 cm bed
(srt) arg sltst / sst, v. thin bed
18 LARGE-SCALE ESTUARINE CHANNEL
Slack water deposit
17 ESTUARINE CHANNEL
Subfacies 2
FACIES C Sd St
16 FACIES RB Sd St
15 ESTUARY MOUTH BAR
Fluvial influence dominant
14 HIGH ENERGY ESTUARY MOUTH SHOAL
?Tidal influence dominant
Subfacies 4
13 ESTUARINE CHANNEL BAR
Slack water deposit
12 ESTUARINE CHANNEL BAR
Fluvial influence dominant
Subfacies 1
FACIES C Sd St
Low-stage deposit
Bar top lag
11 TRANSVERSE BARS
UFR sheet flooding or LFR migration of low-amplitude bars
Low-stage deposit
10 TRANSVERSE BARS
FACIES MSd
sh
msst
cgl

Figure 3.2

in this subfacies (Figs. 3.21, 3.22 & 3.23) comprising "down-current" dipping sets of cross-strata at a variety of scales. The foresets are commonly grainsize graded and inclined at angles of greater than 20^{o} whereas the dipping sets are generally inclined at between 5^{o} and 10^{o}. Inclined sets vary in thickness from a few centimetres up to 1 m but sets of c. 15 cm and c. 70 cm thickness are characteristic of the facies. Variance between foreset dip azimuth and the dip azimuth of the basal boundary of the inclined set may be quite considerable or negligible corresponding to the two types figured by Harms (1975a Figs. 3.5 & 3.6 respectively). Soft-sediment deformation and recumbent-folded foresets are commonly present in the inclined sets of cross-strata. In one example a wedge-shaped set of recumbent-folded deformed cross-strata was observed lying above the uppermost inclined set of a set of compound cross-strata and thickening "down-current" so that its upper surface was approximately flat (i.e. corresponded to the regional dip). The sandstones of the compound cross-strata (including beds with multiple reactivation surfaces) are generally poorly to moderately-sorted but individual foresets may comprise well-sorted sandstones which differ in grainsize from adjacent foresets.

Reactivation surfaces, including convex-upwards surfaces, are present within the sandstone successions lying above the basal scour surfaces (Figs. 3.24 & 3.26), and their presence creates a variety of compound cross-stratification. These compound cross-strata are found within successions of planar cross-strata which occasionally pass upwards into finer-grained bioturbated sandstones (Fig. 3.26, description lower photo-montage). Large-wavelength ripple bedding may also be present at the top of successions of planar cross-strata in this facies (Fig. 3.24). The ripple bedding is developed in well-sorted sandstone which contrasts with the underlying poorly-sorted sandstone.

This subfacies is generally characterised by moderately to poorly-sorted, cross-stratified sandstones with foreset dip azimuth variance of up to c. 150^{o}. In some cases, however, herringbone cross-stratification is clearly developed (Figs. 3.25 & 3.26, see also Fig.3.37) with bipolar palaeocurrents. Beds of planar tabular cross-strata are present in this subfacies which pass upwards into a thin succession of rippled sandstone and siltstone beds from which rippled surfaces also display bipolar palaeocurrents.

Lags of coarse and very coarse-grade quartz grains and

Figure 3.21 COMPOUND CROSS-STRATA AND BROAD SHALLOW SCOURS AT WADI TESHUINAT.

Photo-montage of features of subfacies 1, facies CSdSt (Fig. 3.19); see Figure 3.22 for overlay diagrams.

Figure 3.22 OVERLAY DIAGRAMS OF COMPOUND CROSS-STRATA AND BROAD SHALLOW SCOURS AT WADI TESHUINAT

Upper : compound cross-stratified, fine-grained and well-sorted sandstone in inclined sets of c. 70 m thickness which are composed of foresets inclined at angles of greater than 20°. A large variance exists between the dip azimuths of these foresets and the dip azimuths of the set boundaries.

Lower : broad, shallow scours with overlying fine to medium-grained sandstone (fsu-msl) in sets which vary in thickness from 80-175 cm. Some distortion is present in this overlay diagram as a result of the three-dimensional nature of the exposure.

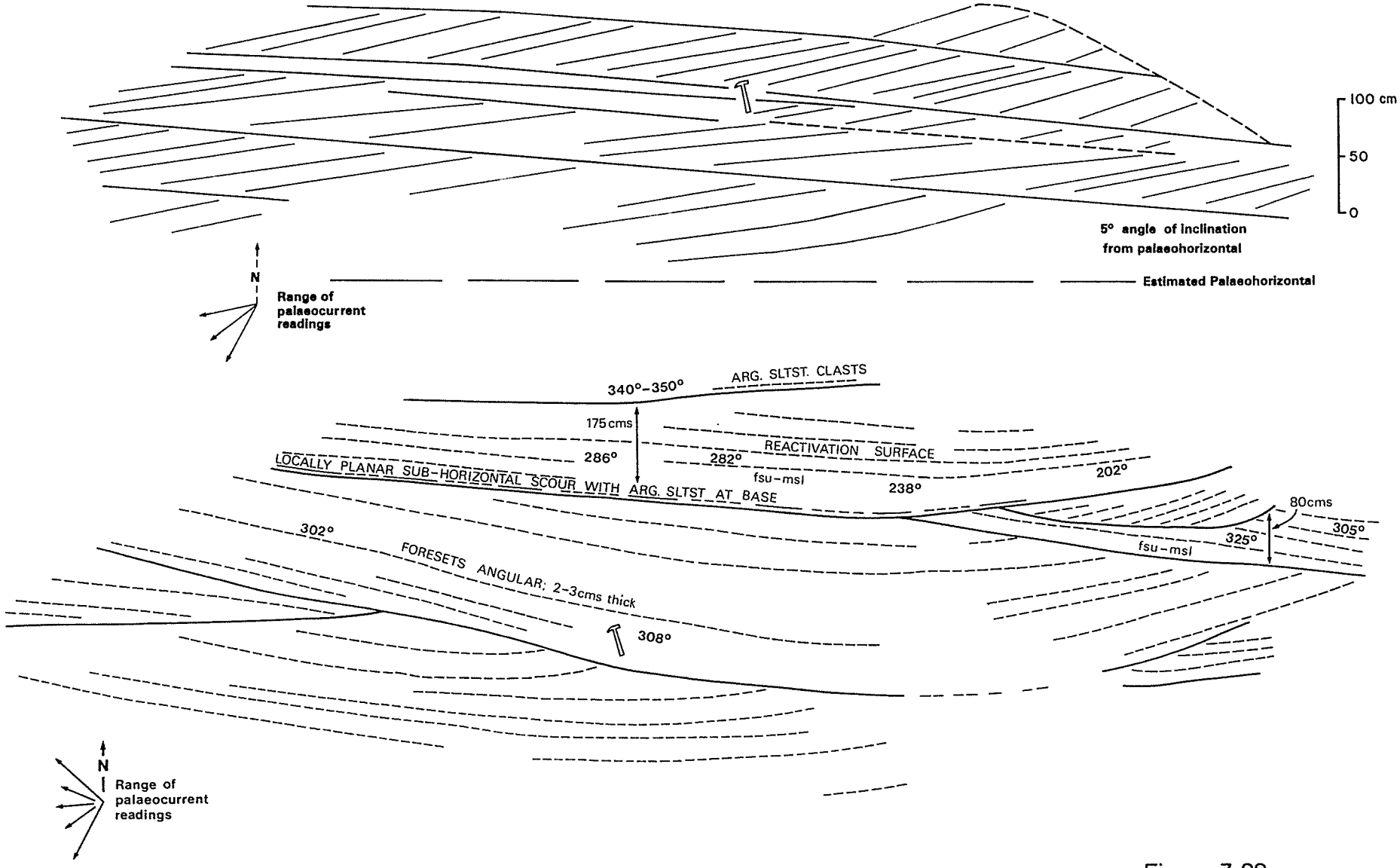
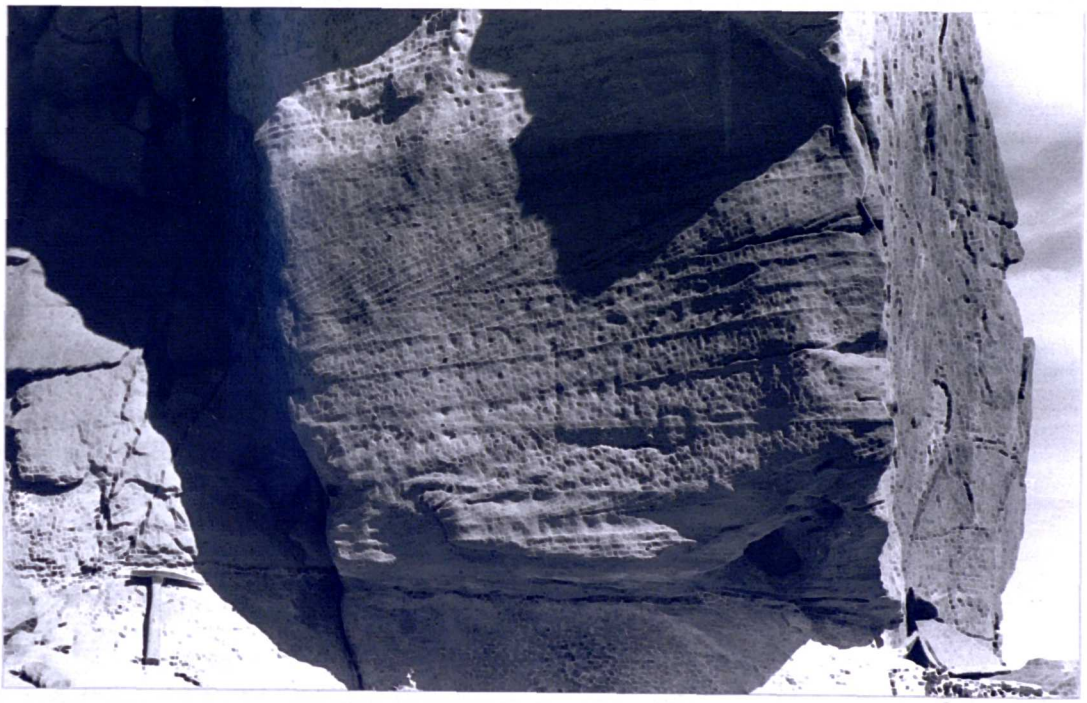

100 cm
50
0
5° angle of inclination
from palaeohorizontal
N
Range of
palaeocurrent
readings
Estimated Palaeohorizontal
340°-350°
ARG. SLTST. CLASTS
175 cms
286°
REACTIVATION SURFACE
282°
fsu-msl
238°
202°
LOCALLY PLANAR SUB-HORIZONTAL SCOUR WITH ARG. SLTST AT BASE
80cms
305°
325°
fsu-msl
302°
FORESETS ANGULAR; 2-3cms thick
308°
N
Range of
palaeocurrent
readings

Figure 3·22

Compound sets of cross-strata, 15-25 cm thick, in fine-grained sandstone at Wadi Teshuinat. These compound cross-strata are characterised by a low variance between foreset and set dip azimuths. The hammer (30 cm long) provides a scale.



Compound set of cross-strata in fine-grained sandstone at Wadi Anshal. The majority of the sets are 4-15 cm thick (typically 6-10 cm thick) i.e. considerably smaller than those of the compound sets illustrated in the upper photograph. These compound cross-strata are characterised by a low-moderate variance between foreset and set dip azimuth. In this example, coarse-grade sand grains are present in grainsize graded foresets. The hammer (bottom left) provides a scale.

Figure 3.23 COMPOUND CROSS-STRATIFIED SANDSTONE AT WADI TESHUINAT AND WADI ANSHAL

Figure 3.24 LARGE-WAVELENGTH RIPPLE-MODIFICATION OF PLANAR CROSS-STRATA, AND CONVEX-UPWARDS REACTIVATION SURFACES

Upper : large-wavelength (60 cm) rippled, well-sorted, medium-grained sandstone, at the top of a succession of planar cross-stratified sets of poorly-sorted, medium- to coarse-grained sandstone at Bir Talauaut. This succession is characterised by recumbent-folded, deformed cross-strata and sets of exceptionally, up to 1 m thickness but characteristically 10-15 cm thickness. The dip azimuths of the foresets of these cross-strata display a variance of 135°. The penknife (8.5 cm long) provides a scale, and a felt-tip pen was used to emphasize the large-wavelength ripple structure. The ripple modification structure is thought to provide evidence of falling flow-stage conditions.

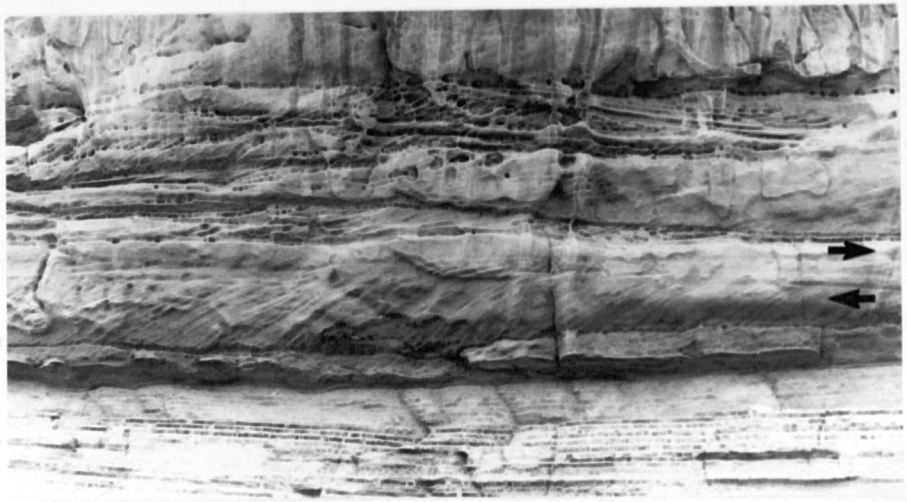
Lower : a variety of compound cross-stratification, present at locality 3, Wadi Teshuinat (Fig. 3.5), in which the inclined set boundaries are also convex-upwards reactivation surfaces. The set in which these reactivation surfaces are present (identified by the position of the hammer) has an irregular basal boundary and angular foresets that are inclined, as in the overlying set, at angles of repose of 25° to 27°. A 5 cm thick, homogeneous bed divides the two sets. See Figure 3.18 (upper) for a proposed process of formation of this structure.

Figure 3.25 EXAMPLES OF HERRINGBONE CROSS-STRATIFICATION

Upper left : sandstones of facies CSdSt lying above sandstones and argillaceous siltstones of facies RBSdSt (lighter coloured beds) at the base of transgressive sequence 3, locality 3, Wadi Teshuinat (Fig. 3.5). Arrows indicate opposite palaeocurrent directions measured from the foresets of two adjacent sets of cross-strata, the lower of which is c. 1 m thick. This is a close-up of part of the exposure illustrated in Figure 3.62 (lower picture). See Figure 3.26 for overlay diagram.

Upper right : planar set of cross-stratified, medium-grained sandstone (identified by the position of the hammer) that has a foreset dip azimuth (190°) with a variance of 125° from the sets above and below; from the same locality as that of the upper left photograph.

Lower : part of a 40 m thick unit of medium-grained, cross-stratified sandstones of facies CSdSt lying entirely within sediments of facies RBSdSt in the upper part of transgressive sequence 4 (214-254 m stratigraphically above the base of the Tadrart Formation) at Bir Talauaut. The right hand arrow marks the position of herringbone cross-stratification lying above a scoured basal boundary. An extensive, low-angle truncation plane is marked by the left-hand arrow. Base to top of photograph is c. 30 m. See Figure 3.26 for overlay diagram and palaeocurrent rose diagram.



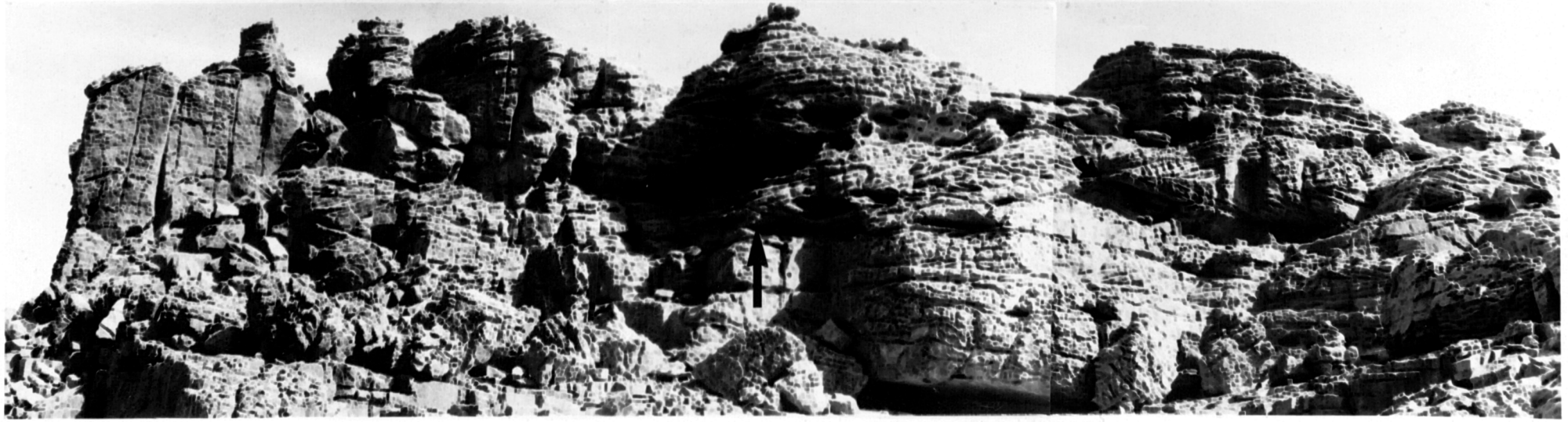
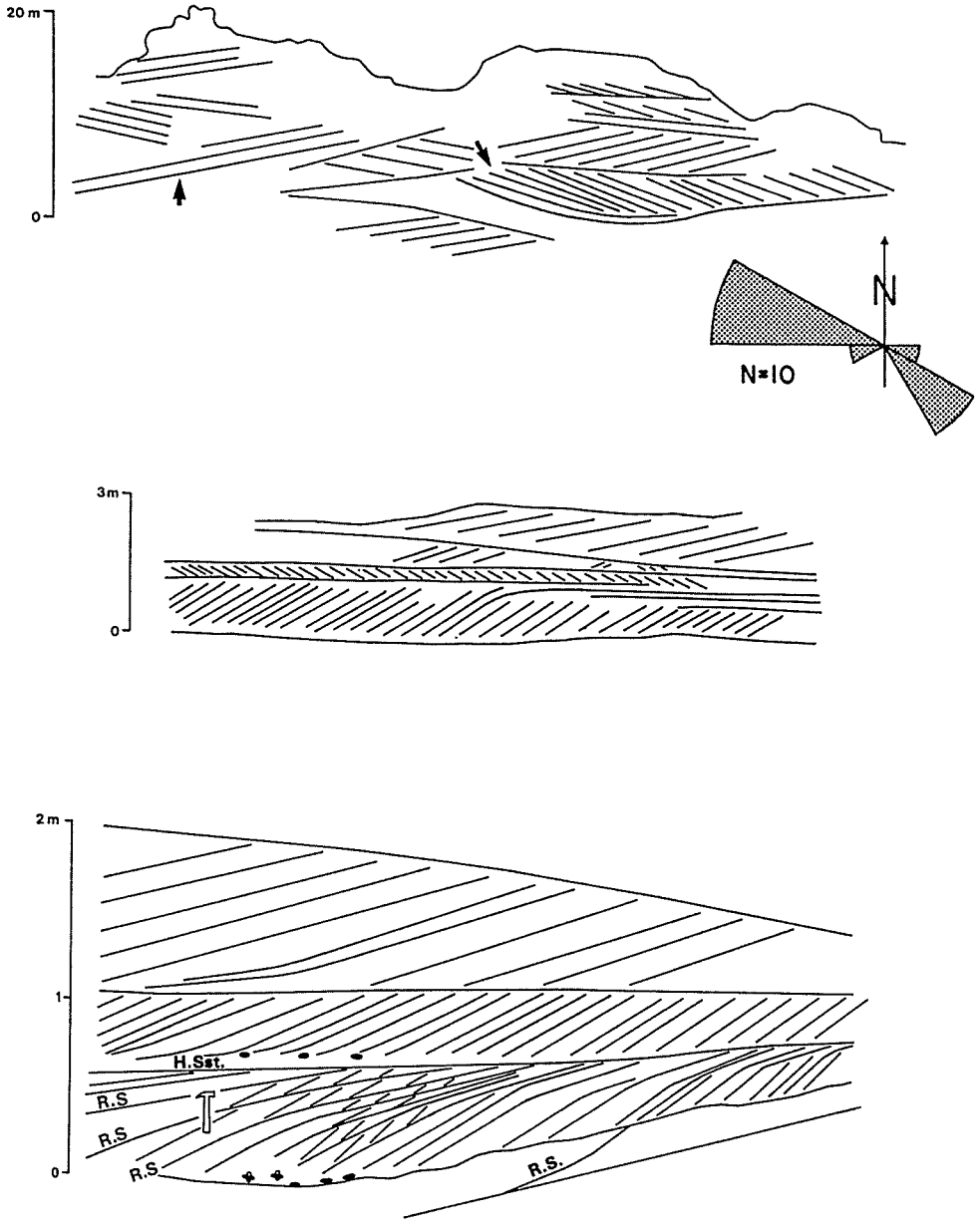


Figure 3.26 OVERLAY DIAGRAMS ON HERRINGBONE CROSS-STRATIFICATION AND CONVEX-UPWARDS REACTIVATION SURFACES

Upper : from photo-montage (Fig. 3.25, lower) to illustrate herringbone cross-stratification above scoured basal boundary (RHS), and large-scale, low-angle truncation plane with strata lying concordantly above (LHS). The bipolar palaeocurrent rose diagram is constructed from foreset dip azimuths measured from the succession illustrated.

Middle : from photograph (Fig. 3.25, upper left) to show herringbone cross-strata. The lowest set illustrated has some impersistent top-set stratification.

Lower : from photo-montage (Fig. 3.24, lower) to show convex-upwards reactivation surfaces with angular foresets lying above these surfaces. Bioturbation structures and shale clasts are present just above the irregular basal boundary of the set. The compound set of cross-strata is overlain by beds of planar cross-strata which pass up (above the succession illustrated here and in Figure 3.24, lower) into bioturbated sandstones with abundant shale intra-formational clasts.



H.Sst Homogeneous sandstone bed
R.S. Reactivation surface

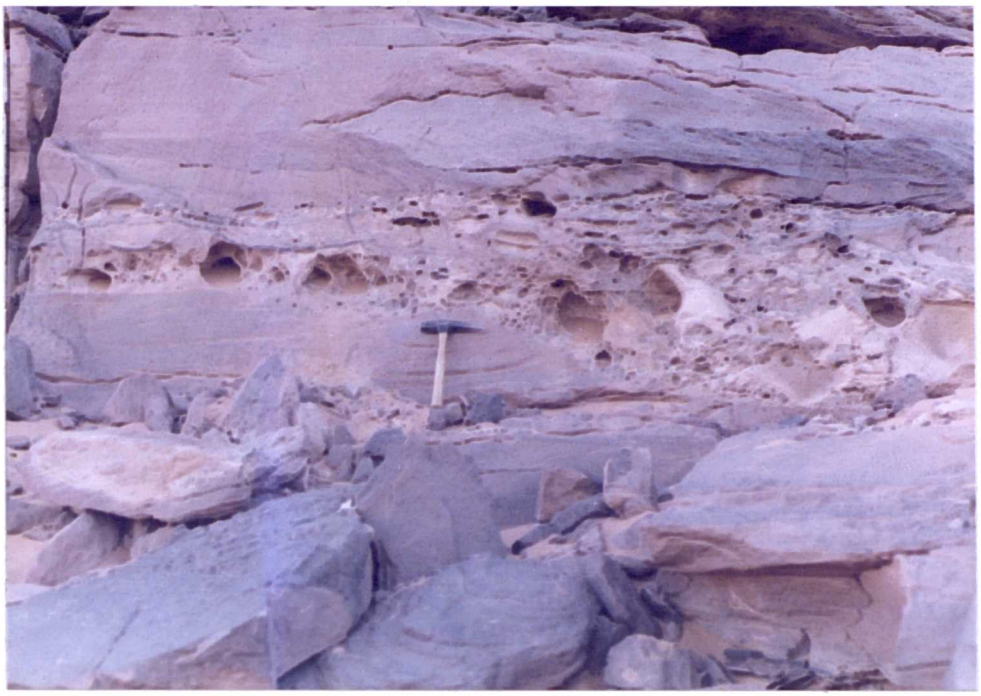
Figure 3·26

argillaceous siltstone intraformational clasts, are sometimes present on the top of beds of horizontal strata and cross-strata (Figs. 3.19, unit 6). Biogenic traces are present in some of the shale and argillaceous siltstone beds, *Arthrophycus* being the only trace fossil identified in these beds (Fig. 3.20, unit 12), whilst *Skolithos* is present in some of the sandstones.

3.3.1.3 Subfacies 2

Bedding sequences comprising trough cross-strata and channel fill cross-strata which pass upwards into planar cross-strata and which lie above scoured, erosive basal boundaries have been assigned to subfacies 2 (Fig. 3.20, units 17 & 18). The basal boundary may be channel-shaped and up to 12 m wide, as in subfacies 1, but it is more commonly irregular or nearly planar. It is generally overlain by a thin argillaceous siltstone bed. The bedding sequences are commonly not complete, being truncated by younger bedding sequences higher in the succession. This subfacies although similar in a number of respects to subfacies 1 of this facies, and to facies MSd, is distinct from both of them by its variety of stratification types and absence of multi-storey cosets of planar cross-strata. As well as the stratification types already mentioned, multiple scour and fill cross-stratification and homogeneous beds are commonly present, the latter generally at the base of the bedding sequence. Current-ripple lamination is also present but lateral-accretion cross-stratification is absent. The bedding sequences, when complete, are c. 2-3 m thick, and show no regular fining-upwards grainsize trend. Where planar cross-stratified sandstone lies above the basal scoured boundary, the foresets are generally inclined at an angle of greater than 25^{o}, and in some cases, greater than 30^{o}. These beds of planar cross-strata are characterised by very thick sets (Fig. 3.27, lower) which are generally isolate and not stacked in a coset of cross-strata. Intraformational clay clasts are commonly present in the cross-strata of this subfacies, and some irregular, wavy, low-angle cross-strata present high up in bedding sequences are thought to indicate the initiation of soft-sediment deformation. An intraformational conglomerate is sometimes present in the basal beds of the channel bedding sequences (Fig. 3.28).

The palaeocurrent variance, as measured from linguoid rippled sandstone bedding planes and foreset dip azimuths, appears generally



Intraformation conglomerate lying 75 m stratigraphically above the base of the Tadrart Formation at Wadi Anshal (Fig. 3.5). The conglomerate, which is monomict and matrix-supported, in places, and clast-supported elsewhere, lies above an erosional basal boundary and is composed of white, fine-grained sandstone clasts and poorly-sorted sandstone matrix. These conglomerates are thought to result from channel collapse events during lateral channel migration.

Figure 3.28 INTRAFORMATIONAL CONGLOMERATE.

to be fairly low although examples of bipolar palaeocurrents were observed from some rippled surfaces. Only a limited amount of palaeocurrent data was available from this subfacies, however.

3.3.1.4 Subfacies 3

Bedding sequences indicative of channel-fill successions but distinguished from those of subfacies 2 by a significantly greater amount of biogenic structures, and by the presence of shell moulds and casts, have been assigned to subfacies 3 (Fig. 3.29). Soft-sediment deformation and ripple lamination are also more abundant in this subfacies. The subfacies is characterised by fine and very fine-grained, moderately to well-sorted sandstones which are present in bedding sequences of planar, trough and channel-fill cross-strata which pass upwards to cross-laminated and bioturbated sandstones. The bedding sequences generally display a poor fining-upwards grain-size gradation although the stacking of channel sandstone bodies commonly obscures this trend.

Near the top of some of the channel sandstone bodies moulds and casts of shell fragments were observed. Similar evidence for dislocated shells and fragments of shells were found in the fine-grained sandstone beds of facies RBSdSt. In many cases the impression left in the sandstone bed was virtually indistinguishable from that left by a clay clast, but in some cases the shape of the umbo was recognisable. The shape and size of these impressions is consistent with their being assigned to *Nucula* cfr. *corbuliformis* Hall. Borghi and Chiesa (1940) observed very similar shell impressions in the Ouan Kasa Formation and assigned them to the same species, an infaunal vagrant bivalve. Evidence for these bivalves had not been observed before in the massive sandstone beds of the Tadrart Formation. Fossil plant fragments are also present in these channel sandstone bodies observed in one case on the upper bedding plane of a medium-grained, trough cross-stratified sandstone bed (Fig. 3.42, upper left).

The trace fossils *Skolithos* and *Rosselia* (see Figs. 2.7 & 2.9) which are present in the sandstones of this subfacies (Figs. 3.29, 3.30, 3.32), provide evidence that suspension-feeding organisms built dwelling tubes in the channels and, also, that organisms fed on the nutrient present within the channel sands. These burrows are present in the upper, finer-grained beds of the bedding sequences (Fig. 3.30).

Figure 3.29 MEASURED SECTION IN FACIES CSdSt, RBSdSt AND VBSd AT TIHERERIN

Notes:

1) The section was measured 224 to 251 m stratigraphically above the base of the Tadrart Formation at locality 17 (Fig. 3.5) in the basal 27 m of transgressive sequence 4 ("Sandstone Marker Bluff").

2) The palaeocurrent readings were measured from trough axes or foreset dip azimuths of cross-strata, except where a ripple sign is shown.

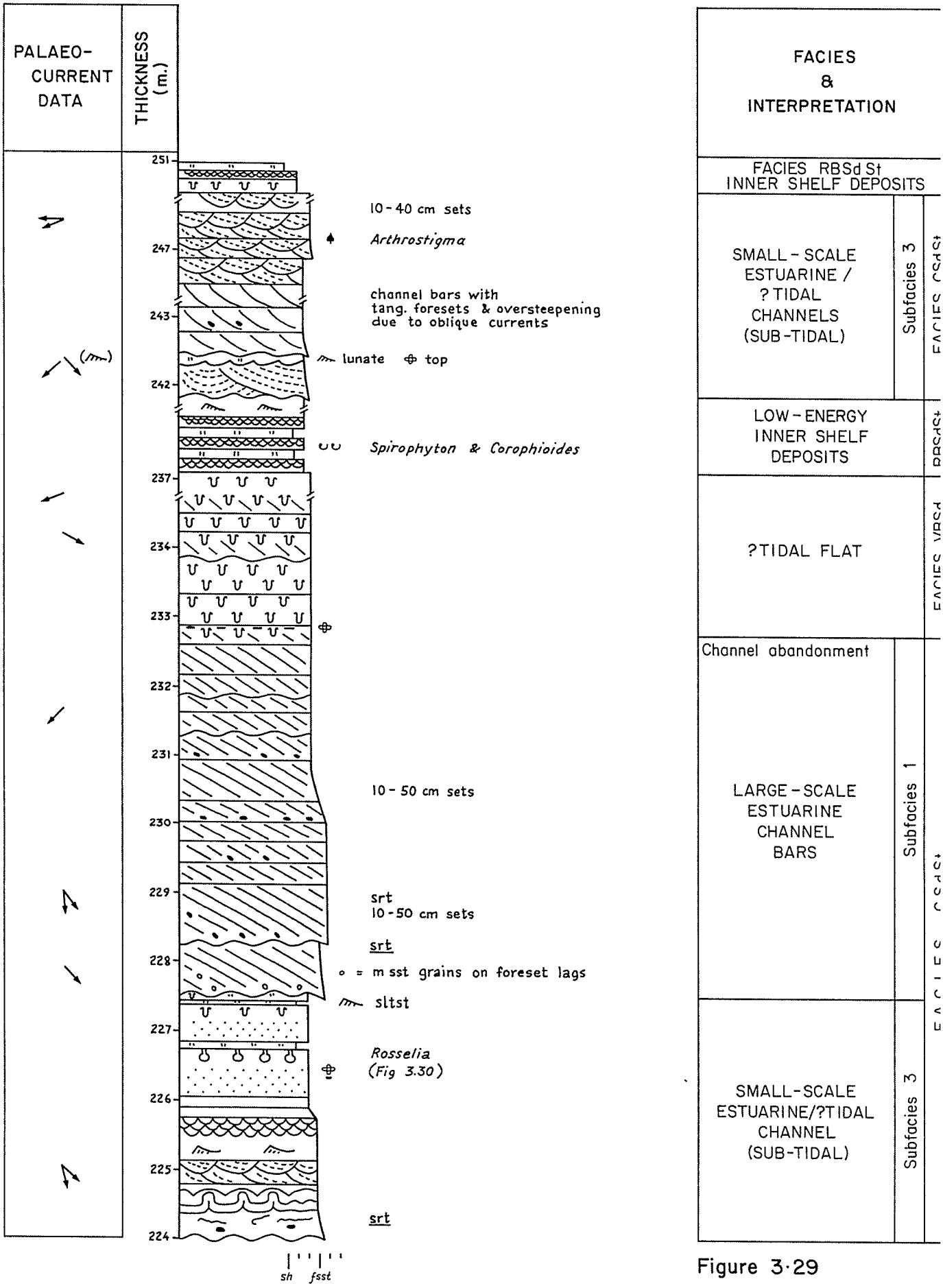
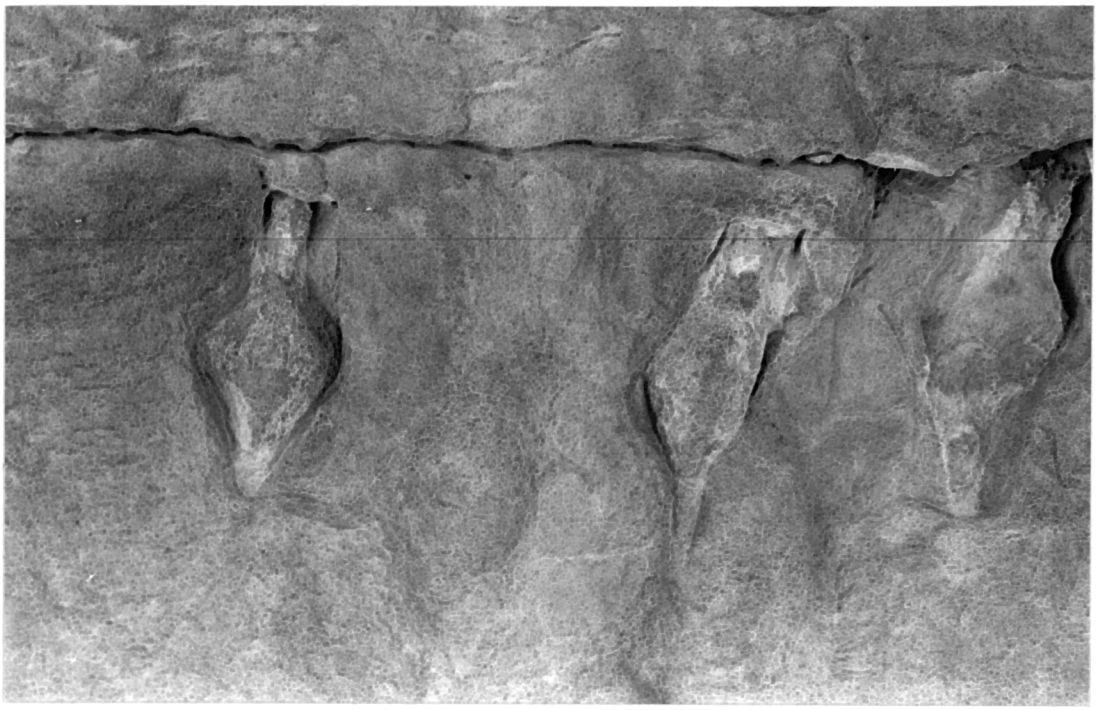

PALAEO-CURRENT DATA
THICKNESS (m.)
FACIES & INTERPRETATION
FACIES RBSd St
INNER SHELF DEPOSITS
SMALL-SCALE ESTUARINE / ?TIDAL CHANNELS (SUB-TIDAL)
Subfacies 3
LOW-ENERGY INNER SHELF DEPOSITS
?TIDAL FLAT
Channel abandonment
LARGE-SCALE ESTUARINE CHANNEL BARS
Subfacies 1
SMALL-SCALE ESTUARINE/?TIDAL CHANNEL (SUB-TIDAL)
Subfacies 3
251
247
243
242
237
234
233
232
231
230
229
228
227
226
225
224
10-40 cm sets
Arthrostigma
channel bars with tang. foresets & oversteepening due to oblique currents
lunate
top
Spirophyton & Corophioides
10-50 cm sets
srt
10-50 cm sets
srt
o = m sst grains on foreset lags
sltst
Rosselia (Fig 3.30)
srt
sh
fsst

Figure 3·29



Rosselia trace fossil burrows in well-sorted, fine-grained sandstone at Tihererin (cf. measured section of Fig. 3.29). In this case the burrows penetrate 8 cm below the upper bedding plane of the bed (shown at the top of the photograph), the surface of which is strongly bioturbated and overlain by a 5 mm thick bed of horizontally laminated siltstone.

Rosellia burrows lying 25-30 cm below the top of the same sandstone bed as that illustrated above. A zone of strongly bioturbated sediment lies above the semi-spherical *Rosselia* burrows.

Figure 3.30 *ROSSELIA* IN SANDSTONES OF SUBFACIES 3, FACIES CSdSt

Just as with subfacies 2, the bedding sequences of subfacies 3 lie above both curved and approximately planar basal scoured boundaries (Fig. 3.31, upper). The subfacies is, however, also characterised by larger-scale major scour horizons of 15-40 m width and of 1-5 m depth (Figs. 3.31, lower & 3.32 upper). These scours and the sandstone succession lying above them, may be present within facies CSdSt and in association with facies VBSd (Fig. 3.32, upper), or they may exist as a large-scale lens of sandstone lying totally within facies RBSdSt (Fig. 3.32, lower). The latter channelled sandstone body shows good evidence for lateral accretion but this is not a common feature of this facies. In some of the channel sandstone bodies lying within facies RBSdSt clay intraformational clasts and shell fragments were found on foreset surfaces, some of which also displayed foreset ripple cross-lamination.

The palaeocurrent data from subfacies 3 display a large variance and are dominated by southwest, south and southeast directions (Figs. 3.29 & 3.32). These orientations are in marked contrast to the prevalent northwesterly foreset azimuth direction obtained from the majority of the cross-strata of facies MSd and CSdSt. However, the beds illustrated in Figures 3.29 and 3.32 are from transgressive sequence 4, the characteristics of which suggest a different palaeogeography. A discussion of the evidence provided by the sediments of transgressive sequence 4 for their palaeogeographic setting is presented in section 3.7 (this Chapter). Although bipolar palaeocurrent patterns were not observed, the range of current readings suggest offshore, along-shore and some onshore currents during deposition of the sediments of this subfacies.

3.3.1.5 Subfacies 4

Laterally impersistent sandstone beds with thin shale and micaceous argillaceous siltstone intercalations have been assigned to subfacies 4 (Fig. 3.20, units 14 & 15). This subfacies is commonly associated with subfacies 1 and 2, and with facies MSd, although it is also present lying above subfacies 3. It differs from the three other subfacies of this facies in not being directly associated with channelling or bedding sequences indicative of channel-fill sandstone bodies. As with all the sediments of this facies, however, it may be associated with large-scale scours and channels. Subfacies 4 commonly

Figure 3.31 CHANNELLED SANDSTONES OF FACIES CSdSt AT THE BASE OF TRANSGRESSIVE SEQUENCE 4

Upper : shallow scours in the basal sandstones of transgressive sequence 4, at Wadi Anshal (Fig. 3.5) ("Sandstone Marker Bluff"). A 2m bar provides scale.

Lower : a major scoured surface overlain by strongly bioturbated sandstones of subfacies 3 in the sandstones of the "Sandstone Marker Bluff", at Wadi Teshuinat (Fig. 3.5). A hammer (bottom right) provides scale. An overlay diagram of this bedding sequence is illustrated in Figure 3.32.

2m

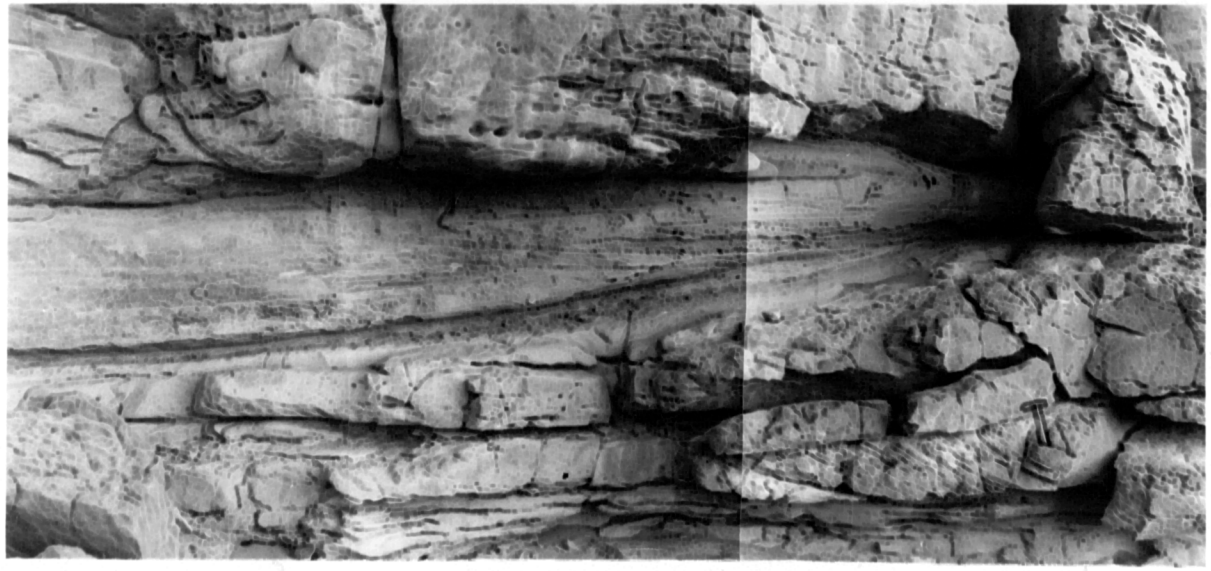


Figure 3.32 CHANNEL-FILL BEDDING SEQUENCES OF SUBFACIES 3 AT WADI TESHUINAT

Upper : strongly-bioturbated, fine-grained sandstones of subfacies 3, from transgressive sequence 4, locality 2 (Fig. 3.5). Horizontal bedding and cross-stratification may be discerned within these sandstones, which lie above a major scour surface that truncates sandstones of subfacies 1 and 2. A channel-sandstone bedding sequence of subfacies 3 lies above and passes upwards into vertically-burrowed sandstones of facies VBSd. The palaeocurrent indicators of the channel sandstones of subfacies 2 which lie above the vertically-burrowed beds indicate that palaeoflows were predominantly towards the south. This section is illustrated in Figure 3.31 (lower).

Lower : channel sandstone body that displays lateral accretion cross-stratification and soft-sediment deformation, and is overlain and underlain by horizontally stratified, bioturbated sandstones of facies RBSdSt, from transgressive sequence 4, locality 2 (Fig. 3.5). The channel is inferred to have migrated towards the right (west) whilst the foreset dip azimuths indicate predominantly southerly-directed palaeocurrents, as in the channelled sandstones illustrated above. Note that three-dimensional distortion in the photographs from which this diagram was taken allows the false impression to be given that the upper beds of facies RBSdSt are not horizontal.

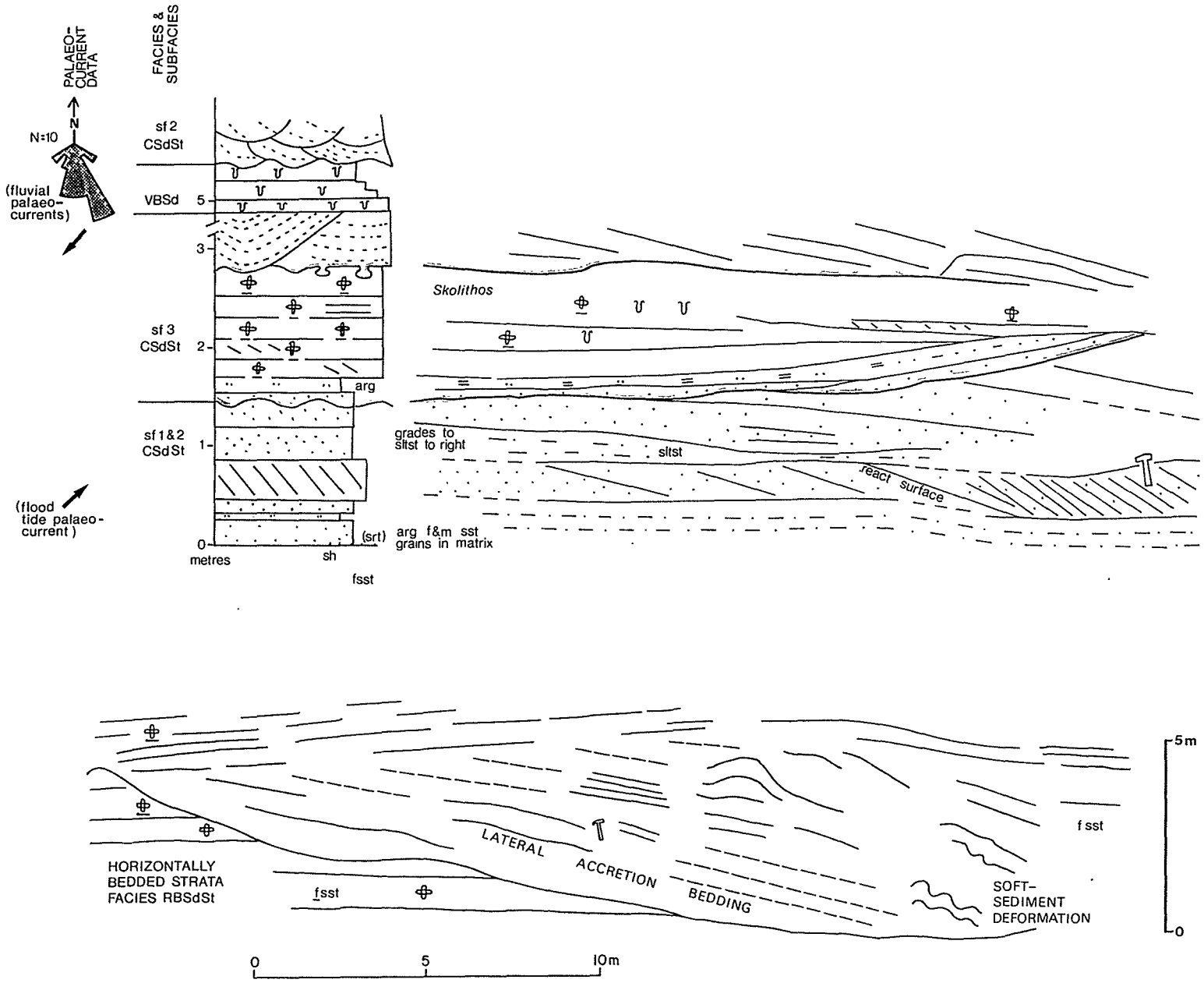

PALAEO-CURRENT DATA
FACIES & SUBFACIES
N=10
N
(fluvial palaeo-currents)
(flood tide palaeo-current)
sf 2 CSdSt
VBSd
sf 3 CSdSt
sf 1&2 CSdSt
arg
(srt)
metres
sh
fsst
Skolithos
grades to sltst to right
sltst
react surface
arg f&m sst grains in matrix
HORIZONTALLY BEDDED STRATA FACIES RBSdSt
fsst
LATERAL ACCRETION BEDDING
f sst
SOFT-SEDIMENT DEFORMATION
5m
0
0
5
10m

Figure 3·32

lies at the top of the thickly-bedded units of the transgressive sequences and is less laterally persistent than the other subfacies. In two cases where the subfacies is well developed, both within the thickly-bedded unit of transgressive sequence 2 in Wadi Teshuinat (Figs. 3.33 & 3.36), the sediments of this subfacies interfinger laterally with adjacent sediments over distances of 1-1.5 km. In one case they interfinger with sediments of facies MSd (Fig. 3.33) and in the other case with sediments of subfacies 1 of facies CSdSt (Fig. 3.36).

The sandstone beds of this subfacies are generally 10-50 cm thick and are composed of very fine to medium-grained, poorly to moderately-sorted sandstones which are generally planar cross-stratified (Figs. 3.33 & 3.35, lower). Grainsize grading and soft-sediment deformation of foresets is common, and shale intraformational clasts are generally present within both the shale beds and the cross-stratified sandstones, generally at the base of sets. Reactivation surfaces and compound cross-strata may be present in the sandstone beds which may also display the following bedding types: horizontal stratification, homogeneous bedding, trough cross-stratification and planar sets of cross-strata with concave foresets.

The sandstone beds thicken and thin laterally (Figs. 3.34, lower, & 3.36, lower). This irregular geometry appears to be of primary origin in many cases, since bioturbation and rippling of the upper surface of these beds suggest preservation, rather than truncation, of the original bed surface. In other cases, however, beds appear to have been truncated prior to deposition of clay and argillaceous silt. The upper surface of some of the sandstone beds is characterised by a long-wavelength (c. 1 m) rippled surface.

Clay drapes lying on the upper surface of foresets and at the base of sets are occasionally present (Figs. 3.27, upper, & 3.34, upper). These are comparable to the structures described by Mowbray and Visser (1982) and Allen (1981).

The fine-grained sediments of this subfacies are generally homogeneous and bioturbated, lying in beds of 5-20 cms thickness. Sandstone and siltstone intercalations and lenses are present within these sediments which display cross-lamination including lenticular and wavy cross-lamination. Bioturbation is common in this subfacies; trails and burrows such as *Fraena* and *Palaeophycus* are often seen on

Figure 3.33 BEDDING SEQUENCES AND SEDIMENTARY STRUCTURES OF SUBFACIES 4

Lower : part of a 7 m thick development of subfacies 4 that lies above an iron-oxide cemented sandstone at the base of transgressive sequence 2, locality 9, Wadi Teshuinat (Fig. 3.5). The sandstone beds, which are generally 15-50 cm thick, are composed of moderately-sorted, fine to medium-grained sandstone which displays planar and compound cross-stratification, and grainsize graded foresets. Foreset dip azimuths are unimodal and northwesterly orientated (i.e. into the plane of the exposure). Recumbent-folded deformed cross-strata are present in sets (generally 5-10 cm thick) of compound cross-strata (see well-developed example above the arrow and also a close-up in the upper left photograph). The top of beds commonly display asymmetrical ripples and are overlain by sandy and silty shale beds of up to 20 cm thickness. These contain abundant clay intraformational clasts, of up to 5 cm in length, and poorly-sorted cross-laminated sandstone intercalations. Bioturbation structures and the straight-burrow variety of *Arthrophycus* (Fig. 2.6) are present in the shale beds. See Figure 3.34 (lower) for overlay diagram.

These beds of subfacies 4 pass by a lateral facies change into sandstones of facies MSd, present at the base of transgressive sequence 2 and illustrated in Figure 3.13. They also pass vertically into sandstones of facies MSd.

Upper left : recumbent-folded deformed cross-strata in compound sets of cross-strata.

Upper right : complex differential compaction structures resulting from major lateral thickening and thinning of shale and sandstone beds.

The hammer provides scale in all three illustrations.

Figure 3.34 OVERLAY DIAGRAMS ON BEDDING SEQUENCES OF SUBFACIES 4.

Upper : argillaceous siltstone and silty shale beds lying above asymmetrically rippled surfaces, and bounding a complex of sandstone beds displaying planar cross-strata with clay drapes on the upper surfaces of some foresets (see Fig. 3.27, upper, for photograph). These structures may provide evidence for tidal conditions although they are not diagnostic.

Lower : lateral thickening and thinning of silty shale and sandstone beds on the base of some of which load structures are present (see Fig. 3.33 for photographs). The geometry of the beds suggest intermittent traction deposition of irregularly-shaped sandstone bodies in a region where suspension deposition and minor ripple-inducing currents were active during periods of limited sediment supply. The truncation of some of the sandstone beds may be attributed to current scouring events prior to the deposition of clay drapes, but other sandstone beds are thought to reflect their original depositional geometry.

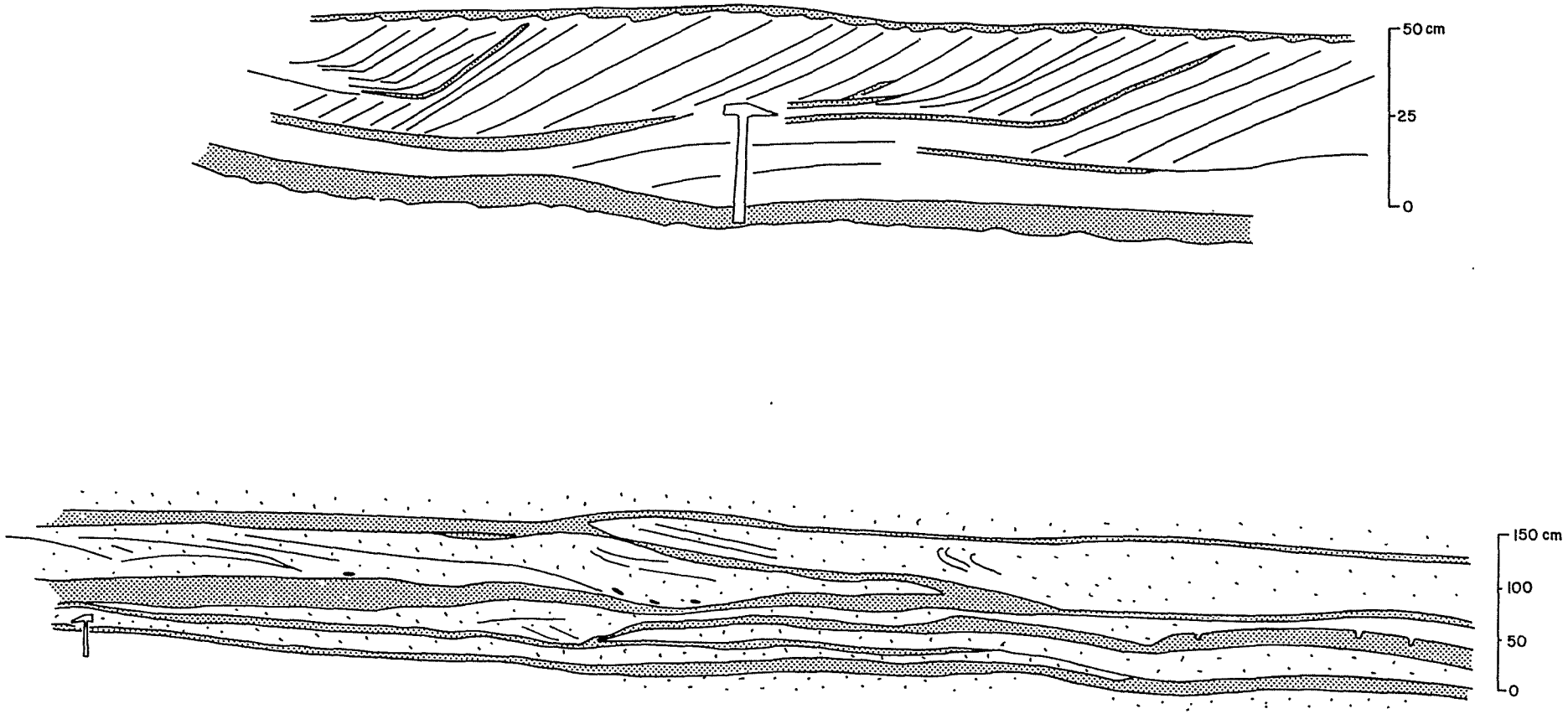
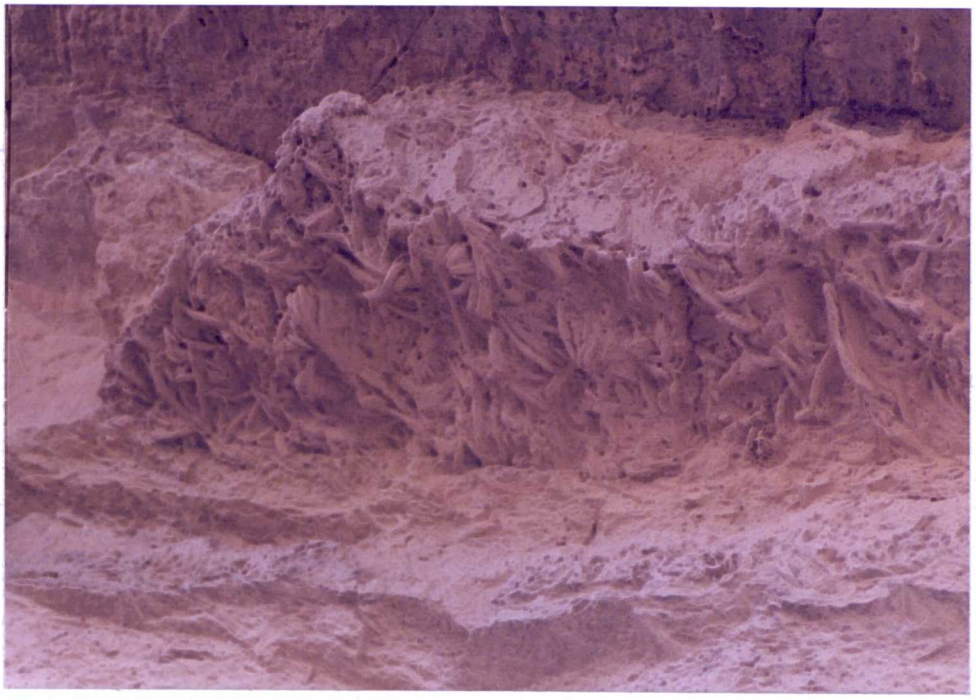


Figure 3·34



Sandstone bed, with abundant *Arthrophycus* on the sole of the bed, lying within a succession of very fine-grained sandstones and siltstones of subfacies 4 at locality 14, Wadi Ahar (Fig. 3.5), near the base of transgressive sequence 1, Acacus Formation (see Fig. 3.37, "Arc de Triomphe").

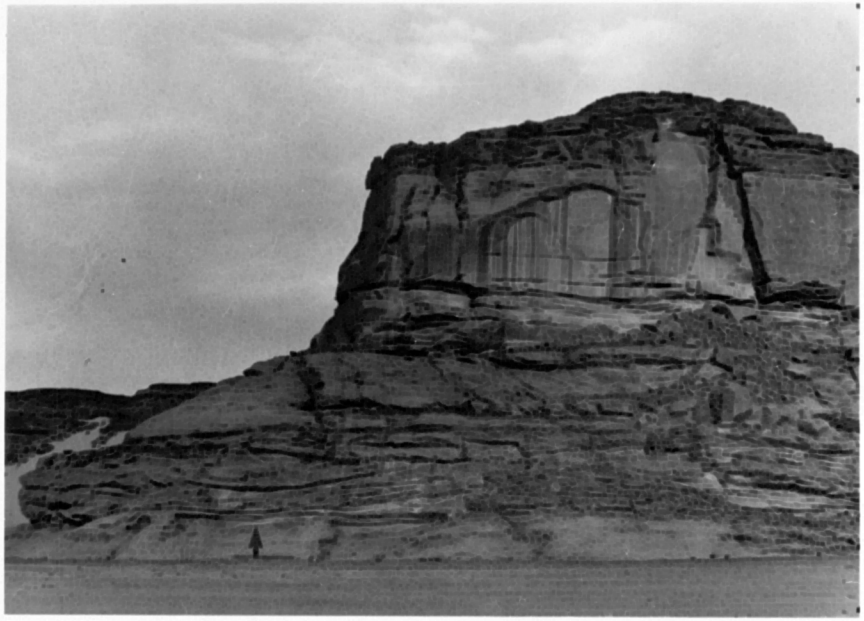
Fine-grained sandstone bed, of 25 cm thickness, displaying planar cross-strata (maximum foreset dip direction is out of the plane of the photograph, towards the left of the photographer), and bounded above and below by shale beds of 2-5 cm thickness. The base of the sandstone bed is loaded into the shale, whilst the top surface is gently rippled. Detail from Figure 3.36, lower.

Figure 3.35 BIOGENIC AND SEDIMENTARY STRUCTURES OF SUBFACIES 4, FACIES CSdSt

Figure 3.36 EXPOSURE OF SANDSTONES OF FACIES CSdSt AND MSd IN ESCARPMENT AT WADI TESHUINAT

Upper : section of irregular scarp forms and slick-rock slopes (above arrow) in a c. 60 m thick exposure of sandstones of facies CSdSt in the upper part of transgressive sequence 2, above the "40 m plateau", at locality 5 (Fig. 3.5). Above this, sandstones of facies MSd and CSdSt of transgressive sequence 3 form a massive slab wall, c. 60 m high, in the upper part of the photograph.

Lower : enlargement of arrowed part of exposure in upper photograph illustrating a major channel 25 m wide and 4 m deep. Below the channel, sandstone beds of subfacies 4, displaying lateral thickness differences are present (see Fig. 3.35, lower).



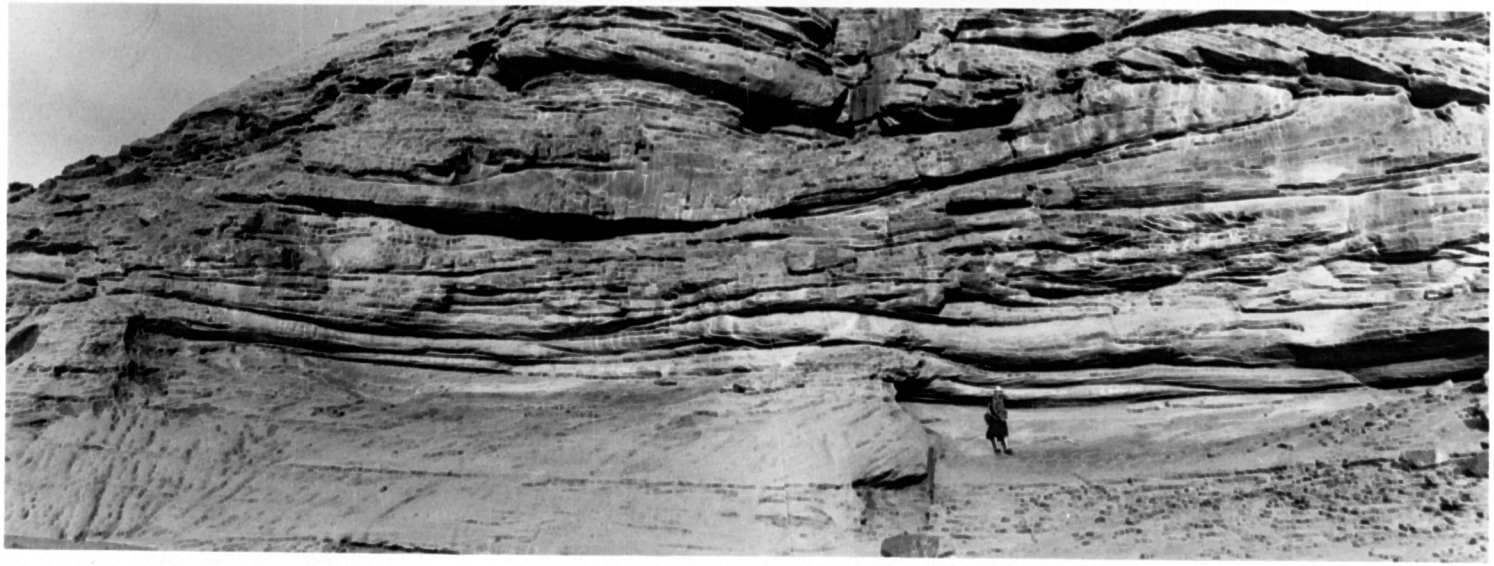


Figure 3.37 NATURAL ROCK ARCH IN SANDSTONES OF FACIES CSdSt AT WADI AHAR

Upper : natural rock arch, or "Arc de Triomphe", sculptured from fine to medium-grained sandstones of unit 14 of the Acacus Formation (see Fig. 3.1) (Klitzsch, 1965a, 1969). The arch, which is more than 100 m high and is located in a region given the name "Bab al Ramlat" ("Pass of Sand" in Arabic) in the Wadi Ahar (locality 14, Fig. 3.5), is composed of cross-stratified and channelled sandstones and subsidiary siltstones, some of which contain abundant *Arthrophycus* (see Fig. 3.35), of facies CSdSt (transgressive sequence 1).

Lower : overlay diagram showing that, even allowing for three-dimensional and perspective effects, many of the breaks in the succession are strikingly subparallel to each other. The palaeocurrent rose shows that foreset dip azimuths measured from cross-strata of subfacies 1 and 2 within this succession display a bipolar distribution. The landrover provides scale.

the rippled top bedding planes of the sandstone beds and *Skolithos* may be present within these beds. *Arthrophycus* is commonly present on the sole of sandstone beds lying above the shale and siltstone intercalations (Fig. 3.35, upper) and *Cruziana* may also be present on these soles. In transgressive sequence 1 (Acacus Formation, cf. Fig. 3.6) the *Arthrophycus* is commonly palm-frond shaped, whereas in the higher sequences (Tadrart Formation) it is usually present in its straight-burrow form (see Fig. 2.6). The *Cruziana* in this facies is similar to that found in facies RBSdSt (see Fig. 2.4, upper central right) and is smaller than the specimens from the Acacus Formation.

From the limited palaeocurrent data available from this subfacies unimodal palaeocurrent roses were plotted with variances of up to 160°.

3.3.1.6 Large-scale scour surfaces and channels

A major characteristic of facies CSdSt is the presence of large-scale scour surfaces and channels. In Wadi Teshuinat a shallow channel 25 m across and 4 m deep is exposed (Fig. 3.36) and, in Wadi Ahar, scours and truncation planes of at least 50 m length are recognisable (Fig. 3.37). A major truncation plane of similar dimensions is present at Bir Talauaut (Figs. 3.25, lower, & 3.26, upper) which has a depth of at least 6 m. The channel in Wadi Teshuinat is lined with a thin siltstone bed and a bioturbated silty shale bed (both 2-5 cm thick). It has similar characteristics, therefore, to the channel described in subfacies 3 (Figs. 3.31, lower, & 3.32, upper). The truncated beds beneath the channel are bioturbated along the scour surface and the channel is filled with cosets of planar cross-stratified, well-sorted, fine-grained sandstones in which reactivation surfaces and bioturbated, argillaceous siltstone drapes and intercalations are present. The foresets of the cross-strata dip into the channel and along its axis. The sedimentary fill of the channel displays a fining-upwards trend, the higher beds of the sequence being flaggy. The channel-fill may be considered a variant of subfacies 3 and 4, therefore. Above the channel-fill, the presence of a bed of planar cross-strata with foresets dipping to the right (Fig. 3.36, lower, top of the photograph) indicates that the channel no longer controlled sedimentation at the time of deposition of this bed.

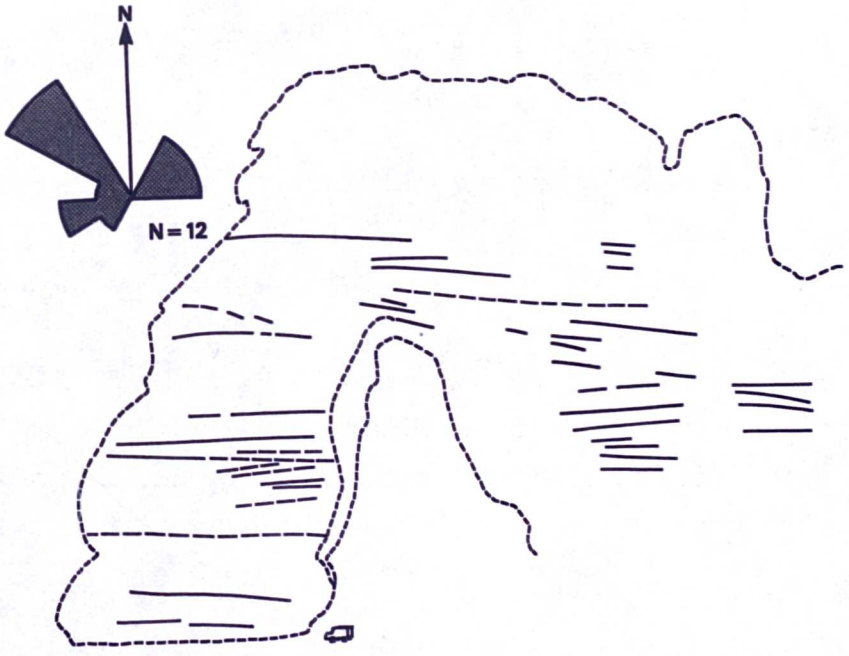


Figure 3.37

3.3.1.7 Petrography

A petrographic study of the sandstones of this facies revealed very similar mineralogies, textures and diagenetic histories to those of facies MSd (see section 3.2.1.4 & Fig. 3.15), although a wider range of degrees of sorting and grain rounding exists, with some examples of well-sorted and rounded textures in medium and fine-grained sandstones. It is interesting to note that the most well-sorted and well-rounded sandstone samples were from the uppermost transgressive sequence (sequence 4, see Fig. 3.6) although poorly-sorted samples were also collected from this sequence. It is possible that this trend in sandstone textures is related to a general increase in marine influence in the sediments of this facies stratigraphically upwards in the succession of the Tadrart Formation (see Fig. 3.6).

Detrital biotite is commonly present in the sandstones of this facies, generally trapped and deformed between sand grains, whilst feldspar is only rarely present. Chloritic clay pellicles are present but chlorite is more abundant in this facies as an authigenic amorphous green to brown interstitial pore fill. This is an important observation since authigenic amorphous chlorite, as opposed to the chlorite of clay pellicles commonly present in facies MSd and thought to develop in non-marine sediments, is considered to provide some limited evidence for marine influence (Rodd, 1980). Both early and late-stage haematite is present in addition to authigenic chlorite, commonly extensively replacing micas and polycrystalline quartz grains.

Scanning electron micrographs of the sandstones of both facies MSd and CSdSt are illustrated in Figures 3.38 and 3.39. From a study of the nature of the quartz overgrowths and the authigenic minerals present, it was concluded that the vast majority of the quartz overgrowths grew in place and that the fractures observed on grains with overgrowths are generally a result of either late diagenetic grain destruction caused by haematite cementation and replacement, or arid weathering processes following exposure. The micrographs illustrate the platey character of the Chlorite Group minerals and the rhomboidal and hexagonal character of the haematite and ilmenite minerals. They also illustrate interlocking authigenic quartz overgrowths, and features resulting from silica dissolution and precipitation on quartz-grain surfaces (Figs. 3.38, 3.39, see Fig. 3.15).

Figure 3.38 ELECTRON MICROGRAPHS OF SANDSTONES FROM FACIES MSd AND CSdSt

Facies CSdSt. Medium-grained sandstone from transgressive sequence 3, Bir Talauaut.

Upper far left : interlocking quartz overgrowths (grains (A) 300 microns diameter).

Upper middle left : detail of diagenetic corrosion of surface (B) of quartz overgrowth (magn. x 800).

Lower far and middle left : pore casts showing nature of the (E,F) interconnected pore space. Large pores (70 microns diameter) are connected by thin tabular radiating pore throats. A residue of iron-titanium oxides is present on the pore cast (magn. x 360, far left; x 200, middle left).

Facies MSd. Fine-grained sandstone, immediately beneath a thin argillaceous, bioturbated siltstone intercalation, from transgressive sequence 2, Wadi Teshuinat.

Upper middle right : surface of quartz grains covered in (C) platey Chlorite Group minerals (magn. x 1300).

Lower middle right : detail of micrograph above, showing (G) presence of platey minerals of the Chlorite Group, and rhombic crystals of haematite and iron-titanium oxides. Micrograph is 40 microns in width.

Upper far right : detail showing hexagonal platey Chlorite (D) Group minerals (2-3 microns diameter) on surface of quartz sand grain (magn. x 4800).

Lower far right : detail showing fractured quartz grain (H) surface, possibly created by grain to grain collision prior to deposition, but more likely by recent grain fracturing during late diagenetic iron-enrichment and/or weathering associated with recent exposure.

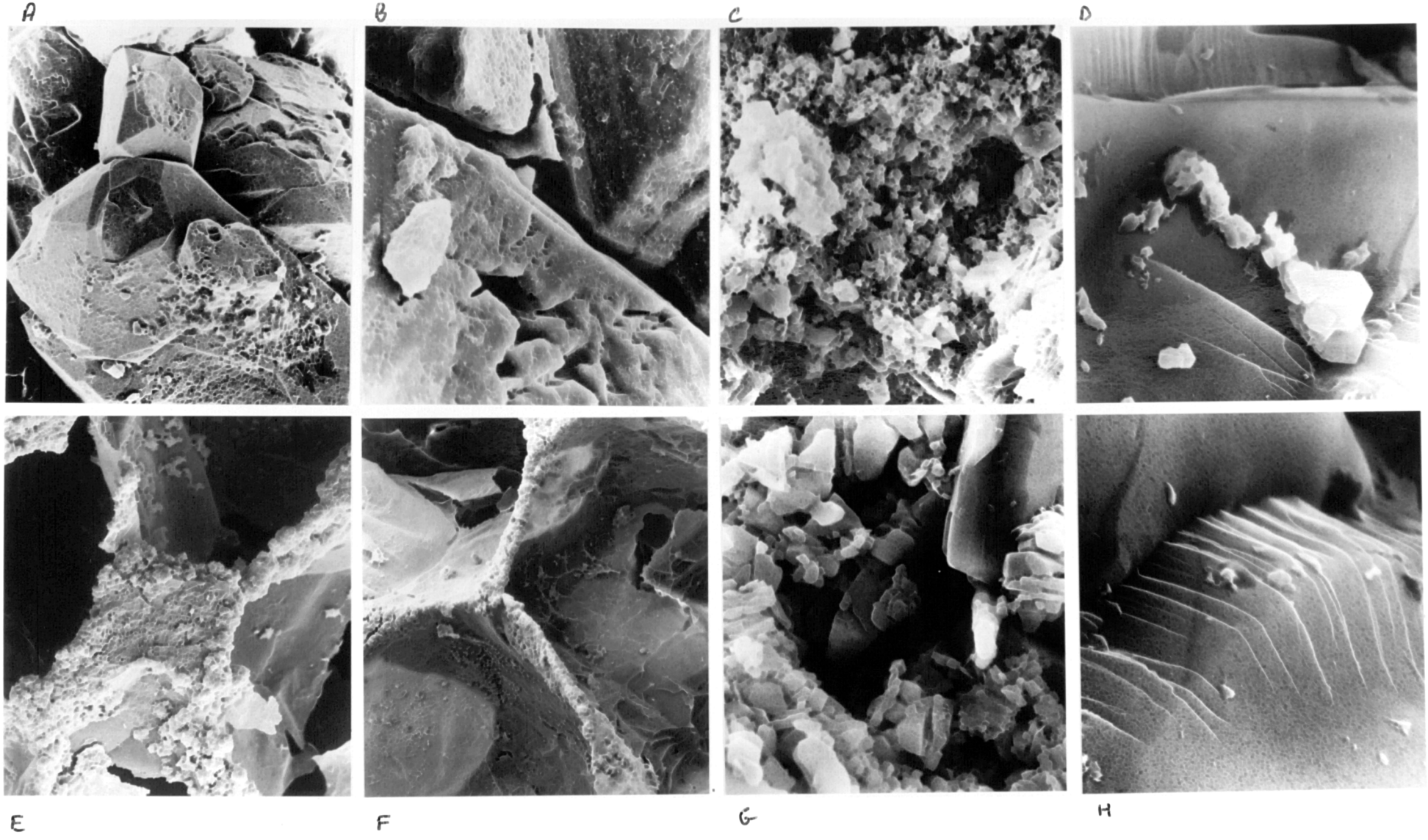

A
B
C
D
E
F
G
H

Figure 3.39 ELECTRON MICROGRAPHS OF SANDSTONE FROM FACIES CSdSt

Poorly-sorted, fine-grained sandstone in a bed displaying recumbent-folded deformed cross-stratification, from transgressive sequence 3, Wadi Anshal.

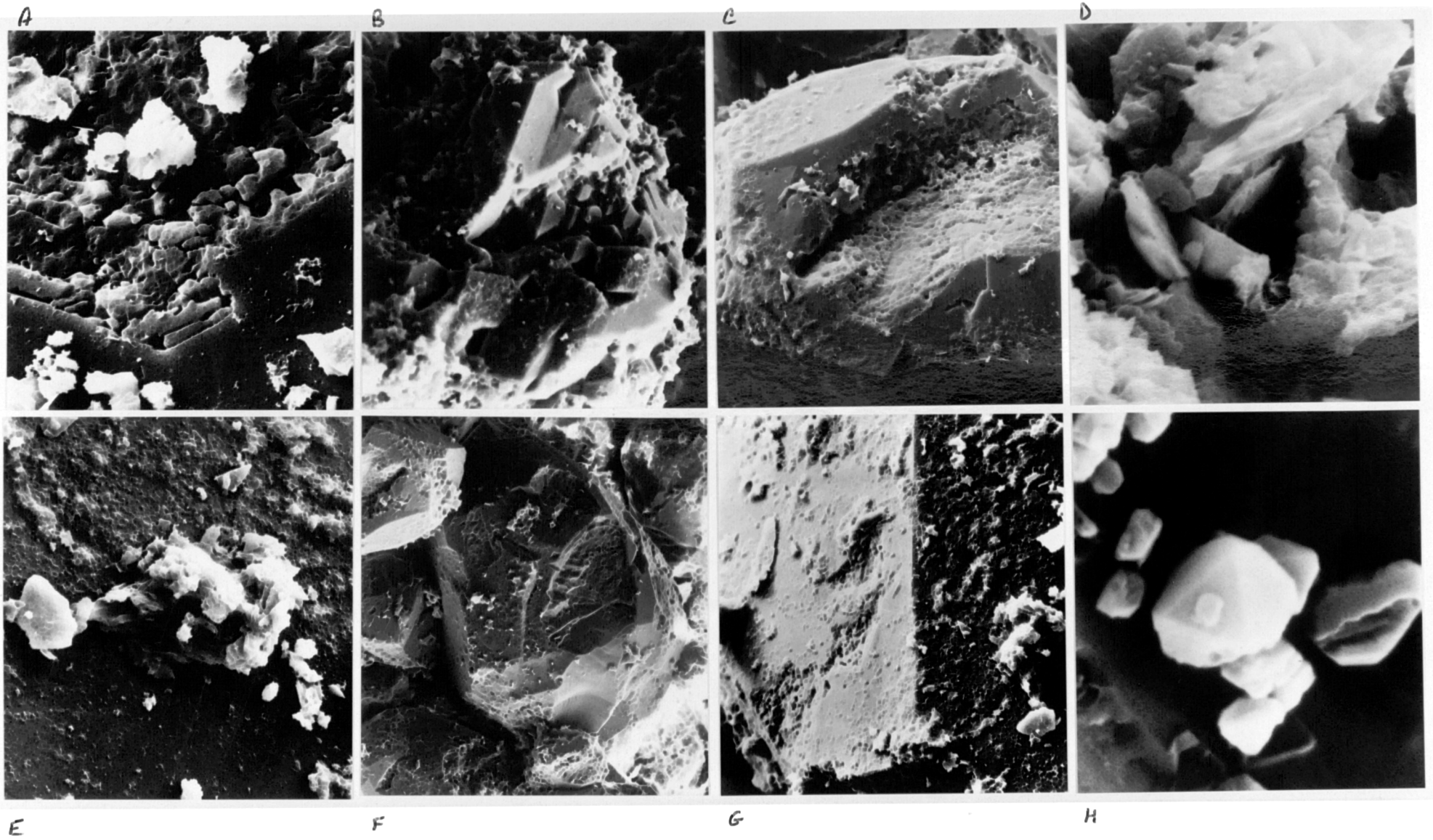
Upper far left : silica dissolution and solution precipitation surface on quartz grain. The light coloured minerals are iron-titanium oxides and Chlorite Group minerals (micrograph is 60 microns in diameter, magn. x 1500). (A)

Lower far left : example of an aggregate of haematite, ilmenite and Chlorite Group minerals (magn. x 940). (E)

Central four micrographs : examples of grain surfaces displaying overgrowths and fractured surfaces. Upper left: magn. x 400. Upper right: magn. x 300. Lower left: magn. x 120. Lower right: magn. x 480. Even at a magnification of x 300 the overgrowth boundary (lower right) is not visibly eroded. The overgrowths, therefore, are thought to have developed within these sandstones rather than to have been inherited. Much of the fracturing seen on the quartz grains is thought to be associated with recent exposure. (B,C F,G)

Upper far right : platey Chlorite Group minerals lying within the pore space of the sandstone (magn. x 4800). (D)

Lower far right : rhombs of authigenic haematite and ilmenite (1-3 microns diameter, magn. x 18000). (H)


A
B
C
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The petrography of this facies, therefore, provides evidence for the same polycyclic, granitic source origin for these sandstones as that determined for facies MSd. The early diagenetic haematite and the chlorite clay rims suggest a continental oxidising environment, whereas the development of abundant interstitial authigenic chlorite may indicate some marine influence.

3.3.2 Interpretation

3.3.2.1 Sedimentary structures and depositional model

The presence of grainsize graded foresets, soft-sediment deformation structures, reactivation surfaces, compound cross-strata and shale intraformational clasts within thick successions of cross-stratified sandstone, some of which is poorly sorted, supports an interpretation for this facies, similar to that for facies MSd, as the deposits of transverse bars in a braided river system. The plant remains suggest deposition in fluvial channels or bars, or in the downstream extension of these fluvial channels. The indications of shell fragments and the abundant bioturbation in subfacies 3, and the presence of clay drapes in subfacies 4, suggest marine influence in this facies, whilst the evidence for some, admittedly limited, herring-bone cross-stratification in subfacies 1 suggests tidal influence (Klein, 1977). These deposits are thought to have been laid down, therefore, on a coastal plain that was dominated by fluvial and estuarine channels. The question of the degree of tidal influence in these channels is discussed below and in section 3.7. It is recognised that insufficient evidence is available to prove that tidal currents were active in the creation of the sedimentary structures of this facies, but it is thought likely that the depositional system was tidal since the deposits of the associated facies RBSdSt and VBSd suggest some tidal influence.

Although subfacies 1, 2 and 3 have been defined as distinct subfacies they are, in nature, representative examples of a gradational range of bedding sequences present in the facies, subfacies 1 being the most similar to the transverse bar deposits of facies MSd, and subfacies 3 displaying the most evidence of marine influence. Just as in facies MSd, evidence for subaerial emergence of estuarine channel bars in the form of desiccation cracks or caliche soil horizons is absent in this facies. Evidence for low-stage/low-tide modification of bedforms is present, however, in the form of compound

cross-strata, convex-upwards reactivation surfaces, and large-wavelength ripple-modification of planar cross-strata.

3.3.2.2 Subfacies 1

The sediments of subfacies 1 have been interpreted as the deposits of large-scale estuarine channel bars. The geometry of the shallow scours, and the presence of argillaceous siltstone drapes overlain by planar cross-strata above these scour horizons, indicate that the scouring events were independent of the filling events, and that a period of slack-water deposition intervened. The cohesive nature of the argillaceous drape explains its preservation beneath planar cross-strata which were deposited over the drape at a later stage under higher-energy conditions (Hjulstrom, in Sundborg, 1956). Algae may also have been responsible for binding the fine-grained sediment. The evidence of argillaceous drapes which pass laterally into beds of broken argillaceous siltstone clasts indicates that flow strength was sufficient in some places to rip up the fine-grained sediments. The presence of lags of coarse sand grains and argillaceous clasts above parallel-stratified sands (Fig. 3.19, unit 6) provides further support for flood-stage scouring and upper flow-regime deposition followed by winnowing of bed surfaces.

The scours typical of subfacies 1 should not be considered the same, therefore, as those associated with scour and fill and trough cross-stratification, where fine-grained drapes are generally absent and where the scouring and filling events were virtually simultaneous. It follows that the scours of this subfacies do not provide evidence of migrating trains of dunes with lee-side scour hollows, but rather evidence of three distinct flow strengths: one sufficiently strong to scour away earlier sediment and to leave channel-shaped scours, one sufficiently weak to allow fine-grained suspension deposition in the scours, and an intermediate flow strength in which transverse bars migrated downstream over an irregular channel bed as a result of slip-face accretion. The range of foreset dip azimuths measured from each set of planar cross-strata lying within a scour (Fig. 3.22, lower), indicates that the transverse bar crests were commonly sinuous.

The fluctuations in flow strength indicated by the characteristics of this facies may be found in both fluvial and estuarine environments (Miall, 1977; Boothroyd & Hubbard, 1976), but the presence of some rare

herringbone cross-stratification (Figs. 3.25 & 3.26) is thought to provide evidence that tidal currents were occasionally active and, hence, that tidal estuarine conditions may have prevailed (Klein, 1977). The rare occurrence of this herringbone cross-stratification, however, is thought to indicate that fluvial flood-stage deposition was the dominant process in these channels.

The scale of some of the sets of cross-strata and of compound cross-strata (Fig. 3.22) indicates that bars of up to 3 m in height existed with sinuous crests up to 40 m in length. The interfingering of foresets bears witness to intermittent avalanching of small-scale bedforms along the crest of the bar in different places at different times. A similar interweaving of foresets was noted by Bluck (1974), in Icelandic braid bars where progradation of lobate bar tails by sporadic slip-face accretion resulted in the interference of two bars and consequently in the interdigitation of their foresets.

The extent to which sinuous bar crests adjust to the scours over which the bars migrate may depend upon the flow strength and water depth at the time of bar migration. In cases where concave-downstream bar crests lie above and are related to underlying channel scours, it is thought that these may have formed during relatively low-stage flows. In other cases, bar-crest sinuosity appears to be unrelated to underlying bed morphology and this is thought to result from the migration of bars over irregular channel bed surfaces during flood stage.

Evidence of fluctuations of flow strength during flood to high-stage flow conditions is thought to be provided by large-scale compound cross-strata (Fig. 3.22, upper) and convex-upwards reactivation surfaces (Fig. 3.26, lower). The former may have formed by the migration of ripple forms, at least 70 cm in height, over an inclined bar margin, whilst the latter is thought to have formed by the process described by McCabe and Jones (1977) where the "washing out" of a bar crest during fluctuating flow conditions results in the creation of convex-upwards reactivation surfaces (see Fig. 3.18). Evidence that slip-face accretion was intermittent is provided by the bioturbation structures at the toe of some foresets. The bar is thought to have built downstream periodically over an irregular surface into a pool of water in which conditions were occasionally sufficiently quiet for the occupation of the sediments of the pool by sediment-injesting organisms.

The small-scale compound cross-strata (Fig. 3.23) and the large-wavelength ripple structures (Fig. 3.24, upper) are thought to provide evidence for modification of bedforms during low-stage flow conditions. In the former case deposition is thought to have occurred on the margins of bars and within channels that were dissecting bars. The migration of small-scale bedforms down-slope or obliquely across the slope of a bar surface is responsible for the inclined sets of cross-strata (Smith, 1970). The large-wavelength ripple modification of planar cross-strata is thought to have been created under high flow strength conditions at the transition to upper flow regime as a result of shallow water sheet runoff (Miall, 1977).

It is common in this facies for the deposits of low-stage conditions to be better sorted and slightly finer grained than the more abundant high-stage deposits. During periods between major fluvial supply of poorly-sorted sediment, therefore, it may be inferred that tidal and fluvial currents partially reworked and sorted the sediment within the channels.

The recumbent-folded deformed cross-strata of this facies are interpreted as in facies MSd (see section 3.2.2.7) by reference to the work of McKee *et al.* (1962), Bluck (1974) and Allen and Banks (1972). Bluck (1974) observed that sediments lying beneath braidplain pools, downstream from migrating bars, were subject to great variation in overburden pressure due to fluctuating water depths and intermittent sediment loading. This results in sudden changes in pore fluid pressure which are compensated for by the expulsion or assimilation of water from the pool, which, in turn, leads to the fluidisation of the sediment and soft-sediment deformation. Since any such deformation would normally be associated with downstream current-induced shear, it seems reasonable to propose this as a mechanism for the generation of the structure, especially as the structure is commonly associated with falling-stage bar tail sedimentary structures (compound cross-strata and reactivation surfaces). The abundance of other soft-sediment deformation structures is attributed to rapid deposition, trapping of a large proportion of interstitial water during flood-stage flow, and the escape of this water during falling-stage conditions.

3.3.2.3 Subfacies 2

Vertical bedding sequences characteristic of the fill of low-sinuosity fluvial/estuarine channels (Collinson, 1978b, pp. 54-55; Boothroyd & Hubbard, 1976) (Fig. 3.20, units 17 & 18) are typical of subfacies 2. The thickness of the bedding sequences suggests the presence of channels of 2-3 m depth, and the preservation of the bedding sequences indicates that this subfacies was deposited in an environment of well-established channellised flow. No lateral-accretion cross-stratification was observed in this subfacies but the fairly flat basal scour surface of some of the bedding sequences may indicate that some lateral channel migration took place. The presence of trough cross-stratification indicates that in-channel dune migration occurred, whilst the linguoid-rippled beds provide evidence for the abandonment of channels as flow transferred to other channels. As in subfacies 1, the presence of argillaceous siltstone lags on channel-scour surfaces indicates that scouring events were distinct and separate from the events responsible for the filling of the channels. The channels which were created, however, appear to have been deeper and to have been more enduring. The presence within the channel sequences of some intraformational conglomerates with competent clasts (Fig. 3.28) suggests that channel banks were fairly well established.

3.3.2.4 Subfacies 3

The plant remains, *Nucula* shell moulds and casts, trace fossils, and the abundant bioturbation within the fining-upwards bedding sequences of subfacies 3, indicate deposition within estuarine channels with considerable marine influence, the coarser-grained sands of which were colonised by filter-feeding and sediment-injesting organisms when fluvial input waned or fluvial supply switched to adjacent channels (Figs. 3.29 & 3.30). The intensely bioturbated fine-grained sediment lying at the top of some bedding sequences (Figs. 3.31, lower, & 3.32, upper) was probably deposited and subsequently bioturbated in quiet-water conditions in abandoned or inactive sub-aqueous channels.

The presence of trough cross-stratification at the base of the channel sequences (Fig. 3.29) is evidence that dunes migrated down the estuarine channels, and the oversteepening of some of the foresets

of these cross-strata suggests that currents passing across the foot of the dune slip face transported sediment away from the foreset toe. This process, as well as the effects of the rapid deposition of sediment in these channels following river floods, and of pore-water pressure changes resulting from water level fluctuations of possible tidal origin, is thought to be responsible for the abundant soft-sediment deformation structures.

The presence of planar and compound cross-stratification in the channel sequences, in addition to trough cross-stratification, is evidence that deposition took place in these channels from bars and/or sandwaves as well as from dunes. Palaeocurrent orientations derived from linguoid-rippled surfaces within the channel sequences show a wide variance which is thought to indicate the former existence of unchannellised marine currents with a great variety of flow directions which were active during periods of limited fluvial input to the subaqueous channels. Where the channelled sandstones of subfacies 3 lie within facies RBSdSt (Fig. 3.32, lower), these are thought to be the subaqueous seaward extension of estuarine channels. Although similar in most respects to the other channels of this subfacies, the evidence of lateral-accretion cross-stratification indicates that some of these seaward channels migrated laterally. The presence of clay intraformational clasts and ripple-laminated beds along foreset surfaces, in some of the large-scale cross-strata of these seaward channels, is attributed to suspension deposition (clay drapes) and bar slip-face modification by marine currents during slack water periods (clay clasts and cross-lamination). The bedding sequences and sedimentary structures of this subfacies bear a strong resemblance to those described from Pleistocene sub-tidal estuarine deposits (de Raaf & Boersma, 1971) and from Tertiary estuarine and tidal, channelled sandstones (Lower Bagshot Beds) (Bosence, 1973). Consequently the sediments of this subfacies have been tentatively interpreted as being the deposits of sub-tidal estuarine channels (Fig. 3.29).

3.3.2.5 Subfacies 4

The geometry of the sandstones and shales of subfacies 4, their trace fossils and the clay drapes on cross-strata foresets, suggest that deposition took place in an environment outside the confines of estuarine channels. These cross-stratified, rippled and bioturbated sandstones may, therefore, have been deposited as estuary mouth bars

and high-energy estuary mouth shoals at the seaward termination of estuarine channels (Fig. 3.20, units 14 & 15). The greater abundance of silty shale in this subfacies, compared with the other subfacies, suggests that traction deposition from channel currents alternated with periods of suspension deposition. The shape and lateral continuity of the sandstone bodies, and the rippling of their upper surfaces, is consistent with a model of periodic deposition of channel-derived sediment as lobate lenses stacked in an imbricate or random fashion on the mouth bar, and subsequently reworked by marine currents. Further, the evidence supports the interpretation that sediment was deposited as lobate lenses in different places at different times on the mouth bar. The presence of cross-stratified beds such as unit 15 (Fig. 3.20; see also Fig. 3.35, lower) provide evidence that, occasionally, channel bars migrated seaward as far as the mouth bar. The structures present in these cross-stratified units (reactivation surfaces, grainsize graded foresets, soft-sediment deformation and shale intraformational clasts), the moderate to poor sorting of the sediment, and the presence of subordinate stratification types (horizontal stratification and homogeneous bedding), are very similar characteristics to those of the channel bar deposits of the other subfacies. This suggests that sediment input to the mouth bar occurred during periods of strong fluvial currents, the deposits of which were only reworked in the uppermost few centimetres by organisms, wave action and marine currents. The palaeocurrent data indicate that the dominant flow direction responsible for the deposition of the sediments of this subfacies was the same as that for subfacies 1 and 2. Since the dominant palaeocurrent orientation of these two subfacies has been interpreted as being an indication of the offshore direction, it may be inferred that the palaeocurrent data of this subfacies provide evidence for deposition from offshore-directed fluvially-derived currents on estuarine mouth bars.

The foreset clay drapes of some of the cross-stratified bar sandstones (Fig. 3.34, upper) may provide evidence for tidal influence during bar progradation (de Raaf & Boersma, 1971). Recent work by Visser (1980), Allen (1981) and Homewood and Allen (1981) has drawn attention to the presence in modern and ancient tidal deposits of tidal bundle structures. These are formed by the action of the subordinate tidal current on the slip-face accretion surface of a

bar or sandwave. This current is responsible for a thin, complex foreset bounded on either side by clay drapes and termed a "tidal bundle". Mowbray and Visser (1982) have shown how the subordinate tidal current may be strong enough to erode the clay drape deposited during the slack water period that preceded reversal of the current, and, consequently, the tidal reversal event results in the preservation of a single clay drape rather than a couplet. In Figure 3.34 (upper left) evidence for erosion of the upper part of a bar, prior to deposition of a clay drape, is clearly visible. Had closer examination of the boundaries and internal structure of the silty clay drapes been undertaken, evidence for ripple-laminated units deposited by the subordinate current might have been obtained. The presence of lenticular and wavy cross-laminated sandstone and siltstone intercalations within the silty shale beds is thought to provide some possible independent evidence for tidal influence in this subfacies (Reineck & Singh, 1973). Further work on this facies might be profitably devoted to this subject since, although the foreset clay drapes are not abundant, the facies may well provide similar evidence for tidal influence on migrating bars as that provided by the Miocene sandstones of Switzerland (Homewood & Allen, 1981). The evidence available does not provide any proof of tidal action, however, since the foreset clay drapes observed may, alternatively, have been deposited during periods of low freshwater discharge.

3.3.2.6 Channel-shaped scour surfaces

The channel-shaped scours, and major angular truncation planes that distinguish facies CSdSt from facies MSd are thought to have been created and preserved by channelling and estuarine processes that were characteristic of the coastal plain environments of these Devonian sediments, but were not active in the Devonian braid-plain environments of facies MSd on account of the less well developed channels and the shallower flows. Although no evidence for a downstream increase in sinuosity has been established from the sediments of facies MSd and CSdSt, except for the rare occurrence of lateral-accretion cross-stratification in seaward estuarine channels, the evidence for more stable channel banks provided by laterally-impersistent, clast-supported, intraformational conglomerates in facies CSdSt suggests that the channels of this facies were deeper than those of facies MSd.

Comparison of the subfacies of facies CSdSt interpreted to be the most landward (subfacies 1) with that interpreted to be the most seaward (subfacies 3) indicates that channels increased both in depth and in width in a seaward direction. Bedding sequences of up to 3 m thickness are characteristic of facies MSd, and of subfacies 1 and 2 of facies CSdSt (estuarine channels) whereas subfacies 3 (?sub-tidal estuarine channels) comprises some examples of channel sandstone bodies of 4-5 m thickness (Figs. 3.32, lower, & 3.36). The channel-shaped scours are generally up to 12 m in width in the seaward facies (1 & 2) whereas they may be up to 25 m in width in the seaward subfacies (3). The larger scour surfaces of at least 50 m width and 6 m depth are thought to be the margins of discrete composite channel systems composed of a large number of erosively-based channels lying within an estuarine channel system. The preservation of such a scour-based estuarine channel complex may have been facilitated by the switching of fluvial input to another estuarine channel system as the first one became choked. This model is analogous to that proposed by Campbell (1976) for channel composite units in the Jurassic of New Mexico, but in facies CSdSt the discrete units of composite channel systems are not as large. Campbell's work, however, was undertaken on braided fluvial alluvium and not estuarine deposits. It is anomalous, therefore, that the channel-shaped scours seen in facies CSdSt were not also present in facies MSd. This anomaly is thought to be explained by the presence of a braid plain of low topographic relief, and the process by which transverse bars were accreted to form lithosomes, in the model proposed for the interpretation of facies MSd (see section 3.2.2.10). It seems unlikely that the sediments deposited and preserved in this braid-plain environment would be divisible into discrete channel-system complexes but, rather, they would form broad units of sheet or blanket geometry.

The combined evidence, therefore, of bedding-sequence thickness, and the scale and character of channel-shaped scours, indicates that channels of c. 2-5 m depth were typical of the depositional setting of this facies, the channels deepening seaward. Bank margins of the estuarine channels may have been more competent than those of the fluvial channels upstream, allowing slightly deeper channels to exist. Breaching of the channel margins, and the consequent development of a new channel complex, may have been responsible for the preservation of

erosively-based estuarine channel complexes. Little evidence for lateral migration of channels exists in the sediments of subfacies 1, whereas some limited evidence for lateral migration was recognised in the more seaward sediments of subfacies 2 and 3. Whilst it seems probable that the majority of the sand bodies of this facies represent vertical accretion under temporally changing flow conditions, a minority of them were probably deposited by laterally migrating channels. Beuf *et al.* (1971) observed in sediments of similar facies to these evidence for a type of periodic lateral migration of channels provided by the presence of a series of parallel channel scours each younger one truncating the older one adjacent to it, and filled with the sediments of migrating transverse bars. Although no similar feature was observed in facies CSdSt this observation does provide some indirect support for the predominance of vertical accretion processes compared with true lateral migration processes.

3.3.2.7 Palaeocurrent data

Although providing evidence of some bipolar current directions, the palaeocurrent data obtained from this facies indicate a strong predominance of major currents (i.e. currents with the power to create bedforms) directed in a northwesterly direction. The palaeocurrent data of the sediments of this facies from transgressive sequence 4 are exceptional, however, in having a large variance and including a southerly orientated component. The general conformity of palaeocurrent directions towards the northwest obtained from both facies MSd and CSdSt is illustrated in Figures 3.55 and 3.56 (see section 3.7, this Chapter). As with facies MSd, therefore, flood-stage fluvial currents appear to have been the agent responsible for the deposition of the majority of the preserved sediment even in the case of subaqueous channels on the inner shelf (sandstones of subfacies 3 lying entirely within sediments of facies RBSdSt). The dominance of northwesterly-orientated currents and the fact that herringbone cross-stratification and foreset clay drapes are not commonly present, suggest that fluvial currents had greater flow strength and were more significant than tidal or any other marine currents. Since ebb-tidal currents have the same flow direction as fluvial currents they are difficult to distinguish from each other. The texture of the sediments, however, provides little support for a flow circulation system involving

extensive reworking of sediment in flood-tidal and ebb-tidal channels. The greater palaeocurrent variance of this facies, compared with that of facies MSd, is thought to reflect the influence of marine currents, of tidal or storm origin, and also a more varied pattern of channel orientations close to the sea.

3.3.2.8 Biogenic structures

The similarity of the trace fossil assemblages of facies CSdSt (see section 3.3.1.1) with those of facies MSd (see section 3.2.2.9) indicates that the organisms responsible occupied broadly similar habitats in the depositional settings of each facies (i.e. lakes, lagoons and abandoned channels). The trace fossils present in this facies may be divided into two categories, those that occur in relatively well-sorted and thickly-bedded, fine-grained sandstone beds (*Skolithos* and *Rosselia*) and those that occur in argillaceous siltstones and shales, and on the soles and surfaces of adjacent sandstone beds (*Arthrophycus*, *Palaeophycus*, *Fraena* and *Cruziana*). The absence of the following distinctive trace fossils, which are characteristic of facies RBSdSt, suggests that the sediments of facies CSdSt were deposited in more restricted conditions than the relatively open-marine inner-shelf conditions of facies RBSdSt: *Spirophyton*, *Corophioides* and *Asterosoma*. Since the shales, argillaceous siltstones and fine-grained sandstones, in which the trace fossils of facies MSd, CSdSt, RBSdSt and VBSd are found, do not differ significantly between facies, it seems unlikely that substrate type played a controlling influence on the trace fossil assemblages present in these facies. On the other hand water salinity differences between the facies may well offer an explanation for the greater abundance of ichnogenera in the more marine facies (RBSdSt and VBSd) than in the non-marine and restricted-marine facies (MSd and CSdSt), the restricted-marine facies meaning, in this context, estuarine channel deposits and channel-mouth shoal deposits that were characterised by low-salinity waters.

The presence of the "straight-burrow variety" of *Arthrophycus* in the sediments of facies MSd and CSdSt only, suggests that, in the rocks under study, the organism responsible for these burrows and the mode of existence of this organism was limited to brackish and fresh-water environments. This conclusion is similar to that reached by Cotter in his study of this trace fossil in the Silurian of North

America (1978, & pers. comm., 1979) (see section 3.2.2.9). Indeed, the "palm-frond variety" of *Arthrophycus* was only found in the well-bedded sediments of facies RBSdSt and VBSd in the upper beds of transgressive sequence 1, whilst it was absent from these facies in sequences 2, 3 and 4 (Fig. 3.6).

Cruziana, unlike *Arthrophycus*, is present in facies CSdSt (subfacies 4) and in facies RBSdSt, but is absent from facies MSd. Seilacher (1969, 1970) and Osgood (1970) argued that *Cruziana* is the crawling trace of a trilobite, an exclusively marine arthropod (Harrington, 1959), whereas others believe that *Cruziana*, defined only at ichnogenus level, provides no more evidence than that of crawling trails of arthropods of both marine and non-marine affinities (R. Bromley, pers.comm., 1982). Selley (1970) and Martino and Zapecza (1978) studied *Cruziana* and *Rusophycus*-bearing sandstone successions which have been interpreted from their sedimentary structures as being of fluvial origin, the former in the Ordovician of Jordan and the latter in the Upper Silurian of North America, and concluded that the trace fossils provided evidence for marine influence in the form of minor transgressions and the occupation of the lower reaches of estuaries by marine organisms during low-stage flow-conditions (Selley, 1970) and the development of marine-influenced interdistributary bays in a high-constructive elongate delta complex (Martino & Zapecza, 1978). In the present study, the possible tidal influence associated with the beds in which the *Cruziana* trace fossils are present (subfacies 4) suggests that these sediments were deposited at the mouths of estuarine/?tidal channels (Fig. 3.20, units 14 & 15) rather than in interdistributary bays. However, it is possible that those sediments of this subfacies that interfinger with sediments of subfacies 1 and 2, and with facies MSd, were deposited as crevasse splay and crevasse channel sandstone bodies produced by sudden incursions of flood-generated currents into tidally-influenced flood-plain regions lying between the major channels and partially open to the sea.

3.3.2.9 Vertical sequence analysis

A study of subfacies transitions in this facies cannot be usefully undertaken without including the associated facies with which it interfingers: facies MSd (braided fluvial), facies RBSdSt (shallow-marine), and facies VBSd (?tidal flat) (Fig. 3.6). A qualitative

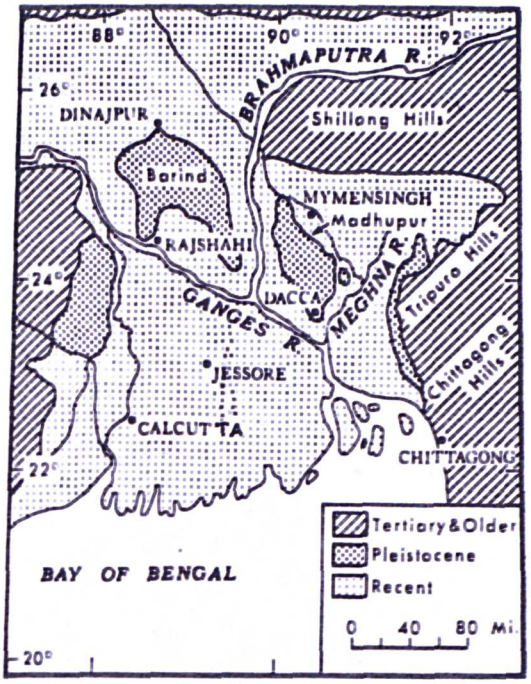
analysis of all these facies in the measured sections provided the following information on facies and subfacies associations. Braided fluvial deposits without channel scour features (facies MSd) are interbedded with possibly tidally-influenced transverse bar deposits lying above broad shallow channel scours (subfacies 1, facies CSdSt). These may be overlain by estuarine mouth bar deposits (subfacies 4, facies CSdSt), or ?tidal flat deposits (subfacies 2, facies CSdSt, and facies VBSd). In the former case a minor marine advance in a large-scale estuarine channel and in the latter case abandonment of an estuarine channel may be invoked to explain these relationships. Deposits of well-developed estuarine channels (subfacies 2, facies CSdSt) are generally stacked vertically, truncating each other. They commonly lie above the estuarine transverse bar deposits of subfacies 1 and pass upwards into ?tidal flat deposits (facies VBSd and RBSdSt). The seaward estuarine channel deposits of subfacies 3 (facies CSdSt) and the mouth bar deposits of subfacies 4 (both of facies CSdSt) are most commonly overlain by shallow-marine deposits of facies RBSdSt.

The complexity of the depositional system is reflected by the fact that seaward estuarine channel and mouth bar deposits are overlain, in some cases, by estuarine transverse bar deposits (subfacies 1, facies CSdSt). Although many of these facies transitions could be explained by fluctuating sea level during deposition, it is not thought necessary to invoke this process except for the major transgressive sequences (see Fig. 3.6). Switching of fluvial input from one estuarine channel to another would probably have been sufficient to substantially change the equilibrium balance between sediment supply, subsidence and compaction which controls minor advances and retreats of the sea along a coastline during a period of stable sea level. Renewed fluvial input to a previously abandoned channel would cause seaward progradation of the fluvial facies. The absence of a prodelta shale facies in the transgressive sequence, and the heterolithic character of facies RBSdSt, suggest that the open marine environment seaward of the estuarine channels was an extensive shallow-marine inner shelf. The palaeogeographic reconstruction which fits the evidence best, therefore, is as follows: fluvial channels of a braid plain passed downstream into a coastal plain environment of estuarine/?tidal channels, which passed

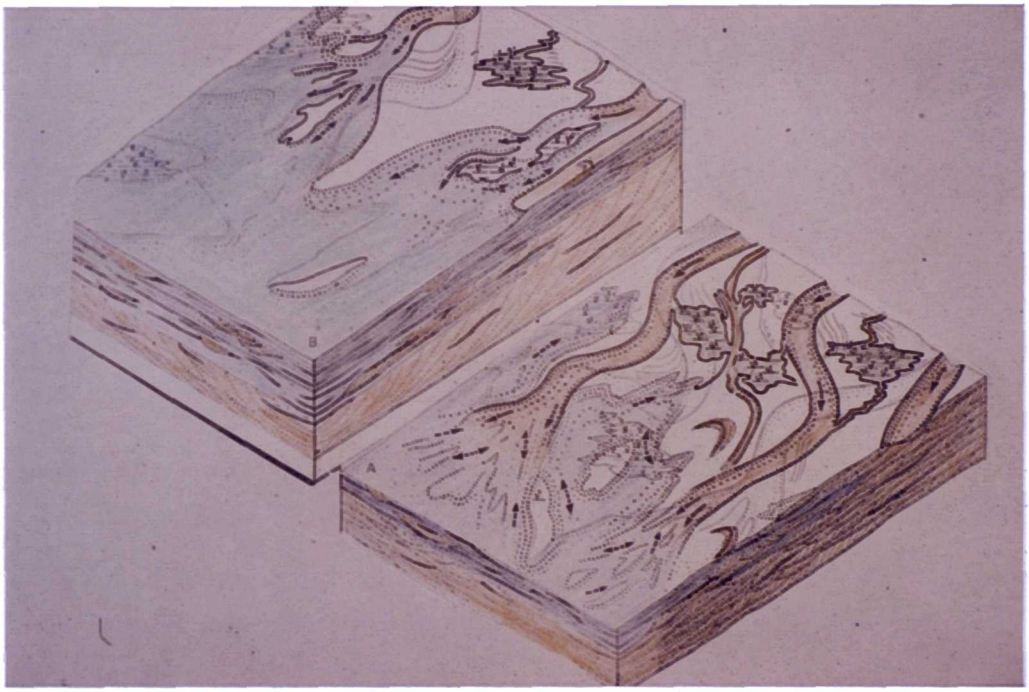
downslope with a very gentle gradient into an environment of shallow-water shelf deposition, subaqueous channels and channel mouth bars.

3.3.2.10 Alternative depositional models

The possibility exists that the absence of evidence for very active low-stage reworking of bars may indicate that the fluvial system responsible for the braid plain and estuarine deposits of facies MSd and CSdSt was characterised by large deep rivers, such as the Brahmaputra River (Coleman, 1969). The low palaeocurrent variance of cosets of planar cross-strata in facies MSd may also be interpreted as providing evidence for deep-water depositional conditions. Faced with similar phenomena, as well as very large-scale bedforms, in the fluvial Triassic Hawkesbury Sandstone of New South Wales, Conaghan and Jones (1975) and Jones and Rust (1981) proposed interpretations based upon the Brahmaputra River model. The Brahmaputra and Ganges rivers together have deposited an extensive Recent sheet sandstone of predominantly fluvial deposits 300 km by 500 km in extent (Fig. 3.40, upper). The mouth of the Ganges River is dominated by sandstone bodies deposited by fluvial currents which extend well out into the Bay of Bengal. The channel floor of the Brahmaputra River is characterised by large-scale migrating bedforms, including sandwaves, which build cosets of planar cross-strata on the channel floor. These channel bars may be bioturbated on their upper surfaces and are rarely emergent. It may be inferred that the palaeocurrent variance of the deposited cosets of planar cross-strata would be very low, and that reactivation surfaces would form within these bars, without their becoming emergent as a result of flow-stage fluctuations (Jones, 1977). This model is not favoured, however, for facies MSd and CSdSt because the evidence for channel depths in both facies is considered sufficiently strong to rule out the possibility that channels were present which were four times deeper than the inferred deepest channel of both facies (6 m deep compared with 25 m deep in the Brahmaputra River). Secondly, the evidence from the planar cross-strata of facies MSd suggests that deposition took place largely by the downstream construction of transverse bars, under conditions of fluctuating flow strength, rather than by the migration of trains of dunes and sandwaves, as in the bed of the Brahmaputra River. It seems probable, therefore, that the lack of evidence of emergence in facies



Recent sheet sandstone deposited by the Brahmaputra and Ganges rivers (after Coleman, 1969).



Depositional model for sandstones of the Tifernine Formation, Lower Devonian, Algeria (after Beuf *et al.*, 1971).

A. (lower) rivers debauching into the sea deposit cross-stratified sandstones. Lacustrine and lagoonal deposits are laid down in interfluve areas.

B. (upper) following reduction in fluvial input, deposition is dominated by marine and littoral processes.

Figure 3.40 POSSIBLE DEPOSITIONAL MODELS FOR FACIES MSd AND CSdSt.

MSd is a result of the low preservation potential of emergence structures in shallow-water braided-river systems.

McKee (1962) studied the Mesozoic Nubian-type sandstones of Libya with the intention of establishing whether an aeolian depositional model would be appropriate. His studies led him to the conclusion that the sediments studied were all deposited subaqueously and that none of them were aeolian. However, Beuf *et al.* (1971) proposed that some of the sedimentary structures of the Saharan Nubian-type sandstones (in this case, those of the Palaeozoic) indicated deposition from aeolian dunes in interfluve areas of braided and meandering fluvial plains. Although some sediment was most probably transported by wind prior to being deposited in fluvial channels, none of the sedimentary structures of facies MSd and CSdSt observed in the present study are thought to provide convincing evidence for subaerial and aeolian deposition. Studies of recent aeolian dunes (McKee, 1966; McKee & Tibbits, 1964; and McKee, 1979), of ancient aeolian deposits (Thompson, 1969) and personal observations of modern and ancient aeolian dune sands (Permo-Triassic sandstones near Penrith, in Arran and at Hopetoun (Moray Firth)) indicate that the following features are characteristic of aeolian dune deposits:

(1) complex inclined truncation planes separating foresets with a wide range of angles of inclination;
(2) sets which thin and thicken laterally in a complex irregular fashion;
(3) soft-sediment deformation;
(4) very large-scale trough cross-stratification;
(5) very large-scale, low-angle truncation planes, often slightly curved and convex-upwards, overlain conformably by thick homogeneous beds;
(6) well-sorted sediment.

None of the subfacies of facies CSdSt and MSd have more than one or two of these characteristics and the abundance of horizontal or sub-horizontal set boundaries, so characteristic of Nubian-type sandstones, is thought to distinguish them from existing aeolian facies models (e.g. Glennie, 1970; Walker & Middleton, 1979).

The trace fossil evidence and the absence of evidence for subaerial emergence suggest that the work undertaken in recent years on off-shore shallow marine sand bodies (Anderton, 1976; Boyles & Scott,

1982; Brenner & Davies, 1974; Hobday & Reading, 1972; Johnson, 1978; Pryor & Amaral, 1971; Spearing, 1975; Stanley & Swift, 1976) may provide a depositional analogy for the sediments of subfacies 3 which are commonly closely associated with shallow-marine sediments of facies RBSdSt. The presence of plant remains and some poorly-sorted sediment does not suggest, however, that the sediments of this subfacies have been extensively reworked by marine currents. Nevertheless many of the same fluctuating and reversing current flow conditions typical of tidal/estuarine environments (Klein, 1977) have been shown to apply to the depositional processes responsible for off-shore sand bodies. The absence in subfacies 3 of recognised large-scale low-angle inclined surfaces such as those described from Precambrian sandstones interpreted as shallow-marine tidal sand bars (Hobday & Reading, 1972), from Eocene sandstones interpreted as shelf sandwaves of up to 20 m height and 2 km length (Nio, 1976), and from linear tidal and storm-dominated sand ridges (Johnson, 1977) as well as the absence of upward-coarsening sequences typical of these sandbodies, indicates that no close similarity exists.

3.3.2.11 Conclusion

The sediments of facies CSdSt provide evidence for a coastal plain and nearshore marine depositional environment dominated by a braided pattern of fluvial and estuarine channels which passed offshore into subaqueous channels and estuary mouth bars and shoals. The palaeocurrent data indicate that this depositional system was dominated by fluvial currents, even in the nearshore environment, a characteristic of some of the mouths of the world's major rivers; e.g. the Ganges (Fig. 3.40, upper; Coleman, 1969), the Mississippi (Elliot, 1978a) and the Niger (Oomkens, 1974). This facies may, therefore, be considered as a deltaic plain/fluviomarine facies, and it bears a strong resemblance to the facies of the Quaternary deposits of the deltaic plain of the Niger Delta (*ibid.*) and to those of the distributary channels of the Rhone Delta (Oomkens, 1970). Where tidal influence on a delta is absent or restricted to a low tidal range a coastal barrier bar is commonly developed and its sediments preserved (e.g. the Rhone Delta, *ibid.*), whereas where the tidal range is microtidal to mesotidal lateral migration of tidal channels ensures that the sediments of the barrier bar have a low preservation potential

(e.g. Niger Delta, Oomkens, 1974; Hayes & Kana, 1976). The absence of recognised barrier bar facies in facies CSdSt and RBSdSt (see section 3.4) is thought to suggest that some tidal influence probably existed in the depositional regime of these facies. The estuarine channels of facies CSdSt have, therefore, been tentatively interpreted as tidal, both inter-tidal and sub-tidal, following comparison with sequences described by Boothroyd and Hubbard (1976) and de Raaf and Boersma (1971). The scarcity of herringbone cross-stratification in the sediments of these channels is not thought to disprove this interpretation since a large number of authors have noted that evidence of the dominant current direction is commonly preferentially preserved over the subordinate current in tidal environments (Bridges, 1982). This is not to suggest, however, that the channels of this facies were dominated by ebb-tidal currents. Evidence has already been presented that suggests that fluvial currents were dominant and that the influence of marine currents was subordinate. Not least of this evidence is the palaeocurrent data from transgressive sequences 2 and 3 (see Figs. 3.56 & 3.57) which indicate a very strong predominance of northwesterly palaeoflows. However, the somewhat enigmatic palaeocurrent data of transgressive sequence 4 (Fig. 3.58) are thought to reflect an increase in influence of marine currents, including alongshore currents (to the southwest), at the expense of fluvial currents. The southeasterly azimuths in this transgressive sequence may be an indication of the activity of flood-tidal currents.

Brackish and low-salinity waters are thought to have characterised the estuarine channels and the perennial lakes, lagoons and abandoned channels of the fluvial/estuarine flood plain. The probable absence of barrier beaches in this environment would have allowed access for marine waters to the restricted flood plain environments, explaining the presence in some of these quiet-water sediments, as well as in those of the estuarine channels, of trace fossils thought to be of marine and restricted-marine affinities.

The absence of desiccation sedimentary structures is thought to indicate that channel water levels generally remained high enough to restrict the amount of desiccation and emergence modification to a minimum. This suggests that a humid climate must have prevailed. The evidence of bedding sequence thicknesses and scoured surfaces indicates that the more landward estuarine channels were commonly

2-4 m deep whilst the seaward channels were commonly 3-5 m deep, but up to 7 m deep, and were present above scoured surfaces of a greater width than those which characterised the landward estuarine channels. No direct evidence was obtained for the semi-enclosed or funnel-shaped coastal body of water that characterises estuaries (Pritchard, 1967), and, given the very widespread distribution of the blanket sandstones of facies MSd and CSdSt (Beuf *et al.*, 1971), it seems probable that the coastal plain was one of very low relief in which estuaries were only temporarily well developed before channel switching led to their abandonment, and in which there was little structural or geomorphological control to allow the development of funnel-shaped estuaries. The environment of deposition proposed for this facies is similar, therefore, in some respects, to that proposed by Beuf *et al.* (1971) for the Lower Devonian sandstones of southern Algeria (Fig. 3.40, lower). However, evidence for meandering channels and for aeolian deposits, such as those recognised by Beuf *et al.* (1971), was not present either in this facies or in facies MSd. This fluvially-dominated coastal/deltaic plain, therefore, may have been similar to that of the Niger Delta and the Ganges Delta (Fig. 3.40, upper) except that the estuarine channels are thought, generally, not to have been deeper than 5 m.

3.4 Rippled and Bioturbated Sandstone and Siltstone Facies (RBSdSt)

3.4.1 Description

3.4.1.1 General

The rippled and bioturbated sandstone and siltstone facies characteristically comprises thinly-bedded heterolithic successions of alternating fine and very fine-grained sandstones, siltstones and subordinate silty shales in beds of 5-30 cm thickness (Fig. 3.6). Repetitive sequences of sedimentary structures are characteristic of the facies, and discordant minor scour surfaces are abundant and commonly present at the base of sandstone beds. These sediments generally weather to rubbly, scree-strewn slopes, and the coarser-grained beds weather to "platforms" or "pavements" which stand out within the slopes.

Trace fossils are abundant in this facies whilst plant remains and shell fragments are present in some beds. Cross-lamination is the most common sedimentary structure, the majority of which is current-formed, although examples of wave-ripple cross-lamination were also observed. Six subfacies were identified in the facies of which subfacies 5 and 6 are only rarely present. Subfacies 1 and 2 comprise alternating thin, bioturbated beds of sandstone, siltstone and silty shale, subfacies 2 being distinguished from subfacies 1 by the presence of thinner, more argillaceous flaggy beds. Subfacies 3, 5 and 6 are characterised by erosively-based vertical bedding sequences indicating decreasing flow strength, whilst subfacies 4 comprises a sequence of bioturbated beds reflecting a vertical increase in flow strength.

3.4.1.2 Subfacies 1

Alternating beds of laterally-persistent bioturbated sandstone and argillaceous siltstone (Fig. 3.41) have been assigned to subfacies 1, which is the subfacies most commonly present in this facies. Complex feeding burrow systems are characteristic of the siltstone beds (*Asterosoma*, see Fig. 2.6, and *Spirophyton*, see Fig. 2.8), whilst the sandstone beds are characterised by shell fragments, vertical burrows (e.g. *Skolithos*, see Fig. 2.7, and *Corophioides*, see Fig. 2.8), trails and resting traces commonly found on rippled surfaces (e.g. *Cruziana*, see Fig. 2.4, *Sagittichnus*, see Fig. 2.9, and *Halopoa*; see also Fig. 2.3 for further examples of arthropod tracks). In most of the

Figure 3.41 MEASURED SECTION OF FACIES RBSdSt AT WADI ANSHAL

Notes:

1) The section was measured 141 to 181 m stratigraphically above the base of the Tadrart Formation at locality 12 (Fig. 3.5) in the upper part of transgressive sequence 3.

2) The palaeocurrent readings were measured from asymmetrical rippled surfaces and cross-lamination. The inferred palaeoslope direction towards the northwest is derived from the palaeocurrent rose diagrams plotted from all the cross-strata dip azimuths available from facies MSd and CSdSt in transgressive sequence 3 (see Fig. 3.57).

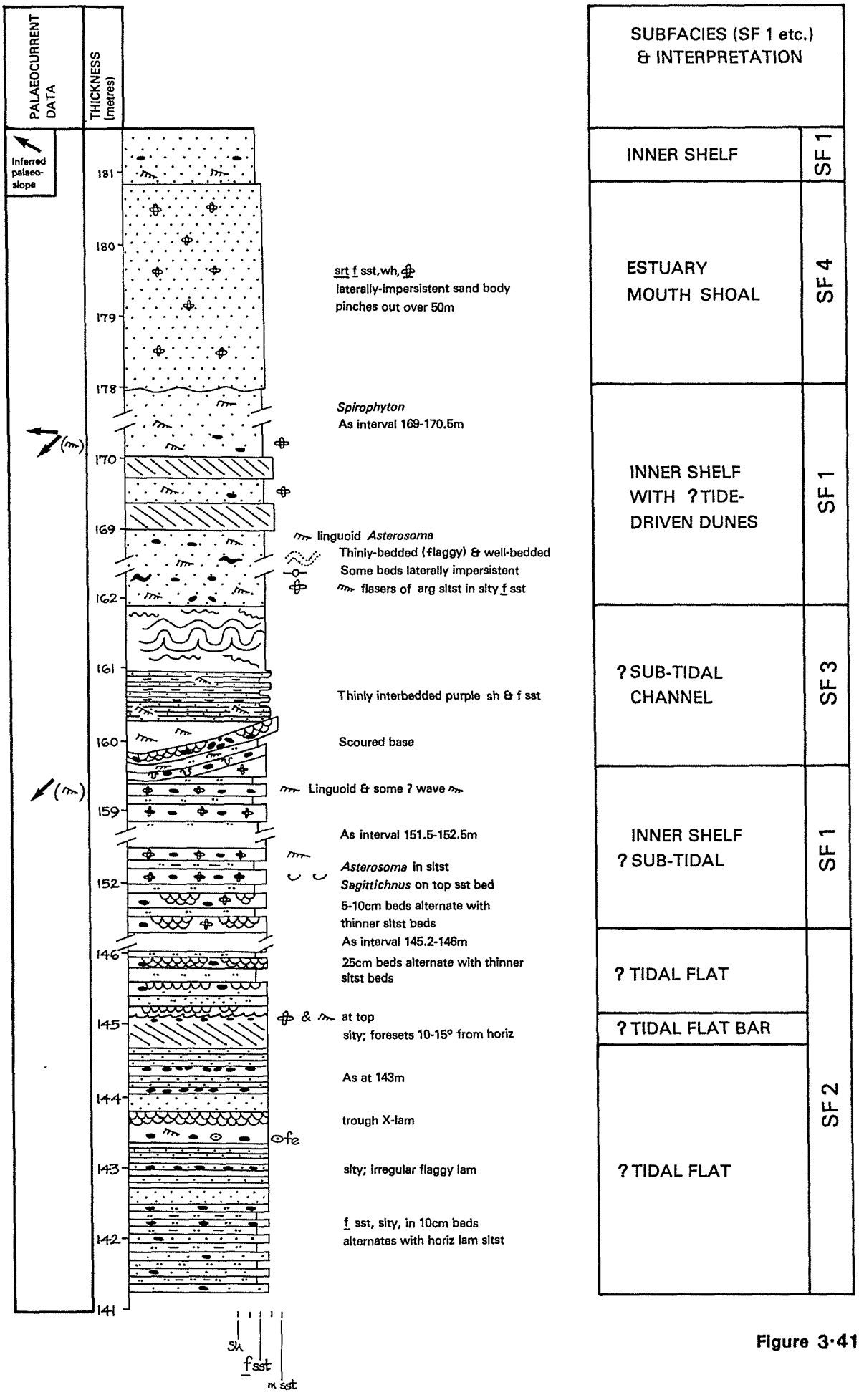

PALAEOCURRENT DATA
THICKNESS (metres)
SUBFACIES (SF 1 etc.) & INTERPRETATION
Inferred palaeo-slope
181
180
179
178
170
169
162
161
160
159
152
146
145
144
143
142
141
srt f sst,wh,
laterally-impersistent sand body
pinches out over 50m
Spirophyton
As interval 169-170.5m
linguoid Asterosoma
Thinly-bedded (flaggy) & well-bedded
Some beds laterally impersistent
flasers of arg sltst in slty f sst
Thinly interbedded purple sh & f sst
Scoured base
Linguoid & some ? wave
As interval 151.5-152.5m
Asterosoma in sltst
Sagittichnus on top sst bed
5-10cm beds alternate with thinner sltst beds
As interval 145.2-146m
25cm beds alternate with thinner sltst beds
& at top
slty; foresets 10-15° from horiz
As at 143m
trough X-lam
fe
slty; irregular flaggy lam
f sst, slty, in 10cm beds alternates with horiz lam sltst
Sh
f sst
m sst
INNER SHELF
SF 1
ESTUARY MOUTH SHOAL
SF 4
INNER SHELF WITH ?TIDE-DRIVEN DUNES
SF 1
?SUB-TIDAL CHANNEL
SF 3
INNER SHELF ?SUB-TIDAL
SF 1
?TIDAL FLAT
?TIDAL FLAT BAR
?TIDAL FLAT
SF 2

Figure 3·41

sandstone beds bioturbation is more pronounced in the upper part of the bed and may be absent in the lower part. The shell fragments were always too broken to be identified, but bivalve shell casts and moulds similar to those described from subfacies 3 of facies CSdSt (see section 3.3.1.4) were observed which have been tentatively assigned to the genus *Nucula*. In addition some samples were collected from sandstones of this subfacies which are possibly referrable to *Conularia* (see Borghi, 1940, plate XVI), a Cambrian to Permian fossil group of possible Cnidarian (hydrozoan) affinities (Clarkson, 1979). The sandstone beds of this subfacies may also contain plant remains, the impressions of both Psilopsida genera, *Arthrostigma* and *Aneurophyton*, being present (Fig. 3.42).

As well as current-formed straight-crested, undulatory and linguoid ripple surfaces and ripple cross-lamination, some "bundle-wise", upward-building, small-scale foresets and bifurcating ripple crests were observed, which are thought to be characteristic of wave-formed ripples (Boersma, 1970) (Fig. 3.43). In addition, climbing-ripple and flaser cross-lamination is present in this subfacies, and loaded ripples are also a common feature (Fig. 3.43). A small number of examples of hummocky cross-stratification (Harms, 1975b) were observed in the fine-grained sandstone beds, and, interbedded with the finer-grained sediments of this subfacies, are planar cross-stratified sandstone beds which, in some cases, lie within bedding sequences thought to be characteristic of tidal creeks (Fig. 3.44) (Reineck & Singh, 1973), and in other cases are isolate, and inter-bedded with strongly bioturbated sediment (Fig. 3.41). Some of the planar cross-strata of the "tidal creek" bedding sequences display recumbent-folded deformed cross-strata. Nearly-homogeneous sandstone beds of irregular shape and displaying complex soft-sediment deformation are also present, lying above scoured surfaces within the well-bedded rippled beds of this subfacies.

Heterolithic bedding sequences characteristic of this subfacies are illustrated in Figures 3.45 (lower right) and 3.47 (lower right). Discordant bedding relationships (Fig. 3.45, upper), loaded channel sandstone bodies (Fig. 3.45, lower) and ferruginised intraformational clay clasts (Fig. 3.47, lower left) are all common features of these sequences. In some cases the beds of this subfacies lie markedly subhorizontally to the regional dip whilst others may be laterally

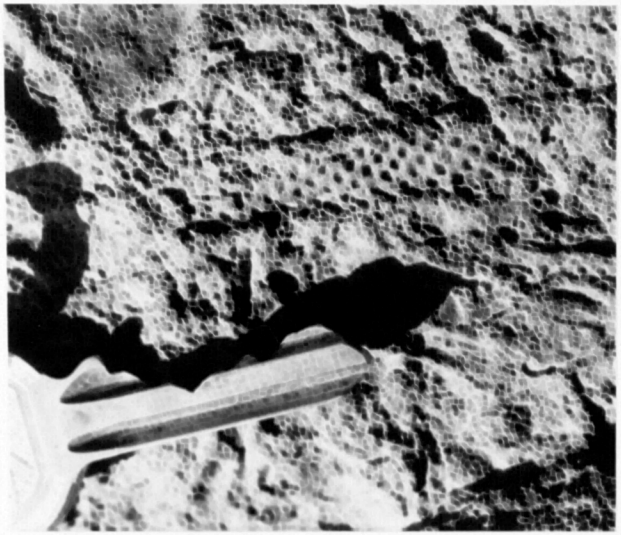
Figure 3.42 PALAEOFLORA OF THE TADRART FORMATION

Scales: lens cap, 5 cm diameter; coin, 2 cm diameter; key, 5 cm in length.

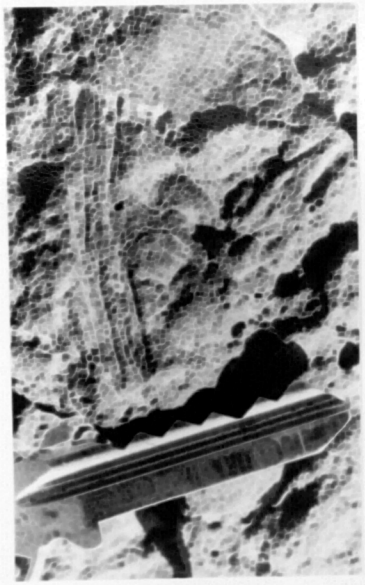
Upper three illustrations : *Arthrostigma*, impressions of stalk and leaf-like fragments with longitudinal ribs, present on the bedding planes of

1) bioturbated, fine-grained sandstones of facies RBSdSt (subfacies 1) at Bir Talauaut, transgressive sequence 4, and
2) medium-grained sandstones of facies CSdSt (subfacies 3), transgressive sequence 4, at locality 17, Tihererin (Fig. 3.5) (left photograph).

Lower two illustrations : *Aneurophyton*, impressions of massive pitted wood, present on the bedding planes of bioturbated, fine-grained sandstones of facies RBSdSt at Bir Talauaut, transgressive sequence 4. These specimens may tentatively be assigned to the species *A. germanicum*, after descriptions in Darrah (1960).



ASAHI
PENTAX



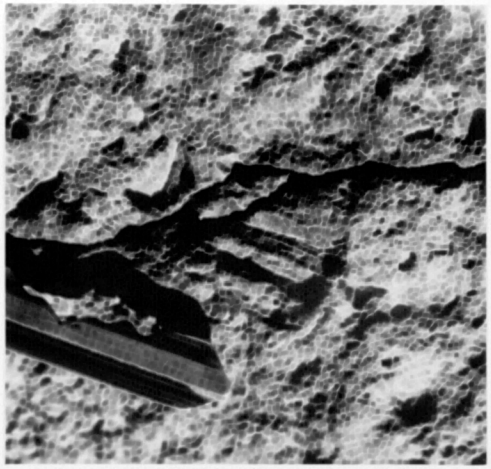


Figure 3.43 SEDIMENTARY STRUCTURES OF FACIES RBSdSt

Scales: lens cap, 5 cm diameter; hammer, 30 cm long.

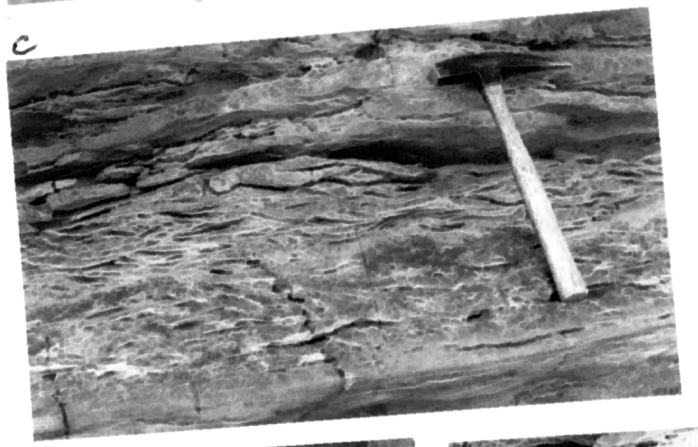
(A, B, C) Upper left and right, and middle left : varieties of cross-lamination. Nearly in-phase climbing ripples, wavelength 20 cm (upper left). Bioturbated very fine-grained sandstone beds with rippled surfaces, faintly cross-laminated or homogeneous internally (upper right). Trough cross-laminated silty very fine-grained sandstone with thin flasers of argillaceous siltstone (middle left).

(E) Lower left : homogeneous and cross-laminated sandstone beds with linguoid rippled surfaces and loaded ripples on the sole of the beds.

(D, F) Middle and lower right : soft-sediment deformation in sandstones of subfacies 3, interpreted as the deposits of subtidal channels. Lower photograph is of sandstones of subfacies 3, illustrated in Figure 3.41.

A

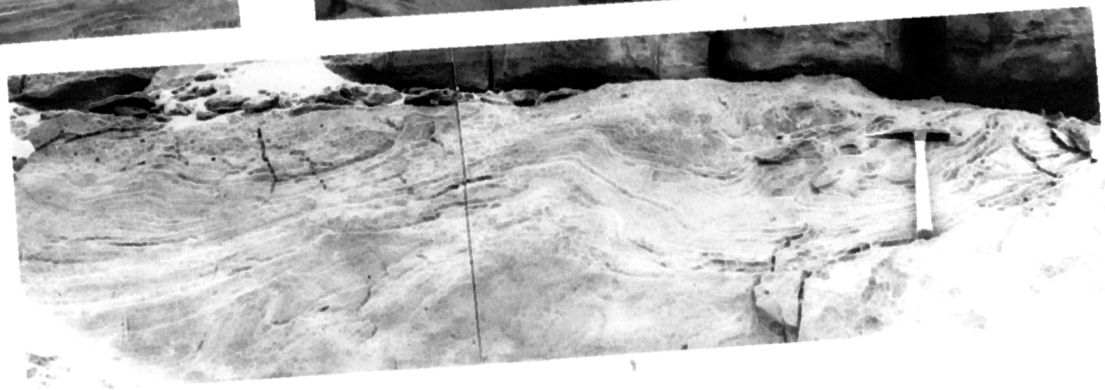
B



C

D

E



F

Figure 3.44 MEASURED SECTIONS OF FACIES RBSdSt AT WADI ANSHAL AND TIHERERIN

Notes:

1) The Wadi Anshal sections are actual bedding sequences that were measured through sediments laterally equivalent to those illustrated in Figure 3.41, 141 to 181 m stratigraphically above the base of the Tadrart Formation at locality 12 (Fig. 3.5).

2) The Tihererin section was measured 150 to 163 m stratigraphically above the base of the Tadrart Formation at locality 17 (Fig. 3.5). For convenience of illustration the vertical succession of the Tihererin section has been foreshortened.

3) All the sections were measured in sediments of the upper part of transgressive sequence 3.

4) The palaeocurrent readings were measured from trough axes and foreset dip azimuths of cross-strata. The inferred palaeoslope direction towards the northwest is derived from the palaeocurrent rose diagrams plotted from all the cross-strata dip azimuths available from facies MSd and CSdSt in transgressive sequence 3 (see Fig. 3.57).

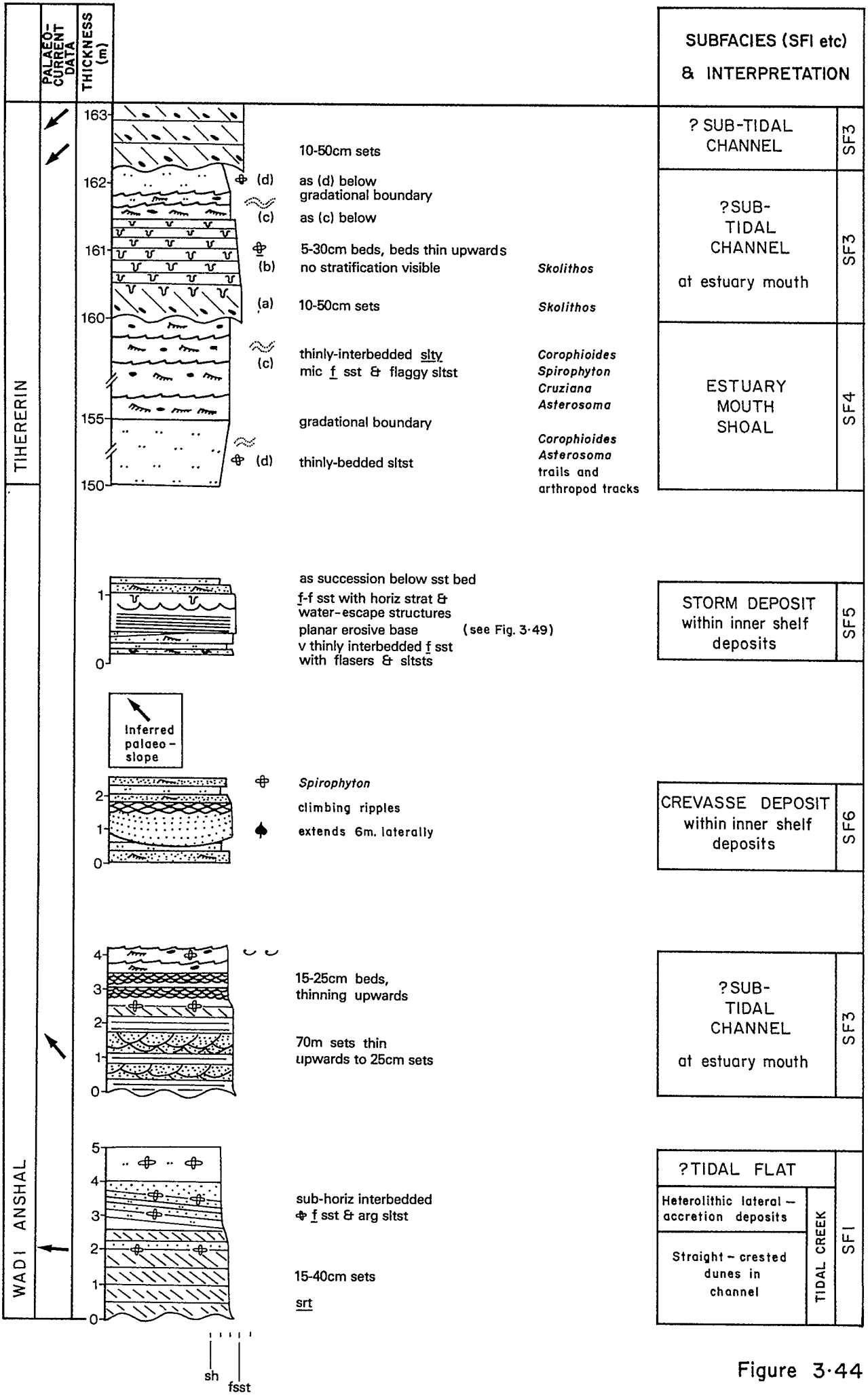

PALAEO-CURRENT DATA
THICKNESS (m)
SUBFACIES (SFI etc) & INTERPRETATION
TIHERERIN
WADI ANSHAL
163
162
161
160
155
150
(d)
(c)
(b)
(a)
(c)
(d)
10-50cm sets
as (d) below gradational boundary
as (c) below
5-30cm beds, beds thin upwards
no stratification visible
10-50cm sets
thinly-interbedded silty mic f sst & flaggy siltst
gradational boundary
thinly-bedded siltst
Skolithos
Skolithos
Corophioides
Spirophyton
Cruziana
Asterosoma
Corophioides
Asterosoma
trails and arthropod tracks
? SUB-TIDAL CHANNEL
SF3
?SUB-TIDAL CHANNEL at estuary mouth
SF3
ESTUARY MOUTH SHOAL
SF4
as succession below sst bed
f-f sst with horiz strat & water-escape structures
planar erosive base (see Fig. 3·49)
v thinly interbedded f sst with flasers & siltsts
STORM DEPOSIT within inner shelf deposits
SF5
Inferred palaeo-slope
Spirophyton
climbing ripples
extends 6m. laterally
CREVASSE DEPOSIT within inner shelf deposits
SF6
15-25cm beds, thinning upwards
70m sets thin upwards to 25cm sets
?SUB-TIDAL CHANNEL at estuary mouth
SF3
sub-horiz interbedded f sst & arg siltst
15-40cm sets
srt
?TIDAL FLAT
Heterolithic lateral-accretion deposits
Straight-crested dunes in channel
TIDAL CREEK
SF1
sh
fsst

Figure 3·44

Figure 3.45 SEDIMENTARY STRUCTURES OF FACIES RBSdSt

Upper : discordant bedding relationships in interbedded siltstones and cross-laminated, very fine-grained sandstones, some of which display wave-ripple cross-lamination and evidence for the former presence of intraformational clay clasts.

Lower left : small loaded channel of very fine-grained sandstone lying within a succession of interbedded *Arthrophycus* and *Cruziana*-bearing sandstones and siltstones. The upper beds of the channel are bioturbated and the sandstones adjacent to the channel provide evidence of current and wave-ripple cross-lamination.

Lower right : succession of beds of facies RBSdSt near the top of transgressive sequence 1 in which the two sedimentary structures illustrated above are present, locality 8, Wadi Teshuinat (Fig. 3.5). The over-hanging wall is composed of the basal sandstones of transgressive sequence 2.

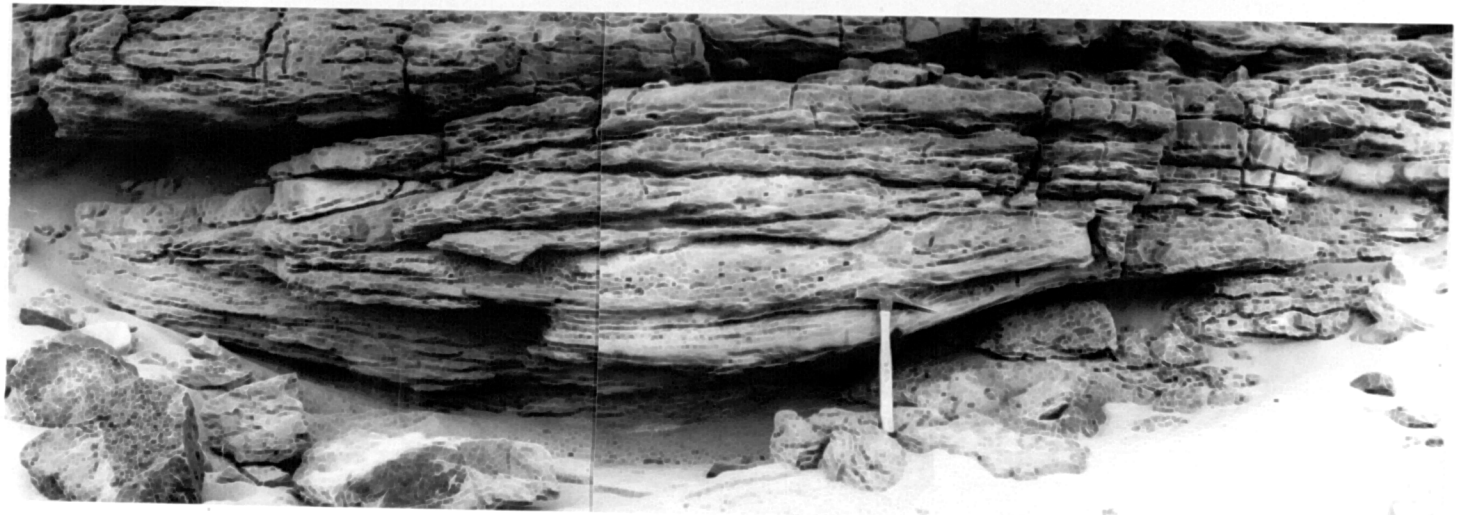


Figure 3.46 MEASURED SECTIONS OF FACIES VBSd AND RBSdSt AT WADI TESHUINAT, WADI ANSHAL AND TIHERERIN

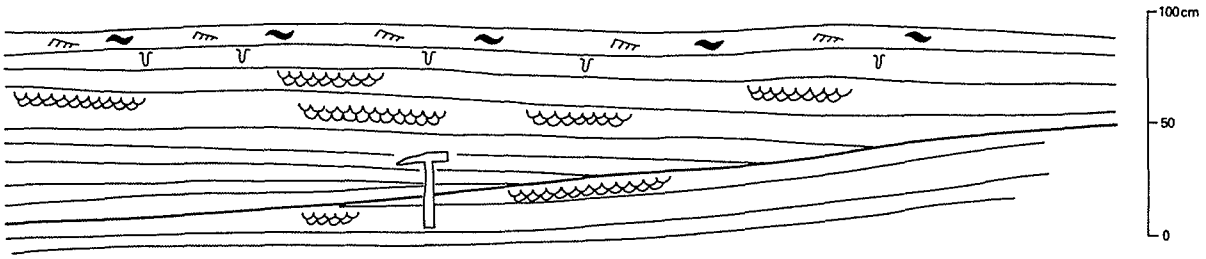
Notes:

1) The section through facies VBSd is an idealised vertical succession based upon measured sections through the sediments of the upper part of transgressive sequence 3 at Wadi Anshal and Tihererin (180 to 210 m, and 207 to 223 m stratigraphically above the base of the Tadrart Formation, respectively) at localities 12 and 17 respectively (Fig. 3.5).

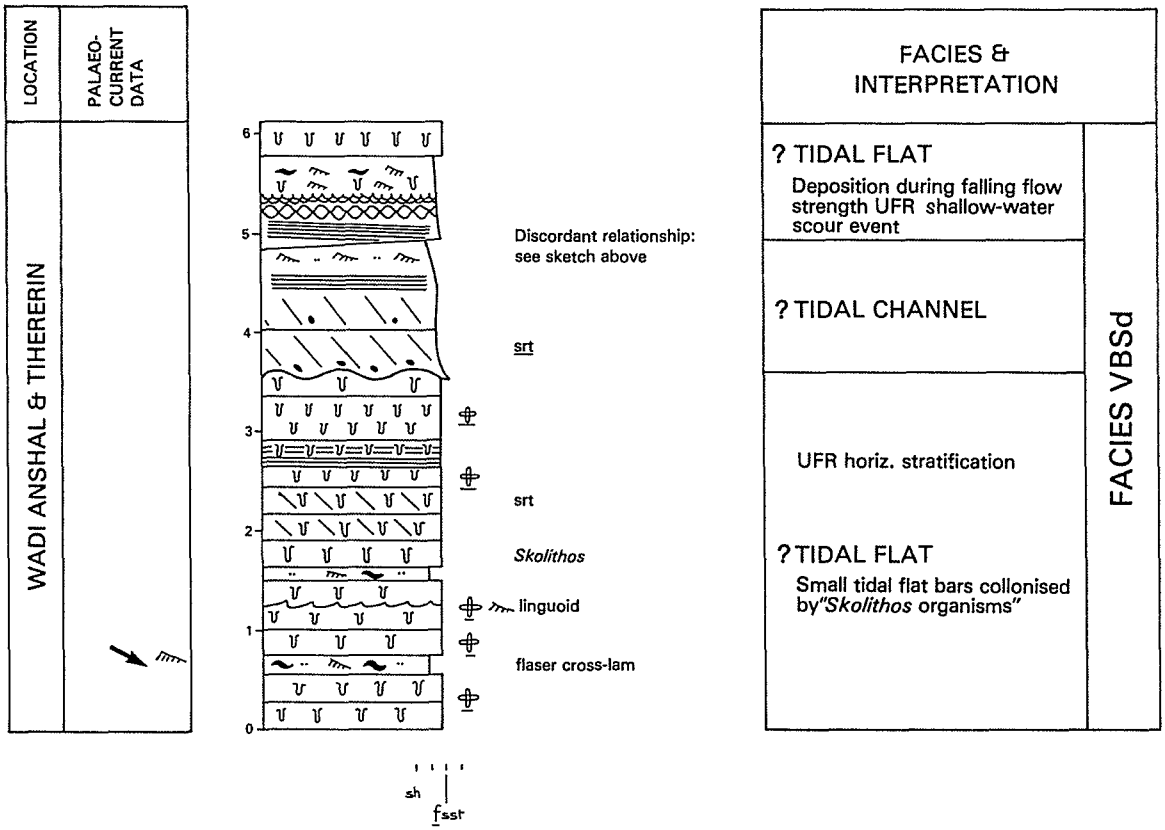
2) The sections measured through facies RBSdSt are actual bedding sequences from the upper part of transgressive sequence 4 (228 to 317 m stratigraphically above the base of the Tadrart Formation) from locality 2, Wadi Teshuinat (Fig. 3.19).

3) The palaeocurrent readings were measured from trough axes and foreset dip azimuths of cross-strata, except where a ripple sign is shown.

4) Abbreviations: UFR = upper flow-regime.



Sketch of discordant bedding relationship of slty f ssts in facies VBSd at Tihererin



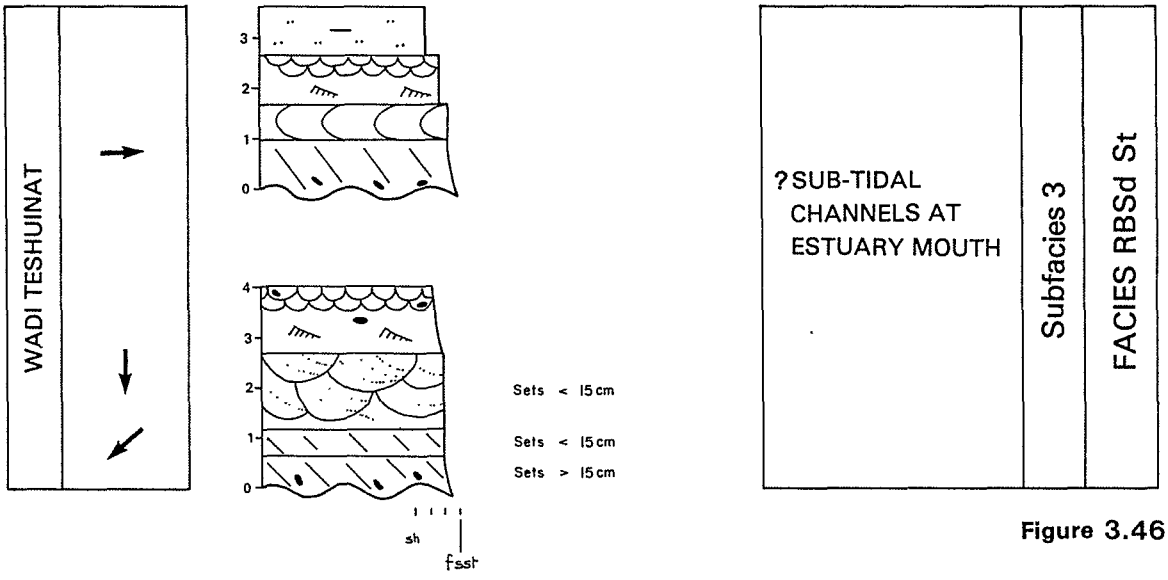


Figure 3.46

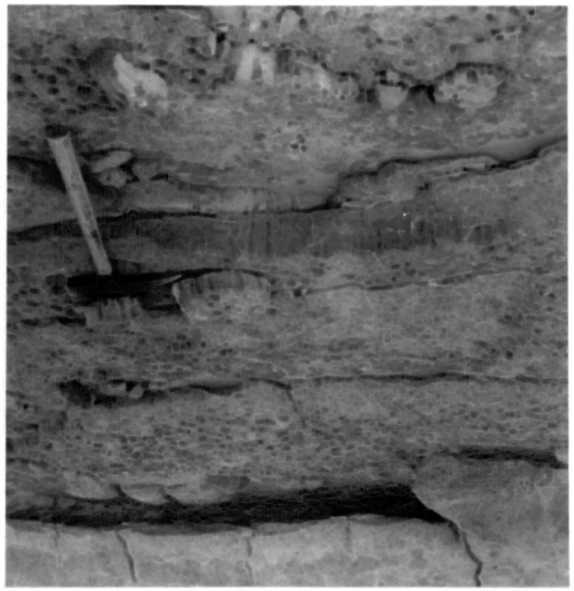
Figure 3.47 SEDIMENTARY AND BIOGENIC STRUCTURES OF FACIES RBSdSt AND VBSd

Scales: lens cap, 5 cm diameter; hammer, 30 cm long.

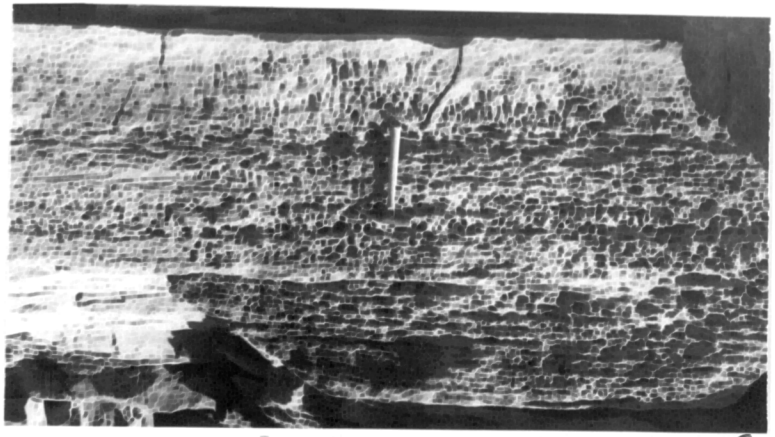
(A, B, D) Upper left and right, and lower middle : vertically-burrowed, fine-grained sandstone characteristic of facies VBSd. Internal stratification is occasionally visible in these beds, usually small-scale planar cross-stratification (upper right).

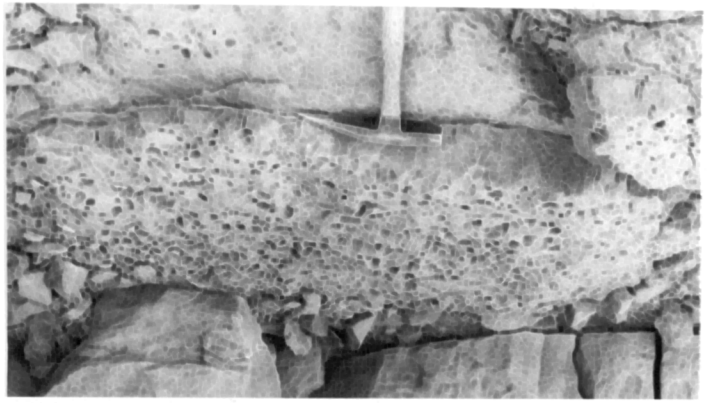
(C) Lower left : ferruginised intraformational clay clast prints in fine-grained sandstone of facies RBSdSt.

(E) Lower right : bedding sequence characteristic of subfacies 1, facies RBSdSt, and composed of alternating beds of cross-laminated sandstone, 8-20 cm thick, and thinner beds of argillaceous siltstone. Nearly in-phase climbing ripple, wave-ripple and linguoid-ripple cross-lamination is present. Some of the sandstone beds are strongly vertically-burrowed.



PENTAX





impersistent. Two varieties of the alternating sandstone/siltstone bedding sequences of this subfacies were observed in which evidence is provided for alternating flow-strength conditions. In the first, cross-laminated and homogeneous sandstone beds, of up to 20 cm thickness and of lenticular appearance in exposure, alternate with thin argillaceous siltstones of up to 5 cm thickness. This variety is characterised by the presence of wave-formed rippled surfaces on some of which *Cruziana* traces were observed. The second variety comprises stacked sequences of horizontally-stratified, and wavy-bedded, very fine-grained sandstone beds, of up to 25 cm thickness, which grade up to ripple-laminated silty sandstone beds with rare *Cruziana*.

3.4.1.3 Subfacies 2

Flaggy alternations of bioturbated, thin and very thin beds (up to 10 cm thick) of cross-laminated, very fine-grained, silty sandstone and horizontally laminated siltstone and silty shale have been assigned to subfacies 2 (Fig. 3.41). Thicker beds of trough cross-laminated and planar cross-stratified sandstone are commonly present, interbedded with the other sediments of this subfacies, some of which display clay-clast lags, rippled surfaces and bioturbation on their upper surfaces. Intraformational clay clasts are commonly present, also, in all the sandstone beds of this subfacies (Fig. 3.47, lower left).

3.4.1.4 Subfacies 3

Channel-fill bedding sequences of sandstones and siltstones lying within channel scours have been assigned to subfacies 3 (Fig. 3.41). The channel floor commonly contains *Skolithos* burrows and is generally overlain by a thin cross-laminated siltstone bed. The bedding sequences in the channel vary considerably, but are all characterised by an upward-fining grainsize trend. The sequence may consist of cross-laminated beds passing upwards into beds displaying soft-sediment deformed stratification (Figs. 3.41 & 3.43, middle and lower right), or it may be characterised by alternations of trough cross-stratification and horizontal stratification in the lower part, passing upwards to bioturbated and cross-laminated sandstones containing shell fragments in the upper part (Fig. 3.44, lower). Typically, however, the bedding sequence of this subfacies is as shown in Figure 3.44 (upper) where bedding types (a) and (b) are dominated by *Skolithos* burrows,

and bedding type (c) by complex internal feeding burrow systems (*Asterosoma* and *Spirophyton*), U-shaped sediment-feeding traces (*Corophioides*, see Fig. 2.8) and trails and tracks including *Cruziana*. At the top of the succession siltstones are present (bedding type (d)) in which the same trace fossil assemblage is present except for *Cruziana* and *Spirophyton*. The cross-stratified, fine-grained sandstones of bedding type (a) may not be present at the base of the sequence or they may be present in a number of sets. Where present they are characterised by clay intraformational clasts. The strongly-bioturbated sandstone and silty sandstone beds of bedding type (b) commonly decrease in thickness upwards and are overlain by the alternating very fine-grained sandstones and flaggy siltstones of bedding type (c). The sandstones are typically very silty and micaceous, and are present in cross-laminated beds with rippled and bioturbated top surfaces. The beds of bedding type (c) are characteristically 5-15 cm thick and commonly contain clay intraformational clasts. These beds pass upwards gradationally into the flaggy siltstones and subordinate silty shales of bedding type (d).

Other typical bedding sequences of this facies are shown in Figure 3.46 (lower) where trough and planar cross-stratified sandstones pass upwards to cross-laminated sediments. Figure 3.47 (upper right) illustrates the character of the *Skolithos* beds (bedding type (b)). This subfacies is similar, therefore, to subfacies 3 of facies CSdSt, differing only in being always entirely associated with beds of facies RBSdSt, containing a greater diversity of ichnogenera, and being composed, generally, of finer-grained sediment. Some large-scale examples of lenticular sandstone bodies of this facies, however, are present in which medium to coarse-grained sandstones are present (Fig. 3.48; see also Fig. 3.32, upper, for an example of subfacies 3, facies CSdSt).

3.4.1.5 Subfacies 4

Coarsening-upwards sequences of bioturbated siltstones to very fine-grained sandstones (Fig. 3.44) which may pass upwards to homogeneous units of strongly-bioturbated, well-sorted sandstone have been assigned to subfacies 4 (Fig. 3.41). Some of the same bedding types recognised in subfacies 3 are also present in this subfacies (bedding types (c) and (d)) and are associated with the same trace fossil assemblages (Fig. 3.44). The sandstone bodies in which this subfacies was recognised are commonly lenticular and laterally impersistent in outcrop.

Channel sandstone body lying within strongly bioturbated sandstones and siltstones of facies RBSdSt, transgressive sequence 4, near the top of the measured section at Bir Talauaut (Fig. 3.5). The sandstone body, which is 6 m thick and is composed of medium to coarse-grained sandstone, is interpreted as a subtidal channel. The stratification of the channel sands is dominated by trough and scour and fill cross-stratification, the sets of which are typically 10-50 cm thick and, on their upper surfaces, display bioturbation structures and clay clast prints. The foreset dip azimuths indicate palaeo-currents orientated southeastwards (116° to 121°). A felt-tip pen was used to emphasize the cross-strata.

Figure 3.48 CHANNEL SANDSTONE BODY OF SUBFACIES 3

3.4.1.6 Subfacies 5

Horizontally-stratified beds in which water-escape structures (Fig. 3.49) and *Skolithos* burrows (Fig. 3.47) are generally present, and which commonly lie above an angular, planar scoured surface, have been assigned to subfacies 5 (Fig. 3.44). This subfacies is only rarely present.

3.4.1.7 Subfacies 6

Scour and fill/channel fill cross-stratified sandstones that pass upwards to climbing-ripple cross-laminated sandstones, and siltstones (Fig. 3.44), have been assigned to subfacies 6. This subfacies is generally present in lenticular sandstone bodies, and is only rarely present.

3.4.1.8 Palaeocurrent data

A limited number of palaeocurrent readings were obtained from the sediments of facies RBSdSt. In transgressive sequence 4 the palaeocurrent orientations derived from asymmetrical ripples created a bipolar palaeocurrent rose and those derived from cross-strata showed a very great variance including palaeoflows to all four points of the compass (see section 3.7, Fig. 3.59). Since the majority of palaeocurrent data obtained for this facies was derived from sediments of transgressive sequence 4, in which the uniformity of palaeocurrent orientations characteristic of transgressive sequences 1, 2 and 3 is absent, it is difficult to ascertain to what extent the palaeocurrents of this facies are similar to those of the coarser-grained facies with which they are associated in each of the transgressive sequences. The impression was gained, however, from the field sections measured in transgressive sequences 1, 2 and 3 that, whilst the palaeocurrent orientations obtained from both ripples and cross-strata of sediments of facies RBSdSt displayed a wider variance than those from cross-strata of facies CSdSt and MSd, the foreset dip azimuths of the cross-strata nevertheless showed a predominant palaeocurrent orientation that was similar to that of the coarser-grained facies, i.e. towards the northwest. The palaeocurrent directions obtained from current ripples in these sequences displayed a wider variance with directions probably testifying to onshore, offshore and along-shore currents. Of particular importance, however, was the observation of superimposed

Subhorizontal bed of homogeneous, fine to very fine-grained sandstone, of 40 cm thickness, with water-escape soft-sediment deformation structures. Moulds and casts of shells, and vertical burrows are present at the top of the bed (not visible) (see measured section, Fig. 3.44).

Detail of water-escape structure in the same bed as that illustrated above.

Figure 3.49 WATER-ESCAPE DEFORMATION STRUCTURES IN SUBFACIES 5

rippled sandstone beds displaying opposite palaeocurrent directions. Although the palaeocurrent data of transgressive sequence 4 are more difficult to interpret, it may be noted that a certain similarity exists between the palaeocurrent rose derived from facies CSdSt (Fig. 3.58) and that of facies RBSdSt and VBSd (see Fig. 3.59).

3.4.1.9 Biogenic structures

A much greater variety of trace fossils is present in facies RBSdSt than in any of the other facies of the External Tassilian Sandstones (Fig. 3.6), the majority of the trace fossils that are characteristic of the *Cruziana* ichnofacies being present (cf. Table 2.6). As in facies CSdSt, the trace fossils may be divided into those that occur in relatively thick sandstone beds and those that occur in interbedded sandstone and siltstone successions. In this facies a further subdivision was recognised, slightly different trace fossil assemblages characterising thin beds of siltstone, and successions of thinly-interbedded very fine-grained sandstone and siltstone. These three trace fossil assemblages were best recognised in the sediments of subfacies 3 and 4, in which bedding types (b), (c) and (d) contain the trace fossil assemblages typical of thick sandstones, interbedded sandstones and siltstones, and of siltstones respectively (Fig. 3.44, upper). In addition to the trace fossils commonly found in the sediments of bedding type (c), *Halopoa* and *Sagittichnus* may also be present, whilst *Scolicia* (see Fig. 2.4) and the palm-frond variety of *Arthrophycus* (see Fig. 2.6) were found in these beds in transgressive sequence 1 only.

3.4.1.10 Petrography

The sandstones and siltstones of this facies are, like those of facies CSdSt, composed of quartz arenites with a range of textures, poorly-sorted samples being as common as well-sorted samples. Well-rounded and spherical grains are only rarely present. Some clay matrix was probably originally present but not in sufficient quantity to class the sediments as quartz wackes. The amount of depositional matrix is difficult to determine since much of the original clay has been diagenetically altered. The sediments are rich in detrital biotite and muscovite but feldspar is rarely present. A heavy mineral assemblage of detrital zircon, ilmenite and tourmaline, and the

following authigenic minerals, chamosite, haematite, ilmenite and geothite are commonly present (Fig. 3.50). Silica dissolution and silica replacement structures are common, as is haematite cementation and replacement of unstable detrital grains, but quartz overgrowths are not so abundant as in facies CSdSt. This is attributed to the greater degree of early diagenetic mineral growth resulting from the alteration of depositional clay matrix.

As well as having a greater percentage of depositional matrix than facies MSd and CSdSt, this facies differs from the coarser-grained facies in having more abundant authigenic chlorite (confidently identified as chamosite in one case, see Fig. 3.50) and in containing very small quantities of spherulitic fibrous kaolinite. Amorphous interstitial chlorite as well as chloritic clay rims on quartz grains are present, whilst the kaolinite is only present where replacing mica. These authigenic minerals suggest that the pore fluids may have had a slightly different acidity than those of the coarser-grained facies. Although this may be some evidence for marine influence, the presence of poorly-sorted samples suggests that these sediments have not been subjected to extensive marine transport and/or sorting processes.

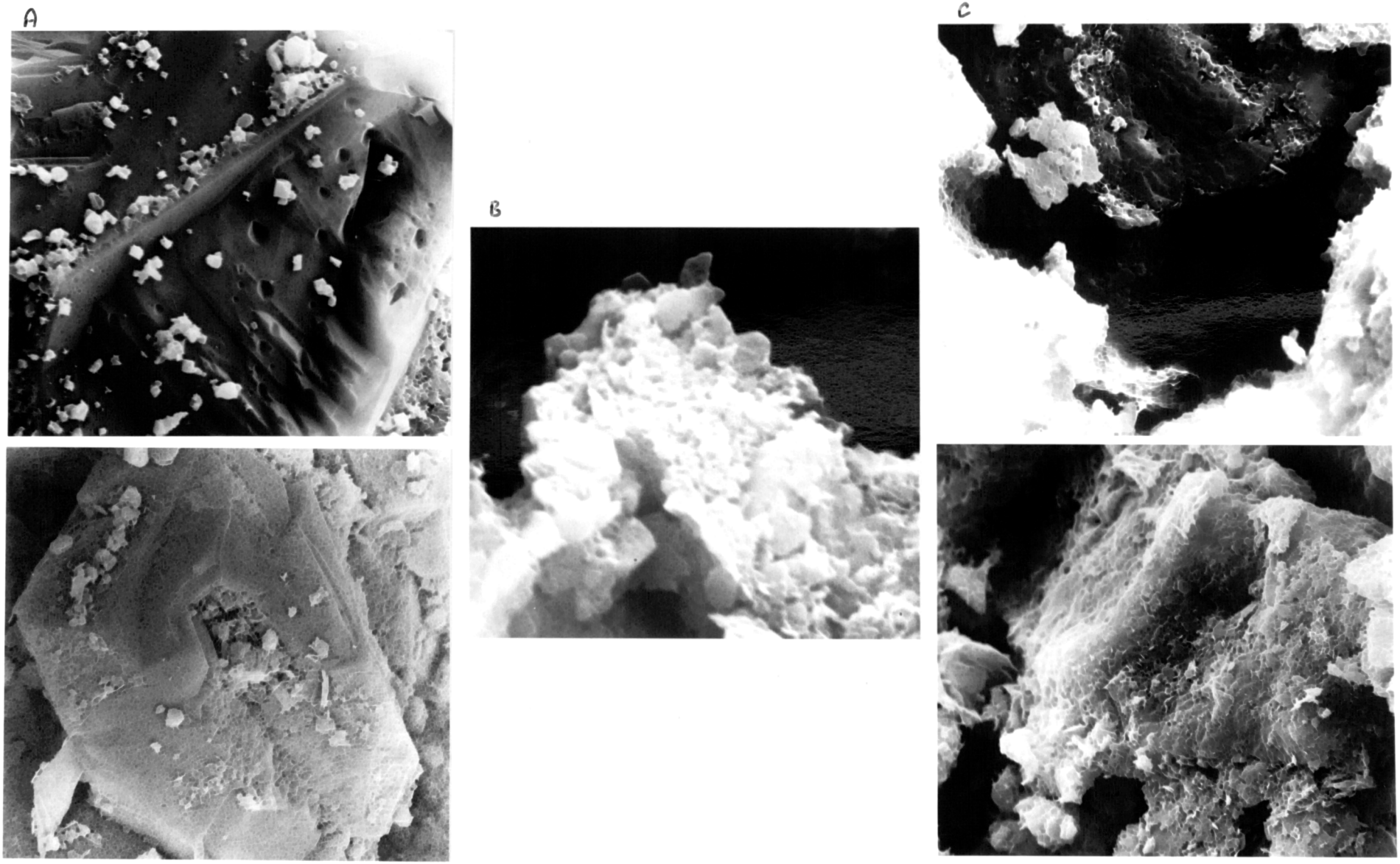
3.4.2 Interpretation

The presence in this facies of plant remains (previously reported in these beds by Lejal-Nicol and Massa, 1980), shell fragments, shell moulds and casts, trace fossils of Seilacher's *Skolithos* and *Cruziana* ichnofacies (see Table 2.6) and texturally-variable but mineralogically-mature sediment in heterolithic beds, provides evidence for a marine depositional environment characterised by spasmodic fluvial detrital supply and the variable flow-strength conditions of shallow-marine environments and particularly tidal regimes (Klein, 1977; Elliott, 1978b). The abundance of current-formed asymmetrical rippled surfaces and cross-lamination, the relatively rare occurrence of wave ripples, and the wide dispersal of palaeocurrent orientations from which clear evidence of bipolar palaeocurrent distributions were sometimes observed, are thought to provide evidence that tidal/wind/storm action predominated over wave action in an environment characterised by a greater variety of flow directions than that of the estuarine channel environment of the associated facies CSdSt. The

Figure 3.50 ELECTRON MICROGRAPHS OF SANDSTONES FROM FACIES RBSdSt

(A, D) Upper and lower left : micrographs of features of an argillaceous very fine-grained sandstone from near the top of transgressive sequence 1, Wadi Teshuinat.
Upper left: silica dissolution surface on angular quartz grain covered with rhombs of authigenic haematite and ilmenite (diameter of micrograph is 40 microns, magn. x 2400).
Lower left: irregular quartz overgrowths on sand grain. The depression in the centre may be the result of inhibition of quartz overgrowth development by the presence of early diagenetic haematite or chlorite (magn. x 1000).

(B, C, E) Upper and lower right, and middle : micrographs of features of a white, micaceous siltstone to very fine-grained sandstone from a bioturbated bed, 7 cm thick, near the top of transgressive sequence 4, Tihererin.
Upper and lower right: authigenic chamosite on irregular grain surfaces (magn. x 2400, upper; x 2000, lower).
Middle: detail of authigenic chamosite.


A
B
C
D
E

presence of abundant clay-clast intraformational conglomerates, plant remains and some texturally-immature sediment, provide evidence that some, at least, of the fluvially-derived sediment was not extensively reworked by tidal or other currents (Elliott, 1978b). The characteristics of this facies, and of associated facies CSdSt, suggest, therefore, that fluvial currents played an important role in creating the sedimentary structures of these two facies, those of facies RBSdSt being more strongly controlled by nearshore marine currents than those of facies CSdSt. It is considered that the sediments of facies RBSdSt provide sufficient evidence for fluvial detrital supply and limited current reworking, however, to indicate that the larger sand bodies of this facies, such as some of those of subfacies 3, are unlikely to have been deposited as offshore sand shoals or bars (Spearing, 1975).

The alternation of finer and coarser-grained beds that characterises subfacies 1, in which sedimentary structure and biogenic evidence of alternating high and low-energy flow conditions exists, is thought to provide evidence for periodic lowering of wave base in a shallow-marine environment and consequent increase in flow-strength at the sediment-water interface. This may have been related to tidal or storm action. The cross-stratified sand bodies within these sequences may have been deposited as tide-driven, shallow-marine dunes or bars (Johnson, 1978), although wind and storm action could also have been responsible.

The palaeocurrent data, the close spatial and temporal association of current-formed structures indicating bi-directional flow, and the presence of the following structures in this facies: scoured surfaces, cross-stratification with sharp set boundaries, clay-clast conglomerates, flaser and wavy cross-lamination, load casts and soft-sediment deformed bedding, provide evidence, especially when taken together with the characteristics of subfacies 4 of facies CSdSt (see sections 3.3.1.5 & 3.3.2.5), for the following processes, processes typical of tidal environments: bedload transport with bipolar-bimodal reversals of flow direction, time-velocity asymmetry of current bedload transport, scouring events, alternation of current bedload transport with suspension settlement during slack-water periods, and differential compaction, loading and hydroplastic readjustment (Dott & Howard, 1962; Klein, 1970, 1977; Reineck, 1963; Reineck & Wunderlich, 1968; Williams, 1960). The absence of evidence for emergence, exposure

and evaporation, such as etch marks, washout structures, interference ripples, flat-topped and double-crested ripples, runzel marks and desiccation cracks (Evans, 1965; Klein, 1977; Reineck & Singh, 1973) suggests that the majority of the sediments of this facies were deposited subaqueously rather than subaerially. Although Desio (1940b) recorded the presence of mud-cracks within the sediments of the Tadrart Formation, this observation has not been supported since. It is proposed, therefore, that a reasonable interpretation for these sediments is that they were deposited predominantly in a tidally-influenced nearshore environment. However, it should be stressed that the evidence indicates that the environment was one dominated by spasmodic but strong fluvial detrital supply and by fluvially-derived currents, suggesting the presence of "fingers" of estuarine channel sands extending into this environment. It should also be stressed that the somewhat limited palaeocurrent data from this facies does not display an unequivocal bipolar current pattern, although some examples exist, but rather a wide range of orientations suggesting the activity of long-shore and wind/storm induced currents, as well as tidal and fluvial currents. The annotations on Figures 3.41, 3.44 and 3.46 should be considered, therefore, as possible interpretations that, on the basis of the data available, are thought to be the most reasonable.

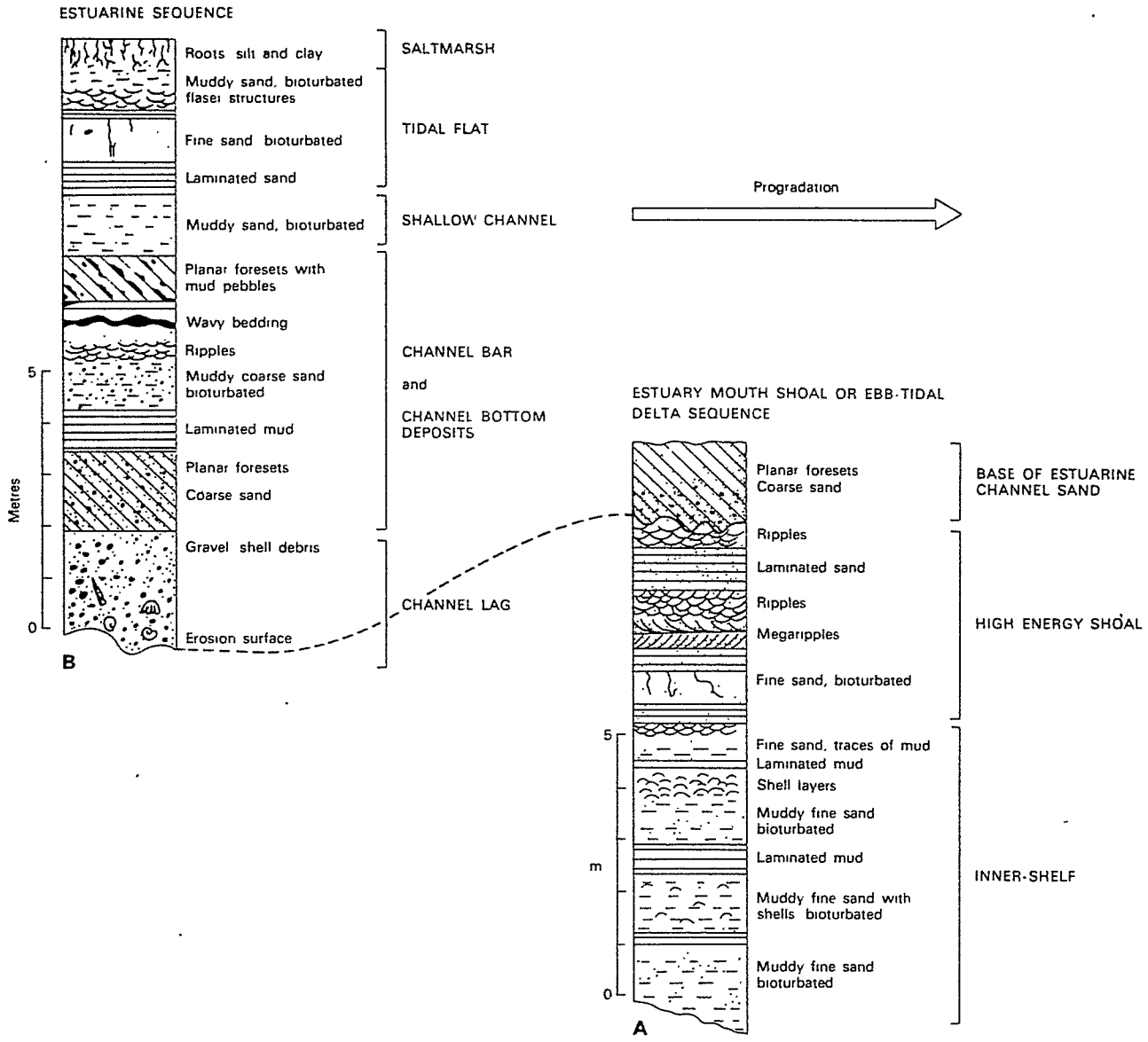
The presence of a diverse assemblage of trace fossils belonging to Seilacher's *Cruziana* ichnofacies (see Table 2.6) and concentrated largely within the thinly interbedded heterolithic sandstone, siltstone and subordinate silty shale sequences, is thought to provide support for a shallow-water marine environment (Seilacher, 1967, 1969), especially as *Scolicia*, one of the trace fossils present, although only present in transgressive sequence 1, is thought to be the trail of a gastropod (Hantzschel, 1975), an organism that first occupied non-marine habitats in the Carboniferous (Clarkson, 1979). The absence of trace fossils other than *Skolithos* in the thick sandstone beds (bedding type (b), see Fig. 3.44) is thought to be a reflection of discontinuous sedimentation in severe environmental conditions in which only suspension-feeding organisms, with burrows that offered protection from scouring currents, could survive, and also a response to substrate type in which the general absence of nutrient ensured that only a filter-feeding mode of existence was viable (Goldring,

1964; Hallam & Swett, 1966; Howard, 1975). On the other hand the differences in the trace fossil assemblages of bedding types (c) (thinly interbedded sandstones and siltstones) and (d) (siltstones) (see Fig. 3.44) are thought largely, to reflect a response to sediment substrate type. No evidence was found to support the proposed existence (Seilacher, 1969) of two separate trace fossil assemblages within the *Cruziana* facies, characterised by the type ichnogenera *Asterosoma* and *Spirophyton* (see Table 2.6), and thought to reflect differences between subfacies within a shallow-marine depositional environment. However, the differences that were recognised between the trace fossil assemblages of facies RBSdSt and CSdSt are thought to reflect differences in salinity and flow-strength conditions between shallow-marine relatively quiet-water environments, and high-energy estuarine/?tidal channel environments.

Considerable effort was devoted in the field to an attempt to recognise within the sediments of facies RBSdSt vertical bedding and trace fossil sequences such as those described from modern and ancient shallow-water shelf and shoreline environments (Goldring, 1962, 1964, 1971; Goldring & Bridges, 1973; Howard, 1972; Howard & Reineck, 1972). Neither the sedimentary structures nor the sequences of trace fossil assemblages typical of bedding sequences generated by migrating beaches or barrier islands was recognised (Howard, 1972; Howard & Reineck, 1972; Davis *et al.*, 1972), in particular the low-angle "swash cross-stratification" that characterises foreshore deposits of beaches (Reineck & Singh, 1973, pp. 93-94) was completely absent. Three types of vertical transitions associated with trace fossil assemblages were recognised, however, at very different scales, two small-scale varieties and one large-scale. The first small-scale variety comprises beds of subfacies 1 with erosional basal boundaries and an upwards increase in biogenic structures, the lower part of the bed commonly being characterised by horizontally-stratified sandstone, and the top of the bed, which is intensely bioturbated, being truncated. These beds are thought to result from storm scouring activity, which erodes the sea floor and resuspends sediment during the storm. This is followed by redeposition as the storm subsides and the re-establishment of an infauna during a period of reduced sedimentation (Goldring & Bridges, 1973; Howard, 1975). The presence of hummocky cross-stratification in some of the sandstones is

additional evidence for storm activity. Although these beds are not uncommon in this facies it is also common to find similar beds in which the preserved original sedimentary structure at the base of the bed comprises cross-lamination, and the upper part of the bed is overlain by a thin silty clay bed with lenticular lenses of sandstone and siltstone. These beds, which comprise the second vertical-transition type recognised, are thought to result from the scouring, variable flow-strengths and the discontinuous deposition of tidal rather than storm currents (Klein, 1977).

The larger-scale variety of vertical bedding sequence, in which a vertical transition of trace fossil assemblages was present, was recognised in the sediments of subfacies 3 and 4 (Fig. 3.44, upper). In these sediments upward-coarsening vertical bedding sequences of 6-12 m thickness exist in which thinly-bedded siltstones, heterolithic beds and overlying cross-laminated siltstones and sandstones, all displaying a diverse ichnofauna, pass upwards to sandstone beds with abundant *Skolithos* burrows and no other recognisable trace fossils. Planar cross-stratification is visible in these *Skolithos* beds, in some cases, and a basal scour surface is generally present. The *Skolithos* sandstones pass upwards in an upward-fining sequence of, typically, a few metres thickness (2-6 m), into the same heterolithic beds with a diverse ichnofauna described from lower in the sequence. These bedding sequences are thought to have been formed by the progradation of an estuary mouth shoal and an overlying estuarine channel such as those described from the mesotidal Georgia Coast of the USA by Greer (1975) (Fig. 3.51). These deposits, therefore, are the distal equivalent of the estuary mouth bar sandstones of subfacies 4, facies CSdSt. In some cases the *Skolithos* sandstone of this subfacies is not erosively based and the vertical upward-coarsening succession may be compared with that described from inter-tidal and sub-tidal shoal sands in a mesotidal embayment of South Carolina (Reinhart & Ruby, 1982). The abundant examples of upward-fining sequences of subfacies 3, in which evidence for in-channel bar and dune migration exists (see Figs. 3.41, 3.44, 3.46), provide evidence for a predominance of channellised flow rather than open-shelf flow, and the lenticular geometry in exposure of many of the sandstone bodies of subfacies 3 is further support for deposition in localised ?tidal/estuarine channels.



Sequence produced by the progradation of an estuary mouth high-energy shoal (A), and an estuarine channel (B); Ossabaw Sound, Georgia Coast, USA (after Greer, 1975; in Reading, 1978).

Figure 3.51 POSSIBLE DEPOSITIONAL MODEL FOR FACIES RBSdSt.

The bedding sequences and sandstone-body geometry of the relatively rare sediments of subfacies 5 and 6 provide evidence for open shelf storm deposition and subaqueous crevasse deposition respectively (Fig. 3.44). The vertical bedding sequence of subfacies 5 (Fig. 3.44) and the water-escape structures of this subfacies (Fig. 3.49) are typical of the rapidly-deposited shelf sediments of storms (Johnson, 1978). The channel fill cross-stratification and climbing-ripple cross-lamination of subfacies 6 also provide evidence for rapid deposition but, in this case, the sedimentary structures are characteristic of crevasse-splays (Elliott, 1978a). The sediments of subfacies 6, therefore, display some similarities to those of subfacies 4 of facies CSdSt, sediments interpreted as the deposits of estuary-mouth bars and possibly crevasse splays (see sections 3.3.2.5 & 3.3.2.8). The association of both subfacies 5 and 6 with subfacies 1 is believed to provide evidence that not only were the storm deposits laid down on the inner shelf but that the crevasse-splay deposits resulted from crevassing of subaqueous levees of seaward channels and were also, therefore, laid down on the inner shelf. However, although the evidence suggests that the majority of the sediments of subfacies 1 were deposited subaqueously, the presence within this subfacies of heterolithic lateral-accretion deposits (Fig. 3.44, lower) that are characteristic of tidal creeks (Reineck & Singh, 1973) suggests that some of these deposits may have been laid down subaerially on tidal flats. Similarly, although no clear evidence exists for subaerial exposure in the sediments of subfacies 2, the presence of planar cross-stratified sandstone beds with bioturbated, rippled top surfaces displaying clay clast lags (Fig. 3.41) which, therefore, have many of the characteristics of inter-tidal sand bars (megaripples) (Klein, 1977; Dalrymple *et al.*, 1978) is thought to provide evidence that some, at least, of the sediments of subfacies 2 may have been laid down inter-tidally. Without further field work on these sediments it is impossible to establish with any greater degree of confidence whether inter-tidal deposits are present.

Beuf *et al.* (1971) studied sediments of a very similar facies to facies RBSdSt in the Algerian lithostratigraphic equivalents of the Tadrart Formation (Oued Karkai and Oued Samene formations) (see Table 1.3) and interpreted them as "transitional-marine" deposits, by which they meant that the sediments displayed both marine and non-marine characteristics. Among the non-marine characteristics they observed

were "fluviatile" cross-stratified and channelled sandstones with unidirectional palaeocurrent indicators, whilst the thinly-bedded sequences displayed a wide variety of palaeocurrent directions and "marine" ichnofacies similar to those described from facies RBSdSt. The "transitional-marine" deposits lay above "fluviatile" deposits in a number of large-scale fining-upwards sequences, the controls for which were considered by Beuf *et al.* (1971) to be a balance of rates of eustatic sea-level rise (possibly glacio-eustatic) and subsidence such that the sea transgressed over a broad, low-relief fluvial plain without the development of a landward-migrating coastline, the sea reworking some but not all of the sediment of the plain (see Fig. 3.40). The predominance of seaward-directed palaeocurrents within facies CSdSt and RBSdSt, indicating the strength of fluvial currents, even within shallow-shelf deposits bearing a full fauna of trace fossils of the *Cruziana* ichnofacies, is a most striking feature of these facies which is of great importance in considering the hydrodynamic regime controlling the development of the facies of the Tadrart Formation (see section 3.7).

3.5 Vertically-Burrowed Sandstone Facies (VBSd)

3.5.1 Description

3.5.1.1 Sedimentary structures, trace fossils and palaeocurrent data

Within the rippled and bioturbated sandstone and siltstone facies, and characteristically towards the top of transgressive sequence 3, a succession of highly bioturbated fine and very fine-grained sandstones in beds 5 to 30 cm thick is present (Figs. 3.6 & 3.46, upper). This association of sandstones with subordinate cross-laminated siltstones is termed the vertically-burrowed sandstone facies on account of the abundant *Skolithos* burrows present within the sandstones (Fig. 3.47). Trace fossils other than *Skolithos* were not observed in this facies, and the regular alternation of sandstones and siltstones, characteristic of facies RBSdSt, is absent. The sandstones are so strongly bioturbated that internal stratification is generally not visible although in some beds, planar tabular cross-stratification and horizontal-stratification may be discerned, whilst on the surface of some beds linguoid ripples may be present. The sandstone beds are laterally persistent and the whole succession is generally well bedded, although these beds may sometimes be clearly subparallel to the regional

dip. Some thin to very thin siltstone beds are present displaying current-ripple and flaser cross-lamination. Channel-shaped sandstone bodies lie within the succession, often only 5 to 10 m wide, which are characterised by scour bases and a fining-upwards sequence of very well-sorted, fine-grained sandstones and cross-laminated siltstones (Fig. 3.46, upper). Fining-upwards bedding sequences from horizontally-laminated sandstones to cross-laminated siltstones in sequences of 40 to 100 cm thickness (Fig. 3.46, sketch) and larger-scale channels of fine to medium-grained, trough cross-stratified sandstone (Fig. 3.6) are also present in the succession.

The palaeocurrent directions obtained from current-rippled surfaces of sediments of this facies in transgressive sequence 3 are bipolar with northwest and southeast orientations.

3.5.1.2 Petrography

The sandstones of this facies are, typically, moderately-sorted quartz arenites with abundant authigenic chlorite, haematite and ilmenite in some samples (Fig. 3.52). Silica dissolution and precipitation structures are common but quartz overgrowths are not well developed, probably because of the original presence of a small percentage of clay matrix which altered during an early diagenetic phase and masked the quartz grain surfaces. Detrital ilmenite, muscovite and biotite are present but feldspar is very rare.

3.5.2 Interpretation

This facies lies within facies RBSdSt, with which it shows some similarities, especially the presence of flaser cross-lamination in thin siltstone intercalations, and bipolar palaeocurrents, characteristics of the fluctuating energy conditions typical of tidal environments (Klein, 1977; Reineck & Wunderlich, 1968). The petrography of the sediments of the two facies is also very similar, providing some sparse evidence for a marine environment, and good evidence for a limited degree of current reworking of the sediment (see section 3.4.1.10). The abundance of *Skolithos* vertical burrows, and the absence of other trace fossils, in the strongly-bioturbated sandstones of this facies, indicates, as in facies RBSdSt, that these sandstones were deposited rapidly, they contained limited nutrient therefore, and were consequently only occupied by filter-feeding organisms. Further, the

Figure 3.52 ELECTRON MICROGRAPHS OF SANDSTONE FROM FACIES VBSd

Note: micrographs are of features in a moderately-sorted, fine-grained sandstone, with abundant green to brown interstitial authigenic chlorite, from a bed with numerous *Skolithos* burrows, lying near the top of transgressive sequence 3, at Tihererin.

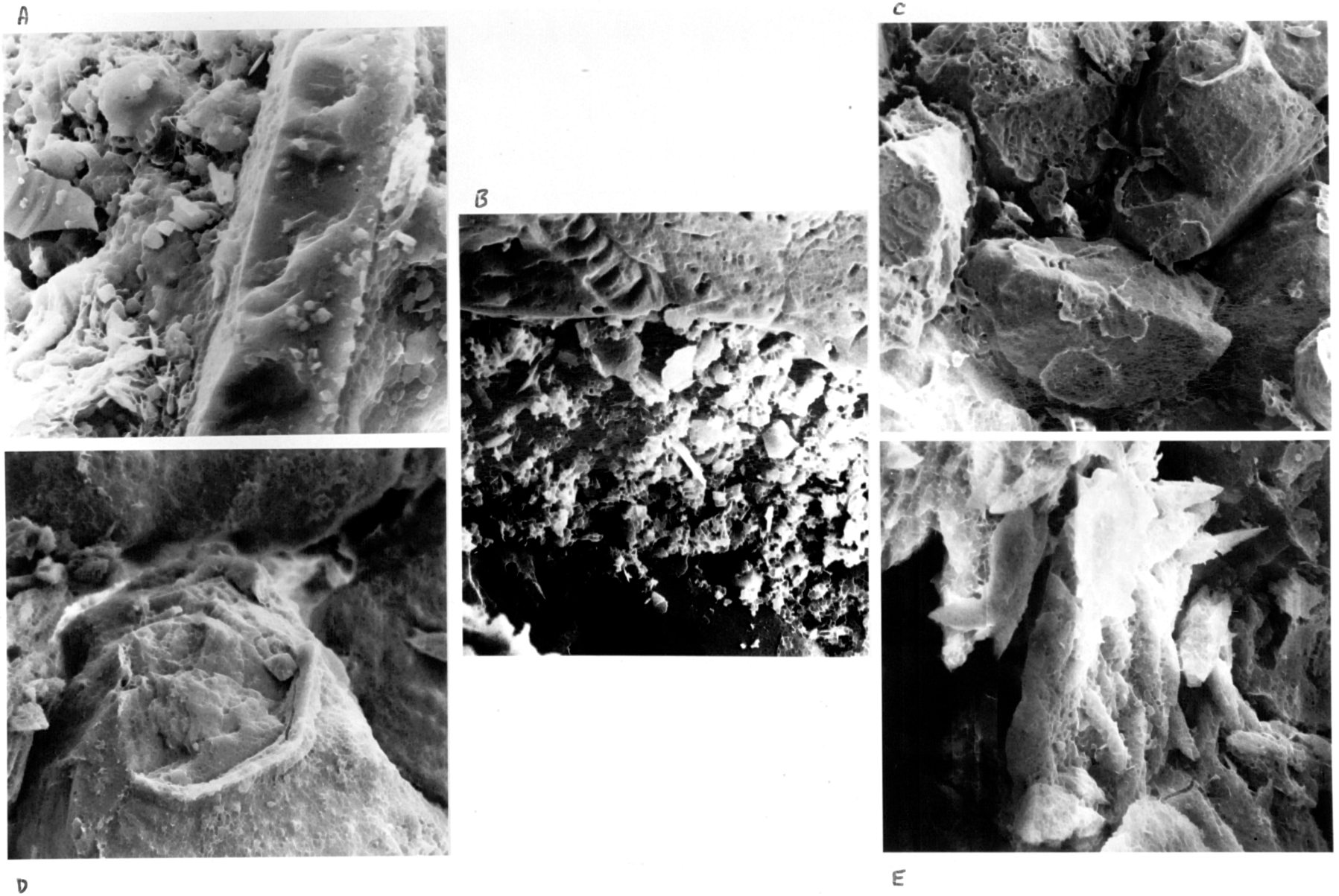
(A) Upper left : tabular detrital grain of ilmenite (central right) surrounded by platey chlorite minerals (micrograph 45 microns diameter, magn. x 2000).

(D) Lower left : dissolution cavity in quartz grain (magn. x 600).

(B) Middle : grain surface with silica dissolution structures (upper) and precipitation surface of authigenic chlorite, haematite and acicular crystals of ilmenite (lower) (magn. x 2000).

(C) Upper right : characteristic quartz grains of facies VBSd. The absence of major quartz overgrowths is thought to be a result of the early presence of abundant diagenetic chlorite, haematite and ilmenite (magn. x 220).

(E) Lower right : interstitial chlorite (magn. x 1300).


A
B
C
D
E

restriction of the fauna to one vertical burrow ichnospecies, is thought to provide evidence for severe environments of periodic strong-current turbulence, scouring and/or subaerial emergence in which specially adapted organisms survived by occupying protected vertical tubes (Goldring, 1964; Hallam & Swett, 1966; Howard, 1975). These conditions apply to tidal sand flats (Evans, 1965; Klein, 1977) and to shoreface littoral environments (Howard, 1972; Seilacher, 1967). The subparallel bedding of the bioturbated sandstones may be the result of deposition on the inclined margins of tidal creeks (Reineck & Singh, 1973) or of deposition in an upper shoreface environment (Elliott, 1978b). The channels within the succession are interpreted as tidal channels of various sizes, whilst the horizontally-stratified sandstone beds are thought to provide evidence for deposition from shallow-water upper flow regime tidal currents on a scoured surface created by the same or an earlier strong tidal current. The vertical transition from these beds to ripple-laminated and bioturbated sediments is thought to reflect the waning flow conditions that succeeded the upper flow regime currents. Insufficient field work was undertaken on the sediments of this facies to establish which of the tidal-flat and the shoreface interpretations is the most valid. Similarly, on the basis of the field work undertaken, it has not been possible to identify the vertical fining-upwards sequences characteristic of some prograding inter-tidal deposits (Sellwood, 1972) or the more complicated vertical grainsize profiles of other inter-tidal deposits (Evans, 1975). Furthermore it is recognised that the inter-tidal flat and tidal channel interpretation represents only one of a number of possible interpretations, and the annotation to Figure 3.46 should be seen in this light. Particularly, the absence of evidence for subaerial emergence and evaporation weakens the proposed interpretation, since many of the structures observed could have been formed subaqueously and/or in energetic shallow-marine environments with no tidal influence. However, as noted in section 3.2.2 for fluvial deposits, the absence of evidence for subaerial emergence by no means proves that it did not occur, and the sedimentary biogenic structures of this facies do bear a strong resemblance to those of inter-tidal flats (Klein, 1977).

3.6 Facies of the Ouan Kasa Formation

3.6.1 Description

3.6.1.1 General

Two facies were recognised in the sediments of the Ouan Kasa Formation exposed at Wadi Wan Kasa (locality 18, Fig. 3.5): the interbedded shale, sandstone and siltstone facies (IShSdSt) and the massive shale facies (MSh) (Fig. 3.6). Three coarsening-upwards cycles are present of between 10 and 18 m thickness, in each of which facies MSh lies at the base and passes upwards gradationally into facies IShSdSt. Only a few palaeocurrent readings were obtained from this formation, from current-rippled siltstones and sandstones. These indicated currents directed both northwestwards and southeastwards. The shales, siltstones and sandstones of this formation contain a fauna of brachiopods, bivalves, styliolines and tentaculitids (Klitzsch, 1969).

3.6.1.2 Massive shale facies (MSh)

The massive shale facies comprises beds of green, grey and purple gypsiferous shales with thin, laterally-impersistent intercalations of very fine-grained sandstone that display sharp basal boundaries and gradational upper boundaries. Trails and other bioturbation traces are abundant on the upper surfaces of these beds which are commonly dissected by *Skolithos* burrows. Horizontal stratification is commonly developed in the lower part of these beds whilst current-formed cross-lamination predominates in the upper part.

3.6.1.3 Interbedded shale, sandstone and siltstone facies (IShSdSt)

A coarsening-upwards trend is discernible in this facies, siltstones predominating at the base, and sandstones at the top. Wave-ripple cross-laminated beds of siltstone, interbedded with thin shales and occasional very fine-grained sandstone beds, form successions of up to 5 m thickness at the base of the facies (Fig. 3.53) in which flaser and wavy cross-lamination, and sole tool marks are characteristic of some beds (Fig. 3.53). Above the siltstones, a succession of bioturbated, cross-laminated fine and very-fine grained sandstones is present in which desiccation cracks may be seen on some rippled surfaces. The thin beds display low-angle cross-stratification where the stratification has not been destroyed by bioturbation. Clay

Tool marks on the sole of a wave-ripple cross-laminated bed of very fine-grained sandstone. Shell fragments and trails are present on the rippled upper surface of this bed.

Alternating beds of cross-laminated ferruginous siltstone and shale. The succession is strongly bioturbated and vertical burrows are visible.

Figure 3.53 SEDIMENTARY STRUCTURES OF THE OUAN KASA FORMATION

clast prints are common in these sandstones whilst plant remains were found in some beds.

The sandstones and siltstones of facies IShSdSt are moderately to well-sorted and composed predominantly of angular to subrounded quartz grains with quartz overgrowths, and of detrital micas. Iron-oxide cementation is abundant with both haematite and limonite present. The micas are generally partly replaced by haematite. Authigenic chlorite is also present in some samples as well as some possible glauconite.

3.6.2 Interpretation

The presence of the following fossils in the Ouan Kasa Formation suggests an open-marine environment of deposition (Borghi & Chiesa, 1940): *Nucula corbuliformis*, *Spirifer corbuliformis*, *Chonetes orientalis* and *Strophomena carianata* (Fig. 3.6). The sediments of facies MSh are thought to have been deposited largely below wave base from suspension in an open-marine shelf environment where occasional storms were responsible, as a result of lowered wave base, for the presence of sharply-based, fining-upwards sandstone beds, with basal tool marks (Johnson, 1978). The coarsening-upwards cycles of facies IShSdSt reflect an upwards change from deposition below to deposition above wave base and suggest either a gradual upwards change in base level, or progradation of the depositional unit. This can occur as a result of eustatic changes or by progradation of estuary mouth shoals locally reducing water depth (Elliott, 1978b). The presence of intraformational clay clasts within the sandstones of facies IShSdSt is thought to provide evidence for seasonal reworking of very thin beds of fine-grained suspension deposits, and the desiccation cracks present in these sandstones indicate that progradation events and/or sea level changes were sufficiently pronounced to allow occasional subaerial emergence of the depositional surface. Although an open marine environment, therefore, is indicated in a general sense for deposits of the Ouan Kasa Formation, this evidence, as well as that of the plant remains and flaser cross-lamination, suggests that these deposits were laid down not far from the coastline in a region where sediment was derived from an estuary, and currents, probably of tidal origin, were directed primarily onshore and offshore. The sediments showing evidence of emergence may have been deposited on tidal flats.

3.7 Conclusions

3.7.1 Correlation between Composite Sections

3.7.1.1 General

In this study of the facies of the Tadrart and Ouan Kasa formations it was found that the sediments of the Tadrart Formation could be conveniently divided into thickly-bedded, relatively coarse-grained units, generally characterised at their base by horizons of Liesegang rings and iron-oxide cementation, and thinly-bedded, finer-grained units (Fig. 3.1). The thickly-bedded units are composed of sediments of the multi-storey cross-stratified sandstone facies and the channelled sandstone and siltstone facies, whilst the thinly-bedded units are composed of sediments of the rippled and bioturbated sandstone and siltstone facies and the vertically-burrowed sandstone facies (Fig. 3.6). It was observed that these facies are also present in the upper beds of the Acacus Formation (units 14 & 15, Fig. 3.1), which led to the identification of four sequences each of which was interpreted as being broadly transgressive, since the sequences are characterised by the presence of the coarser-grained facies at the base, interpreted as fluvial and estuarine channel complex deposits, and finer-grained facies, interpreted as shallow-marine and ?tidal-flat deposits, lying above (Fig. 3.6). The interbedded shale, sandstone and siltstone facies, and the massive shale facies of the Ouan Kasa Formation (Fig. 3.6) provide evidence for relatively open-marine depositional conditions, which, since this formation is thought to lie conformably above the Tadrart Formation, represent the culmination of the depositional processes responsible for the fourth transgressive sequence. Although the sediments of the upper beds of the Acacus Formation have been studied in less detail than those of the Tadrart Formation, these beds have been included in this study by reference to Klitzsch (1965a, 1969) and to a section measured by the author at Tanesfert (see Fig. 3.2). This was considered necessary because it was felt that any interpretation of the sediments of the Tadrart Formation that attempts to place them in the context of their lithostratigraphical equivalents in the Central Sahara must embrace all of the sediments of the External Tassilian Sandstones, since these form the only unambiguous lithostratigraphical unit recognisable throughout the Eastern Hoggar and Western Murzuq Basin in the Upper Silurian

to Lower Devonian strata (see Tables 1.3 & 3.1; Fig. 1.6; Freulon, 1964).

The thickly-bedded units of the four transgressive sequences (Fig. 3.6) were correlated between the composite sections at Bir Talauaut, Wadi Teshuinat, Wadi Anshal and Tihererin (Fig. 3.2) with some degree of confidence, as a result of the good exposure of these beds in the wadis and cliffs of the Jabals Akakus and Tadrart. The sandstones at the base of transgressive sequence 1 of the Jabal Akakus escarpment (see Fig. 1.12) and the sandstones of the "Sandstone Marker Bluff" at the base of transgressive sequence 4 (see Figs. 3.29 & 3.31, upper), are both particularly well exposed and may be traced laterally over large distances. In the north of the study area the correlation of these thickly-bedded sandstone units proved more difficult as a result of the complete truncation by the sediments of overlying sequences of some of the thinly-bedded units characteristic of the upper part of the transgressive sequences, resulting in the vertical juxtaposition of very similar thickly-bedded sandstones of different transgressive sequences. Nevertheless the sections measured in the north at Wadi Auis, Wadi Telia (north and south), Al Awaynat and Tanesfert (Fig. 3.2) have been tentatively correlated, by the recognition of major scour surfaces and horizons of Liesegang rings and iron-oxide cementation, with the major composite sections measured further south, in order to establish a lithostratigraphic correlation of all the sections measured in the External Tassilian Sandstones of the Jabals Akakus and Tadrart (Fig. 3.54). The resulting stratigraphic cross-section, which has been constructed using a datum at the base of the External Tassilian Sandstones, i.e. at the base of the thickly-bedded sandstones of transgressive sequence 1 (exposed in the Jabal Akakus escarpment and at Wadi Ahar; Figs. 3.5 &3.37), is orientated approximately north-south which is oblique to the inferred depositional strike. From palaeogeographical reconstructions, based on the facies studies and palaeocurrent data of each of the transgressive sequences (see section 3.7.4 below), the depositional strike for the first three transgressive sequences is inferred to have been northeast-southwest (with a sediment source area in the southeast), whilst the evidence for that of the fourth transgressive sequence is enigmatic, but is thought to indicate a depositional strike also aligned northeast-

southwest. Consequently the stratigraphic cross-section (Fig. 3.54) displays features characteristic of both dip and strike-sections. An example of the former phenomenon is the lateral decrease in thickness of the thickly-bedded sandstone unit of transgressive sequence 1 in a northward direction, thought to reflect increased distance from sediment source area. An example of the latter phenomenon is the interdigitation of sediments of facies CSdSt, and of facies RBSdSt and VBSd, particularly well developed in transgressive sequences 3 and 4, and thought to reflect lateral facies changes along depositional strike related to the position of major estuarine channel complexes lying within finer-grained shallow-marine deposits. The stratigraphic cross-section also illustrates that not only are the External Tassilian Sandstones truncated northwards by the Upper Devonian "Chatti Formation" and the Lower Carboniferous Mrar Formation (see Fig. 1.9), but their thickness is also controlled by three internal erosion surfaces, lying at the base of transgressive sequences 2, 3 and 4, the truncation effects of which are to some extent compensated for by thickness differences within sequence 4.

3.7.1.2 Control on stratigraphic thicknesses of measured sections

Accurate correlation of measured sections to form composite measured sections, and between composite measured sections to create a lithostratigraphic cross-section (Fig. 3.54), was dependent, in part, upon the accuracy of stratigraphic thicknesses measured (see section 3.1.2). A brief account of the thicknesses involved provides an idea of error limits present in the measured sections (see Fig. 3.5). The thickest virtually-continuous vertical succession was measured at Wadi Anshal (248 m). Further north, the Wadi Teshuinat composite section was constructed by the recognition of stratigraphic markers, from three separate measured sections in the lower, middle and upper Wadi Teshuinat. The correlation between those in the lower and middle Wadi Teshuinat was sufficiently uncertain that an error limit of ±20 m was established for the total estimated thickness of the section (318 m). In the south, the Tihererin composite section was the only section in which the uppermost beds of the Tadrart Formation were measured. Using the "Sandstone Marker Bluff" of transgressive sequence 4 (Fig. 3.54) as a marker, sections at Tihererin and Selfufet were correlated to establish a total thickness for the Tadrart Formation in this area of 331 m,

for which an error limit of ±30 m was established. The discrepancy between this thickness and that established by Klitzsch (1965a, 1969) for the type section of the Tadrart Formation measured in this area (317 m) is well within the estimated error limit. This encourages some confidence, therefore, in the thicknesses established for the composite sections illustrated schematically in Figure 3.54.

3.7.2 Environments of Deposition

From the interpretation of the six different facies recognised within the External Tassilian Sandstones (Fig. 3.6) and from their position within the four transgressive sequences identified within this succession the following observations may be made about the characteristics of the vertical sequences present within this succession. The deposits of each of the four transgressive sequences typically comprise fluvial and fluvial/estuarine sediments at the base, which pass upwards to shallow-marine and/or ?tidal-flat deposits. Estuarine channel sandstone deposits typically pass upwards to estuary mouth bar and high-energy shoal deposits which, in turn, pass upwards to inner-shelf deposits (see section 3.3.2.9). In transgressive sequence 4 the inner shelf deposits pass up to relatively open-marine deposits which display evidence for minor storm activity and occasional shallowing and emergence, possibly as a result of estuary-mouth shoal progradation. Transgressive sequence 2 displays the greatest evidence for continuous fluvial conditions, whilst in all the other sequences the fluvial deposits are interbedded with estuarine deposits (Fig. 3.54). On the other hand transgressive sequence 4 is characterised in the three southerly composite sections by the presence, at the base, of the "Sandstone Marker Bluff", a sandstone body generally containing no fluvial deposits but composed of estuarine transverse bar and seaward channel deposits in which estuarine scour surfaces are prominent and abundant evidence of marine influence is displayed in the form of shell fragments and bioturbation (see Fig. 3.31).

It has been suggested, in the proposed interpretations of the sedimentary structures of facies CSdSt, RBSdSt and VBSd (sections 3.3.2, 3.4.2 and 3.5.2, this Chapter), that the estuarine channels and the inner-shelf environment may have been subjected to tidal influence. The evidence from these facies for a dominance of fluvial currents, and for the absence of barrier bars, and tidal and storm-

dominated sand ridges (typical of macrotidal environments, such as that of the Ord River Delta; Wright *et al.*, 1976), suggests that a deltaic plain and shallow-marine delta-front depositional environment existed, similar in some respects to those of the Niger and Ganges deltas. Given the difficulty of obtaining unequivocal evidence for ancient tidal deposits, attempts have been made in the literature to apply indirect arguments to support a tidal interpretation (Bridges, 1982). Since tidal currents are considered one of the most effective marine agents for reworking of sediment, tabular bed geometry, abundant cross-stratification and beds of clay clasts may be considered as some indirect evidence of tidal action (*ibid.*). However, the presence of some texturally relatively-immature sediment in facies CSdSt and RBSdSt, and the strikingly unidirectional composite palaeocurrent roses plotted from a majority of the sediments of facies CSdSt (see Figs. 3.56 & 3.57), indicate that, whatever the merits of a tidal interpretation, fluvial currents were pre-eminent over all other currents. A further argument has been used to support tidal interpretations based upon the indirect evidence of the effect of the width of the continental shelf on tidal currents (*ibid.*). Since tidal waves can only be generated in ocean basins, the tidal range at the coastline is dependent upon the controls on propagation of the tidal wave across the shelf. Although, in a general sense, an amplification of the tidal wave can be shown to take place across a shelf of decreasing water depth, recent work has shown that no simple general relationship can be established between width of continental shelf and coastal tidal range (Howarth, 1982). The evidence from the blanket geometry and the facies of the External Tassilian Sandstones for a wide, shallow continental shelf, and from the palaeocontinental reconstructions of Scotese *et al.* (1979) for proximity to a major ocean basin on the margin of Gondwanaland (see Fig. 1.2), neither supports nor denies a coastal tidal interpretation, therefore. The problem must remain unresolved in this thesis. However, the sedimentary structures of facies CSdSt and RBSdSt indicating a component of tidal influence in the estuarine channels and on the inner shelf, and the association of these facies with facies VBSd, tentatively interpreted as being of tidal flat origin, taken together with the absence of delta-front barrier bar facies, is considered sufficient evidence to propose that tidal currents with a moderately low tidal range

(microtidal or mesotidal) may have been active in the littoral environment. Consequently tidal terminology has been applied to the littoral environments discussed below and in the cross-section (Fig. 3.54).

The facies study of the sediments of the External Tassilian sandstones has provided the following generalised palaeogeographic reconstruction whilst the vertical sequences present provide evidence that these environments migrated landward during successive periods of sediment accumulation, i.e. the sequences are retrogradational or transgressive. Starting with the most landward depositional environment evidence exists from facies MSd for a perennial braided fluvial environment in which the main channels were typically 2-4 m deep and were poorly supported by weak channel banks. These channels lay within a braid plain of abundant wide, shallow channels, sandflats, abandoned channels and widespread lakes. The braid plain, which was extensive and of low topographic relief, experienced occasional minor marine transgressions resulting from periodic changes in the balance between sea level, subsidence and sediment supply, with the result that lagoonal and estuarine channel conditions occasionally prevailed. Passing seawards, facies CSdSt, RBSdSt and VBSd provide evidence for a fluvially-dominated estuarine environment of straight to sinuous ?tidal/estuarine channels, flanked by ?tidal-flat and shallow-marine relatively quiet-water deposits, and passing downstream to subaqueous channel and mouth bar deposits lying in a shallow-water inner-shelf setting. The coastal plain is thought to have been one of very low relief in which estuarine channel complexes were only temporarily well developed before channel switching led to their abandonment. The estuarine channels increased in depth seawards, the subaqueous channels being typically 3-5 m deep, whilst the evidence of the scour surfaces suggests that the width of the channels also increased seawards. The shallow-water inner-shelf environment appears to have been dominated by fluvial/estuarine currents which deposited land-derived sediment both on channel mouth bars and further seaward, some of which sediment was reworked into ?tide-driven dunes and some of which was resuspended and deposited as sheet sands probably as a result of storms. Passing further seawards facies MSh and IShSdSt provide evidence for a relatively open-marine environment dominated by suspension deposition and influenced by occasional storms and shallowing events.

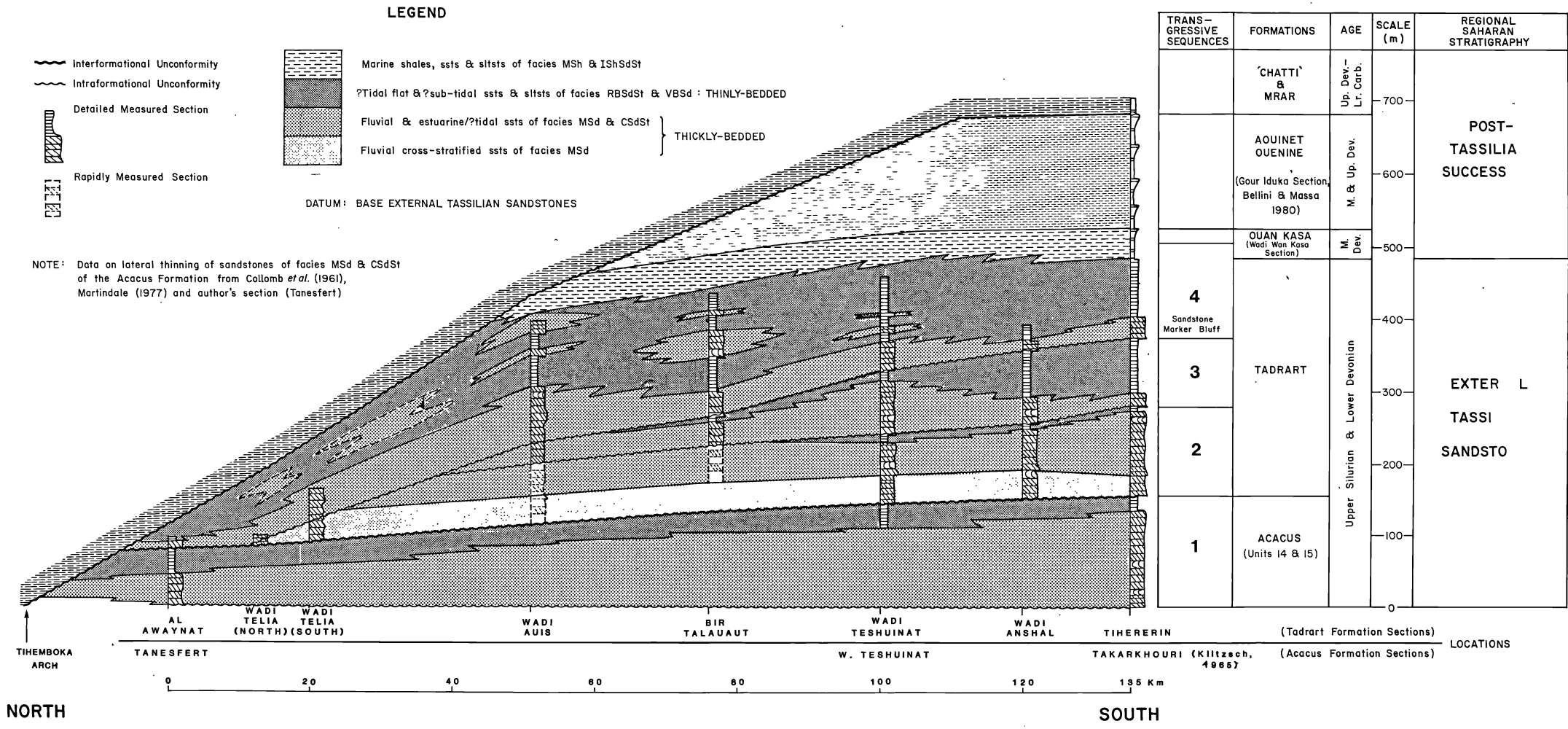


Figure 3·54
LITHO-STRATIGRAPHIC CROSS-SECTION NORTH-SOUTH THROUGH JABALS ACACUS AND TADRART

Figure 3·54

The transgressive sequences recognised are probably best understood by reference to fluvial, estuarine/delta top, and shallow-marine depositional models. The Platte River fluvial model and the mesotidal shallow-marine models of Georgia and South Carolina, USA (see sections 3.2.2 & 3.4.2 respectively) are thought to provide the best modern analogues for the sediments of facies MSd and RBSdSt respectively, whilst the sediments of facies CSdSt and VBSd have been compared with those of modern microtidal and mesotidal estuaries/deltas and tidal flats respectively. No good modern analogue appears to exist for these estuarine/deltaic deposits which are thought to have been laid down in a coastal setting of very low relief characterised by a dynamic equilibrium of strong fluvial sediment supply, subsidence and relative sea level rise that resulted in a gradual landward migration of the coastal depositional setting. The coastline is thought to have been poorly developed as in the case of a fluvially dominant delta, and interfluve areas between estuarine/?tidal channels are thought to have been the site of inter-tidal, lagoonal and lacustrine deposition.

3.7.3 Fossil Evidence

The study of the trace fossils of the External Tassilian Sandstones has shown that trace fossils of Seilacher's *Scoyenia*, *Skolithos* and *Cruziana* ichnofacies are present (see Table 2.6), and that they are present in three distinct assemblages (Table 3.2). These assemblages may broadly be assigned to the *Skolithos* and *Cruziana* ichnofacies but this does not imply acceptance of the environmental interpretations originally inferred for these ichnofacies (Seilacher, 1967). The "*Skolithos* assemblage" is present in brackish, fresh-water and shallow-marine sandstones thought to have been deposited in environmentally severe conditions of scouring currents and periodic turbulence. This assemblage is particularly characteristic of the ?tidal flat deposits of facies VBSd. The "*Arthrophycus* assemblage" is present in the low-salinity, brackish and fresh-water deposits of facies MSd and CSdSt, and is characterised by the "straight-burrow variety" of *Arthrophycus* which is absent from the "*Cruziana/Asterosoma* assemblage" found in the marine sediments of facies RBSdSt. Indeed, although the "*Cruziana/Asterosoma* assemblage" is characterised by a much larger number of ichnogenera than the "*Arthrophycus* assemblage",

ICHNOFACIES	ASSEMBLAGES	LITHOLOGY AND FACIES	GENERA
Skolithos	*Skolithos* (Endogenic vertical burrow and dwelling trace-fossil assemblage of fresh-water brackish, and marine environments)	Sandstones of facies MSd, CSdSt, RBSdSt & VBSd	*Rosselia** *Skolithos*
	Arthrophycus (Exogenic and endogenic trace-fossil assemblage of fresh-water and brackish environments. Crawling traces and feeding burrows predominate)	Argillaceous siltstones, silty shales and adjacent sandstones of facies MSd & CSdSt	*Arthrophycus* *Cruziana** *Fraena* *Palaeophycus*
Cruziana	*Cruziana/Asterosoma* (Exogenic and endogenic trace-fossil assemblage of shallow-marine environments. Crawling traces and feeding burrows, both simple and complex, predominate)	Interbedded sandstones, argillaceous siltstones and silty shales of facies RBSdSt	*Arthrophycus** *Asterosoma* *Corophioides* *Cruziana* *Halopoa** *Sagittichnus** *Scolicia** *Spirophyton*

* NOTES:

Rosselia - only observed in sandstones of facies CSdSt.

Cruziana - not present in sediments of facies MSd.

Arthrophycus - only the "palm-frond variety" of this trace fossil was observed in the "*Cruziana* assemblage", and then only in sediments of transgressive sequence 1.

Halopoa & Sagittichnus - only rarely present.

Scolicia - only observed in transgressive sequence 1.

Table 3.2 TRACE-FOSSIL ASSEMBLAGES OF LIBYAN EXTERNAL TASSILIAN SANDSTONES

only the "palm-frond variety" of *Arthrophycus* was found in the sediments of facies RBSdSt and this occurrence was found to be restricted to the sediments of transgressive sequence 1. Howard (1972) referred to the presence of *Arthrophycus* within an offshore facies of the Upper Cretaceous Blackhawk Formation of the USA. Considerable differences exist between Howard's specimens (1972) and both the "straight-burrow" and "palm-frond" varieties of the Palaeozoic of North Africa (Turner & Benton, 1983) and it is considered, therefore, that the environmental significance of the Mesozoic ichnogenus has little relevance in considering that of the Palaeozoic ichnogenus. The presence of *Cruziana* in the "*Arthrophycus* assemblage" was found to be characteristic of facies CSdSt, but not facies MSd, suggesting that this trace fossil, regardless of the question of its possible trilobite genesis, is present in brackish and restricted, as well as shallow-marine, environments, but is not present in non-marine environments. The "*Cruziana/Asterosoma* assemblage" contains the largest number of ichnogenera of the three assemblages, including all those of the "*Arthrophycus* assemblage" and is present in the heterolithic, ?tidal and inner-shelf deposits of facies RBSdSt. A generalised subdivision of this assemblage into trace fossils commonly present in different lithologies proved possible in the sediments of channel mouth shoals (see section 3.4.2) but no evidence was obtained to support the proposed division of the *Cruziana* ichnofacies of Seilacher (1967) into *Asterosoma* and *Spirophyton* assemblages (Seilacher, 1969; see Table 2.6 & section 3.4.2).

The fossil evidence of the External Tassilian Sandstones may be summarised as follows therefore. The fluvial sediments (facies MSd) are nearly barren, with only some of the trace fossils of the "*Skolithos* and *Arthrophycus* assemblages" present (Table 3.2), indicating the former existence in these sediments of some burrowing soft-bodied invertebrates. The estuarine channel sediments (facies CSdSt), deposited in low-salinity and brackish waters, contain plant remains, *Nucula* bivalves and all the trace fossils of the "*Skolithos* and *Arthrophycus* assemblages", indicating the presence of burrowing soft-bodied invertebrates and of crawling arthropods. The shallow-marine heterolithic, ?tidal and inner-shelf deposits (facies RBSdSt) also contain plant remains and *Nucula* bivalves, but are characterised by an abundance of ichnogenera (the "*Cruziana* assemblage"). This is thought to provide evidence for the presence in the normal marine salinities of the depositional environment of these sediments of abundant arthropods and some gastropods, as well as many soft-bodied organisms. The

relatively open-marine sediments of facies MSh and IShSdSt contain plant remains, *Nucula* bivalves and a number of genera of brachiopods as well as tentaculites, styliolines and an abundance of trace fossils of the "*Cruziana* and *Skolithos* assemblages".

The fossil assemblages of these sediments bear some similarities to those of the alluvial and marine facies of the Old Red Sandstone of Southern Britain (Allen, 1979) in which plant remains and the trace fossils of invertebrates characterise alluvial sediments whilst, in addition to these fossils, deltaic, coastal and tidal sediments are characterised by the presence of brachiopods, bivalves, gastropods, arthropods and lingulids.

3.7.4 Palaeocurrent Evidence

The palaeocurrent data obtained from all the measured sections have been plotted according to their location for each of the transgressive sequences, except for transgressive sequence 1, where insufficient data were available (Figs. 3.55-3.59). The vast majority of the palaeocurrent data in this study was obtained from foreset dip azimuths of cross-strata of facies MSd and CSdSt. These sediments are present in the basal coarse-grained or thickly-bedded units of each transgressive sequence (Fig. 3.54; see Fig. 3.1). The palaeocurrent data from these facies have been plotted separately for transgressive sequence 2 (Figs. 3.55 & 3.56), those plotted in Figure 3.55 being from the sediments of facies MSd lying below the "40 m plateau", and those in Figure 3.56 being from the sediments of facies CSdSt (and subsidiary interbedded sediments of facies MSd) lying above the "40 m plateau" (see section 1.3.4, Chap. 1). The palaeocurrent data from these two facies in transgressive sequence 3 have been plotted together (Fig. 3.57) because the sediments of these facies are complexly interbedded, whilst in transgressive sequence 4 facies MSd is absent, except in the north of the study area, and the majority of the readings are from facies CSdSt (Fig. 3.58). Only in the case of transgressive sequence 4 were sufficient measurements obtained from the fine-grained or thinly-bedded unit of the sequence (composed of sediments of facies RBSdSt and VBSd) to allow a plot of palaeocurrent readings.

The palaeocurrent data from the fluvial deposits at the base of sequence 2 (Fig. 3.55) are thought to provide good evidence for a

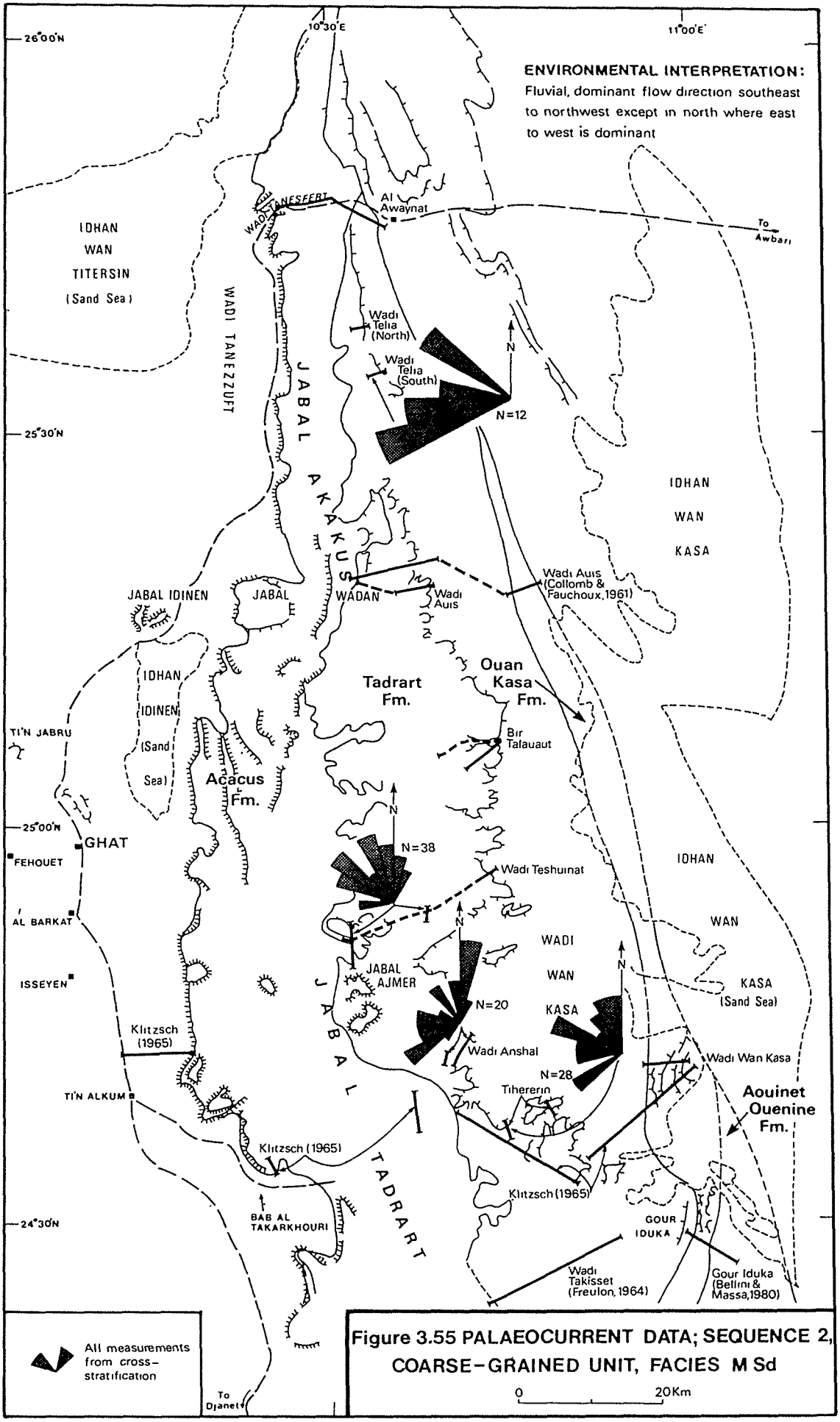


Figure 3.55 PALAEOCURRENT DATA; SEQUENCE 2, COARSE-GRAINED UNIT, FACIES M Sd

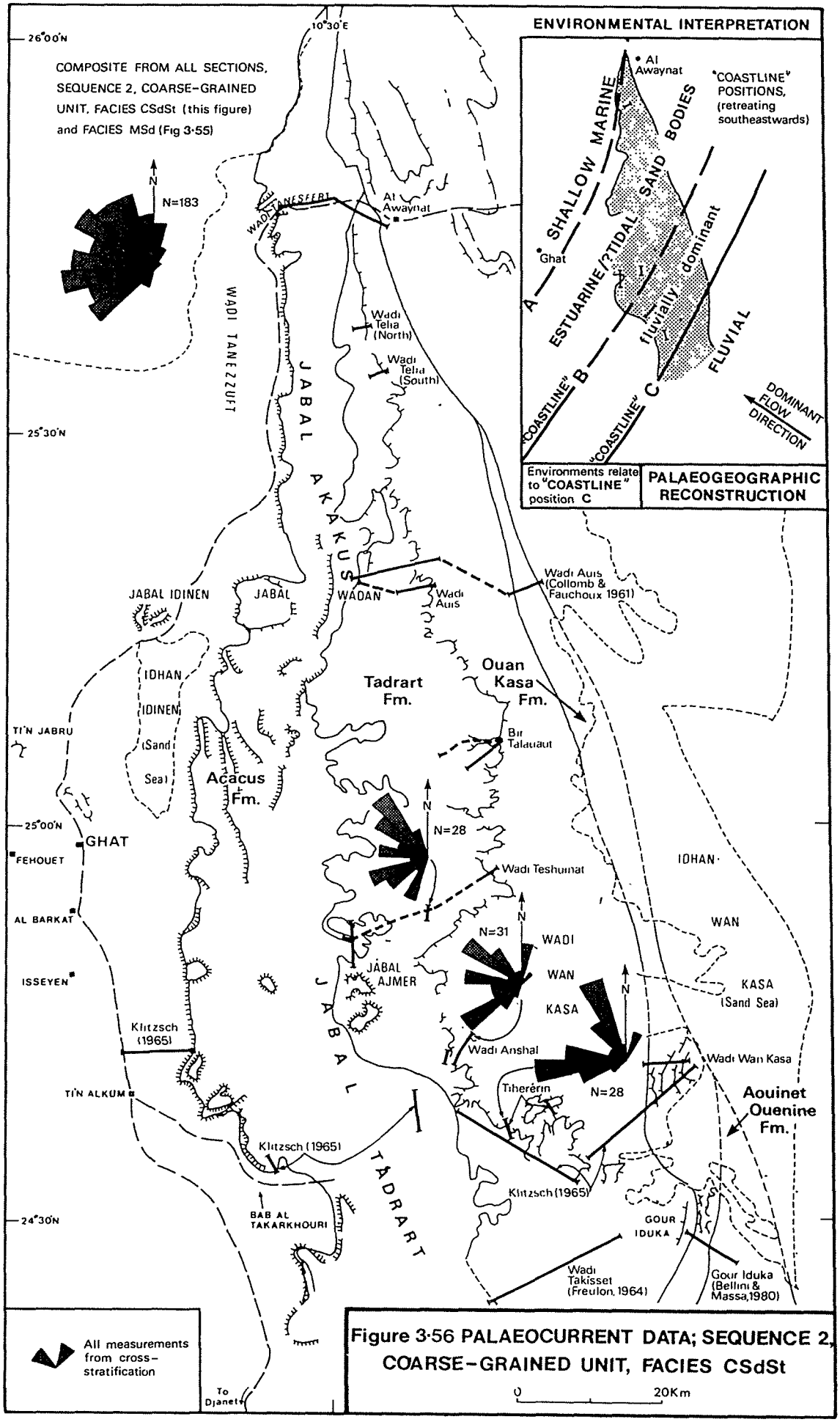


Figure 3·56 PALAEOCURRENT DATA; SEQUENCE 2, COARSE-GRAINED UNIT, FACIES CSdSt

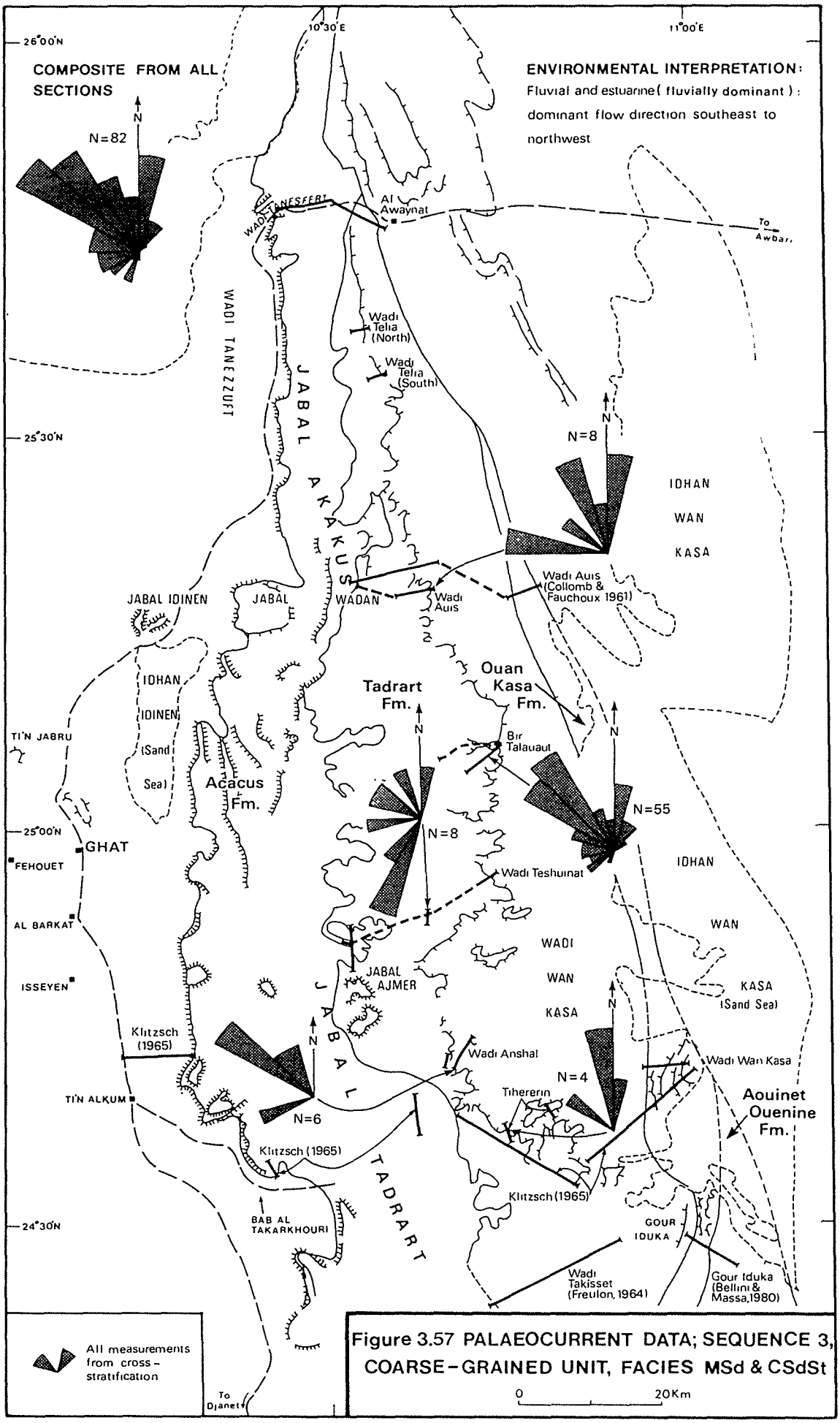


Figure 3.57 PALAEOCURRENT DATA; SEQUENCE 3, COARSE-GRAINED UNIT, FACIES MSd & CSdSt

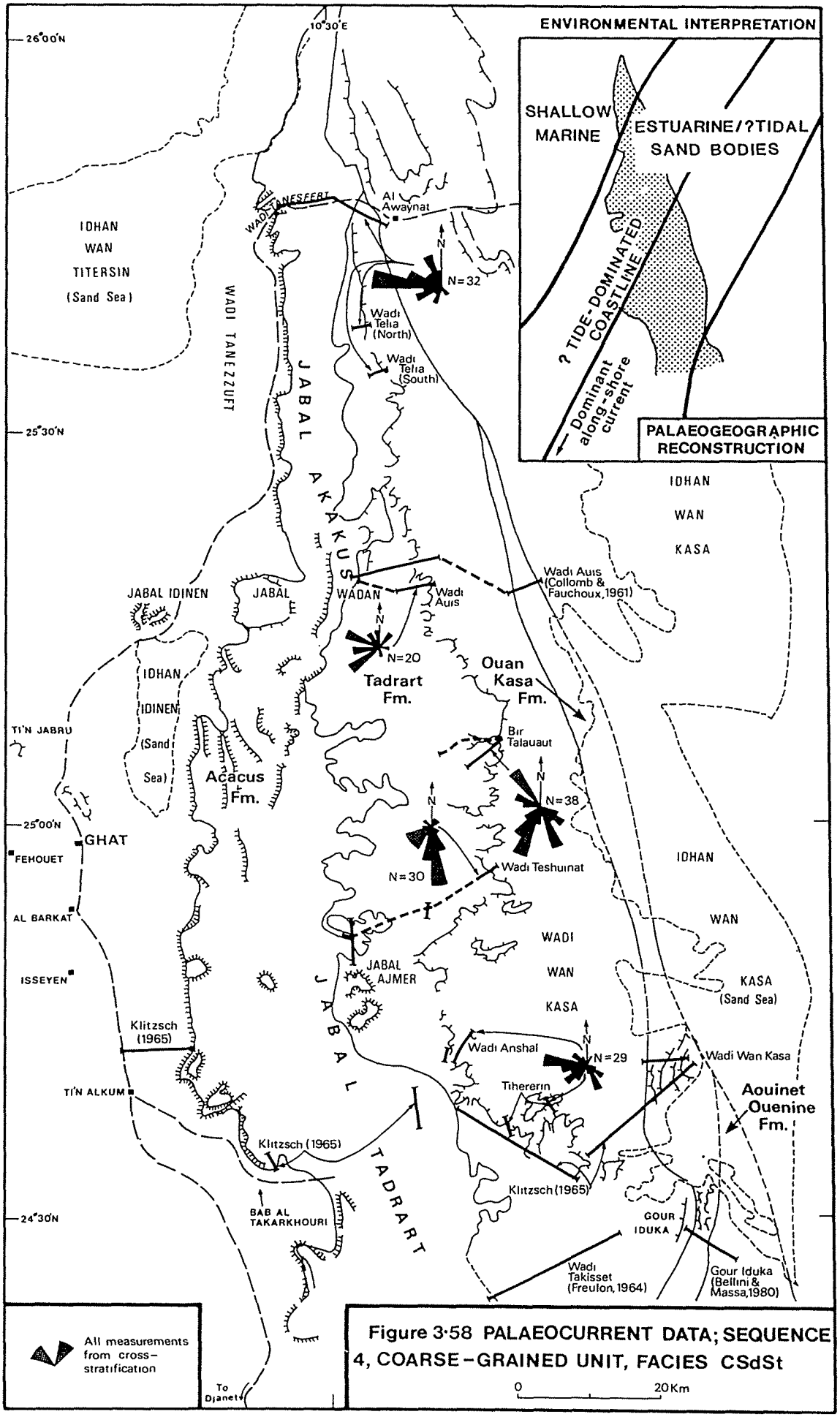


Figure 3·58 PALAEOCURRENT DATA; SEQUENCE 4, COARSE-GRAINED UNIT, FACIES CSdSt

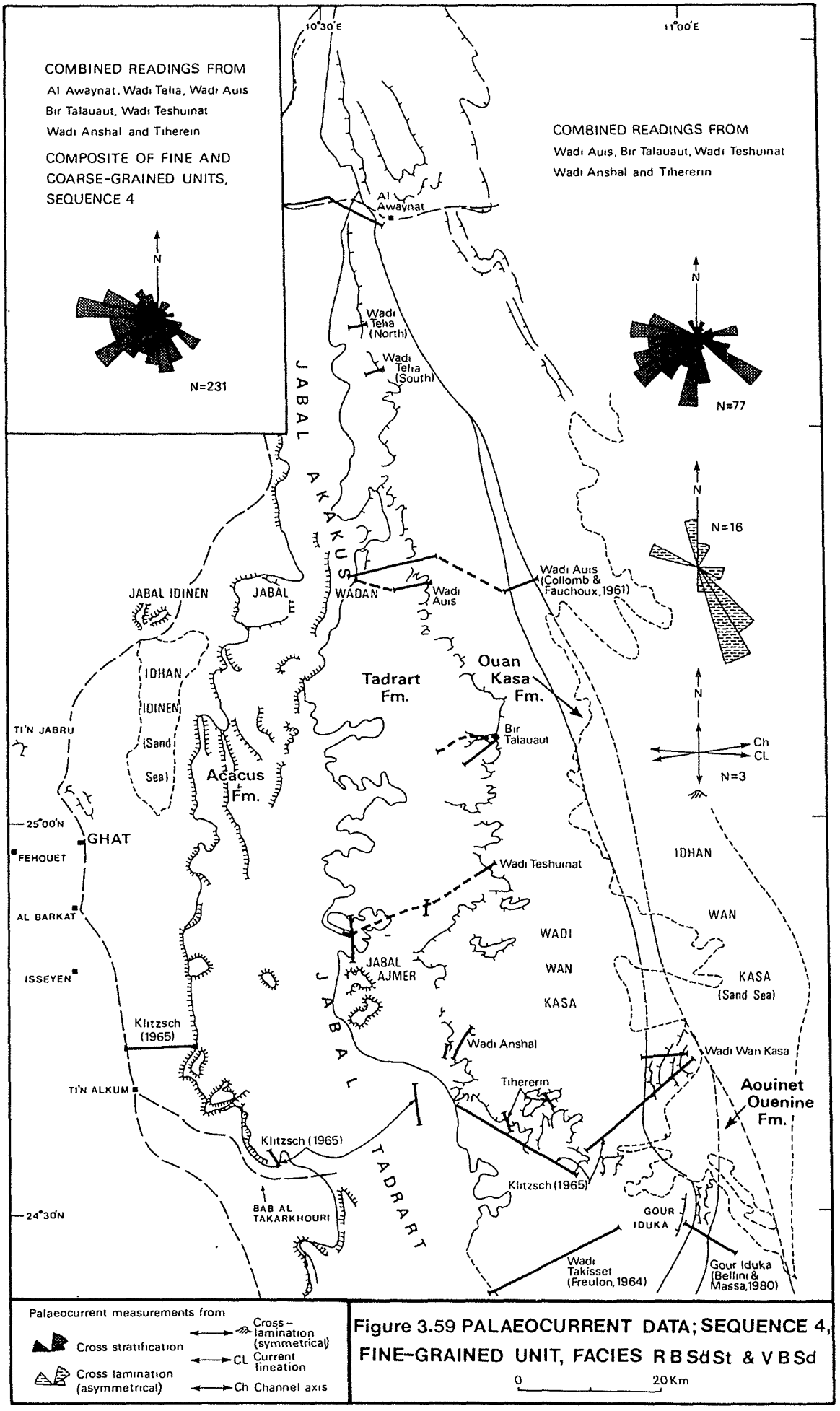


Figure 3.59 PALAEOCURRENT DATA; SEQUENCE 4, FINE-GRAINED UNIT, FACIES R B Sd St & V B Sd

depositional palaeoslope towards the northwest whilst the evidence from the overlying estuarine (and subordinate fluvial) deposits suggests that similarly orientated fluvial currents were dominant in this facies. A composite palaeocurrent rose of all the readings obtained from facies MSd and CSdSt in transgressive sequence 2 (Fig. 3.56, upper left) emphasises the unidirectional nature of the palaeocurrent data, and displays the wide variance. The significance of this large variance is difficult to assess because the readings come from a large area and from different stratigraphic horizons. The evidence of the facies study and of Figures 3.55 and 3.56, therefore, suggests that the coarser-grained sediments of transgressive sequence 2 were deposited by fluvial and fluvially-dominated estuarine/?tidal channel flows, directed predominantly towards the northwest (Fig. 3.56, upper right). The finer-grained sediments were deposited in shallow-marine inner-shelf environments which retreated southeastwards with time (Fig. 3.56, upper right) resulting in the deposition and preservation of a transgressive sequence. From the limited palaeocurrent data available from facies MSd and CSdSt of transgressive sequence 1 (Fig. 3.37; Martindale, 1977) a similar palaeogeographic setting may be inferred, with northwesterly directed channel palaeoflows predominating, and with some subordinate easterly directed flows of possible flood-tidal origin. The data from transgressive sequence 3 (Fig. 3.57) also provide evidence for a very constant palaeogeographic setting dominated by northwesterly directed palaeocurrents. The polymodal palaeocurrent data of transgressive sequence 4 (Figs. 3.58 & 3.59), on the other hand, are thought to provide evidence for stronger influence of marine currents in this sequence, with a strong southwesterly-orientated along-shore current strongly developed. As already noted, facies MSd is absent from sequence 4, except in the north of the study area, and facies CSdSt in this sequence is dominated by the more seaward subfacies (?tidal subaqueous channels and mouth bars). A more typically marine palaeocurrent pattern might reasonably be expected, therefore. The absence of the previously pervasive unidirectional northwesterly palaeoflow is enigmatic, however, and may be related to a possible change in coastal physiography associated with the major uplift of the Tihemboka Arch that preceded the deposition of this sequence (see section 3.7.5 below).

3.7.5 Correlation with Southern Algeria

Evidence for an uplift of the Tihemboka Arch during the early Devonian is provided by a correlation of the External Tassilian Sandstones across the Tihemboka Arch (Fig. 3.60; Table 3.1). This correlation illustrates that, whereas the coarse-grained units of sequences 1, 2 and 3 decrease in thickness towards the northwest, the coarse-grained unit of sequence 4 appears to thin towards the southeast. Figure 3.60 was constructed by correlating the composite section measured in Wadi Teshuinat and the composite section measured approximately 250 km away at Wadi Samene (Dubois *et al.*, 1969; Bennacef *et al.*, 1971). Correlation was made possible by the fact that, although the transgressive sequences are similar to each other in many respects, certain characteristics distinguish them. Major scoured surfaces are common at the base of each of the transgressive sequences but the largest-scale scours with the greatest relief are commonly at the base of sequence 3. This suggested a correlation of the coarse-grained unit (composed of facies MSd and CSdSt) of sequence 3 with the "Middle Hard Bed", a coarse-grained, thickly-bedded unit in the middle of the Oued Karkai Formation (Fig. 3.60), at the base of which Dubois *et al.* (1969) recorded the presence of a major palaeo-topography. Using this initial correlation it was found that the description of the Oued Tifernine and Tamelrik formations corresponded remarkably accurately with those of sequences 2 and 3 respectively of the Wadi Teshuinat succession. The description of the "pavements" at the top of the Tamelrik Formation bears a close resemblance to that of the thick successions of facies RBSdSt found at the top of sequence 3 in the Wadi Anshal and Tihererin composite sections (Fig. 3.54). The absence of a "Hard Bed" in the Atafaitafa Formation corresponding to the coarse-grained unit of sequence 1 is explained by the northward thinning of this unit illustrated in Figure 3.54. It is thought probable that the fluvial and estuarine/?tidal deposits of this unit pass northwestwards by a lateral facies change into the more marine deposits of the Atafaitafa Formation. The report of marine fossils of Gedinnian age in the sediments of this facies (Dubois *et al.*, 1969) is consistent with this correlation, and provides a tentative approximate age for the first transgressive sequence.

The validity of this correlation of coarse-grained units over a

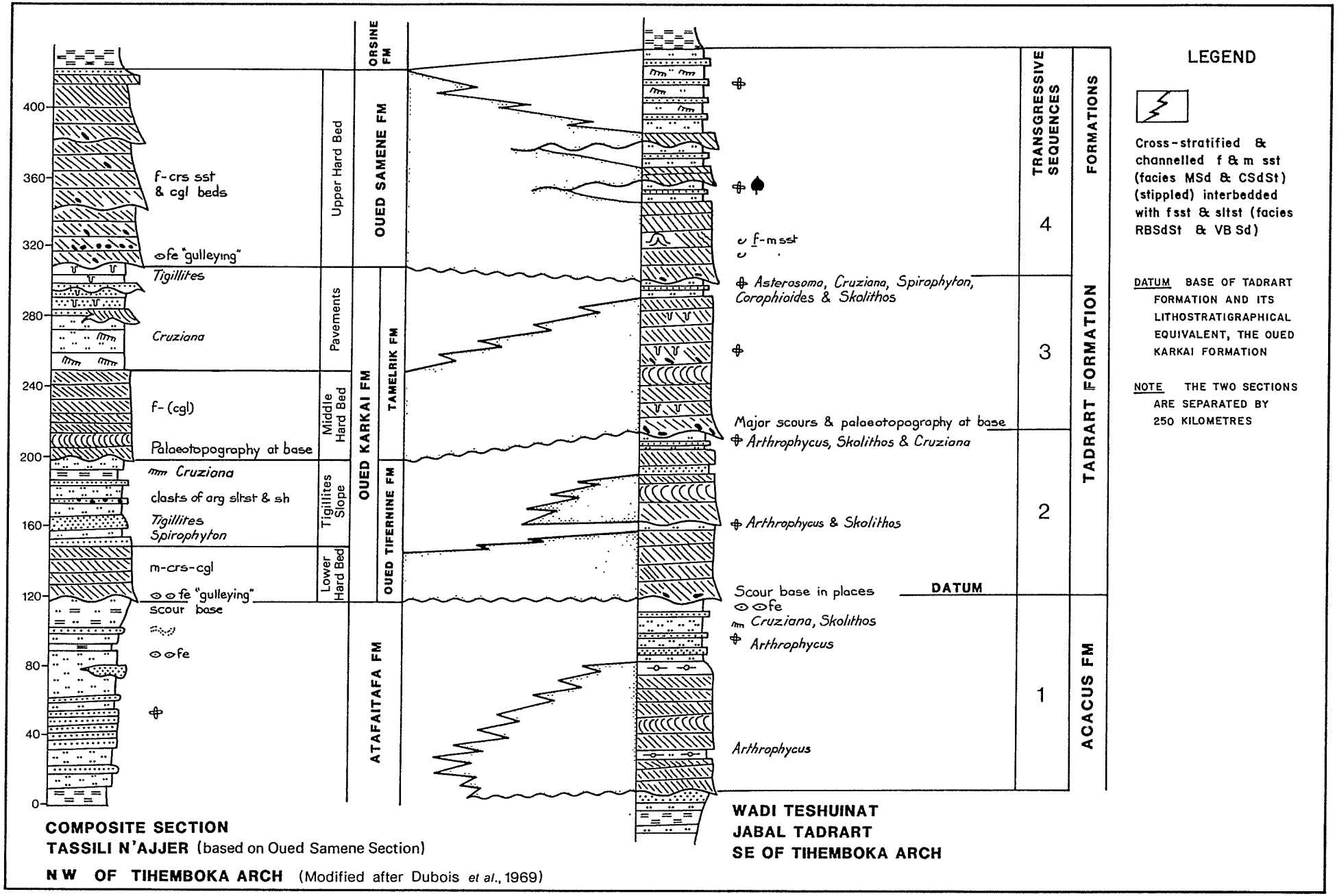


Figure 3. 60 CORRELATION OF EXTERNAL TASSILIAN SANDSTONES ACROSS

distance of 250 km may be reasonably questioned. It was found, however, that correlation of the transgressive sequences in the Jabals Akakus and Tadrart could be established with confidence over distances of up to 100 km, and Dubois *et al.* (1969) and Beuf *et al.* (1971) have shown that the coarse-grained units of the Oued Karkai and Oued Samene formations may be correlated for distances of up to 800 km west of the Tihemboka Arch (Fig. 3.61). This correlation also shows that each successive coarse-grained unit is more laterally extensive towards the west. It seems reasonable to assume, therefore, that the lowest of the coarse-grained units seen in the Jabal Tadrart may have passed laterally into finer-grained facies east of Wadi Samene.

The correlation established in Figure 3.60, therefore, indicates that the Oued Samene Formation is the lateral equivalent of transgressive sequence 4. Unlike the stratigraphically lower successions, however, the sediments of these lateral equivalents bear little resemblance to each other, those of the Oued Samene Formation being the thickest and most continental in character of the coarse-grained units, whilst those of sequence 4 are the thinnest and the most marine in character of the four sequences in the Jabal Tadrart. It seems possible, therefore, that the sediments of the Oued Samene Formation and of sequence 4 are not closely related palaeogeographically, whereas the lower sequences appear to be directly genetically related.

Although the preservation of transgressive sequences 1, 2 and 3 appears to have been controlled by the Tihemboka Arch (Figs. 3.54 & 3.61) it has been found that the activity of the arch cannot be related to the facies of these sequences. The palaeocurrent and outcrop evidence indicates that the coarse-grained units (facies MSd and CSdSt) of sequences 1, 2 and 3 represent sheet sandstones of considerable extent which were spread over the Central Sahara by channels flowing predominantly towards the northwest and depositing detritus derived from a region southeast of Jabals Akakus and Tadrart. Correlation of the composite sections measured in the Jabals Akakus and Tadrart with a section at In Ezzan (Beuf *et al.*, 1971) (Fig. 3.61) indicated that the In Ezzan region may have acted structurally and topographically as a positive region during much of the period of time represented by sequences 1, 2 and 3. Some evidence for a northwestward decrease in grainsize towards the Tihemboka Arch in the study area is provided by the sandstones of facies MSd at the base of transgressive sequence 2,

the succession at Tihererin (Fig. 3.11) being coarser-grained than that at Wadi Teshuinat (Figs. 3.13 & 3.14). A similar trend is discernible in the coarse-grained unit of sequence 1. The Tihemboka Arch, therefore, appears to have acted as a structural axis which was subject to periodic uplift, causing localised erosion of the sediments deposited over the arch, whilst in the intervening periods it may have been subject to the same subsidence as adjacent regions.

The stratigraphic correlations (Figs. 3.54, 3.60 & 3.61) indicate that an uplift of the Tihemboka Arch took place prior to the deposition of the sediments of sequence 4 which was of greater significance than the earlier epeirogenic movements of this structural alignment, whilst the facies evidence suggests that this uplift may have been responsible for a change in the regional palaeogeography applying at the time of deposition of the sediments of sequence 4. In Figure 3.61 it is shown that the lateral extent of the coarse-grained blanket sandstone of sequence 4 is greater than any of the underlying coarse-grained units in Algeria. Given the absence of a similar thick coarse-grained sandstone unit of predominantly non-marine character in sequence 4 in Libya, it is thought probable that the Algerian blanket sandstone represents deposits eroded from a newly-uplifted sediment source area possibly located in the Eastern Hoggar.

3.7.6 Transgressive Sequences

The processes responsible for the presence of three very similar transgressive sequences, and a succeeding transgressive sequence which differs in a number of respects from the underlying sequences, in the Upper Silurian to Lower Devonian sediments of Southern Algeria and Southwest Libya were considered by Beuf *et al.* (1971) to be dominated by eustatic sea level changes and variation in sediment supply (Fig. 3.40). In recent seismic stratigraphic work (Vail *et al.*, 1977; Brown & Fisher, 1982) it has been argued that regional unconformities and their correlative disconformities are chronostratigraphic time horizons such that all the sediments lying above one of these surfaces are always younger than those below. The unconformity/disconformity surfaces are termed "sequence boundaries" because they may be used to define "seismic sequences" the sediments of which are all depositionally genetically related but are not necessarily synchronous. The unconformity/disconformity surfaces may be attributable to world-wide

sea level changes or to localised interplay between eustacy and tectonics, but in any event they are thought to represent a relative fall in sea level with respect to the depositional surface. It has been argued that world-wide sea level falls were relatively rapid events whilst sea level rises took place relatively slowly (Vail *et al.*, 1977). It follows, therefore, that the characteristics of transgressive sequences 1, 2 and 3 may be explained by proposing that the depositional surface which formed a part of the continental margin of Gondwanaland was tectonically stable and represented a surface on which onlapping transgressive sequences were deposited during each eustatic sea level rise, and over which a surface of unconformity/disconformity was created during each sea level fall.

During the period of erosion related to these sea level falls structurally-positive regions such as the Tihemboka Arch appear to have undergone a greater degree of erosion than relatively structurally-negative regions (Fig. 3.61). During the period of erosion preceeding the deposition of sequence 4, however, evidence exists that epeirogenic activity was more marked and a new, more proximal sediment source area may have been established. Nevertheless this period of erosion was also succeeded by the deposition of a transgressive sequence. Klein (1982) argued that cratonic sedimentation is characteristically strongly tidally-influenced and dominated by transgressive sequences where epeirogenic/orogenic activity is limited, whereas regressive sequences were thought to be a response to active tectonics. This would suggest that, despite the increased epeirogenic activity associated with sequence 4, the stratigraphic succession of the External Tassilian Sandstones was largely controlled by eustatic changes, a view which is consistent also with the work undertaken on cratonic sedimentation by Sloss and Speed (1974) and Sloss (1978). A slightly different emphasis was adopted by Busson (1978) to explain the stratigraphy and sedimentology of Mesozoic Nubian-type sandstones of the African margins. He argued that widespread vertical movements of the stable craton were responsible for the development of these continental blanket sandstones, an argument which is not significantly different from those quoted earlier since it also depends upon relative changes in sea level.

It was noted in section 3.2.2.10 that Selley (1978) proposed a retreating-escarpment hypothesis of alluvial fan deposition (see Heward, 1978) to account for the coarser-grained Nubian-type sandstones

of Jordan, and the cyclicity characteristic of the Cambro-Ordovician succession of Jordan. This hypothesis is not thought to provide a satisfactory explanation for the transgressive sequences of the External Tassilian Sandstones because it does not provide an explanation for the presence of medium and coarse-grained sandstones of one cycle overlying fine and very fine-grained sandstones of an earlier cycle where these older beds are very widespread. In areas of uplifted arches, such as the Tihemboka Arch, coarse-grade sediment may have been derived from a range of eroded older rocks, but, in areas of little epeirogenesis where sediments of facies RBSdSt and VBSd were deposited over a very large area, a downward change in base level that did not result in the removal by erosion of all of these beds, and which caused the development of a retreating escarpment, would not allow the deposition of anything coarser than fine-grained sandstone. If, on the other hand, a depositional model is invoked to explain the cyclicity of the External Tassilian Sandstones in which sediment may have been derived from a far-distant source area then the recurrence of coarse-grade sediment is less problematical. Furthermore, work undertaken on the provenance of Cretaceous Nubian-type sandstones of Southwest Egypt by Mazzullo and Ehrlich (1980) showed that the coarser-grained units of recognised transgressive sequences were characterised by fluvial deposits containing two quartz grain families, one of "Piedmont" and the other of relict coastal plain origin. The presence of quartz grains thought to have been derived from a "Piedmont" crystalline terrain, as well as reworked grains from earlier Nubian-type sandstones, indicates that, in this case a retreating escarpment hypothesis would be inapplicable.

One of the problems of interpretation posed by cratonic sediments and emphasised by the study of the Cambro-Ordovician Orthoquartzite-Carbonate Suite of North America (Dott & Byers, 1980) is the common absence of major beds of shale or claystone in these successions. Harms (1980) proposed for the Mesozoic Nubian-type sandstones of Egypt that this phenomenon was a result of humid tropical weathering conditions in which all minerals other than quartz were intensely chemically weathered and were transported in suspension far beyond the fluvial domain onto the continental shelf. The general turbulence characteristic of the fluvial flow conditions, therefore, was thought to be a cause of the absence of thick shale beds. In this study inner

shelf deposits pass upwards to thick shale beds only in the case of transgressive sequence 4. In all earlier sequences thick shale beds are absent. This is thought, similarly, to be a result of the turbulent flow conditions and also the lateral migration of slightly-sinuous channels (Cotter, 1978) that are characteristic of the fluvial, ?tidal/estuarine and shallow-marine environments of the sediments of these sequences. It is not possible to invoke the activity of humid tropical weathering, however, to explain the textures of these Palaeozoic sediments, since they are thought to have been deposited in a humid temperate climate of middle latitudes (see Fig. 1.2). The absence of shales at the top of transgressive sequences 1, 2 and 3 suggests that, even at the highest sea-level stand of the transgression, deep water conditions were not developed on the shelf.

The interpretation considered most plausible for the transgressive sequences of the External Tassilian Sandstones is that they represent retrogradational deltaic deposits laid down during periods of sea level rise accompanied by a rate of sediment supply which, although possibly very great, was insufficient to maintain the coastline at a constant position. Curtis (1970) showed that the typical vertical succession of such a retrogradational sequence was represented by fluvial facies, overlain by deltaic plain facies overlain, in turn, by prodelta and marine clays. In a delta characterised by 1) a braided-fluvial upper deltaic plain, 2) a lower deltaic plain of low-sinuosity estuarine/?tidal channel complexes and 3) an offshore shallow-marine environment of strong fluvially-derived currents and other subordinate marine currents, the vertical succession of sediments preserved in a transgression would be a fining-upwards arenaceous succession, such as those of the External Tassilian Sandstones. Similar sequences, resulting from marine transgressions over deltaic plains, have been recorded from the Rhone and the Niger (Oomkens, 1970, 1974) although, in these cases, the presence of fine-grained deposits on the deltaic plain led to a different vertical grainsize profile. In the Rhone delta, where a coastal barrier bar is developed, landward advance of this sand body led to the development of an intrasequential disconformity at its base. No similar disconformity was seen within the transgressive sequences of the External Tassilian Sandstones and consequently the deltaic model of the Niger Delta which is dominated by channel sandstones is considered more appropriate. On a stable

shelf the deposits of transgressions have a greater preservation potential than those of regressions and this is thought to explain the dominance of transgressive sequences in the External Tassilian Sandstones, periods of regression generally, although not always (see below), being characterised by erosion and the development of surfaces of unconformity/disconformity.

The presence at the base of the transgressive sequences of major scour surfaces, iron-oxide cemented beds and fluviatile and estuarine channel sandstones displaying abundant evidence for erosive scouring events is thought to indicate that the rocks present were deposited over a long period of time during which only very little sediment was preserved. The scour surfaces or "palaeotopographic features" characteristic of the base of sequence 3 are well displayed in Wadi Teshuinat (Fig. 3.62) at which locality evidence exists for a minor regressive sequence in which two unconformity/scour surfaces are developed. The regressive sequence is composed of inner-shelf deposits in which nose-shaped crevasse splay or mouth-bar deposits are present (facies CSdSt, subfacies 4), overlain erosively by ?tidal/estuarine deposits which, in turn, are overlain by fluvial sandstones lying within very major concave-upwards scour surfaces. This succession indicates that, whereas generally one surface of unconformity is present dividing the sediments of successive transgressive sequences, in some cases, sediments deposited during regressions are preserved as highly truncated regressive/prograding sequences. Elsewhere the concave-upwards scours truncate down to the sediments of facies RBSdSt of the preceding cycle, completely removing any regressive deposits that may have been present (Fig. 3.63). Even when only one major erosion surface is present, however, at the base of a transgressive sequence, it is not uncommon to find that the basal beds of the sequence are sediments of estuarine/?tidal channels which pass upwards into fluvial sediments, indicating that a thin offlapping prograding succession is present in some places at the base of the major transgressive sequences (Fig. 3.63). In the "Sandstone Marker Bluff" a similar trend is discernible (Fig. 3.29), whilst at the base of transgressive sequence 2 sediments of facies CSdSt, subfacies 4, are impersistently present with fluvial sediments of facies MSd (Figs. 3.33 & 3.64). This association has been used as evidence that some of the sediments of subfacies 4 may be crevasse-splay deposits lying

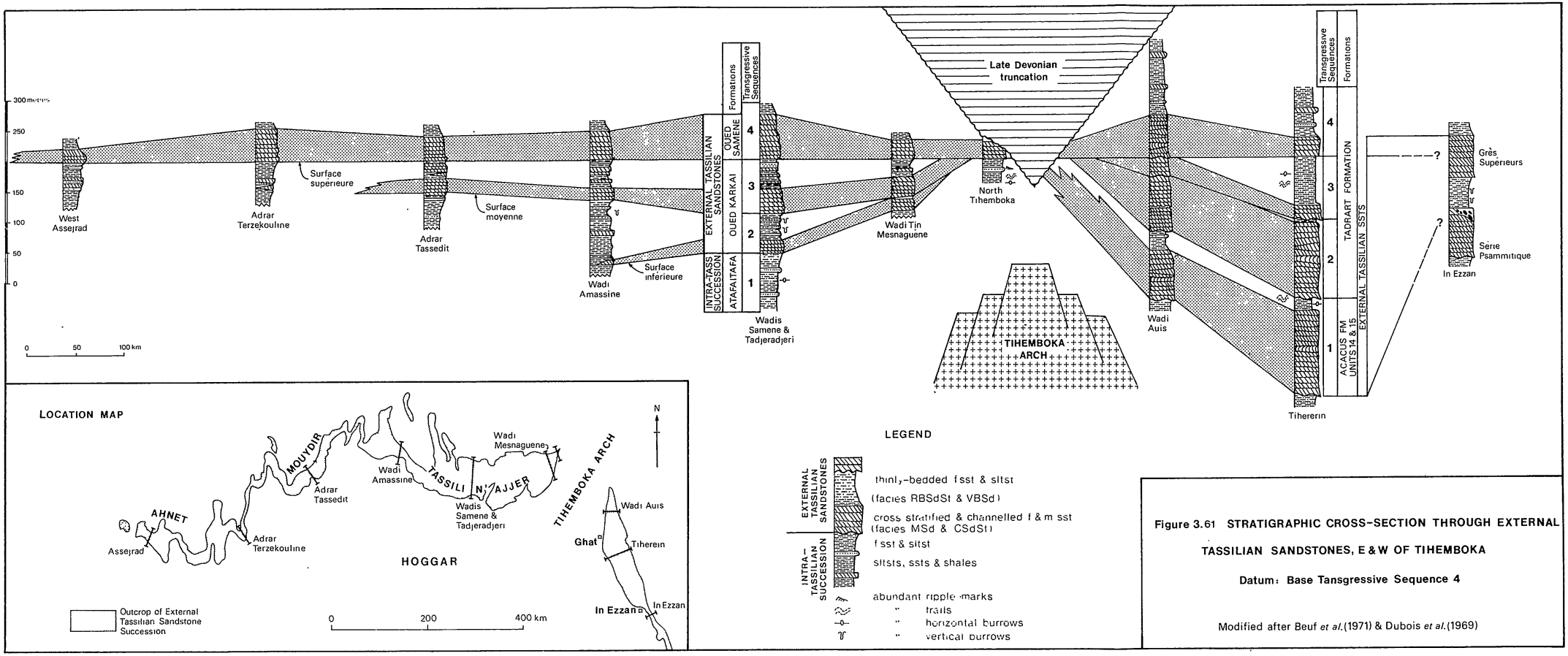


Figure 3.61 STRATIGRAPHIC CROSS-SECTION THROUGH EXTERNAL TASSILIAN SANDSTONES, E & W OF TIHEMBOKA

Datum: Base Tansgressive Sequence 4

Modified after Beuf *et al.*(1971) & Dubois *et al.*(1969)

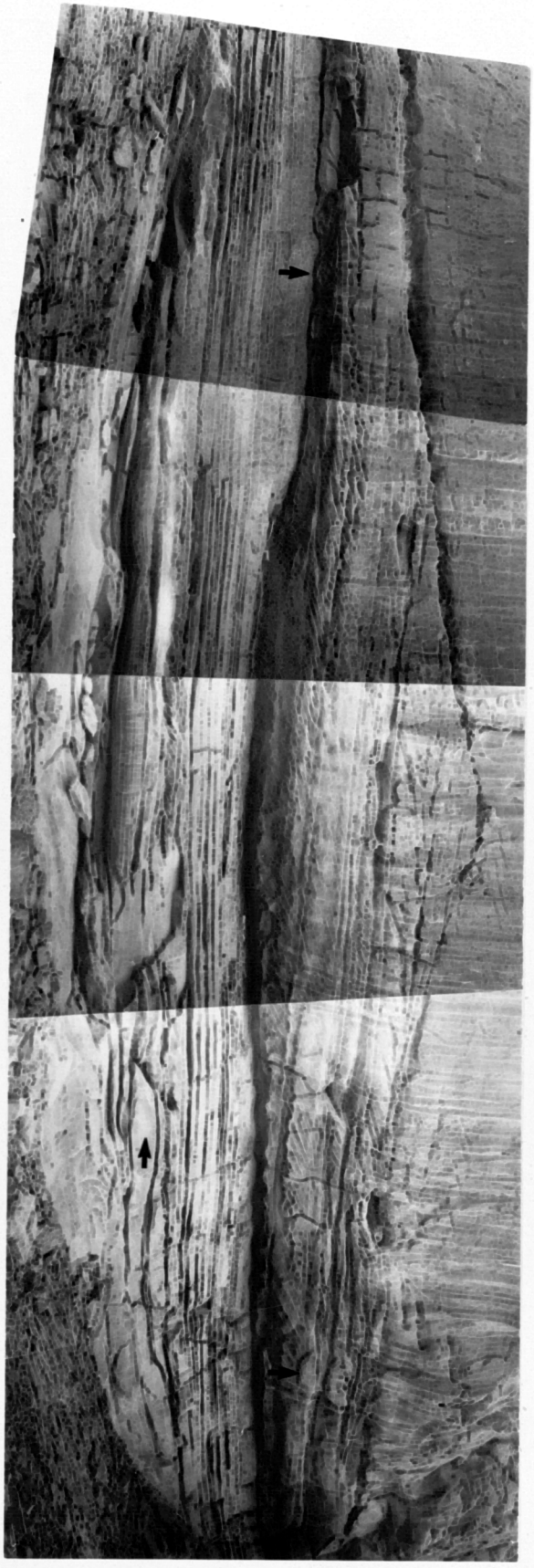
Figure 3.62 LARGE-SCALE SCOUR HORIZONS AT THE BASE OF TRANSGRESSIVE SEQUENCE 3

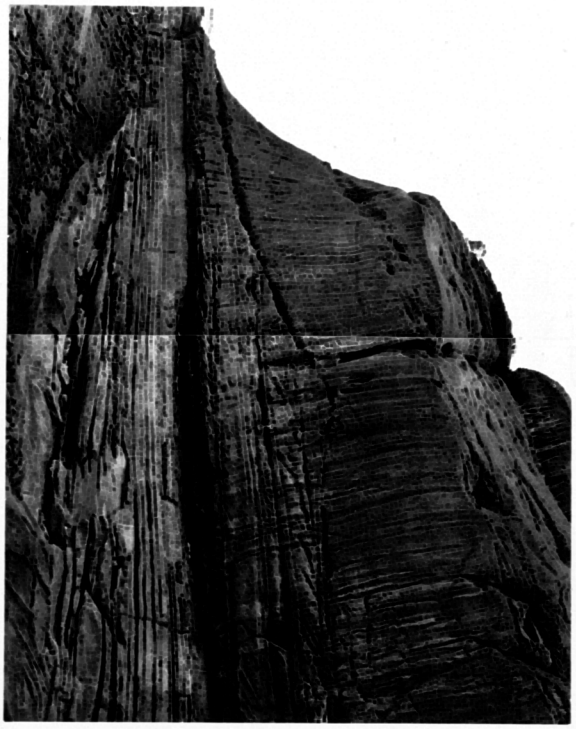
Note: all three illustrations show exposures of the lower beds of transgressive sequence 3 and the upper beds of transgressive sequence 2 in a jabal at locality 3, Wadi Teshuinat (Fig. 3.5). Another illustration of this jabal, a view from the southeast, is presented in Figure 3.3.

Upper left : two distinct scour surfaces are present within, and at the base of the sandstones of facies MSd and CSdSt which make up the basal beds of transgressive sequence 3. The lower of these is nearly planar whilst the upper scour is channel-shaped, aligned broadly east-west, and has a depth of 15 m and a width of 40 m from crest to trough.

Upper right : exposure in slab walls and subordinate slick-rock slopes of the complete succession (c. 75 m thick) of sandstones of facies MSd and CSdSt which make up the "coarse-grained unit" of transgressive sequence 3. The scour surfaces shown in the upper left illustration are visible at the base of the sandstone bluff.

Lower : close-up photo-montage of the two scour surfaces illustrated above and the beds with which they are associated. The lower scour surface is nearly planar but can be seen locally to truncate the underlying beds (see arrow on left hand side), whilst the upper surface clearly truncates the underlying beds throughout this exposure. The near-planar scour surface divides beds of facies RBSdSt, at the base, from those of subfacies 1 of facies CSdSt, whilst the channel-shaped scour surface divides beds of facies CSdSt from those of facies MSd. Characteristic features of these facies are visible and include herringbone cross-stratification (see arrow on right hand side, see Fig. 3.25) and a nose-shaped sandstone body lying within well-bedded sediments of facies RBSdSt (see lowermost arrow).





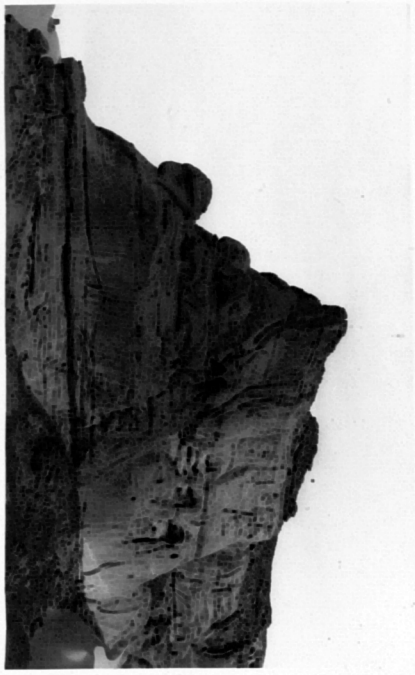
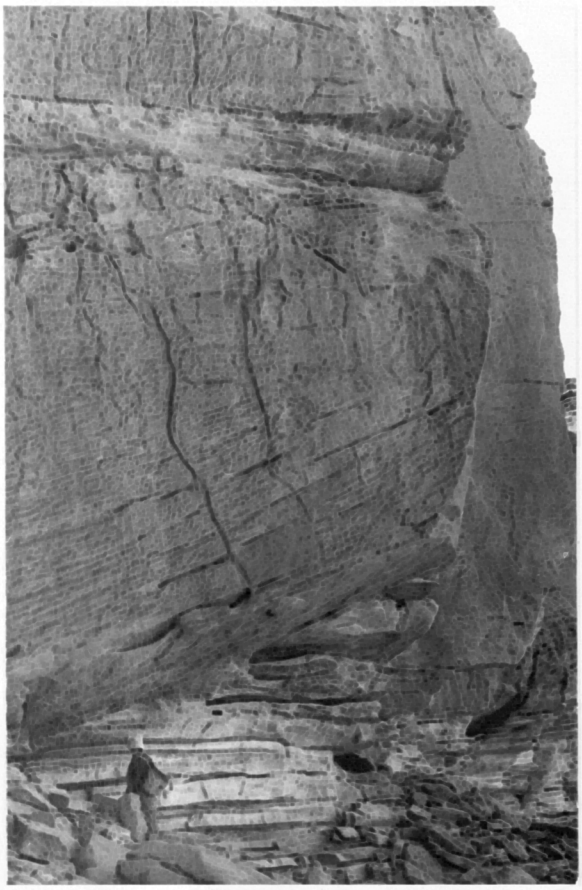


Figure 3.63 CHANNEL SANDSTONES AND SCOUR SURFACES AT THE BASE OF TRANSGRESSIVE SEQUENCES 3 AND 4

Upper : large channel sandstone body (lying above a scoured surface of 6 m relief) composed of sandstones of subfacies 2, facies CSdSt, which is exposed at the base of transgressive sequence 3, locality 5, Wadi Teshuinat (Fig. 3.5). The sandstone body truncates sediments of facies RBSdSt and is immediately overlain by sandstones of facies MSd. The foreset dip azimuths of the channel sandstone cross-strata indicate northwesterly-orientated palaeocurrents. The person at the bottom of the photograph provides scale.

Lower : channel sandstone body (lying above a scoured surface of 2.2 m relief) composed of sandstones of subfacies 3, facies CSdSt, which is exposed at the base of transgressive sequence 4, in the "Sandstone Marker Bluff", locality 13, Wadi Anshal (Fig. 3.5). The sandstone body, as in the illustration above, truncates sediments of facies RBSdSt but is, in this case, overlain by sandstones of subfacies 1, facies CSdSt.



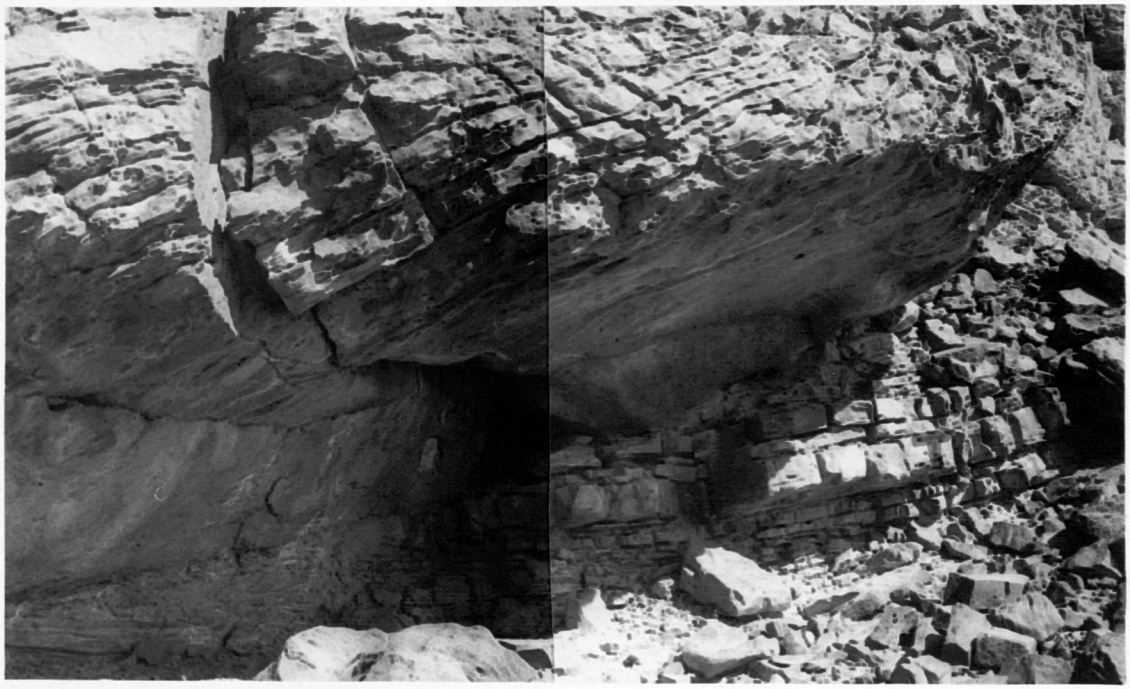
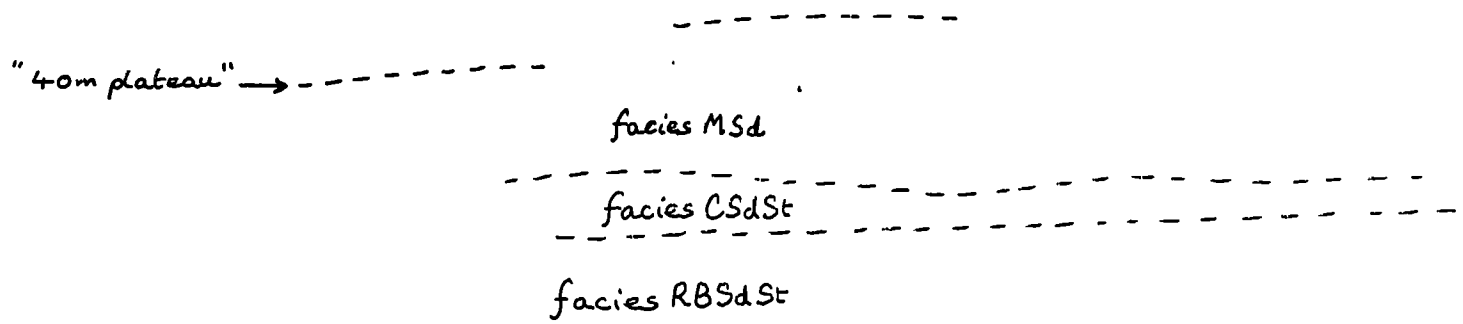
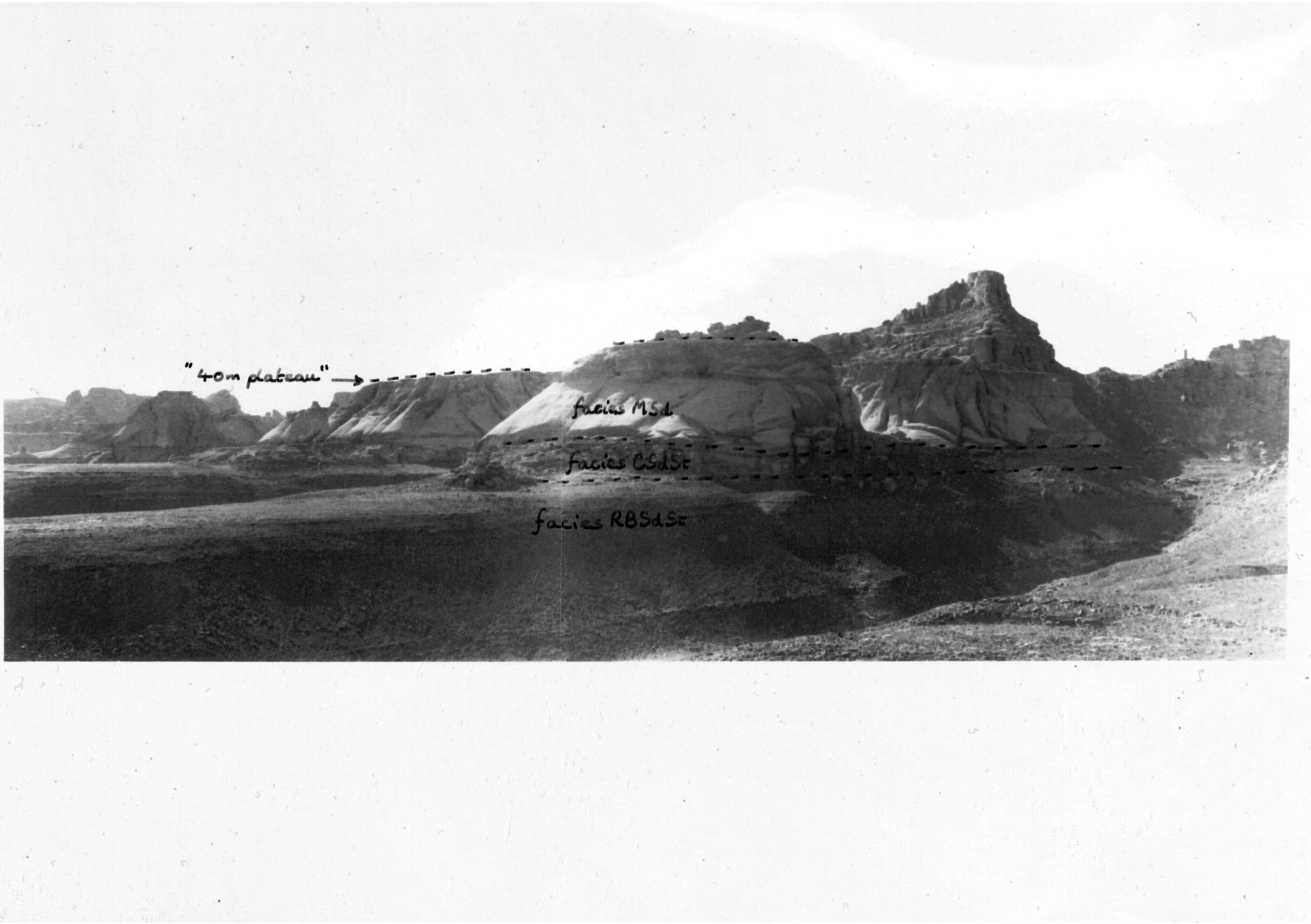


Figure 3.64 CHARACTERISTICS OF THE BASE OF TRANSGRESSIVE SEQUENCE 2

The basal sandstones of transgressive sequence 2 in the Upper Wadi Teshuinat (this view is from locality 11, Fig. 3.5) are predominantly weathered to light-coloured slick-rock slopes providing a sharp contrast to the rounded dark-coloured exposures of the underlying well-bedded sediments of facies RBSdSt. An apparent angular unconformity at the base of the slick-rock slopes is actually an irregular but nearly-horizontal geomorphological boundary that is subparallel to the bedding. The bedding sequence at this boundary, measured at a position in the middle of the photo-montage, comprises a conformable succession of sandstones, and subordinate siltstones of facies RBSdSt, CSdSt (subfacies 4), and MSd from the base upwards. The sediments of facies CSdSt which lie at the base of the slick-rock slope comprise very thin sandstone beds of subfacies 4 in which rare *Arthrophycus* is to be found. On the skyline the "40 m plateau" is visible.


"40m plateau"
facies MSd
facies CSdSt
facies RBSdSt


"40m plateau"
facies MSd
facies CSdSt
facies RBSdSt

within lacustrine deposits rather than mouth-bar deposits.

In some cases the succession at the base of the transgressive sequences shows no scour horizon or unconformity. This is characteristic of the base of sequence 2 (base of the Tadrart Formation) in Wadi Teshuinat (Fig. 3.64). Geomorphological features give the impression in some places of the presence of an unconformity when in reality only a disconformity exists. Similar disconformities and unconformities, characterised by scour surfaces, and evidence of large-scale channelling, were recorded by Ramsbottom *et al.* (1979) from the Carboniferous of England. These features lie at the base of transgressive sequences which have been interpreted as being the result of eustatic sea level changes.

Strong iron-oxide cementation of beds near the base of the transgressive sequences, and the development of Liesegang rings (Fig. 3.65), has been used as evidence for hiatuses and the existence of palaeosoils and subaerial palaeo-erosion surfaces (Beuf *et al.*, 1971; Freulon, 1964). It was found that many of the iron-rich beds studied petrographically had suffered such strong haematite cementation and replacement that it was difficult to determine how much of the haematite was of early diagenetic origin. It seems probable, however, that much of the iron enrichment is a late diagenetic phenomenon caused by recent subaerial exposure, and the precipitation of iron minerals from meteoric waters along fine-grained intercalations. The presence of these iron-enriched beds at the base of the transgressive sequences, often along fine-grained argillaceous horizons, may be related as much to their geomorphological position at the base of massive porous sandstone units, through which meteoric waters have passed, as to their stratigraphic position. Indeed the presence of *Arthrophycus*-bearing, strongly iron-oxide cemented, fine-grained beds in facies MSd indicates that these beds do not provide evidence either for a hiatus or for subaerial exposure.

The evidence, therefore, of the basal characteristics of the transgressive sequences suggests that the sediments lying between the major unconformities and correlative disconformities of the External Tassilian Sandstones were deposited both during periods of relative sea level fall, at which time preservation potential was extremely low, and succeeding periods of relative sea level rise, during which time preservation potential was much greater. The transgressive

Figure 3.65 IRON-ENRICHMENT AND LIESEGANG RINGS IN SANDSTONES NEAR THE BASE OF TRANSGRESSIVE SEQUENCES

Scale is provided in these illustrations by a hammer the position of which is indicated, where necessary, by arrows.

(α, β, ε) Upper, and lower right : Liesegang rings developed in sandstones at the base of transgressive sequence 1, Upper Wadi Teshuinat (Fig. 3.5) (lower right illustration), and in sandstones immediately beneath the base of transgressive sequence 2, locality 15, Tihererin (Fig. 3.5) (upper two illustrations).

(c) Lower left : extensive iron-oxide enrichment of sandstones 8 m above the base of transgressive sequence 2, Wadi Anshal (locality 12, Fig. 3.5).

(b) Lower middle : "tramline crusts" developed by iron-oxide enrichment and typically present at a number of different horizons in the basal sandstones of the transgressive sequences. This example is from 10 m above the base of transgressive sequence 2, locality 11, Wadi Teshuinat (Fig. 3.5).

A

B

C

D

E

nature of the sequences recognised has been invoked as evidence for stability and tectonic inactivity of the craton during the late Silurian/early Devonian, and it has been argued that the cyclicity of the succession cannot be convincingly explained by a retreating escarpment hypothesis. The palaeocurrent evidence indicates that the sediment source area for the sandstones of sequences 1, 2 and 3 lay to the southeast of the Jabals Tadrart and Akakus, possibly, at least in part, in the area of In Ezzan (Fig. 3.61). This suggests that at the time of deposition of the most northwesterly-located sandstones of facies MSd of these sequences (i.e. during low sea level stands) a fluvial plain must have extended over distances greater than 1000 km along depositional strike and 500 km down palaeoslope from the sediment source area (Fig. 3.61). During the ensuing sea level rise the coastal depositional setting is thought to have migrated landward over this extensive low-lying fluvial plain, leading to the preservation of fluvial deposits overlain by estuarine and shallow-marine sediments. At maximum sea level high stand the cratonic platform, which had been dominated earlier by blanket fluvial deposition, became an extensive shallow-marine environment. Subsequent sea level fall resulted in widespread erosion and the deposition of thin regressive sequences in some localised areas of greater relative subsidence. At sea level low stand fluvial conditions were re-established and a renewed cycle of transgressive sedimentation commenced. The depositional settings that appear to be characteristic of these Nubian-type sandstones are difficult to illustrate by reference to modern analogues, but it is hoped that, despite the equivocal nature of much of the evidence, this study of the processes by which these sediments were deposited, their facies, stratigraphy and regional distribution, may have justified the conclusions drawn.