Palynofacies and sedimentology of some Late Jurassic sediments from the British Isles and northern North Sea

Thesis

Palynofacies and Sedimentology of some Late Jurassic Sediments from the British Isles and Northern North Sea

by

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Earth Sciences

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I dedicate this work to the fond

memory of Dr. Robert Flett M.D.

For the kindness and hospitality shown
to me by him and his family during my
stay at the Links Hotel, Brora in the
autumn of 1978

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Abstract

The variation of physical, chemical and ecological conditions through stratified water bodies is shown to be a useful integrating tool in the palaeoenvironmental modelling of epeiric sea facies. Particular emphasis is placed on the implications of oxygen deficiency in meromictic water bodies for the deposition of 'black shales'. In epeiric settings preservation is probably a more important determinant of black shale deposition than primary productivity. Optimal preservation of marine carbon occurs in dysaerobic to anoxic bottom waters. Such facies can be identified by a synthesis of sedimentological, geochemical and palaeoecological data. In dysaerobic to anoxic environments the kerogen is dominated by amorphous material of marine, but indeterminate, origin. The proportion of terrestial organic matter correlates with grain size, proximity to fluvial sources and level of oxygenation. Palynomorph assemblages vary along gradients of hydrodynamic energy and proximity to fluvial sources. Dinocysts can be directly related to hydrographic stability; they are rare in basinal, bottom water, sediments except when redeposited.

The Oxfordian Piper Formation was deposited in an aerobic, hydrographically unstable, shallow shelf regime. Palynofacies trends suggest an approximately NW-SE onshore-offshore gradient. Apart from scale the Piper sequence is analogous to coeval onshore sediments at Brora. The <u>eudoxus-pectinatus</u> zones of the Type Kimmeridge Clay were deposited in a distal basin with an alternating mixed (aerobic-dysaerobic) and meromictic (dysaerobic-anoxic) watermass. The resulting cyclicity is analagous to Quaternary sapropel sequences. Coccolith limestone intercalations are interpreted as indicative of partial mixing of a meromictic watermass. A formal subdivision of the Type Kimmeridge Clay Formation is proposed. Kimmeridgian sediments of the Brora-Helmsdale outlier comprise dysaerobic-anoxic basinal shales with interbedded redeposited sandstones. A similar, but more proximal, facies association is observed in the Toni-Thelma

area. Late Jurassic distal, basinal, dysaerobic-anoxic shales occur in the Maureen field.

Acknowledgements

I am indebted to Robertson Research for the considerable resources that they placed at my disposal during the period of my C.A.S.E. Studentship. In carrying out all the geochemical and palynological processing involved in this project they enabled my research to be of a greater scope and depth than would have otherwise been possible. I also acknowledge N.E.R.C. for their general financial support, and in particular a six month extension of my grant period which enabled me to participate on Legs 76A and 77 of the Deep Sea Drilling Project. Grateful thanks are also due to Ian Chaplin and the staff of the thin section lab. at the Open University for their considerable efforts on my behalf.

I thank Occidental Petroleum and Phillips Petroleum and their partners for participating in my studentship by allowing me access to North Sea core material, and for approving the release of the findings of the relevant parts of this study. The opinions and interpretations expressed in Chapters Four and Seven of this thesis are entirely my own and do not necessarily reflect or agree with those held by employees of Phillips or Occidental Petroleum.

On a more personal note, I owe much to the continuing help, advice and hospitality of Dr. Les Riley, formerly of Robertson Research. I acknowledge the patience of my supervisor, Dr. Chris Wilson, and the encouragement and understanding of Prof. Brian Funnell has also been much appreciated. Last, but not least, I should like to pay my small tribute to some extraordinarily nice people who hardly ever get the credit or acknowledgement they deserve. My warm thanks go to Mrs. Hazel Owen (and Doreen!) and Mrs. Anne Wood and Sue Julier of the inter-library loans departments of the O.U. and U.E.A. libraries respectively. They may never share the academic limelight but their contribution is a very real one. On their dull and repetitive labours is my understanding built.

LIST OF ILLUSTRATIONS

CHAPTER ONE

Introduction

There are two main reasons why this thesis is of an unusual scope and length. Firstly, Robertson Research relieved me of the monotonous and repetitive task of palynological sample preparation. The time which would have otherwise been spent preparing samples was employed in making several extensive literature reviews on a number of topics, not all of which are included in this thesis. Secondly, although this studentship was initially confined to the Piper cores, it had, by the time of the 'first circular', already expanded to include the Type Kimmeridge Clay, and eventually incorporated the Kimmeridgian of the Brora-Helmsdale outlier and several Phillips cores from the northern North Sea. The involvement of Robertson Research allowed me to examine an unprecedented number of samples and provided me with valuable corroborative geochemical data. This work therefore combines the results of detailed field work, core logging, extensive microscopic investigations and a considerable amount of literature synthesis. I have attempted to give as much emphasis to the theoretical considerations and the interpretation of the data as I have to the purely descriptive aspects of my research.

The principal aim of this research project was to demonstrate the value of the multi-disciplinary approach in palaeoenvironmental analysis. More specific goals were to examine the potential of the palynofacies technique, to explore the relationship between the palynologist's and geochemist's view of kerogen, and to develop a coherent model to explain some of the major features of Late Jurassic sedimentation in the U.K. and northern North Sea. The object of the exercise was to develop models and techniques which could be successfully employed to assist in petroleum exploration by the oil industry. In developing useful models there is an inevitable trade-off between detail and perspective. I have attempted to provide a conceptual framework of the neritic marine environment which utilises the concept of watermass stability. The flexibility of this framework is suited to multi-disciplinary investigations and its fundamental basis provides predictive capability. The complexity of the model may be built up progressively depending upon the quantity and diversity of the available data, but the basic framework should be recognisable in any marine shelf sequence. Although particular emphasis has been given to epeiric sea facies and to source-rock sediments, the modelling has considerably wider applications than the aspects developed here.

From my early reading on modern anoxic basins and plankton ecology I rapidly became convinced that the concept of watermass stability had great potential as an integrating force in multi-disciplinary investigations of marine sediments. More extensive reading has only served to confirm this view. The strength of this concept lies in its ability to link benthic and pelagic ecology and to provide a meaningful correlation of physical, chemical and biological aspects of the marine environment. I have attempted to convey my belief in this approach and its significance in Chapter Two. A complementary review and synthesis of the literature relating to the palaeoenvironmental applications of palynology and kerogen studies is given in Chapter Three. An initially unexpected bonus was that modern advances in the ecology of cyst-forming dinoflagellates made it possible to relate palynofacies studies directly with the watermass stability model.

The scope of the Piper study was inevitably curtailed by the general expansion of the area of research. Considered in context, Chapter Four should be regarded only as an example of the inter-relationships between palynofacies and sedimentology in shallow marine sediments. The study of the 'middle part' of the Type Kimmeridge Clay Formation was undertaken for a number of different reasons (see Chapter Five). The variety of existing data available for the section allowed a partial test and demonstration of the multi-disciplinary technique. Furthermore, the convenient scale of the cyclic sedimentation of the organic-rich sediments and coccolith limestones was very useful in helping to formulate the conceptual links subsequently incorporated into the watermass stability model. Chapter Five is primarily a review, synthesis and reinterpretation of existing data, most of which has apparently escaped the attention of the geologic community - despite the great commercial interest in the Kimmeridge Clay! I hope it may encourage further work.

Chapter Six on the Kimmeridgian part of the Brora-Helmsdale outlier is almost a thesis in itself. The bulk of the chapter consists of rather routine outcrop descriptions. These might appear to be over-long, but anyone familiar with the complexity of the section will, I am sure, appreciate the value of the systematic locality accounts. Introduced to this section during a Robertson Research field course, I was impressed by the contrasts it exhibited with other Upper Jurassic sequences, and the almost complete absence of any recent account. The section was studied largely for comparative purposes and partly because some geologists were tempted to liken the Piper Formation with the Allt na Cùile and Lothbeg sandstones. The question was posed as to whether a palynofacies investigation could discriminate between the palaeo-environments of these sandstones. The answer is yes (see Chapter Eight). The study of the Phillips cores almost automatically followed on after that on the Brora-Helmsdale section. The Phillips cores, which had only recently been studied by Robertson Research, were obviously of a very similar fault-scarp related facies whose examination promised interesting comparisons and contrasts. The principal conclusions of this research project are presented in summary in Chapter Eight.

It could be convincingly argued that the eventual scope of this research project never represented a practical proposition for completion within the allotted time. However, all but the last quarter of Chapter Six and Chapters Seven and Eight <u>were</u> completed within the three year period. Delays in completion were predominantly of a logistical nature. Chapters Two and Five, and most of Six, were written during 1979 to 1980. Sadly they do not appear to be as prescient now as they might have been when written. Conceptually, however, the major reviews in Chapters Two and Three have aged very little; they were never intended to be simply catalogues of recent research. More up to date thoughts can be found in Tyson 1984a, 1984b*. I largely resisted the temptation to include new material once I had completed the text, but I occasionally substituted more recent references where it seemed appropriate and useful to do so. I have tried to cite only those references that I found particularly valuable and which would help the reader to reconstruct my line of thought for themselves. During the writing of this thesis I attempted to separate the data from the interpretation wherever it was practical to do so. This inevitably results in a degree of repetition but is probably better science.

Palynological data

Although a quantitative method was used to describe the palynomorph and kerogen assemblages of the sediments analysed during this study, the data is best thought of as semi-quantitative. This is because of the considerable uncertainties involved in the counting procedure which are discussed in part two of Chapter Three. Although the data so derived is very useful for delineating various trends, and can, for example, be qualitatively correlated with geochemical results, I do not believe it is suited to statistical analysis. If no clear trend appears to exist when the data is examined 'by eye' it is safest to assume that there is no trend. I believe the probability of creating artifacts through statistical treatment is very high when dealing with data of this kind of variability. This would be particularly true in the absence of organic carbon and other geochemical data, since there is always a temptation to think of the numbers as representing proportions of different kinds of organic matter rather than percentage particle abundances.

I should like to emphasise that the numbers in the tables and appendices are arithmetic means and that no significance should be attached to fractions or values less than 1%, other than that they confirm the presence of that particular component.

* Included here.

Abbreviations used in text

A.O.M.	Amorphous organic matter
B.O.D.	Biochemical oxygen demand
cmt	Cement (petrographic)
C.T.G.	Coarse-tail graded
D.O.M.	Dissolved organic matter
O.M.	Organic matter
Org.C.	Organic carbon (weight percent)
R.P.D.	Redox potential discontinuity.
S.D.	Standard deviation
Tp	Total palynomorphs (% of kerogen)
Tw	Total phytoclasts

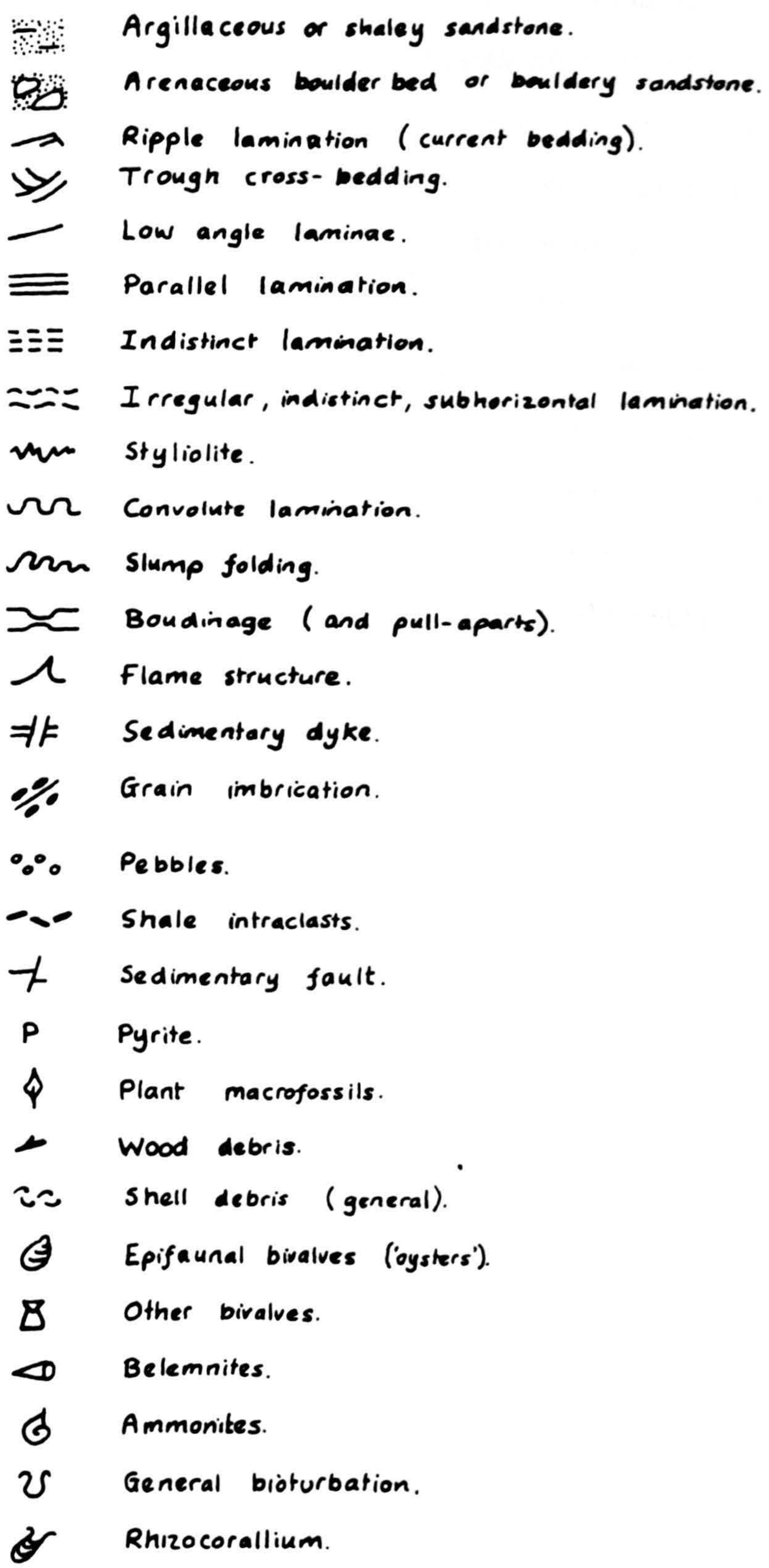


Fig. 1 Key to symbols used on lithological sections

CHAPTER TWO

The geological implications of watermass stratification and the nature of black shale palaeoenvironments

TERMINOLOGY

The terms aerobic, dysaerobic and anaerobic are used as defined by Rhoads and Morse (1971) to describe the various levels of oxygenation to be encountered in anaerobic basins. The three terms provide a scheme in which each level of oxygenation can be defined by its approximate range in dissolved oxygen concentration or by its faunal characteristics (biofacies). The term 'dysaerobic biofacies' is here used in a sense which encompasses both stable dysaerobic conditions (as in oxygen minima) and unstable conditions where the level of oxygenation shows small scale fluctuations around a dysaerobic mean value and the redox potential discontinuity (R.P.D.) is maintained very close to the sediment surface. The differences between the stable and unstable forms of the dysaerobic biofacies are discussed in later sections.

The widely used adjective 'oxic' is herein considered to be equivalent to "aerobic" (as defined above) and the term "hypoxic", which is most commonly encountered in the ecological and biological literature, is equated with dysaerobic and anaerobic conditions. The use of the word "anoxic" is here restricted entirely to waters devoid of dissolved oxygen and containing free H_2S (see Richards, 1965). The important distinction betweeen anaerobic and anoxic should be noted; while anoxic conditions are always anaerobic the converse is not necessarily true. In ecological terms anaerobic conditions are those with insufficient oxygen to support organisms with aerobic metabolisms; geologically they are conditions with insufficient oxygenation to support benthic metazoan organisms with fossilisable hard parts or organisms which are capable of bioturbating the sediment. Anaerobic is thus an ecological term,while anoxic is based more on chemical criteria. The normal consequence of sulphate reduction in surficial marine sediments is that anaerobic conditions are also normally anoxic; the only common exception to this is the core of oxygen minimum layers. The distinction between anaerobic and anoxic is probably more crucial in non-marine environments with limited sulphate availability (see anoxic sulfidic and non-sulfidic classification of Berner, 1981).

The well established genetic terms "gyttja" and "sapropel" are used here to describe the depositional conditions of black shales (see Hansen, 1959; Krejci-Graf, 1964; Calvert, 1976). Gyttjas are organic-rich deposits formed under aerobic, dysaerobic or anaerobic conditions, while sapropels are deposited only under anoxic bottom waters (corresponding only to the anaerobic biofacies). The terms gyttja and sapropel do <u>not</u> take into account the type of organic matter present in the sediment except by implication. The relationship of the various terms is demonstrated for a theoretical anoxic basin in Fig. 2.1.

There is no accepted nomenclature for organic-rich sediments and the terminology of fine grained rocks is confused in general. The largely undefined terms black shale, bituminous shale, kerogenous shale, oil shale, carbonaceous shale and sapropelite are in common usage but are used in different contexts by different authors. One of the major problems in creating a classification of organic-rich rocks is the variety of purposes for which a classification is required and hence the number of potential criteria involved (e.g. lithological, geochemical, economic, genetic, palaeoenvironmental, palaeoecological, etc.). Organic content itself is so variable that it is not suitable for constructing an elaborate terminology and a purely lithological classification is not, in my opinion, what is required. Since most geologists think of a "black shale" as being more than simply a shale which is black, it is apparent that the genetic aspect must be emphasised. The following scheme represents an attempt to construct a multi-disciplinary framework for the terminology of organic-rich sediments (Fig. 2.2).

Argillaceous rocks with over 3% organic carbon ($\sim$4% organic matter) are termed kerogenous mudrocks (the term mudrock is used as defined by Blatt et al. 1972). These kerogenous mudrocks are divided by their

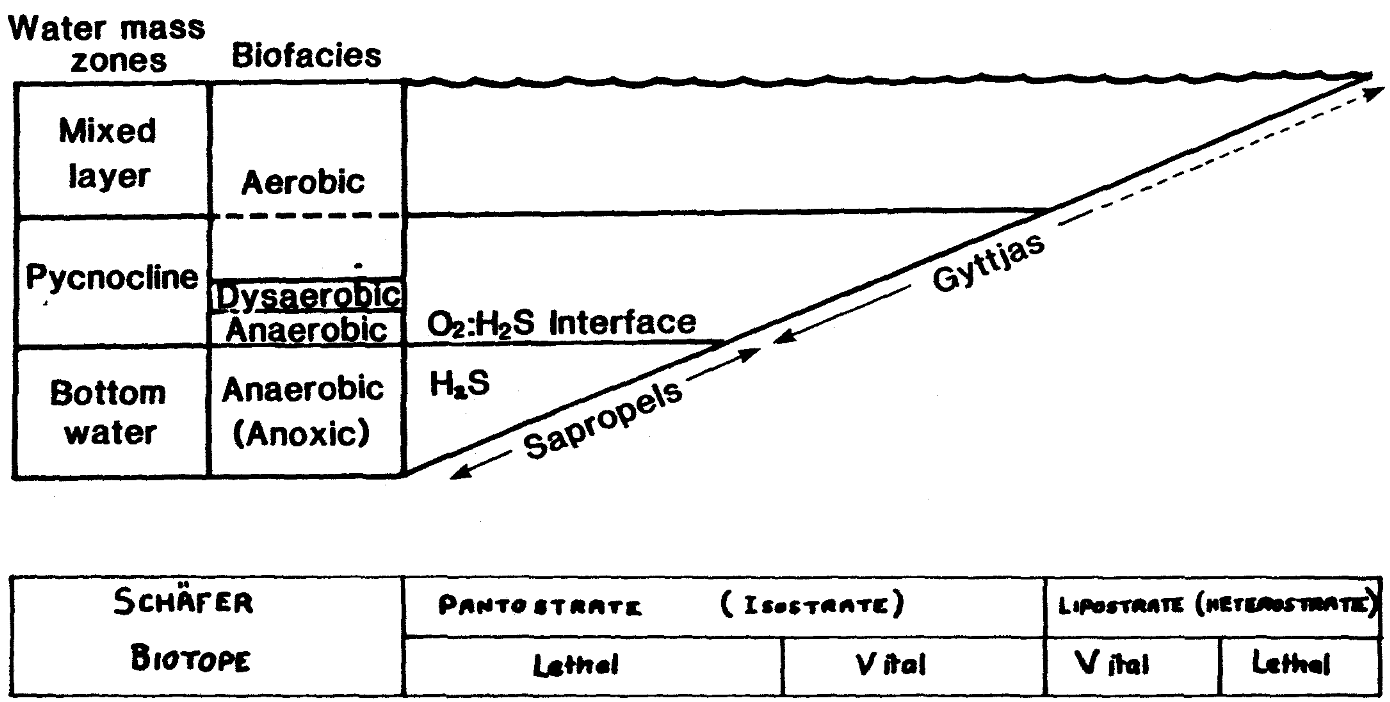


Fig. 2.1 Terminology applied to stratified, oxygen-deficient basins.

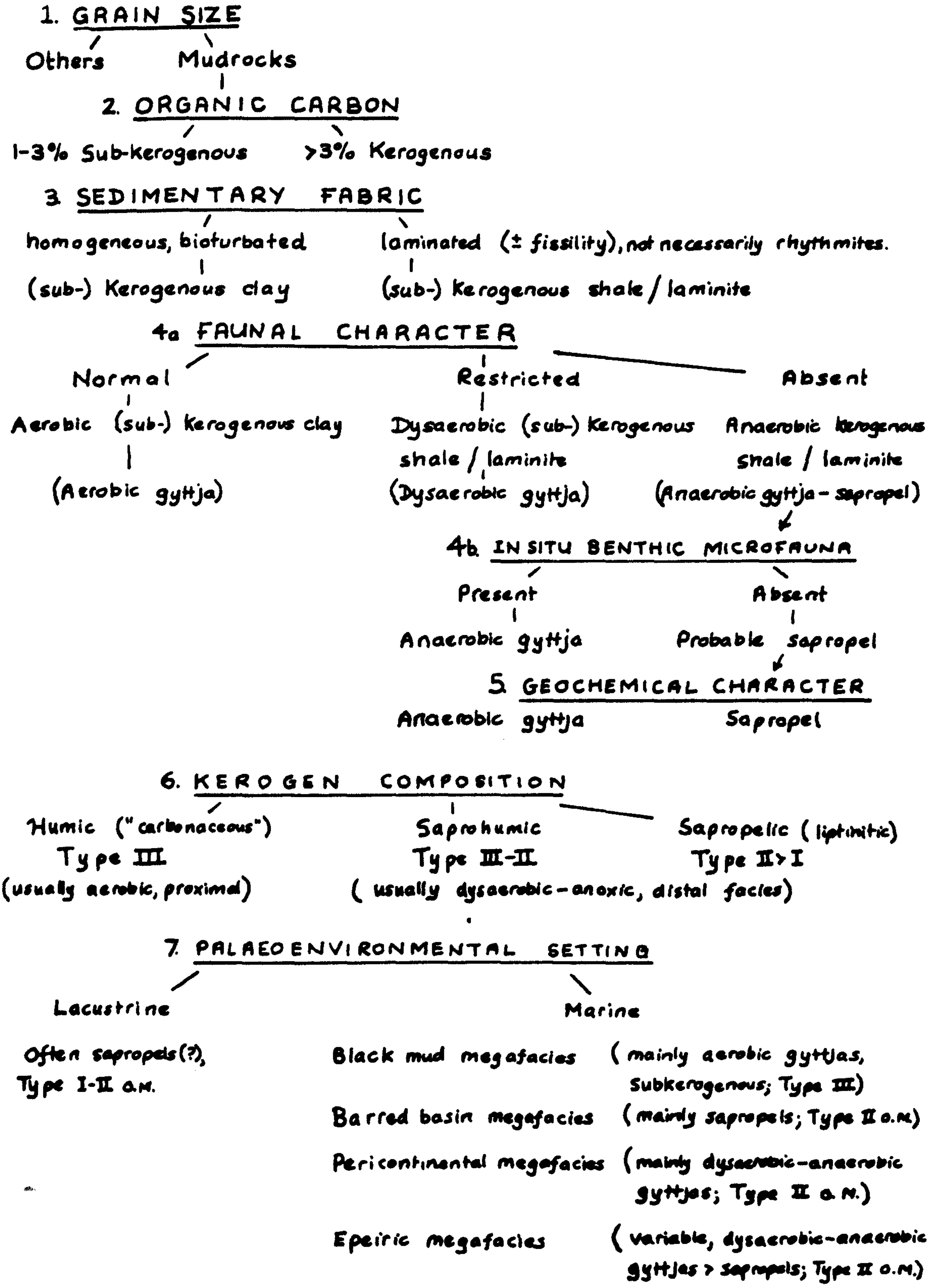


Fig. 2.2 An attempted genetic classification of organic-rich mudrocks.

fabric into kerogenous shales and kerogenous clays. The kerogenous shales are distinguished by their planar fabrics and fissile character, and the clays by their homogeneous nature and lack of fissility. Byers (1974) has shown that fissility in these rocks is determined by the presence of laminations,which in turn is determined by the presence or absence and extent of bioturbation. The black (greyish, brownish or greenish black) kerogenous shales may be termed black shales (or more precisely black kerogenous shales). The kerogenous clays are clays because they have been bioturbated and hence were deposited under aerobic conditions and are gyttjas. The kerogenous shales include both dysaerobic gyttjas (fairly well laminated, fissile, minor bioturbation), anaerobic gyttjas and (anoxic) sapropels (entirely laminated, no benthos or bioturbation). The fissility of highly kerogenous anaerobic-anoxic sapropels may in fact be impaired by the large organic content which has a binding effect on the rock,and so lamination is the principal criterion. In the case of organic rich calcareous rocks (limestones or marls) the suffix "shale" may be replaced by "laminite" and (black) kerogenous mudstone laminites or (black) kerogenous micrite laminites substituted for (black) kerogenous shales, although their essential genetic equivalence should not be forgotten. Lithological prefixes such as silty, pyritic, etc. may be added where appropriate but it is not the intention of this classification to provide a precise lithological definition.

The remaining criterion used is kerogen type. This is very informative for numerous reasons which are discussed in full in Chapter Three. Kerogen is the organic component of sedimentary rocks which is insoluble in ordinary organic solvents. The amount of kerogen in a rock is not the same as the total organic matter content,although the two are sufficiently closely related to allow the use of organic carbon values as the basis for giving a sediment the prefix "kerogenous". There are three principal types of kerogen:

(i) Sapropelic kerogens: largely derived from plankton and spores, much of it is amorphous; it also includes waxy cuticle material from higher plants. Rich in hydrogen.

(ii) Humic kerogens: derived from the tissues of higher plants, including vitrinite (wood), semi-fusainite (partly carbonised wood) and colloidal humic gels. Poor in hydrogen.

(iii) Inertinite: inert carbon (totally carbonised wood).

This three fold division is very simplistic and more detailed classifications are discussed in Chapter Three. Note that cuticle debris, although like wood derived from higher plants, is grouped with the sapropelic kerogens because of its geochemical affinities. From palynological examination the relative proportions of these different kerogens can be used to derive a prefix for the rock which describes its kerogen content (see Fig. 2.3). This situation is, however, complicated by the variable composition of the amorphous materials present in organic-rich rocks, which must be regarded only as potentially sapropelic in composition. Care must be taken not to confuse the term sapropel in its genetic sediment sense and sapropel as a kerogen term; the use of the same word is unfortunate but this double-usage is strongly entrenched in the literature, particularly since sapropel sediments often contain sapropelic kerogens. The prefix "sapropelic" is only used here in its kerogen sense. Sapropel as a sediment term should be restricted to its genetic definition and not applied in a mere lithological sense (note misuse by D.S.D.P.).

Detailed definitions and discussions of the terms aerobic, dysaerobic and anaerobic are given later in the chapter where the environmental "megafacies" classification of organic-rich sediments shown in Fig. 2.2 is also explained.

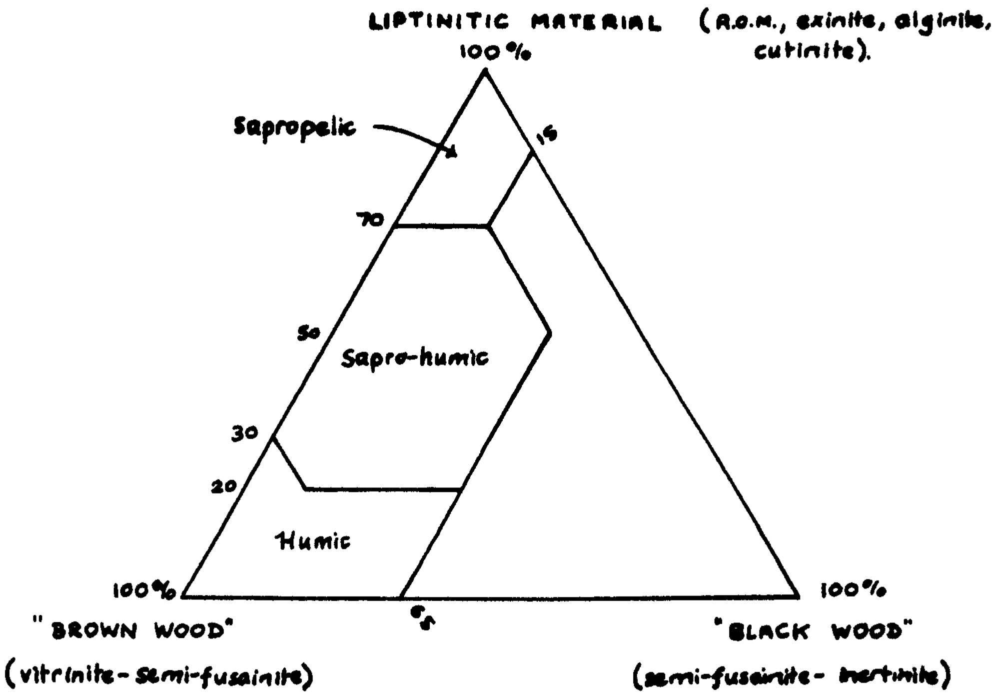


Fig. 2.3 Derivation of prefix describing kerogen character in immature sediments.

SHELF SEA HYDRODYNAMICS

A basic familiarity with shelf sea hydrodynamics is essential if we are to understand the conditions under which neritic black shale basins evolve, for as molecular diffusion alone is a very inefficient process, the initial distributions of temperature, salinity, dissolved gases and nutrients in the water column are primarily controlled by the distribution of turbulent mixing processes. Swift (1976) has recently shown that a convenient way to approach shelf sea hydrodynamics is to consider it in terms of the three major flow strata: the upper boundary layer (i.e. the mixed surface layer), the core flow and the lower (benthic) boundary layer (see Fig. 2.4).

In the upper boundary layer there are several potential sources of turbulent mixing which maintain the surface mixed layer in its characteristically homogenous condition:

(i) the breaking of waves

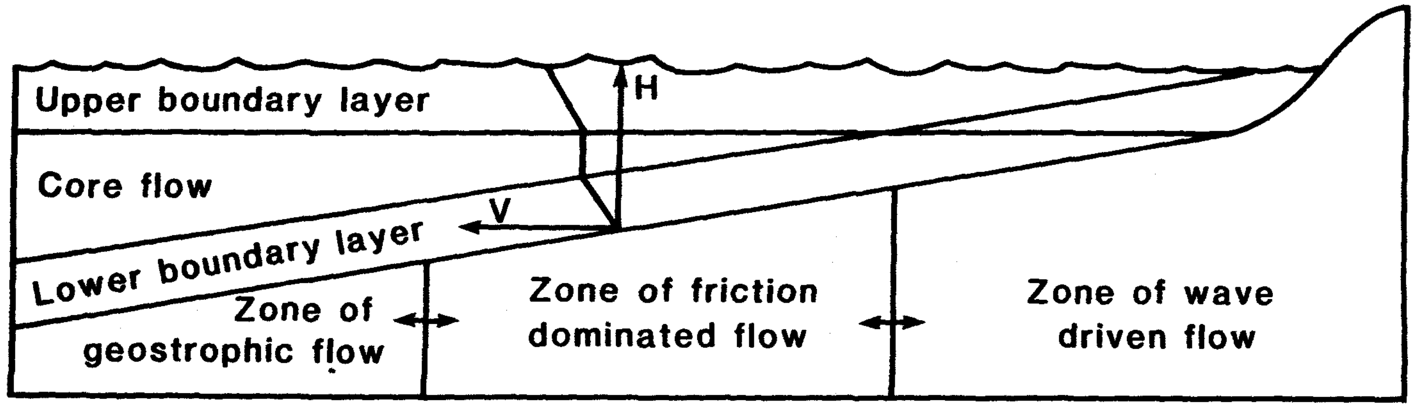
(ii) vertical shear associated with Ekman transport

(iii) wind driven Langmuir circulation

(iv) seasonal (and diurnal) thermal convection

Wind generated orbital wave motion is negligible at depths equivalent to the wavelength, and is less than 5% of its surface value at half this depth (Zenkovitch, 1967). Since the maximum wavelength of open ocean waves is about 300m, the influence of wave activity rarely exceeds 150m (except perhaps in severe storms) and in seas of restricted fetch, such as the Baltic and Black Seas, is generally less than 50m (Kukal, 1971). Wind stress at the sea surface also results in a vertical veolocity gradient in the upper boundary layer and a net transport of the surface water layer (Ekman transport). Under the influence of the Coriolis force the surface water motion trends at 45° to the right of the wind direction and, as this movement is transmitted downward by vertical shearing stresses, it decreases in velocity and deviates successively further to the right (an effect known as the Ekman

A



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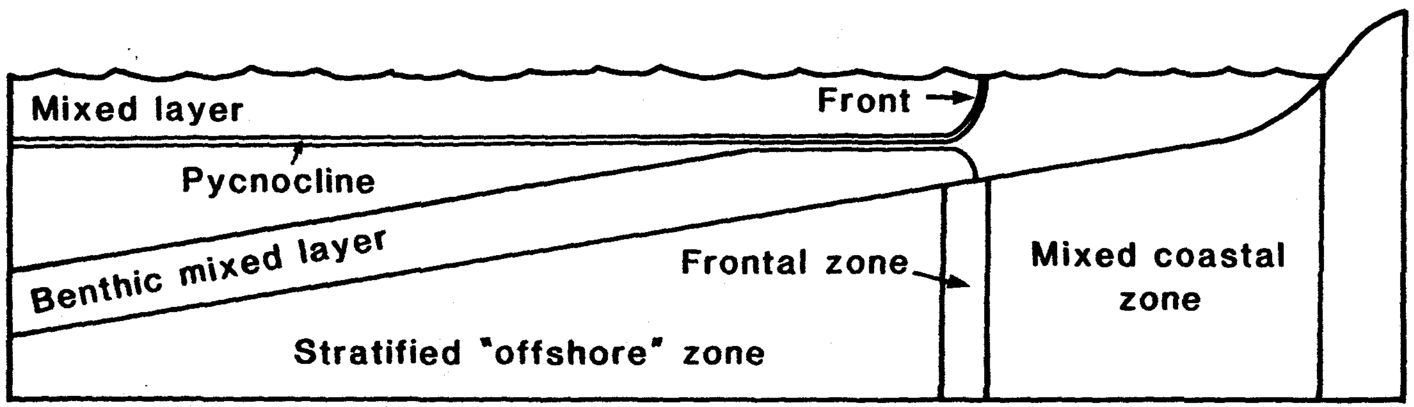


Fig. 2.4 Hydrodynamic structure of shallow shelf seas. A after Swift (1976), showing velocity profile with depth.

spiral, see Sverdrup et al. 1942). The base of the upper boundary layer is defined as the "depth of frictional influence" where the direction of flow has been so deviated that it is directly opposed to that at the surface. Shear instabilities associated with the Ekman velocity structure are an important source of turbulent mixing.

Above a certain wind speed the velocity structure described above becomes unstable and alters to a system of horizontal helical vertices (aligned more-or-less parallel to the mean wind direction) which is known as Langmuir circulation (Langmuir, 1938). Thermal convection resulting from density inversions due to diurnal and seasonal variations in insolation also contributes to the mixing in the upper boundary layer. Below the upper boundary layer is the core flow in which the water flows in a slab-like fashion with little vertical shear and hence little turbulent mixing (Swift, 1976). The lower boundary layer is formed by the frictional retardation of the core flow as it moves across the sea floor and exhibits a reverse Ekman spiral (where the deflection due to the Coriolis effect decreases away from the sea bed). Flow instabilities and bottom roughness effects in the lower boundary layer generate turbulence which can lead to appreciable mixing of the water column in shallow nearshore areas, e.g. tidal mixing in the English Channel (Pingree, 1978).

Swift (1976) has used the relative importance of the three flow strata in the water column to define three lateral 'hydraulic provinces' (see Fig. 2.4). The boundaries of these hydraulic provinces are determined by the flow conditions and therefore show a great deal of temporal and spatial variation, including a displacement seawards during storms. Readers are recommended to consult Swift (1976) and other papers in Stanley and Swift (1976) for a more comprehensive treatment of these aspects of shelf sea hydrodynamics.

Since (a) the distribution of physico-chemical properties within the watermass is controlled by the distribution of turbulent stirring processes, and (b) the character of the surface layer is modified by contact with the atmosphere, water columns which are deeper than the boundary layer usually show a marked vertical heterogeneity or stratification. As can be seen from Fig. 2.4, this stratification will be inhibited in the nearshore zone by wave turbulence and tidal mixing in the benthic boundary layer (Pingree, 1978) leading to the development of shallow sea frontal systems between the stratified and non-stratified waters (Simpson & Pingree, 1978). The factors influencing the development of stratification may be demonstrated by considering the seasonal stratification cycle typical of all water bodies in temperate latitudes (see LaFond, 1954: Tully, 1964; Sundaram & Rehm, 1971, 1973; Dietrich et al. 1980; and Fig. 2.5).

Under the present climatic conditions, mid-latitude shelf seas are characterised by a seasonal thermal stratification. Heating of the surface layer beginning in spring produces a warm surface layer which expands downward due to the turbulent entrainment of the underlying water. During the early summer increased insolation and reduced turbulence result in the formation of a strong thermal discontinuity (or thermocline) at the base of the mixed zone. Eventually a point is reached where the buoyancy contrast between the warm, low density surface layer and the underlying colder, denser water,reaches an equilibrium with the turbulent mixing processes and the water column is stabilised. As the density gradient (or pycnocline) at the base of the mixed layer becomes more intense it is the site of increasing current shear, reinforcing the tendency for the surface circulation to be decoupled from the bottom water. In the stratified condition the pycnocline marks the base of the upper boundary layer; only slight, patchy turbulence occurs in the pycnocline associated with the breaking of internal waves (Gregg, 1973). The average depth of the seasonal shelf pycnocline-thermocline ranges from 15-40m but varies regionally due to local variations in the intensity of mixing processes (La Combe, 1974).

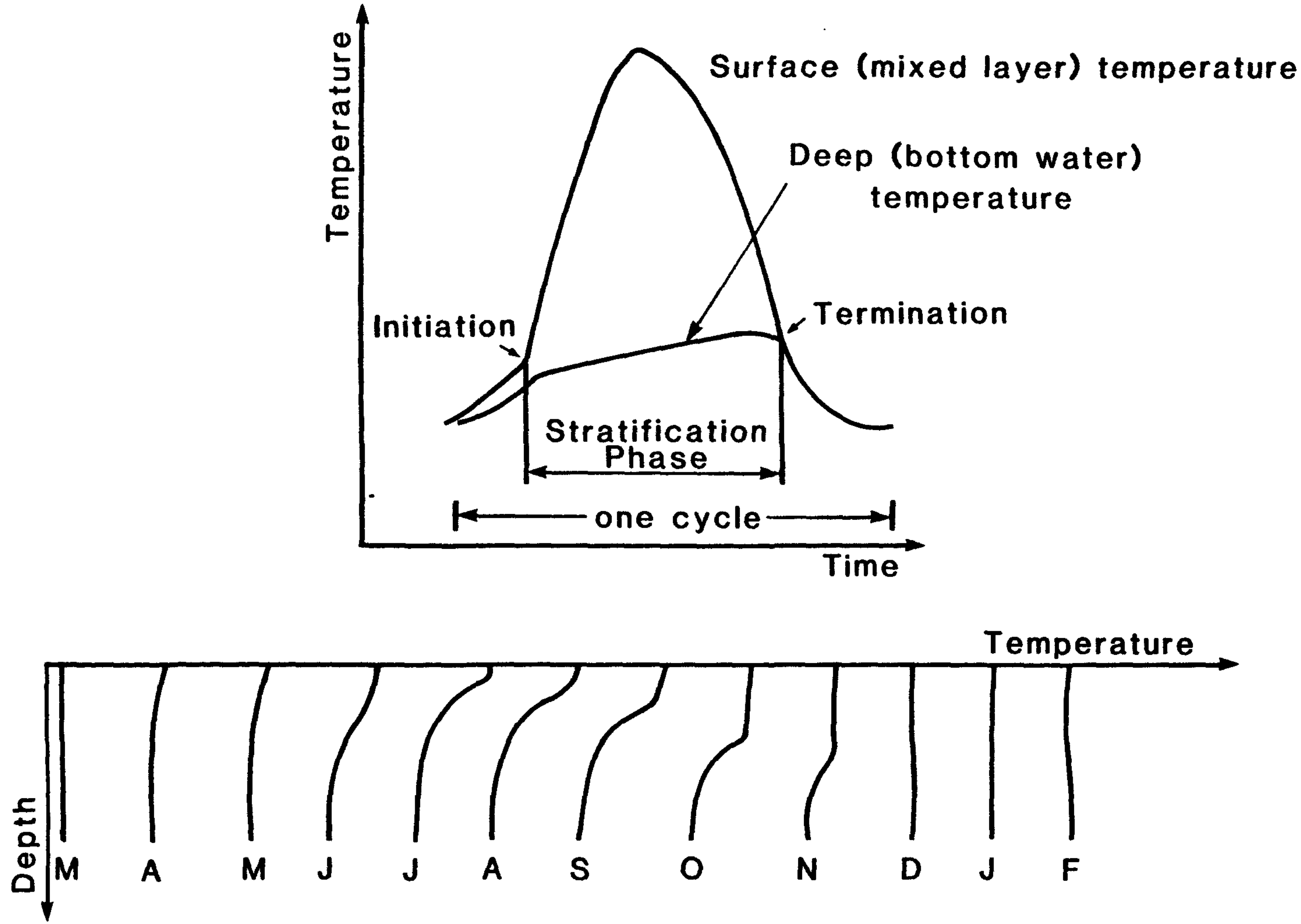


Fig. 2.5 Annual thermal stratification cycle typical of shallow temperate water bodies (after Sundaram & Rehm, 1971, 1973).

In the autumn storm mixing and Langmuir circulation combine with convectional overturn produced by surface cooling to produce a deepening of the mixed layer and erosion of the thermocline-pycnocline. In a normal water column density increases with depth and the situation is stable; the stronger the vertical positive density gradient the more work that must be done to overturn the watermass. Since temperature is the main determinant of the density gradient in most water columns the interplay of turbulent mixing and heat input is of prime importance (see Fig. 2.6). Although the temperature distribution is initially determined by the degree of turbulent mixing, the stabilising effect of the buoyancy input due to surface heating represents a feedback mechanism which actually leads to the reduction of the mixed layer depth during stratification phases. The development of stratification is thus an autocatalytic phenomenon and if it were not for seasonal cooling and storm mixing, stratification would be a permanent feature of water bodies in temperate latitudes.

In a stratified watermass three main layers may be distinguished (Fig. 2.7).

(a) The mixed layer (homogenised by turbulent mixing)

(b) The transition or discontinuity layer (characterised by sharp gradients in temperature, density, salinity, water chemistry, etc.)

(c) The bottom water layer (generally characterised by only slight changes with depth; a benthic mixed layer may or may not be present).

Because the depth of the discontinuity layer is hydrodynamically determined, the thermocline, halocline and pycnocline will all tend to occupy more-or-less the same position in the water column (unless for example the halocline has an independent origin). However, because of the non-uniform relationship between the temperature and density of water, at higher temperatures pycnoclines may occur which are not represented by obvious thermoclines. Since the density distribution influences the depth of turbulent mixing, which in turn controls the distribution of

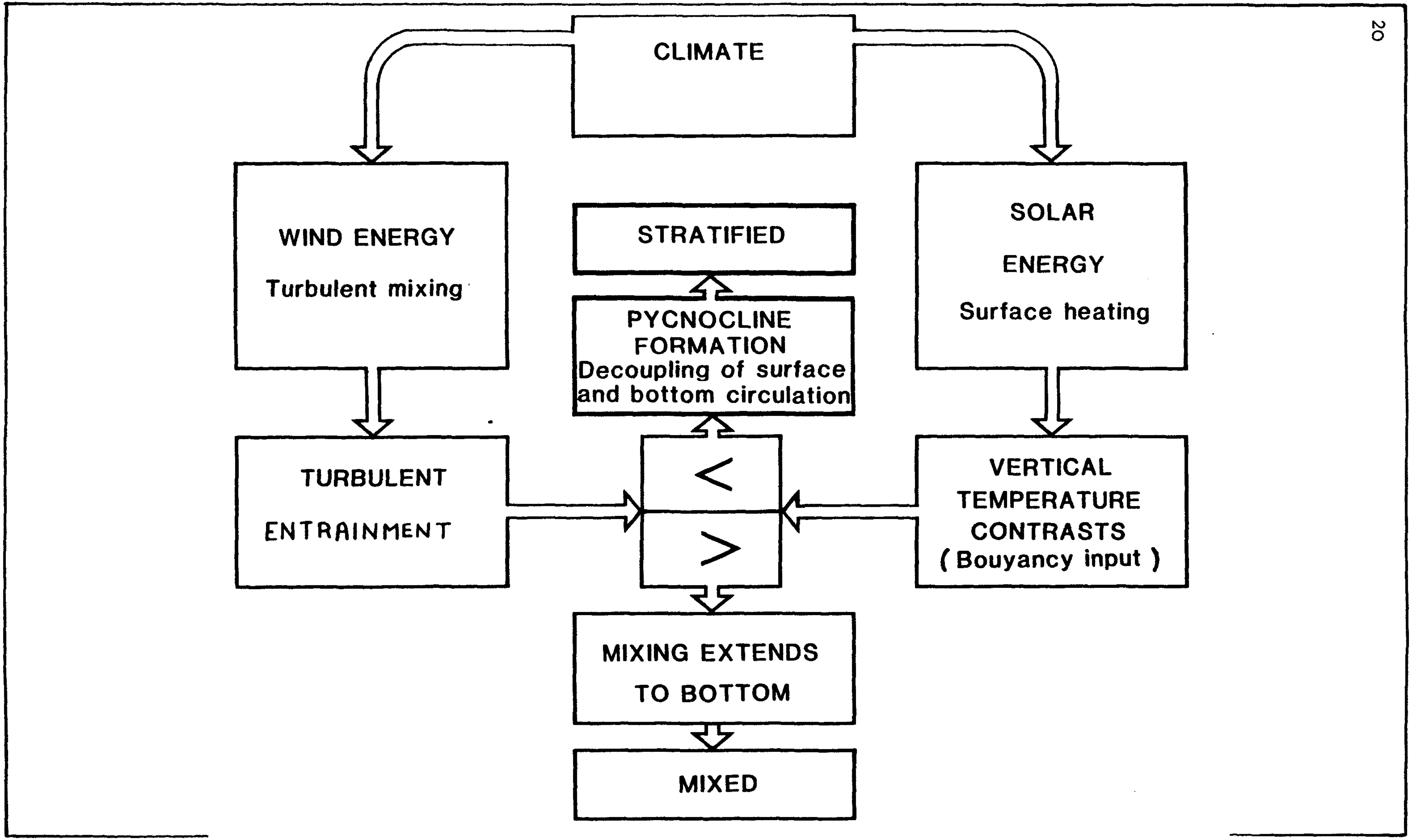


Fig. 2.6 Schematic of factors controlling thermal stratification in shallow temperate water bodies.

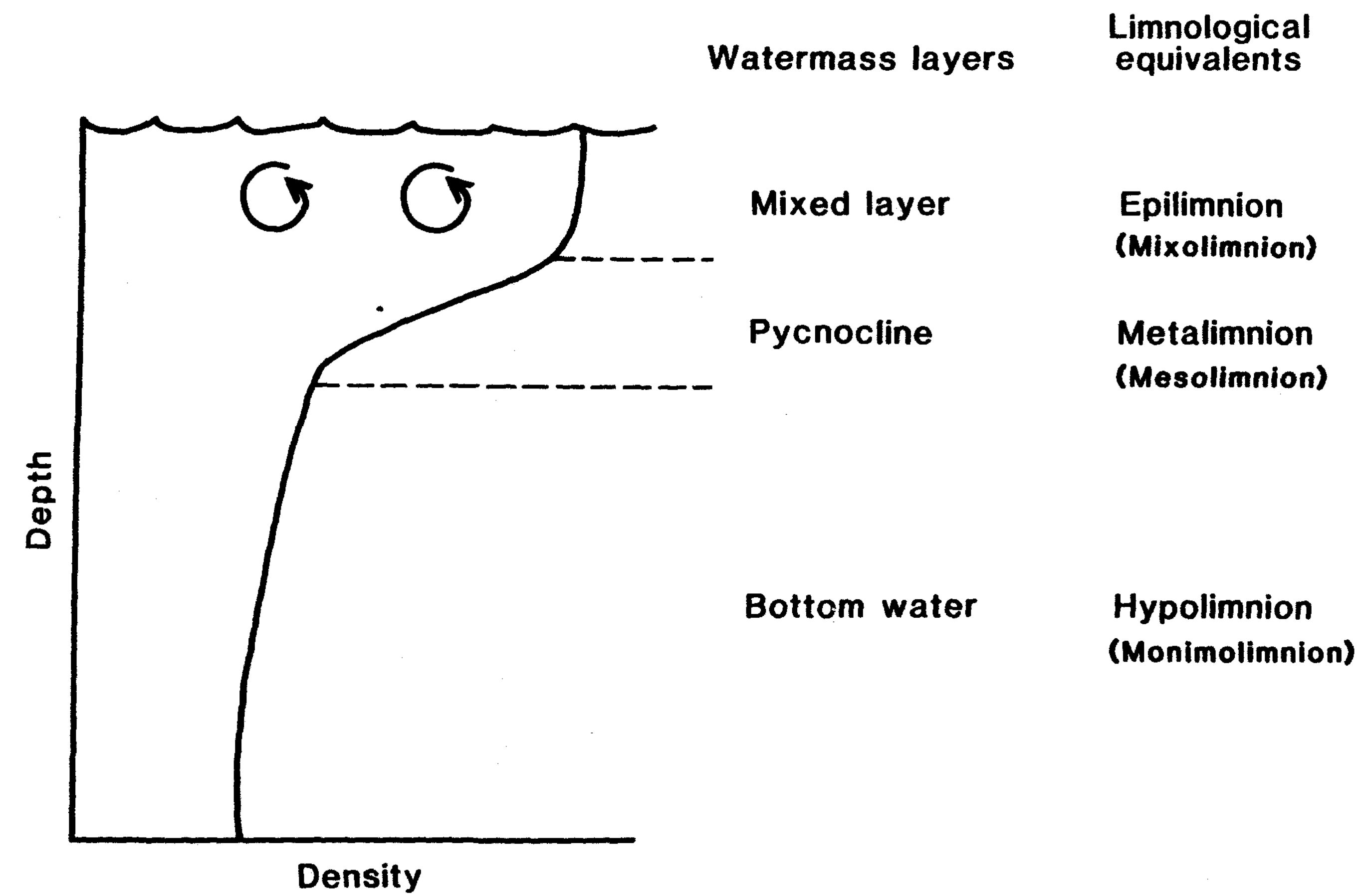


Fig. 2.7 Terminology of stratified water bodies. Limnological terms in parentheses are those applied to meromictic water bodies (from Hutchinson, 1957).

other physico-chemical properties, the pycnocline is by far the most significant level in the water column. The pycnocline may be defined as the layer corresponding to the depth interval with the highest (positive) density gradient. The strongest pycnoclines are usually those associated with haloclines, but since these tend to be restricted to temperate land-locked basins and low salinity plumes off river mouths, thermo-pycnoclines are generally more important. The pycnoclines of modern anoxic basins located in temperate latitudes result from strong haloclines (which may be either augmented or diminished by temperature) while those in tropical latitudes may result almost wholly from thermoclines (Richards, 1965).

The increase in the stability of the water column produced by the presence of a pycnocline greatly decreases the efficiency of the vertical mixing processes, significantly reducing the vertical transfer of heat, solutes and oxygen (Armstrong & LaFond, 1966; Pingree & Pennycuick, 1975). If vertical circulation is sufficiently restricted (depending on the intensity and stability of the pycnocline), and lateral circulation impeded (e.g. by sills), the rate of utilisation of oxygen due to organic decay in the water column and at the sediment-water interface may exceed the rate at which oxygen is mixed downward from the surface. Progressive deoxygenation of the bottom water will result. Such deoxygenation is a common and well documented phenomenon in the hypolimnion of stratified eutrophic lakes (e.g. Beadle, 1974), and has been utilised in the palaeoenvironmental interpretation of lacustrine black shales (e.g. Bradley, 1948; McLeroy & Anderson, 1966; Lowe, 1976; Boyer, 1981).

Permanent stratification (i.e. where there is only partial vertical mixing throughout the year) is termed meromixis (see Hutchinson, 1957). In a meromictic watermass the anoxic bottom water zone is termed the monimolimnion and the overlying mixed part the mixolimnion (see Fig. 2.7). The mixolimnion itself may be seasonally (thermally) stratified and divisible into a epi-, meta- and hypolimnion above the monimolimnion. This is the case in both the Baltic Sea (Fonselius, 1970) and the Black

Sea (e.g. Dickman & Artuz, 1978) where the permanent stratification results from salinity contrasts and the mixolimnion is thermally stratified during the summer. As such, the temperature structure controls the stability of the mixolimnion. Two types of meromixis are recognised by Walker and Likens (1975):-

(a) Ectogenic meromixis: external inputs of saline or freshwater or turbidity produce stratification.

(b) Endogenic meromixis: thermal stratification (rendered permanent by location and situation of basin) plus "chemical stratification contributed by biological processes in deep water".

Several authors (e.g. Kuenen, 1968; Schmalz, 1969; Seibold, 1970; Brongersma-Sanders, 1971; Grasshoff, 1975) have emphasised that water balance and circulation pattern also play important roles in the formation and maintenance of semi-enclosed, marine anoxic basins. There are three types of circulation pattern which are relevant to the disucssion of black shale environments (see Fig. 2.8).

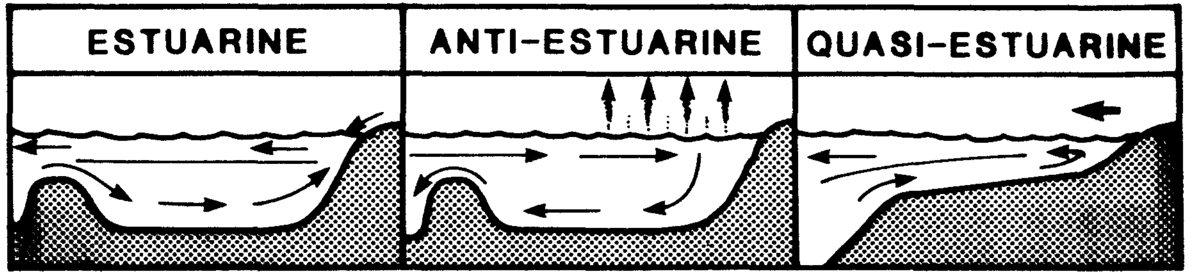
(i) Estuarine circulation

(ii) Anti-estuarine circulation

(iii) Coastal upwelling (coastal Ekman divergence) or quasi-estuarine circulation

Estuarine circulation occurs in estuaries and in semi-enclosed basins with a positive water balance where the combined rates of precipitation and runoff from the surrounding catchment area exceed loss by evaporation plus inflow of oceanic water (Grasshoff, 1975). This situation produces an outflowing, lower density, lower salinity surface layer which is compensated by the subsurface inflow of oceanic water which is drawn upward at the basin head. The strong halocline at the base of the surface layer inhibits vertical circulation,and where a shallow sill sufficiently restricts lateral circulation with the open sea, the basin will tend to become anoxic. Most of the world's best known anoxic basins (e.g. see Table 16.1 in Deuser, 1975), including the

Figure 2.8 Major circulation types of marginal seas (after Seibold, 1970; Grasshoff, 1975)



	ESTUARINE	ANTI-ESTUARINE	QUASI-ESTUARINE
CLIMATE	Humid/temperature	Arid/tropical-subtropical	Tropical-subtropical
WATER BALANCE	Positive	Negative	Wind forced
WATER EXCHANGE	Surface outflow, bottom inflow	Surface inflow, bottom outflow	Offshore surface drift and upwelling
SURFACE WATER CHARACTER	Lowered salinity, high nutrients	High salinity, low nutrients	Slightly lowered salinity, high nutrients
BOTTOM WATER CHARACTER	Normal salinity, high nutrients	'High'salinity, moderate nutrients	Slightly lowered salinity, high nutrients
BOTTOM OXYGENATION	Moderate to poor	Good	Poor
WATERMASS STRATIFICATION	Strong halocline	Poorly defined	Strong inclined pycnocline
PRIMARY PRODUCTIVITY	High	Low	Very high
BOTTOM ORGANIC CONTENT	High	Low	High to very high
CLASTIC SEDIMENTATION	High (but variable)	Low	Low
BASIN TYPE	Marginal land-locked seas	Marginal land-locked seas	Eastern margin of ocean basin

Baltic and Black Seas, are salinity stratified, land-locked basins with shallow sills and estuarine circulation.

Anti-estuarine circulation occurs in semi-enclosed basins with a negative water balance, i.e. where loss by evaporation exceeds the total freshwater input (Grasshoff, 1975). Evaporation increases the density of the surface seawater which sinks (despite being warmer) and forms a saline outflow from the basin which is compensated by a surface inflow from the open ocean. In this situation nutrients are being constantly removed from the euphotic zone and carried out of the basin by the underflow so primary productivity is usually low (e.g. the modern day Mediterranean). Since little organic matter is produced in the basin and little derived from the generally arid coastal areas, these conditions do not really favour black shale deposition. Schmalz (1969), however, suggested that the ponding of very saline bottom water in a silled basin may sufficiently increase the stability of the water column to result in stagnation and produce a pre-evaporite euxinic phase of sedimentation (see also Roth, 1978; Arthur & Natland, 1979). Stagnation will, however, only result in a black shale deposit if sufficient amounts of organic matter are being provided to the sea floor. The Orca Basin in the northern Gulf of Mexico is in fact anoxic due to the stabilising effect of dense, hypersaline bottom water, though in this case geochemical evidence indicates the brine was produced by solution of a sub-seabed salt deposit and not by evaporation (Shokes et al. 1977; Trabant & Presley, 1978). Jordan (1974) proposed a model for European Jurassic and Cretaceous bituminous shales which used dense hypersaline bottom waters formed through halokinesis and sea bed solution of salt deposits to produce anoxic bottom conditions and hence black shales (in much the same way as in the present day Orca Basin). Although this model has recently been revived in a slightly modified form by Degens and Paluska (1979) it is considered to be over-elaborate and its general applicability is in doubt.

Unlike estuarine and anti-estuarine circulation, coastal upwelling is not restricted to semi-enclosed basins. The circulation pattern is similar to that in estuarine circulation but is due to wind forcing rather than water balance and is often called quasi-estuarine circulation (Brongersma-Sanders, 1971). Sustained, wind-driven coastal upwelling occurs in subtropical latitudes on the west coasts of continents where persistent equator-ward winds produce an offshore movement of water in the upper boundary layer (the Ekman transport described earlier - see Fig. 2.4). The mass deficit produced at the coastline by the offshore movement of the surface water is compensated by the upwelling of water (from depths of $\sim$100-200m) into the nearshore surface layer (Smith, 1968; Barber & Smith, 1981). Mooers et al. (1978, p.53) have conveniently summarised the relationship between wind pattern and upwelling as follows: "A wind pattern favouring upwelling generally occurs on a seasonal basis as the atmospheric subtropical high pressure zone intensifies over the ocean and moves poleward in summer. As a consequence, there may be at least weak upwelling year round in low latitudes, but only seasonal upwelling in mid-latitudes, which spreads poleward as the summer season advances. The poleward edge of the sub-tropical high constitutes a boundary between the so-called westerlies and the easterlies, and is consequently a locus for storm tracks; i.e. atmospheric cyclones and anticyclones. Correspondingly, the poleward extremity of the seasonal coastal upwelling zone is marked by low persistence or high intermittency. This intermittency results in transient coastal upwelling or so-called coastal upwelling wind event cycles". Additional useful comments on the relationship between meteorological patterns and upwelling may be found in Cushing and Dixon (1976) and Hartline (1980).

The upwelling water is generally nutrient-rich and results in a high level of primary productivity which, with other facts favourable, may lead to the deposition of an organic-rich sediment (e.g. off S.W. Africa:

Calvert & Price, 1971). The subsurface water tapped during coastal upwelling is often oxygen depleted itself, originating from an oxygen minimum layer formed at levels of low eddy diffusion by the oxidative demand of the organic debris settling from the highly productive surface waters. Where these oxygen minimum layers intersect the continental slope they commonly result in a belt of organic rich, laminated sediments (Calvert, 1964, 1966; Murty et al. 1969; Gross et al. 1972; Von Stackelberg, 1972; Dow, 1978; Mattick et al. 1978; Diester-Haass, 1978). Coastal upwelling systems, though commonly producing anaerobic conditions, do not usually result in persistent anoxic bottom water. Both oxygen minimum zones and coastal upwelling have been used extensively in the interpretation of black shale palaeoenvironments in the Palaeozoic (e.g. Brongersma-Sanders, 1965, 1966, 1969; Heckel, 1977; Berry & Wilde, 1978; Legget, 1978) and the Mesozoic (e.g. Frush & Eicher, 1975; Schlanger & Jenkyns, 1976; Thiede & van Andel, 1977; van Andel et al. 1977; Fischer & Arthur, 1977; Arthur & Schlanger, 1979). For further discussion see later under "pericontinental megafacies".

PRODUCTION, SEDIMENTATION AND PRESERVATION OF ORGANIC MATTER

Factors controlling the accumulation of organic matter

The most conspicuous characteristic of black shales is that they are abnormally enriched in particulate organic matter. Black shales commonly contain 3-15% and in extreme cases, as much as 70% by weight of organic matter. The factors which influence the organic content of sediments have been discussed at some length (e.g. Bitterli, 1963; Hallam, 1967; Gross et al. 1972; Arthur, 1979; Cornford, 1979) and may be listed as follows:-

(1) The total amount of organic matter potentially available in the basin of deposition. This may be divided into:-

(a) the primary production of marine organic matter within the basin (essentially plankton and its degradation products plus marine macrophytes)

(b) the amount and source (and therefore character) of allochthonous organic matter entering the basin of deposition.

(2) The rate of sedimentation of the particulate organic matter (the likelihood of organic material being incorporated in the sediment increases as its transit time through the water column and active zone of the sea floor decreases).

(3) The rate of consumption of organic matter within the depositional environment and during early diagenesis; a function of:-

(a) chemical oxidation (Eh)

(b) biological consumption (e.g. feeding of zooplankton, nekton and benthos and bacterial degradation).

(4) The rate of clastic sedimentation (relative to that of the organic matter). The so-called 'clastic dilution factor'.

(5) The stability of the environment (i.e. the length of time for which favourable conditions persist).

Primary productivity

The controls on primary productivity may be discussed with reference to only three principal parameters: light intensity, nutrient availability and water column stability (e.g. Lorenzen, 1976; Pingree et al. 1977; and also Strickland, 1965; Whittle, 1977 and Berger, 1976 for general aspects). The water layer in which there is sufficient illumination to support active photosynthesis is know as the euphotic zone and varies from 5-45m in relatively turbid nearshore waters up to 150m in clear, tropical waters. The base of the euphotic zone is called the compensation depth and is the level at which the amount of oxygen produced by photosynthesis only equals that required by plankton respiration. Neither compensation depth or mixed layer depth (the latter also controlling nutrient availability) are constant and temporal and spatial variations must be taken into account.

The interplay between light intensity, nutrient availability and water column stability may be demonstrated by discussion of the seasonal plankton cycle typical of temperate shelf seas (which is intimately related to the stratification cycle described earlier). Only the main features of this cycle are described below (see also Fig. 2.9) and further details may be found in Strickland (1965); Wood (1965); Tait, (1968); Cushing and Walsh (1976); Reynolds (1980).

In the winter mixing occurs throughout the water column, but because of the poor light intensity and the turbulence (which results in the phytoplankton cells being carried out of the euphotic zone before they have chance to divide), the phytoplankton cannot exploit the available nutrients. In the spring (i.e. late February to early April) increased insolation and reduced wind strengths lead to the development of the seasonal thermocline-pycnocline (as described earlier) which increases the stability of the water column. This increase in stability allows the phytoplankton to remain near the sea surface utilising the improved light intensity and nutrients mixed into the surface layer during the winter to produce the 'spring bloom'. The rapid growth of the phytoplankton soon depletes all the available nutrients in the surface layer, which with extensive zooplankton grazing, leads to a delcine in primary productivity. The products of the phytoplankton bloom settle through the pycnocline and are remineralised, the regenerated nutrients trapped and accumulated. Higher levels of productivity will be maintained in shallow sea frontal systems (see Fig. 2.4) where sub-pycnocline water rises to the surface (Pingree et al. 1975; Simpson & Pingree, 1978). During the summer months the phyto-plankton may concentrate in the vicinity of the thermocline (unless light is limiting) to exploit the limited amount of nutrients that may be mixed upward through the pycnocline, e.g. associated with internal wave turbulence (Pingree et al. 1977; Jacques, 1974). During the onset of increased turbulence in the autumn partial mixing may provide enough nutrients to the surface layer, yet maintain sufficient stability to

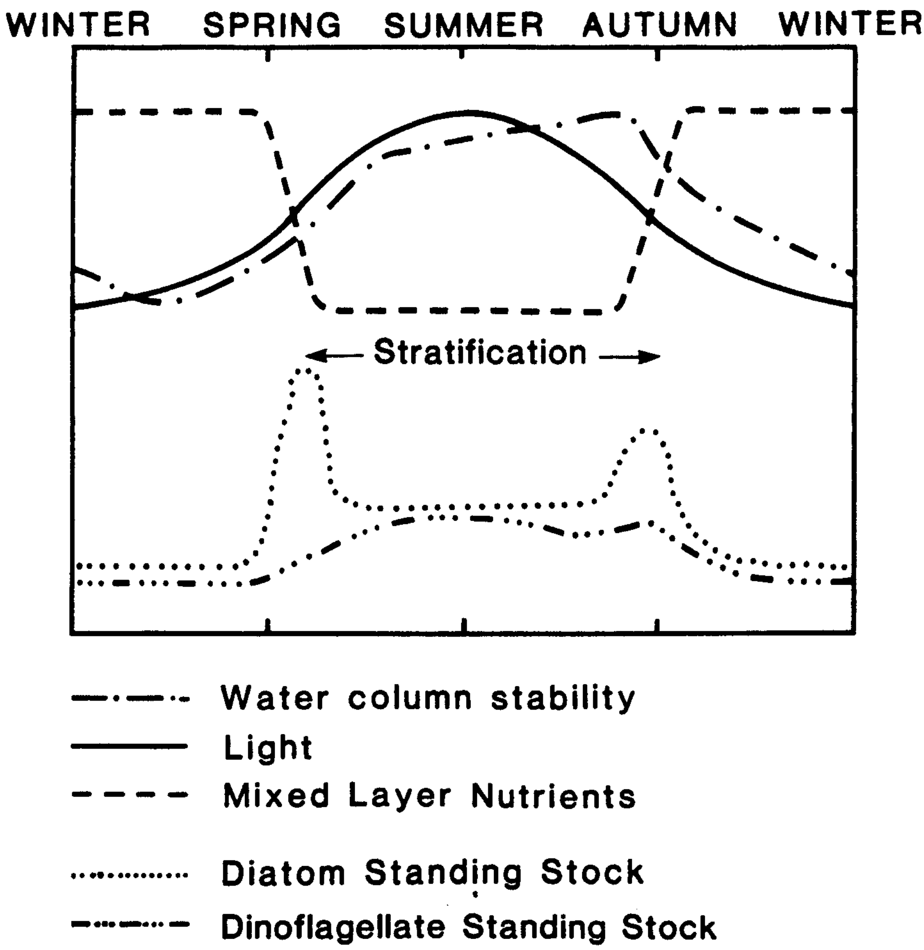


Fig. 2.9 Annual productivity cycle of temperate water bodies.

permit an autumn bloom. If such blooms do occur they are usually very short lived because of the weakening light and increased turbulence and expansion of the mixed zone. During the whole cycle aerobic conditions are maintained on the sea floor due to turbulent mixing in the benthic boundary layer.

The relative importance of each of the three parameters light intensity, nutrient availability and water column stability, varies with latitude. Light, for example, is more-or-less constant in the tropics, but like water column stability is seasonally limiting in temperate and high latitude seas. In stable, clear, tropical waters (where the euphotic zone is at its deepest) the main control on productivity is nutrient availability (Riley & Chester, 1971; Berger, 1976). Because the seasons are less marked in the tropics there is low,but more-or-less continuous,plankton growth throughout the year, maintaining the surface layer in a constant state of nutrient depletion. Most tropical water columns, other than those in areas of divergence or upwelling, are low in nutrients and are said to be oligotrophic. As in temperate seas in summer, phytoplankton and bacteria are often concentrated near the pycnocline rather than in the top few metres of the sea (Hobson & Lorenzen, 1972; Jacques, 1974; Lorenzen, 1976; Sorokin, 1978). If we set aside areas of upwelling as special cases then temperate seas are generally the most productive. Coastal waters are always more productive than offshore waters, a phenomenon known as the 'landmass effect' which is partly due to the supply of nutrients from the land and partly due to the shallower water depths (which imply greater mixing and better nutrient recycling).

Some authors (e.g. De Buisonje, 1972; Gallois, 1976 and Hemleben, 1977) have proposed that massive red tide-type phytoplankton blooms may be significant processes in the deposition of certain types of sediments, including coccolith limestones and black shales. Because these blooms can be associated with the mass mortality of fish and shell fish they are of considerable commercial importance and much has been written on them (see Ryther, 1955; Brongersma-Sanders, 1957; Wood, 1965; Rounsefell & Nelson, 1966; Wyatt & Harwood, 1974; and Steidinger, 1973, for reviews upon which the following discussion is based). The principal characteristics of red-tide type phytoplankton blooms may be summarised as follows:-

(1) they are unusually large blooms of phytoplankton which usually consist of dinoflagellates but may also be due to diatoms, blue-green algae (such as Oscillatoria = Trichodesmium; see Sournia, 1970; Qasim, 1972; Jayaraman, 1972) or sulfur oxidising bacteria (e.g. Caumette & Baleux, 1980).

(2) they are most common in sub-tropical seas.

(3) they generally occur in the summer months under conditions of calm and high light intensity.

(4) they are short term events (on a time scale of weeks rather than months).

(5) some bloom organisms secrete metabolic exotoxins which may lead to mass mortality among fish or shell fish (paralytic shell fish poison or PSP).

Although there are a great many observations recorded in the literature there seems to be a lack of consensus over what factors control the development of red tides. There do, however, appear to be two common factors, these are:-

(a) their association with periods of increased water column stability, and

(b) their association with various scales of upwelling.

The necessary water column stability to allow these blooms to develop may be provided by haloclines associated with river mouths or by thermoclines developing under calm, warm conditions of reduced circulation. Nutrient enrichment may occur in association with upwelling or supply of nutrients from land in river waters. Ryther (1955) has suggested that the plankton are significantly concentrated by currents and do not represent a bloom in equilibrium with the nutrient conditions where they are found- they are often found concentrated in long wind rows which suggests Langmuir circulation may play an important role in this respect.

The exact conditions required to produce a red-tide type bloom will vary according to the particular ecological requirements of what ever bloom organism is involved. Dinoflagellates, which are the most common causes of red-tides, are autotrophic, motile plankton and are therefore better adapted to low nutrient conditions (Steidinger, 1973; Landry, 1977). This explains why they are most abundant in temperate seas during the summer and why they are the most numerically abundant group of plankton in tropical and sub-tropical seas (Wyatt, 1976). As we have noted above, red tides are associated with high water column stability (including the cessation or reduction of upwelling and development of stratification), and stratification is associated with low nutrient availability. The hydrologic conditions resulting in red tides therefore favour dinoflagellates, which because they are motile plankton, are also more capable of exploiting subsurface concentrations of nutrients (e.g. see Packard et al. 1978). To allow massive phytoplankton blooms to develop zooplankton grazing must be suppressed. The rapid development of red tides and their short duration may give little opportunity for a large zooplankton population to develop; zooplankton may also be limited by dinoflagellate toxins or other ecological mechanisms (e.g. see Wyatt & Harwood, 1973).

From the literature which has been examined it would appear that these blooms are frequently associated with regions where stratified, stable water adjoin relatively turbulent nutrient-rich upwelling water (e.g. at zones of coastal Ekman divergence or in shallow sea frontal zones; see Fig. 2.10). The following model may be proposed. In the absence of other plankton, dinoflagellates multiply in the stratified areas and may migrate (or be moved by currents) towards frontal zones

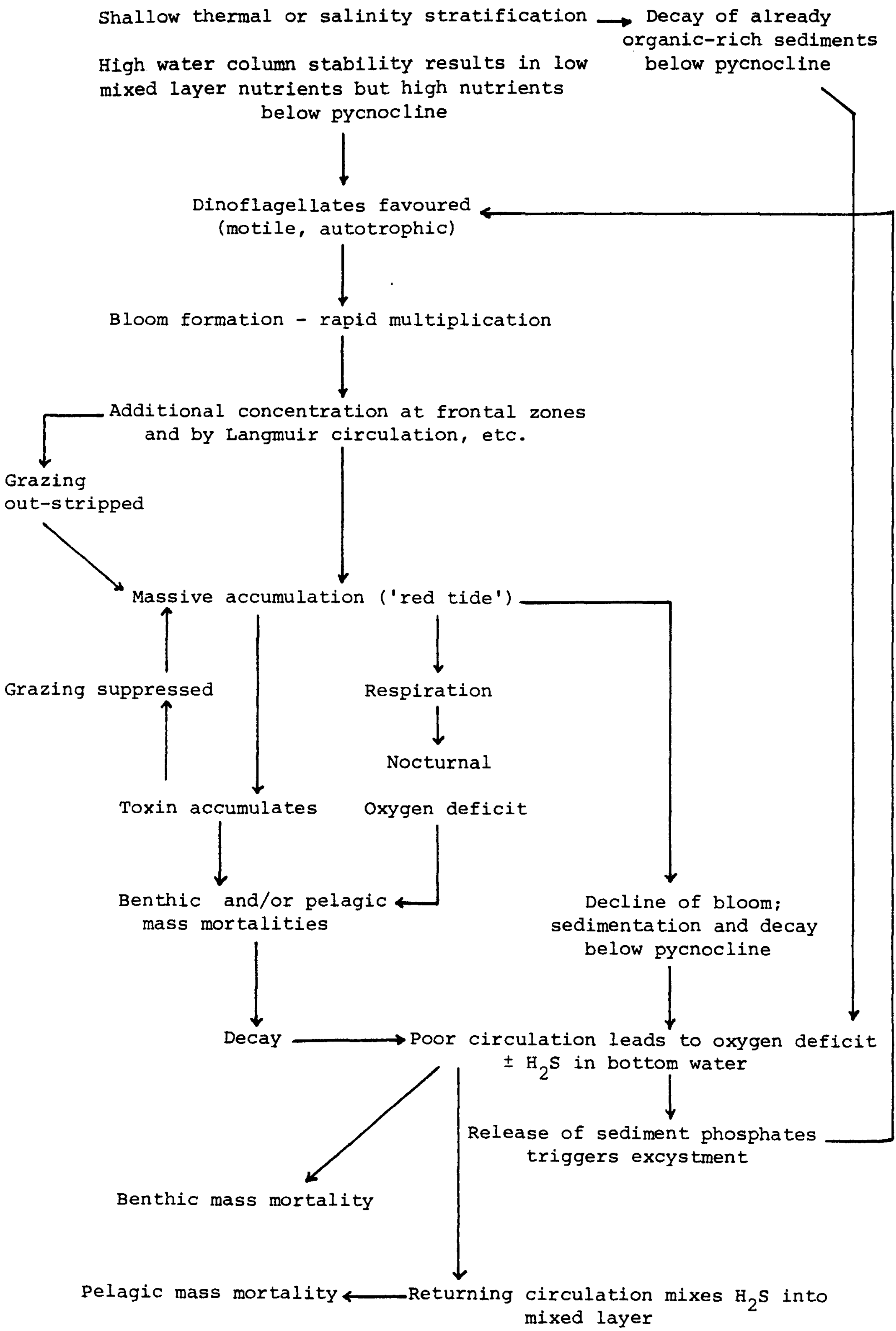


Fig. 2.10 Schematic of inter-relationship between hydrography, massive plankton blooms and mass mortalities

where they may become concentrated. The vicinity of frontal zones (where thermoclines or haloclines 'outcrop' at the surface) are well known to contain denser populations of plankton (e.g. Pingree et al. 1975; Simpson & Pingree, 1978; Bowman & Iverson, 1978). Apparently such blooms may eventually become so large that exotoxin concentration or nutrient or light limitation cause the death of the bloom, but a change in circulation pattern and water column stability may have the same effect. The decay of the bloom plankton (and other organisms which may have been poisoned by exotoxins) may produce widespread deoxygenation and further mass mortality due to the mixing of H_2S into the surface layer (Brongersma-Sanders, 1957). The calm conditions that result in the initial stability of the water column may also act to prevent the dispersal of H_2S generated from the decay of previously deposited organic-rich sediments. The returning circulation may mix this H_2S into the surface layer and there cause mass mortality of a second origin (this appears to occur in Walvis Bay - see Copenhagen, 1953). Mass mortalities due to rapid overturn and mixing of H_2S-rich bottom waters have been well documented in some Norwegian fjords (Strom, 1939) and stratified, anoxic tropical lakes (Beadle, 1966, 1974; Denny, 1972). The scale of the mass mortalities will depend upon the initial degree of anoxicity and the rate of overturn. Bottom water deoxygenation during red tides appears to be enhanced by the rapid deposition of readily biodegradeable planktonic organic matter via zooplankton faecal pellets (see Seki et al. 1974; Tsuji et al. 1974). However, bottom water deoxygenation may be partly responsible for the initiation of the red tide in the first place. Some degree of deoxygenation is a common feature of bottom waters during summer stratification (e.g. see Armstrong, 1977; Gade & Edwards, 1980; Taft et al. 1980) which coincides with a rise in the R.P.D. in the sediments. Lowered Eh conditions, the presence of sulphides and the release of ferric bound phosphates may trigger the excystment of red tide dinoflagellate hypnozygotes present in the sediment and hence the red tide itself (see Ilsuka, 1972; Ishio & Kondo, 1980).

In normal temperate seas the alternation of stratification with periods of mixing produces an overall high but strongly seasonal primary productivity. The nutrients which have been regenerated and accumulated during the winter (because of low light intensity) and summer (because of stratification) are utilised in the spring and autumnblooms respectively. The summer stratification phase in such systems is normally one of low productivity dominated by dinoflagellates. In meromictic systems, however, nutrients are accumulated and remain trapped in the bottom waters (and sediments) reaching much higher concentrations than in normal systems. During mixing episodes, when the pycnocline will be slightly lowered, the amount of nutrients released and the corresponding peak in primary productivity are, therefore, potentially higher in meromictic systems. Primary productivity will be determined by the scale and frequency of mixing events and by the relative positions of the nutricline (corresponding to the chemocline and usually associated with the main pycnocline) and the photosynthetic compensation depth. When high nutrient concentrations occur below the main pycnocline <u>and</u> in the euphotic zone the conditions are optimal for the dinoflagellates and the blue-green algae (or cyanobacteria, cyanophytes) and eutrophication and high productivity are the usual result provided light is not limiting. The planktonic blue-green algae are particularly good at exploiting such conditions; they concentrate at the level of the main pycnocline utlising light from above and nutrients from below (Fogg, 1972) as do dinoflagellates, but they are also capable of nitrogen fixation and therefore less dependent on mixing for their essential nutrient supply. Heavy summer blooms of blue-green algae are typical of eutrophicated water bodies (as in the present day Baltic Sea; Jansson, 1980; Leppakoski, 1980) and to a certain extent alternate with dinoflagellate blooms, the two forms probably capable of ecologically and chemically inhibiting each other depending upon which ever is dominant at any one time. Since the decay of these large blooms causes oxygen depletion and recycles the nutrients, eutrophic meromictic systems will tend to be self maintaining and the anoxic conditions sustained for as long as the hydrographic regime remains the same. Where the nutricline (which may either correspond with, or be below, the main pycnocline) occurs below the euphotic zone, nutrient trapping will be extreme and surface productivity very low and dependent on mixing events. Clearly it is the hydrography which controls the hydrochemistry and plankton dynamics and not the reverse, and if anything, massive plankton blooms are a symptom rather than the cause of 'black shale conditions'. It should be appreciated that regardless of whether they are eutrophified or not, meromictic systems are always less efficient than their normal counterparts since nutrients are being largely trapped unused in bottom waters or sequestered in sediment.

Sedimentation of organic matter

The consistent relationship between fine grained sediments and high organic contents is due to the hydrodynamic equivalence of most organic particles to silt and clay grade clastics (Trask, 1939). Most organic debris either bypasses, or is winnowed out of, coarser nearshore sediments and deposited further offshore, as is partly reflected in the offshore increase in the ratio of organic to inorganic material in the total suspended matter of shelf waters (e.g. Manheim et al. 1970). In high energy, turbulent areas where coarse grained sediments are deposited, the combination of sorting, mechanical comminution and high oxidation result in very low organic contents (<1%), but where hydrodynamic conditions permit the accumulation of fine grained sediments in the coastal zone, organic contents are locally high, particularly in estuaries or similar situations where there is a large input of allochthonous organic matter.

The organic content of modern shelf sediments generally increases offshore as the grain size of the sediments decreases, then declines as the supply of organic matter diminishes seawards. This relationship can be seen in many of the world's modern shelves where the maximum organic contents occur on the outer shelf and upper continental slope as a result of textural parameters and also due to the presence of oxygen minima (e.g. Dow, 1978; Gross et al. 1972, etc.). Kulm et al. (1975) in their work on the Oregon shelf, suggested that an important way in which organic matter bypasses the inner shelf may be offshore movement over inclined isopycnal surfaces, a mechanism which may explain why the most organic-rich sediments often occur seaward of the most productive waters (see also Pak et al. 1980). This is probably a significant process, as organic matter (both as detritus and plankton) is often observed to accumulate in the vicinity of the pycnocline. Stratification has already been shown to have a significant effect on the site of deposition of organic particles in lakes (Davis & Brubaker, 1973).

Many black shales show micro-laminations which consist of couplets of mineral-rich laminae (clastic particles or micro-plankton skeletons) and darker organic-rich laminae,which clearly demonstrate regular changes in the transport and or production of material. Reineck and Singh (1973) have termed such microlaminated sediments rhythmites. Bradley (1931) has interpreted organic-mineral couplets as annual increments of sediment (i.e. 'varves'), where one laminae reflects the background sedimentation and the other a seasonal peak in primary production. However, Calvert (1966) has pointed out that there is an alternative possibility where biogenic laminae represent the background sedimentation and darker laminae represent seasonal sediment inputs. Strongly seasonal productivity is characteristic of temperate latitudes while uniform productivity is more characteristic of tropical regimes. Where the predominant plankton are coccolithophores, diatoms or radiolarians the light coloured planktic laminae are often organic-poor because of the dilution factor due to the skeletal material and the darker clay-rich laminae are relatively organic-rich. Where only organic-walled plankton are present rhythmites consist of kerogen-clay couplets (e.g. see Bitterli, 1963). The seasonal precipitation of inorganic aragonite that occurs in some lakes can also lead to rhythmite deposition (see Kelts & Hsü, 1978).

Micro-laminations produced by the above mechanisms will only be preserved where ecological factors (and dissolved oxygen in particular) inhibit the bioturbating activities of the macro-benthos. Consequently rhythmites are most common in distal, basinal sediments deposited under anaerobic conditions where physical and biological reworking is absent or insignificant. The preservation of micro-laminations under anaerobic bottom conditions has been excellently demonstrated by Calvert (1964) for sediments deposited within the oxygen minimum zone in the Gulf of California. Hulsemann and Emery (1961) show similar sindings for anaerobic silled basins in the same area. Simola (1979) gives an example of diatomaceous rhythmites from a stratified anoxic lake.

Dickman (1979) and Dickman and Artuz (1978) have suggested a somewhat different hypothesis for the origin of dark organic-rich laminae in rhythmites deposited in anoxic basins. They consider seasonal expansions of the mixed layer (and hence lowering of the pycnocline) to result in the mass mortality of pelagic anaerobic sulphur bacteria which are known to proliferate in the uppermost parts of the anoxic zone (i.e. at the O_2:H_2S interface; see Sorokin, 1978). The subsequent sedimentation of these dead bacteria is believed to be a significant contribution to the organic content of dark layers in rhythmites. If this hypothesis is correct sedimentation of the declining phytoplankton stock in late summer would be followed by a more terrigenous layer rich in dead bacteria during the autumn to spring.

Preservation of organic matter

Since the earliest days of black shale palaeoenvironmental modeling geologists have realised that the areas of maximum accumulation of organic matter coincide with areas of low circulation and oxygen deficient bottom waters (Woolnough, 1937; Strom, 1939; Twenhofel, 1939). Although there is incontestable evidence for the increased preservation of sedimented organic matter under anaerobic and anoxic conditions, certain points of controversy remain as to the exact mechanism by which organic matter is preserved. Richards (1970) could find no evidence to suggest that there is greater preservation of organic matter within anoxic water columns (though this may in part be due to the difficulties of intra-water column sampling), even though the underlying sediments generally have above average organic contents. This implies that the process controlling the enrichment of sediments in organic matter occurs as the sediment water interface and contradicts (if only quantitively) the belief that anoxicity decreases predation and bacterial degradation in the water column significantly enough to result in a higher flux of organic matter to the bottom. In oxygenated water columns, intra-water column losses will only be significant in deep oceanic areas but this may in part be compensated by the rapid sedimentation of organic material via the agency of zooplankton faecal pellets.

Richards (1970) and Deuser (1975) have expressed the view that the higher organic contents of sediments in anoxic basins may be due to faster rates of sedimentation and accumulation rather than slower rates of organic decomposition. The effect of a high sedimentation rate is to rapidly remove the sedimented organic matter from the bacterially active top few centimetres of the sediment, allowing less time for bacterial degradation and consequently resulting in higher preservation (Heath et al. 1977; Didyk et al. 1978; Curtis, 1978; Coleman et al. 1979; Muller & Suess, 1979). Although in oxygenated nearshore waters high sedimentation rates and a high supply of organic matter can combine to result in the deposition of organic-rich black muds (gyttjas), the sedimentation rate of true basinal anaerobic and anoxic black shales (i.e. sapropels) is probably too low for this mechanism to apply. If sedimentation rate were the only control then the clear cut correlation between anaerobic bottom conditions and organic-rich sediments should not exist.

Increased preservation of organic matter at the sediment-water interface in anoxic or anaerobic basins could result if, as the common view supposes, anaerobic bacteria are less efficient than aerobic ones, but this is not supported by experimental data (Foree & McCarty, 1970). As anoxic conditions tend to occur in deeper waters it might be suggested that lower temperatures are responsible for lowering the metabolic rate of the bacteria in such environments but this is not supported by recent work on the Black Sea (Karl, 1978). Foree and McCarty (1970) believe that anaerobic conditions might result from the large scale accumulation of organic materials rather than the reverse and the hydrodynamic concentration of organic material may indeed play a more important role than previously supposed.

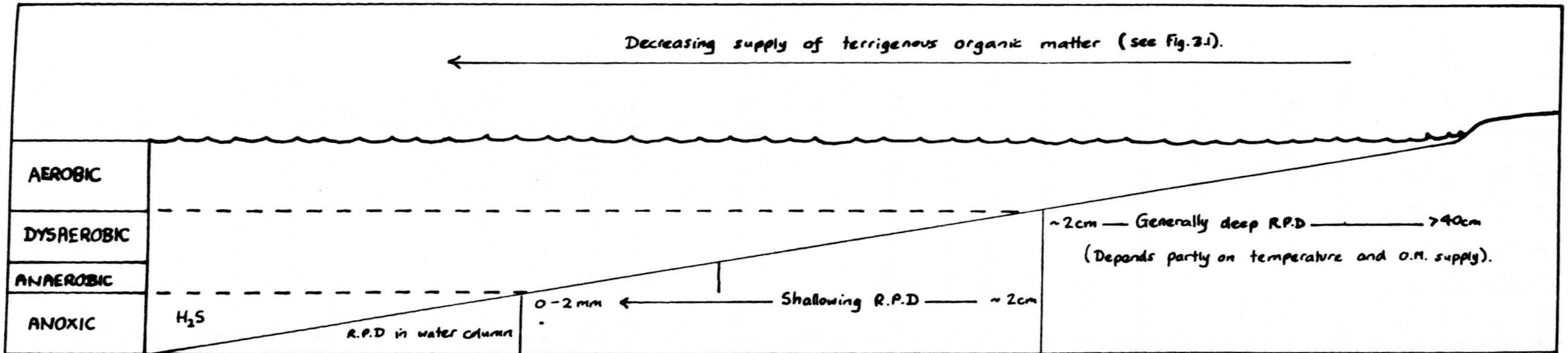
Degens and Mopper (1976) have suggested that most of the consumption of organic matter occurring at the sediment-water interface is due to the activities of the metazoan benthos (including the microbial faunas in their digestive tracts). They also commented that the low rate of organic matter consumption in anaerobic environments was probably a reflection of the lower biological activity due to the absence of metazoans. This is here considered a reasonable inference on several grounds. Organic-rich muds in aerobic conditions normally support a large number of metazoans, though toxic pore waters and the low mechanical strength of the sediment may exclude some faunal elements. If anaerobic conditions were to develop there would be no metazoan benthos and the detrital food chain would be disrupted leaving a lot of organic matter un-utilised except by anaerobic bacteria, allowing more to pass through the sediment-water interface. Sediments deposited under anaerobic water conditions should therefore contain more organic matter than sediments deposited under aerobic conditions at the same rate.

Secondary diagenetic consumption of organic matter is also a very important factor in controlling the organic content of sedimentary rocks. The organic content of gyttjas is progressively reduced during early diagenesis by bacterial oxidation, sulphate reduction and fermentation processes (Curtis, 1978; Coleman et al. 1979); these processes are all modified by anoxic and/or anaerobic bottom waters (see Fig. 2.11). The importance of the oxidative processes of the sediment water interface will obviously diminish as bottom oxygenation decreases and will reach a minimum under anoxic bottom conditions; the other processes are discussed below.

In the marine environment the bacterial metabolism of organic matter occurs in a step like manner (see Table 2.1) where each stage is characterised by different bacterial assemblages, processes and conditions (Claypool & Kaplan, 1974; Fenchel & Jørgensen, 1977). The first stage is characterised by aerobic oxidation and occurs above the redox discontinuity in gyttja sediments or above the chemocline-pycnocline in stratified anaerobic water bodies. This stage is occupied by a variety of heterotrophic organisms (including metazoans) which are all capable of catalysing the complete oxidation of organic matter. Once all the molecular oxygen is used up (i.e. below the R.P.D. or chemocline) the next stage, anaerobic denitrification, occurs where nitrate ions are used as oxygen donors to metabolise sugars, organic acids and alcohols (Manheim, 1976; Fenchel & Jørgensen, 1977). This process often occurs where the dissolved oxygen level falls below 0.15 ml/l and is characteristic of some oxygen minimum layers (Sorokin, 1978).

In the sulphate reducing zone heterotrophic, obligate anaerobes reduce sulphate to sulphide by utilising organic compounds such as lactate, succinate, malate and pyruvate (e.g. see Fenchel & Jørgensen, 1977; Postgate, 1979). Apparently the sulphate reducing bacteria prefer

Fig. 2.11 Features of organic matter deposition in stratified, oxygen-deficient basins.



SAPROPEL

- Aerobic oxidation in upper water column only (a function of water depth).
- Peaks in sulphate reduction at $O_2:H_2S$ interface and at the bottom.
- Sulphate reduction dominates and results in low lipid utilisation.
- Sediment sulphate reduction limited by sulphate supply (and suitable substrates?).
- Methanogens inhibited by uncombined H_2S?
- Sedimentation rate primarily a dilution effect?
- Organic carbon contents frequently >3%.
- Kerogen dominated by A.O.M. Type II or I.
- High hydrogen indices.
- Type I kerogens denote eutrophication and abundant cyanobacterial OM supply to sediment.

GYTTJA

- Progressive reduction of aerobic oxidation.
- Increasing dominance of sulphate reduction.
- Sulphate reduction tending to become limited by sulphate supply rather than metabolisable O.M.
- If sedimentation rate is high more rapid sulphate depletion in pore water results in early methanogenesis.
- Organic carbon content partly related to sedimentation rate (although less so?).
- Organic carbon values frequently >3%.
- Kerogen usually of a mixed (saprohumic) or sapropelic composition with increasing preservation of A.O.M. Type II.
- Increasing hydrogen indices.

GYTTJA OR ORGANIC-POOR SEDIMENT

- Extensive O.M. oxidation above R.P.D.
- Generally high sulphate reduction rates (limited by supply of metabolisable O.M. - i.e. extent of oxidation prior to burial).
- Open system sulphate reduction → extensive FeS formation depending on Fe availability.
- If sedimentation rate is high, rapid sulphate utilisation results in early methanogenesis.
- Organic content partly a function of sedimentation rate.
- Organic carbon values generally <3%.
- Kerogen dominated by refractory phytoclasts and palynomorphs, A.O.M. mostly destroyed. Type III.
- Low hydrogen indices.

Table 2.1 Bacterial zones in sediments and stratified water bodies (data from Fenchel & Jorgensen, 1977; Curtis, 1978; Postgate, 1979)

BACTERIAL ZONE	Eh mV	O_2 ml/l	PRINCIPAL BACTERIAL SUBSTRATES	REACTIONS	DIAGENETIC MINERAL PHASES (δC^{13} range $^o/oo$ P.D.B)
AEROBIC ZONE	+400	8-0.15	Various, Lipids destroyed	$(CH_2O) + O_2 \rightarrow CO_2 + H_2O$	
DENITRIFICATION ZONE	-100	0.15	Sugars, organic acids and alcohols	$(CH_2O)(NH_3) + 4NO_3^- \rightarrow 6CO_2 + 6H_2O + 2N_2 + NH_3 + 4e^-$	CALCITE AND
SULPHATE REDUCTION ZONE	-200	0.0	Lactate, succinate, malate, pyruvate	$2(CH_2O) + SO_4^{2-} \rightarrow 2CO_2 + S^{2-} + 2H_2O$	SULPHIDES δC^{13} 0 to -25
CH_4 FERMENTATION ZONE	-300	0.0	Acetate	$2(CH_2O) \rightarrow CH_4 + CO_2$	Fe-CALCITE Fe-DOLOMITE SIDERITE δC^{13} -10 to +15

only small molecules (Sorokin in Price, 1976) and are not capable of degrading more complex compunds such as lipids (Foree & McCarty, 1970). The extent of organic matter consumption by bacterial action in this zone is controlled by the supply of suitable metabolisable organic matter (Aller & Yingst, 1980; Sholkovitz, 1973; Goldhaber & Kaplan, 1974) and sulphate is not usually a limiting factor in most marine sediments (Berner, 1971). The CO_2 produced during sulphate reduction produces increased pore water alkalinity and may result in the precipitation of diagenetic carbonates (Shishkina, et al. 1971; Lein, 1978; Suess, 1979). Some of the CO_2 may be utilised by sulphur oxidising autotrophs in the vicinity of the RPD (Kepkay et al. 1979). Of the sulphide produced during sulphate reduction, part reacts with iron (to form monosulphides and subsequently pyrite; Berner, 1971; Goldhaber & Kaplan, 1974) and part diffuses upward and is oxidised (chemically or by photosynthetic sulphur bacteria) at the RPD or chemocline. The rate of sulphate reduction falls off rapidly with depth presumably due to the diminishing amounts of suitable organic substrate.

The final zone is the anaerobic fermentation zone (or methane zone) which is inhabited by methanogenic bacteria. These are a specialised group of obligate anaerobes which occur only under very reducing conditions (Eh -300mV) and utilise a small number of simple substrates of which acetate (a major organic product of sulphate reduction?) is the most important (Fenchel & Jørgensen, 1977). The methanogenic bacteria are apparently inhibited in the sulphate reducing zone by the presence of sulphate and/or sulphide and competition for hydrogen and thus the fermentation zone always underlies the sulphate zone when ever the two co-occur (see Foree & McCarty, 1970; Manheim, 1976; Lein, 1978; Curtis, 1978; Fenchel & Jørgensen, 1977). Bernard (1979) believes that the methane bacteria are active in the sulphate zone but are limited by the amount of hydrogen available for CO_2 reduction and that the small amounts of methane produced may be consumed by some sulphate reducing bacteria.

The CO_2 produced during the fermentation phase produces high pore-water alkalinity and may lead to the precipitation of diagenetic carbonates which are characteristically isotopically heavy (Curtis, 1978). The usually small amount of refractory organic material that survives all these bacterial processes is incorporated into the sediment and converted to kerogen.

In stratified anaerobic basins the aerobic bacterial oxidation zone occurs only above the chemocline-pycnocline and organic matter will spend less time in this zone that it would do if it occurred on the sea floor. This means that the extent of the initial degradation of the organic matter is less important and more lipids (which are normally destroyed in the aerobic zone) will survive to reach the anaerobic zone. In anaerobic basins the dominant bacterial process is sulphate reduction. Since anaerobic decomposition under anoxic conditions is characterised by high lipid preservation (Foree & McCarty, 1970; Didyk et al. 1978; Eglinton & Barnes, 1978), this indicates that the sulphate reducing bacteria cannot metabolise lipids (or at least not efficiently), probably reflecting their preference for low molecular weight substrates. The result is that lipid-rich material (especially 'amorphous sapropel') becomes preferentially concentrated in sapropels and subsequently leads to a higher content of organic matter than under aerobic conditions (Kendrick, 1979). Since sulphate reduction in sapropels is possibly limited by the lack of suitable substrates, fermentation processes will also be restricted because the methane bacteria depend partly upon the acetate produced by sulphate reducers. If the sediments contain little available iron to 'mop up' the H_2S by sulphide formation (e.g. Shishkina et al. 1977; Jørgensen, 1980) the accumulation of H_2S will probably also tend to inhibit the extent of bacterial fermentation. In stratified basins with anoxic bottom water the main site of pyrite formation (at the R.P.D. in gyttjas) may shift to the O_2:H_2S interface such that little uncombined iron reaches the sea floor (Borchert, 1965; Berner, 1971) and high concentrations of free sulphide are maintained in the pore waters. The low supply of detrital iron to distal sapropelic black shales and the low supply of iron through the chemocline may account for the low pyrite observed in "bituminous shales" (q.v. Borchert, 1965).

General discussion

From the preceding account it is possible to construct two scenarios for the deposition of organic rich sediments:-

A Organic rich black muds (kerogenous clays, aerobic to dysaerobic gyttjas)

This variety of sediments occur mainly in the lower mixed layer zone of coastal areas characterised by large supplies of allochthonous organic matter. Since the bottom conditions are normally oxidising only relatively refractory terrigenous organic matter survives sedimentation and early diagenesis with the consequence that organic contents are usually only high in the vicinity of major fluvial inputs (e.g. deltas, estuaries, and also salt marshes). Much of the labile organic matter which survives to pass through the R.P.D. is effectively 'converted' to pyrite by the activities of the sulphate reducing bacteria, resulting in a dark but relatively organic poor sediment. High accumulation rates favour higher preservation of organic carbon; in most settings high accumulation rates go hand-in-hand with increased dilution of the organic component but deltas and estuaries are a notable exception (see Chapter Three).

B Black Shales (kerogenous black shales; sapropels and sapropelic anaerobic and dysaerobic gyttjas)

This type of sediment is deposited mainly in the anoxic or hypoxic bottom waters of meromictic stratified basins. Increased preservation of organic matter results from:

(i) the absence or marked reduction of oxidation processes at the sediment-water interface

(ii) the absence or reduced activity of metazoan benthos at the sediment-water interface

(iii) the lower residence time of organic matter in the aerobic zone (especially if this is entirely within the water column)

(iv) the sedimentation of higher proportions of marine sapropelic (lipid-rich) organic matter

(v) the substrate limitation of sulphate reducing bacteria (partly as a consequence of iv) and its subsequent effects through the bacterial 'food chain'

(vi) the persistence of pore water sulphides and their inhibitory affect on bacterial fermentation processes

(vii) the fact that much of the sulphate reduction which does take place does so within the water column rather than in sediment (Indrebø et al. 1979) may also decrease the efficiency of the methane bacterial zone

(viii) the low clastic dilution characteristic of the distal parts of stratified basins.

From the point of view of petroleum source rocks it is clear that preservation is generally a far more important control on the organic content than is the supply of organic matter. Outside of the areas of coastal Ekman divergence there is generally a negative correlation between areas of high marine productivity and areas of accumulation of authochthonous organic matter because the productivity is based on the recycling of nutrients (e.g. Rowe et al. 1975) and not their sequestering in sediments. Although meromixis can improve preservation by leading to bottom water oxygen depletion it does result in a general decrease in basin producitivy except in coastal areas, local areas of upwelling or following short term mixing events. The present day oceans are stratified and primarily oligotrophic except where the nutrients accumulated below the mixed layer are made available in the zones of divergence and upwelling. It follows that most black shale basins had only low or moderate overall primary productivity but excellent preservation (see also Tucholke & Vogt, 1979, p.797). Where the organic content of a sediment consists primarily of refractory terrestial (woody) debris the first scenario applies. Although this situation normally only arises in coastal sediments it is clear that redeposition can result in bottom water 'black mud' deposits (e.g. the bulk of the Cretaceous Hatteras Formation in the central Atlantic - see also Chapter Three).

The character and source of organic matter in kerogenous mudrocks

Since the type of organic matter in the environment of deposition has a significant influence on the preservation path and accumulation of organic-rich sediments, the origins of kerogenous mudrocks cannot be discussed unless the character of their kerogen is known (see for example comments in Roberts & Montadert, 1979; Arthur, 1979). The main groups of kerogens found in sedimentary rocks are listed in Table 2.2 (for full account see Chapter Three). Woody debris (vitrinite, semi-fusainite and cutinite) is generally less than 50% of the total kerogen in organic-rich sediments except those formed in nearshore regions and in the vicinity of deltas. Degens and Mopper (1976) claim that the contribution of terrestial material is apparently small compared to the marine, autochthonous component regardless of whether the sediments are deposited during transgressions or regressions. It is clear, however, that planktic material (which contains relatively little structural organic matter in comparison with macrophyte debris) is much more easily decomposed (Godshalk & Wetzel, 1977). Very few animals possess the necessary enzymes to hydrolyse the structural polysaccharides and other polymers (e.g. lignin) that compose macrophyte tissue (Fenchel & Jørgensen, 1977) and this might be expected to lead to a greater relative preservation of terrestial organic matter. Most of the more reactive marine organic matter is apparently only preserved under 'intensively

Particulate organic matter (POM)				Kerogen							
	Category	Source	Constituents	Coal maceral group	Coal maceral	A	B	C	D	Kerogen type: General	Kerogen type: E
Structured	Palynomorphs	Plankton	Blue-green algae and prasinophycean phycomas	Exinite or Liptinite	Alginite	Vegetable (MOV)	Algal	Phyrogen	Aqueous	Sapropelic (oil prone)	I
			Dinocysts and acritarchs		Sporinite		Herbaceous				II
		Sporomorphs	Spores and pollen						Spores and pollen		
	Phytoclasts	Higher plant debris (macrophyte (tissues)	Cuticle		Cutinite				Terrestial		
			Ligno-cellulosic material	Vitrinite	Telinite	Tracheal (MOT)	Woody	Hylogen		Humic (gas prone)	III IV
			Carbonized material	Inertinite	Fusainite	Lignitic (MOL)	Coaly	Melanogen	Charcoal		
Unstructured	Amorphous organic matter (AOM)	Polygenetic and heterogeneous	Organic aggregates and flocs formed from dissolved organic matter, products of bio-chemical degradation of POM, and faecal pellets	Variable, but often with components of the following macerals:-	Collinite	Colloidal (MOC)	Amor-phous	Amor-phogen	Amor-phous	Variable: Humic	III
					{Liptodetrinite Bituminite}					Variable: Sapropelic	II/I

Tables 2.2/3.1 A classification of particulate organic matter found in marine sediments and the approximate equivalence of the most commonly used kerogen terms. Column A after Correia (1971) and McLachlan & Pieterse (1978); B after Staplin (1969) and Hunt (1979); C after Bujak _et al_. (1977); D after Masran & Pocock (1981); E after Tissot & Welte (1978). Coal maceral terminology (much abbreviated) after Stach _et al_. (1975). Fusainite should read fusinite.

reducing conditions' (Kendrick, 1979). Spores and pollen and algal cyst materials are very resistant to degradation.

The most abundant type of organic matter in dysaerobic gyttjas to sapropels is that usually referred to as amorphous sapropel (Staplin, 1969; Correia, 1971; Burgess, 1974; Correia & Peniguel, 1975). The amorphous, structureless nature of this material has led to a great deal of conjecture concerning its origins and affinities. Much of this amorphous material is relatively lipid-rich and may legitimately be called amorphous sapropel but some morphologically identical material has a humic composition. Many workers have noted that some of the organic material present in recent clay grade sediments is quite often undistinguishable from the amorphous organic matter (AOM) in thermally unaltered ancient sediments, which implies that the amorphous material is formed within the depositional environment and only modified during diagenesis.

Amorphous organic material, known to marine biologists as organic aggregates, is in fact very abundant in the modern marine environment (for full discussion see Chapter Three). Parsons (1975) has estimated that, except during algal blooms, this 'organic detritus' often exceeds phytoplankton carbon by a factor of ten or more. Much of the organic aggregate material is believed to form from the vast reservoir of dissolved organic matter (DOM) present in the sea by flocculation, adsorption and bacterial activity, and so is only indirectly related to primary productivity (see Riley, 1963, etc.). The composition of this modern AOM is very variable and appears to depend upon the extent to which it has been biodegraded; its initial composition probably resembles that of plankton (e.g. Eppley et al. 1977) but is altered to refractory carbohydrate by the loss of the more labile compounds. AOM is also derived from deposit feeder and zooplankton faecal pellets (Johnson, 1974; Honjo & Roman, 1978).

If relatively unbiodegraded AOM (initially derived from DOM or plankton) is sedimented into an anoxic environment its lipid content will be preserved (and perhaps supplemented by bacterial lipids) and sapropelic, AOM-rich sapropels will be deposited (as described earlier). In an aerobic environment the organic aggregate material will be destroyed or degraded to refractory material with a more 'humic' composition. Correia (1971) has noted that where black shales are rich in AOM they normally have very low amounts of plankton (and commonly low plant debris as well) and this has been confirmed by the author's own studies. The lack of woody kerogens may be explained by a distal depositional environment and the sparcity of plankton by the permanent stratification of the basin waters which produces anoxic bottom conditions, low overall primary productivity in the basin centre and the cessation of cyst production (except on the margins). However, in some cases this low in situ productivity of marine organic matter may only serve to accentuate the importance of the allochthonous component, and this fact, taken with the more resistant nature of terrestial kerogens and the effects of pycnoclines on the offshore transport of suspended materials, may result in many black shales having a 'sapro-humic' kerogen composition. This effect is particularly pronounced in rather small basins or basins fed by major rivers draining well vegetated areas. Although sapropelic components are at their greatest abundance in sapropels it does not necessarily follow that the total kerogen composition is dominantly sapropelic. The general relationships between bottom environment sedimentation, preservation and character of organic matter are summarised in Fig. 2.11.

BLACK SHALE MEGAFACIES

After undertaking an extensive reivew of the 'black shale literature' I have come to the conclusion that there are only four main environmental settings in which organic-rich argillaceous marine sediments are deposited. As each of these environmental situations can incorporate a whole spectrum of different sediment types these are here referred to as 'megafacies'. The four megafacies are:-

1. The black mud megafacies
2. The barred basin megafacies (euxinic megafacies)
3. The pericontinental megafacies
4. The epeiric megafacies

The main features of these megafacies are summarised in Table 2.3, and are described below.

1. The black mud megafacies

The sediments of this megafacies are not true black shales in the normal usage but include argillaceous sediments deposited in nearshore regions such as estuaries, distal parts of deltas and lagoons,where there is a large supply of organic matter and a relatively high sedimentation rate. Since fully oxygenated conditions normally prevail above the sediment-water interface (with the possible exception of some lagoons) the sediments are gyttjas. They usually contain prolific benthic faunas and are intensively bioturbated (the structures belonging to the Cruziana ichnofacies; Seilacher, 1967). Depending on the particular facies involved,the faunas may show a low salinity influence. The presence of interbedded rippled silt and sand layers, occasional erosional surfaces and sorting of shell debris are common indicators of current activity which testify to a 'shallow' depth of deposition. Primary laminations will only be preserved if and when the sedimentation rate is so high that the benthos cannot adapt to it; kerogen-clay couplets will occur only very rarely. Sediments of this type usually contain significant amounts of humic kerogens (i.e. plant debris) and inertinite. Geochemically they show the characteristics of gyttjas (see Krejci-Graf, 1964, 1972). Because these sediments have abundant benthos, are bioturbated and contain clear evidence of 'shallow water' deposition they do not correspond with most geologists preconceived

Table 2.3 Megafacies classification of black shale environmental settings

Megafacies	BLACK MUD	BARRED BASIN	PERICONTINENTAL	EPEIRIC
Location	Normally the nearshore zone	Marginal, land-locked seas or submarine topography	Outer continental shelf and slope on eastern margins	Internal areas of epeiric seas
Geometry	Narrow belts and lobes	Irregular: extensive to limited sheets	Narrow belts	Extensive sheets ($\geq$100,000 km^2)
Sediment type	Gyttja	Sapropel	Gyttja	Gyttja (localised sapropels)
Biofacies	Aerobic	Anaerobic (anoxic)	Dysaerobic-anaerobic	Dysaerobic-anaerobic
Control on organic content	Input of allochthonous O.M. + rapid burial	Preservation	Production + preservation	Preservation
Mechanism	Proximal sedimentation, hydrodynamic equivalence and rapid burial	Sill + pycnocline	Upwelling + oxygen minimum	Regional watermass stratification + palaeogeography
Occurence in time	Ubiquitous	Variable (freaks of palaeogeography)	Probably ubiguitous but of variable scale	Only during transgressive peaks and warm global climates
Modern examples	Estuaries, lagoons, prodelta deposits	Black Sea, Baltic, Cariaco Trench, various fjords	Walvis Bay shelf, S.W. Africa, Peruvian Shelf	None

genetic conception of black shale. However, they are often dark grey to black (due to their high organic contents and appreciable sulphides) and when compacted are sometimes fissile enough to warrant the name black shale in the purely descriptive sense. The black mud megafacies is essentially equivalent to the 'nearshore macrofacies' of Timofeev and Bogolyubova (1979).

2. The barred basin megafacies

Because all the modern anoxic basins of the world are barred basins (Fonselius, 1963; Richards, 1965; Deuser, 1975), nearly all the black shale deposits in the sedimentary record have at some time or another been interpreted as deposits of restricted, silled, basins (see for example Woolnough, 1937). In these basins haloclines and/or thermoclines (which reduce vertical circulation) and topographic sills (which impede lateral circulation) combine to produce anoxic bottom waters and lead to sapropel deposition. The modern examples include the Black Sea (Ross & Degens, 1974), the Baltic (Fonselius, 1970), the Cariaco Basin (Richards, 1974) and numerous fjord-like basins (see Strøm, 1939; Gucluer & Gross, 1964; Grasshoff, 1975; Deuser, 1975). Barred basins are essentially freaks of palaeogeography and range in scale from small fjords to the very large basins like the Black Sea and Eastern Mediterranean. In each case watermass stratification and basin circulation pattern (both influenced by climate) are just as crucial as sill depth (see Fig. 2.8). We saw earlier that estuarine circulation favours the deposition of organic-rich sediments in restricted basins and so most black shale barred basins will occur in humid climates. These basins represent the last site of sapropel deposition when all other environments are unfavourable; with the possible exception of upwelling areas (where transient anoxic episodes occur), this is the situation which occurs at the present day. The character of the organic matter in barred basin sediments will depend upon the size of the basin, the climate and the vegetation on the basin margins.

3. The pericontinental megafacies

This megafacies consists of those organic-rich sediments of the upper continental slope and outer shelf which occur where oxygen minimum layers intersect the sea floor along the continental margins. Present day occurrences include those of the continental margins of California (Calvert, 1964), South West Africa (Calvert & Price, 1971), Oregon (Gross et al. 1972; Kulm et al. 1975) and India and Pakistan (Murty et al. 1969; Von Stackelberg, 1972).

Wyrtki (1962) has shown that the presence of an oxygen minimum is due to the consumption of oxygen by organic decay and that the position of the minimum is determined by the position of the layer with minimal advection of oxygen (i.e. the layer with least horizontal circulation) within the water column. The initial oxygen concentration of the water at this depth is, however, determined by its salinity and temperature at source and its subsequent history (e.g. see Menzel & Ryther, 1968). Since the level of least advection normally occurs just below the main ocean thermocline, where settling organic matter may become concentrated at the associated pycnocline, the oxygen minimum and the thermocline show the same latitudinal depth distribution; they shallow towards the equator (due to the equatorial divergence) and towards the eastern margins of the oceans. Over 90% of the organic matter produced in the euphotic zone is recycled above the ocean thermocline (Whittle, 1977) and below this level the majority of the organic matter present in the water column is apparently refractory and resistant to oxidation (Wyrtki, 1962; Menzel & Ryther, 1968; Riley, 1970). This means that a supply of organic matter sufficient to cause an oxygen minimum can only normally occur in the top 300-500m of the ocean (Menzel & Ryther, 1968; Suess, 1980); oxygen deficient waters may, however, spread to other depths by advection.

The processes which form the oxygen minimum are essentially the same as those which lead to deoxygenation in restricted basins with the exception that the abyssal circulation of the oceans provides oxygen to the base of the water column and that deoxygenation occurs within the water column (rather than from the sea floor upwards), since most organic matter is recycled long before it reaches the bottom. At the eastern margins of the oceans, coastal Ekman divergence leads to upwelling and high productivity and the decay of organic matter on the upper continental slope (and within the sub-pycnocline watermass) results in a strong intensification of the oxygen minimum. Where the oxygen minimum layer intersects the sea floor sediments of the pericontinental megafacies are deposited with the following characterists:-

(1) The sediments are organic-rich, often containing up to 10% or more of organic matter (Diester-Haass, 1978). They are rich in amorphous organic matter (Davey & Rodgers, 1975).

(2) The biofacies range from aerobic to anaerobic and are commonly dysaerobic. Temporary anoxic conditions may occur above the sediment water interface (see earlier under the discussion of red tides and algal blooms). The sediments are gyttjas (Calvert, 1976).

(3) Anaerobic bottom areas show micro-laminated sediments (Diester-Haass, 1978; Von Stackelberg, 1972). May be absent where no seasonality.

(4) Plankton assemblages are of low diversity and high density and often dominated by siliceous forms, especially diatoms (which are characteristically large; Landry, 1977).

(5) Benthos (where it occurs) is generally of low diversity and high density; the macrobenthos is dominated by worms, shrimps, sea pens asteroids and echinoids (Sanders, 1969; Thiel, 1978; Calvert, 1964). Bioturbation decreases with diminishing oxygenation. Foraminifera may be abundant (Phleger & Soutar, 1973).

(6) The pH of interstitial waters is usually alkaline (7.5-9 in Walvis Bay sediments and 7.3-7.5 off Peru; Shishkina et al. 1977) but is lower in the overlying waters (7.4-7.9) due to high partial pressures of CO_2 caused by organic decay (Manheim et al. 1975; Piper & Codispoti, 1975).

This means that carbonate dissolution may occur above the sediment-water interface but material in the sediment may be excellently preserved (e.g. Phleger & Soutar, 1973). Sholkovitz (1973) found good preservation of calcareous fossils in sediments deposited in waters of 0.05-0.1 ml/l and poor preservation in sediments deposited in waters of $\geq$ 0.4 ml/l of oxygen.

(7) The sediments are generally highly reducing; Shishkina et al. (1977) recorded Eh values of -120 to -260 mV in Walvis Bay sediments and +69 to -227 mV in Peruvian sediments. Gallardo (1977) notes the presence of a thin surface oxidised layer (of a few mm) even in waters of typically less than 0.1 ml/l of oxygen.

(8) The sediments of upwelling regions are often enriched in various trace elements (e.g. see Calvert & Price, 1971; Brongersma-Sanders et al. 1980, and the section on trace elements later in the chapter) and are classically associated with phosphorite deposition (see later).

(9) Because of the link between ocean circulation patterns and climate the adjacent land areas are generally arid and contribute little clastic material to the shelf sediments (e.g. Cross et al. 1966; Brongersma-Sanders, 1971; Calvert & Price, 1971; Milliman, 1971).

It must be appreciated that the actual oxygen concentrations observed in oxygen minimum layers are very variable (see Table 2.4) and that although anaerobic or dysaerobic conditions may be encountered on eastern margins, in other areas the whole of the minimum is aerobic. It is clearly insufficient to talk of an oxygen minimum without specifying particular oxygen concentrations and depth ranges. The most intense oxygen minimum occurs in the N.E. Arabian Sea where values commonly lie between 0.09 and 0.15 ml/l (Vinogradov & Voronina, 1961). Russian oceanographers have reported the presence of H_2S in the minimum but this is disputed by other workers (McGill, 1973). Hydrogen sulphide

TABLE 2.4 Oxygen minimum layers

Location	Dissolved Oxygen	Depth limits	Source
Western Atlantic	~4.0 ml/l	300- 500 m	Emery & Uchupi, 1972
Wn. Gulf of Mexico	2.2 -2.5 ml/l	?	Emery & Uchupi, 1972
S.W. African margin (Walvis Bay)	~1.0 ml/l	40- 400 m	Brongersma-Sanders, 1969
Brazilian margin 7°S	~2.0 ml/l	300- 400 m	Magliocca, 1978
Brazilian margin 22°S	4.0 -4.5 ml/l	600- 800 m	Magliocca, 1978
Oregon margin	~1.0 ml/l	500-1800 m	Gross et al. 1972
Mexican margin (Pacific)	~0.1 ml/l	100- 900 m	Douglas et al. 1976
Gulf of California	~0.5 ml/l	~100-1200 m	Parker, 1964
Gulf of California	<0.2 ml/l	450- 800 m	Donegan & Schrader, 1982
Peruvian margin	0.1 -0.2 ml/l	50- 500 m	Pak et al. 1980; Judkins, 1980
Chilean margin	~1.0 ml/l	150- 450 m	Ingle et al. 1980
Chilean margin	~3.0 ml/l	900-1700 m	Ingle et al. 1980
Pakistan margin	0.1 -0.6 ml/l	200-1000 m	Haq et al. 1973
N.E. Arabian Sea	0.09-0.15 ml/l	~125-1250 m	Vinogradov & Voronina, 1961
Red Sea (Atlantis II Deep area)	0.9 -1.3 ml/l	300- 650 m	Weikert, 1980

N.B. Oxygen concentrations and depths vary, particularly on margins with seasonal upwelling. Midwater anoxia has been occasionally observed in the Indian Ocean and off Peru (Dugdale et al. 1977). Additional data can be found in Dietrich et al. (1980).

has also been reported from the oxygen minimum off Peru where unusual current conditions in the Peru-Chile undercurrent apparently occasionally allow complete denitrification and production of H_2S by sulphate reduction (Dugdale et al. 1977; Codispotti, 1981). Seasonal upwelling on the Somali, Pakistan and Indian margins (during the south west monsoon season) produces the high productivity characteristic of the Arabian Sea.

The present author believes that a fair degree of restraint is required in the usage of oxygen minima and upwelling in the modeling of black shale palaeoenvironments. Several points should be born in mind:

(a) Upwelling is often only a seasonal event and its impact on the sediments may be less significant than the background conditions that exist for the remainder of the year (see Diester-Haas, 1978).

(b) The oxygen minimum-upwelling model may only be used for black shales with the appropriate characteristics which are located on (or near) present or palaeo-continental margins.

(c) To validate an oxygen minimum model, oxygenated sediments <u>must</u> be shown to occur over the same time interval in the deeper parts of the basin.

(d) At the present day, when we may expect upwelling to be relatively intense (Fischer & Arthur, 1977), upwelling only involves 0.1% of the oceans (Ryther, 1969).

4. <u>The epeiric megafacies</u>

Any model which is proposed to account for epeiric black shales must be capable of explaining their fundamental characteristics, namely: their essentially 'mid-continent' location, their intercalation within neritic sediment sequences, their large lateral extent (often in excess of 100,000 km^2), their gross uniformity, the general absence of features indicating bottom currents, the widespread evidence of dysaerobic to anoxic bottom water conditions,and the stability and persistence of these conditions often for considerable periods of time.

Examples of this type of sedimentation are most common among the extensive marine platform sequences of the North American and Eurasian continents. Some of the well-documented epeiric black shales of these areas include: the Upper Devonian Chattanooga-New Albany-Ohio-Sonyea shales of North America (Conant & Swanson, 1961; Byers, 1979; Cluff, 1980), the Pennsylvanian black shales of North America (Heckel, 1977), the Permian Kupferschiefer of N.W. Europe (Deans, 1950; Wedepohl, 1964), the Toarcian Jet Rock-Posidonienschiefer-Schistes Cartons of N.W. Europe (e.g. Morris, 1979), the Volgian (Upper Kimmeridgian - Ryazanian) black shales of N.W. Europe and Russian basins (see Chapter 8), the Late Albian Mowry Shale of North America (Davis, 1970; Byers & Larson, 1979) and the Upper Cretaceous Colorado Group, also of North America (Hattin, 1971; Frush & Eicher, 1975; Simpson, 1975; Byers, 1979; Kauffman, 1979). No adequate modern analogue exists for these extensive mid-continent black shale deposits. The black mud and pericontinental megafacies models do not satisfactorily account for the characteristic features of these sediments and the classic barred basin model has long since been abandoned for the epeiric black shales. The mechanism proposed here utilises a known combination of climate, sea-level and palaeogeography to create a variably meromictic, regionally stratified epeiric sea. Before outlining this mechanism the existing models for epeiric black shales are briefly discussed.

Existing models for the epeiric black shales

Several authors have explained epeiric black shales by suggesting the extension of an expanded oxygen minimum layer into the shelf sea interior (e.g. Heckel, 1977; Fischer & Arthur, 1977; Ryan & Cita, 1977; Thierstein & Berger, 1978). There are a number of considerable problems related to this hypothesis which are discussed below.

(i) Expansion of the oxygen minimum layer is said to occur at times of decreased abyssal circulation resulting from lowered latitudinal temperature gradients (Fischer & Arthur, 1977). Should this be the case, one would have anticipated that the expansion of the minimum would be predominantly downward rather than upward and outward into the shelf seas.

(ii) During 'polytaxic episodes' (when expansion of the oxygen minimum is supposed to occur) oceanic mixing and nutrient recycling are of dminished efficiency (Fischer & Arthur, 1977). Increased stability will undoubtedly result in decreased primary productivity. Since the presence of the oxygen minimum depends upon organic matter supply, decreased productivity will in no way favour its expansion or intensification. Berger (1979, p.307) notes that "if oxyty drops, organic matter is locked up in sediments together with nutrients, thus consumption drops and oxygen depletion is retarded". It should always be remembered that significant (at least dysaerobic) oxygen minima only occur below highly productive surface waters.

Müller and Suess (1979, p.1347) note that "it is a characteristic feature of many hemipelagic (continental margin) sediments, accumulating fast enough to resolve the last glacial and interglacial periods, that glacial sections contain considerably more organic matter (by the factors of 3-7) than interglacial sections". They show that although these differences correlate with bulk sedimentation rates they probably reflect chages in primary productivity. It is tempting to argue that the more extreme (glacial) global climates promote greater mixing and efficiency of nutrient recycling, higher productivity in upwelling areas, and potentially more intense or broader oxygen minima (see also De Vries & Schrader, 1981). From the distribution of fossil Quaternary phosphorites on the Peru-Chile continental margin, Burnett (1980, p.761) notes that, since earlier in the Quaternary, the oxygen minimum has "either shifted its location or <u>contracted</u> in size". My tentative conclusion is, that since epeiric black shales depend upon preservation (via meromixis) while pericontinental black shales depend upon sustained high productivity, the climatic conditions which favour one do not favour the other. Equable global climates clearly favour epeiric black shales during transgressions but probably did not favour the optimum development of the oxygen minimum and pericontinental megafacies.

(iii) Ocean and shelf watermasses are usually decoupled at the continental margin by hydrodynamic frontal systems (see earlier) and turbulent edge effects (e.g. see Fig. 6 in Walsh, 1977). Anti-estuarine cells on the scale conceived by Heckel (1977) are considered highly unlikely for the above reason. Modern upwelling cells are intimately related to the eastern boundary currents at the continental margin (e.g. see Brongersma-Sanders, 1971) and cannot be shifted ad hoc to the interior of an epeiric sea, although 'upwelling' in the general sense can occur virtually anywhere given the right conditions. The significance of the atypically narrow shelf widths in modern upwelling areas should not be underestimated; the shelf is only 15 km wide off Peru, 25 km off Baja, California and 45 km off North West Africa (data from Walsh, 1976).

(iv) While epeiric black shales are being deposited in the shelf sea interior, sediments characteristic of fully oxygenated bottom conditions (e.g. reefal limestones and shallow water carbonate facies) are often found along the palaeo-margin. This evidence is clearly incompatible with an oxygen minimum extending <u>across</u> the margin.

Another model of epeiric black shale sedimentation was proposed by Hallam (1967-1975) utilising calculations made by Keulegan and Krumbein (1949). He hypothesised an extensive, shallow shelf sea with a very low oceanward gradient (of less than 1:600) which he claimed would result in the offshore dissipation of turbulent wave energy and lead to stagnation in the interior. This model is inadequate because, although such a gradient might result in the dampening of oceanic wave energy, it does not account for turbulent mixing generated by wind stress within the epeiric sea, and because it suffers from a too simplistic palaeogeography. Integral to this model was the belief that the black shale sediments represented shallower conditions than overlying aerobic sediments! Hallam (1978) has reverted to his pre-1967 views and abandoned this model (see also Hallam & Bradshaw, 1979). The important point made by this hypothesis was that it suggested large distances may be almost, if not as efficient, at restricting circulation as are sills. Byers (1977) states that sills (with pycnoclines) are an essential requisite for the development of all black shales, whereas in fact large distances combined with shallow water depths may sufficiently decouple circulation patterns so as to isolate the more internal regions of a shelf sea,which, if stratified, may become prone to meromixis.

The epeiric sea stratification model

The great lateral extent and synchronous development of epeiric black shale indicates that they are deposited at times of global transgression during which there is an effective deepening of the water column (that may be reinforced or moderated by local tectonics). If the increase in depth of the water column results in the depth of the mixed layer (which is climatically determined) being less than the total depth, and in the absence of significant tidal stirring, stratification will be initiated. As the transgression continued the increase in the ratio of the bottom water volume to mixed layer volume would enhance the stability of the stratification.

Several authors have suggested that the increase in the sea:land ratio during transgressions would result in significant climatic amelioration (Hays & Pitman, 1973; Fischer & Arthur, 1977). This would have several important consequences, including:

(i) Warmer global climates which would favour thermally induced stratification in the relatively shallow epeiric seas (with subsequent reduction and/or stabilisation of the mixed layer).

(ii) Greater climatic stability and equitability which would favour the establishment of meromixis in (seasonally) stratified basins (see also

Degens & Stoffers, 1976).

(iii) Increased average water temperatures resulting in higher rates of organic decay and biological oxygen demand, lower dissolved oxygen concentrations at saturation, and the maintenance of effective thermal pycnoclines due to the relationship between water density and temperature.

During the still-stand phase of the transgression cycle large areas may have become subject to seasonal bottom water oxygen depletion and parts of the epeiric seas may have become persistently de-oxygenated (dysaerobic-anaerobic). Anoxic bottom waters may have developed in the deeper, actively subsiding depocentres. The spatial distribution of the anaerobic, dysaerobic and aerobic biofacies will depend upon the interaction between bottom topography and the watermass layering, the extent and distribution of tidal mixing and the degree to which the epeiric sea is hydrographically decoupled from the oceans. The widespread uniformity of epeiric black shales indicates that the principal control occurs within the watermass (i.e. from the sea surface downward) rather than being related to bottom topography (sea floor upwards). The restriction of black shale facies entirely to rapidly subsiding depocentres (Hallam & Bradshaw, 1979) does not seem a suitable solution to epeiric black shale formation in general, but may account for localised sapropel deposition amidst widespread dysaerobic to anaerobic gyttjas.

It is very difficult to accurately assess the likely role and importance of salinity in the epeiric sea stratification model. For the boreal basins it has been widely conjectured that meromixis may have resulted from increased runoff and a reduction in the salinity of the mixed layer (e.g. Kauffmann, 1975, 1979). While obviously attractive because of the well documented possible modern analogues, this hypothesis suffers from the absence of any reliable absolute criteria for determining palaeosalinities. This problem is magnified by the fact that in a stratified water column only the non-demersal nekton and the plankton have any potential for recording the nature of the surface layer salinity, since the bottom water is likely to remain at essentially normal marine salinities. Furthermore, since the presence of meromixis in itself has profound effects on the pelagic ecosystem, it may be too easy to confuse the effects of changes in watermass structure with those that might result as a response to lowered salinity. Stress responses often involve rather general strategies which are not specific to the nature of the stress applied.

The consistent presence of what are presumed to have been relatively stenohaline organisms (such as ammonites) in the Mesozoic boreal seas suggests that any freshening of the surface water that may have occurred was slight or geographically restricted. Workers on the European Mesozoic epeiric seas have now generally rejected lowered salinity as the key characteristic of the boreal faunal province (see end of chapter) in favour of more sophisticated concepts of environmental stability, but salinity variations are still considered to have been of major importance in the Cretaceous western interior basin of the U.S.A. For the latter, Kauffman (1979, p.A426-7) has suggested that subnormal salinities were periodically characteristic throughout the middle Cretaceous except at times of peak transgression (latest Cenomanian to earliest Turonian, Coniacian and parts of the Santonian) and sometimes occurred throughout the whole water column. However, it is apparent that oxygen deficiency and possible meromixis also occurred during those episodes of normal salinity (e.g. Hattin, 1975, 1981, etc.) as well as during times of possible freshening, and that isotopic data may not be entirely reliable.

It is obvious that the palaeogeographic setting of epeiric black shales is completely different from the modern salinity stratified anoxic barred basins. They were much more extensive and did not have very narrow and shallow sills which could have helped to pond-up low salinity surface waters, although in some cases the 'oceanic' connections of the black shale basins were relatively narrow seaways (e.g. the Albian Mowry Sea of the U.S.A., and the Volgian West Siberian basin). Surface mixing and evaporation would have tended to equilibrate the surface and bottom water salinities in all areas except the low salinity wedges developed in the immediate vicinity of river mouths. In addition, since the principle epeiric black shale episodes correlate with major transgressions one would have suspected from a priori reasoning that the land areas were less extensive and less likely to include major drainage systems, particularly if they were peneplained and denuded during the preceeding regression and were not tectonically rejuvenated. It is possible that the warmer global climates, through enhancing thermal stability of the water column, might have allowed slight differences in salinity to have developed between the surface and bottom water, which would have normally have been prevented by more extensive mixing under shallower or less stable conditions. The main causes of epeiric black shale sedimentation (i.e. of meromixis in epeiric seas) are therefore the major transgressions (which result in greater water depths and warmer and more stable climates) and palaeogeographic situations which decouple the oceanic-marginal shelf sea and epeiric sea circulations. Uncompensated tectonic subsidence is also clearly an important factor in the case of some epeiric black shale sequences,but its main effect appears to be supplementary rather than primary. The specific role of salinity has yet to be proved.

Comments on the coupling between shelf and ocean black shale episodes

All true black shales results from various scales of meromixis. The tranisition from the 'normal' to the 'black shale mode' is brought about by a combination of a palaeogeography and/or a palaeobathymetry which restricts lateral bottom water circulation and one or several of the following:

(a) The development of stable stratification (meromixis) where none previously existed.

(b) A change in the character and/or intensity of the principal basin pycnocline.

(c) An increase in the temporal stability of the previously existing watermass stratification.

Although they both require stable palaeogeographic configurations, the controls on bottom water de-oxygenation in epeiric seas and oceans are somewhat different. In epeiric seas the palaeogeography is responsible for restricting lateral circulation and climate controls vertical mixing via the stability and intensity of watermass stratification. In the oceans all but the cold polar seas are permanently stratified and oxygenation is controlled by the intensity of the thermohaline circulation which in turn is controlled by climate. Since climate is a major determinant in both cases,epeiric and oceanic black shale events will tend to be coupled, but differences will result from:

(1) The relative degree of palaeogeographic restriction in the ocean basin as compared with its epeiric counterparts

(2) The differing stabilities inherent between a deep ocean and a 'shallow' epeiric basin

(3) A potential phase-lag between climatic effects on epeiric watermass stratification parameters and oceanic thermohaline circulation.

(4) The relative latitudinal position of the ocean basin and the epeiric sea.

If the ocean basin is in an early opening phase it will not be participating in the true global thermohaline circulation but may have lateral circulation with other oceanic basins. Such a restricted ocean basin may become a true barred basin if it becomes salinity stratified (due to hyposaline surface waters or hypersaline bottom waters, e.g. see Arthur & Natland, 1979; Weissert, 1981) or it may show cyclic alternations in bottom water oxygenation due to climatically induced changes from estuarine to anti-estuarine circulation patterns without any pronounced salinity anomalies.

The importance of climate as a major control on black shale episodes has several interesting implications when considered in the general context of environmental stability (see later). Since tropical and subtropical seas and oceans are generally fairly stabily stratified at the present day, the potential for significant changes in the stability of watermass stratification is greatest in temperate and polar latitudes, i.e. while the tropical and subtropical climate remains fairly constant, climatic conditions in temperate and polar latitudes have varied considerably through geologic time. This implies that with regard to black shale deposition, the importance of having a favourable palaeogeography increases towards the equator and the importance of having a favourable climate increases towards the poles. It also implies that climatic cycles will have their strongest impact on sedimentation in polar, temperate and subtropical latitudes since it would be anticipated that low latitude climatic variations would be of low amplitude. In epeiric seas cyclic black shale sedimentation may result from climatically induced variations in the stability and intensity of watermass stratification, while in ocean basins similar cyclicity may result from overall shifts from estuarine to anti-estuarine circulation patterns. In a north-south ocean crossing several latitudinal climatic belts, climatic changes in the temperate, subtropical and polar parts of the ocean would presumably produce alterations in the water balance and circulation which would effect the whole ocean and hence produce cyclic sequences in the low latitudes even though the in situ climatic conditions may have changed very little.

SEDIMENTOLOGICAL IMPLICATIONS OF WATERMASS STRATIFICATION AND MEROMIXIS

We have previously seen that one of the consequences of the development of a pycnocline is to decouple the surface layer circulation from the bottom water, creating what might be called a 'false bottom effect'. This means that in the absence of a strong benthic boundary layer current, velocities capable of transporting sand grade clastics or producing current structures will be limited predominantly to that part of the sea floor above permanent pycnocline depth. Quartz sands and hydrodynamically equivalent debris will be essentially trapped in the nearshore zone,whose width will depend on the mixed layer depth and bottom gradient. This relationship occurs in the present day Baltic Sea where sediments above the main pycnocline are sandy and those below are muddy (Exon, 1972), as is the case in stratified lakes (Sly, 1978). Only aeolian, turbidity current and glacially transported quartz sand will occur in the distal parts of stratified basins. Current structures are not a totally dependable criterion since they may be produced by turbidity currents, sub-pycnocline tidal currents and geostrophic bottom water currents (especially contour currents, see Stow & Lovell, 1979).

It should be possible to interpret progressive offshore changes in facies in terms of zones of hydrographic stability (see beginning of chapter and following section on the biological implications of watermass stratification). Proceeding from onshore to offshore there will be a zone of continuous turbulence (i.e. above the fairweather wave base at about 10-15m; Kukal, 1971; Ingle, 1975), a zone which is only seasonally affected by strong turbulent wave activity (between fairweather and storm wave bases), and a zone below the main pycnocline which is essentially an area of quiet mud deposition. Where tidal influences are minimal sand bodies will be limited to the foreshore zone, but where they are strong, tidal sand bodies may be widely distributed above the main pycnocline, with only turbidity current sands occurring below. One aspect of this rather fundamental zonation is that the only syndepositional reservoirs in meromictic black shale basins will be turbidites, mass flows and submarine fan channel sands. Having established which hydrographic stability zone applies to any set of sediments (using all or any of the various criteria outlined in this Chapter) it is therefore possible to predict the likely geometry and character of potential reservoir units in the same zone and up and down the depositional dip. A classification of environments based on hydrographic stability zones is clearly more flexible and more realistic than any attempts to define specific bathymetric limits to any particular facies. Since such zones provide an accurate description of the depositional environment (through their sedimentological, biological and geochemical implications), they represent a very satisfactory way of side-stepping the whole time consuming and largely irrelevant question of palaeobathymetry. The general organisation of the classification is shown in Table 2.5; for the sake of simplicity the presence of salinity stratified zones, haloclines and fronts within the neritic mixed layer (e.g. estuaries and deltas) has been ignored.

As well as tending to trap sand grade clastics in the 'nearshore zone' stratification also has a very significant effect on the dispersal of clay grade sediments. Such sediments are added to the basin via the suspended sediment load of rivers and the offshore movement of fine material resuspended by wave and current action in coastal areas (see reviews by McCave, 1972 and Drake, 1976). In their works on the Oregon continental shelf Harlett and Kulm (1973), Kulm et al. (1976), McCave (1979) and Pak et al. (1980a) have shown that suspended sediment becomes concentrated at the seasonal and permanent thermoclines and is transported offshore (and partly onshore) across isopycnal surfaces. Bottom nepheloid layers may form low density turbidity currents (c.f. Moore, 1969) and flow downslope into the bottom water zone (and possibly 'cascading' through isopycnal surfaces, see Drake, 1976) and there deposit turbiditic mud layers (e.g. Rupke & Stanley, 1974; Hess, 1975; Stow & Shanmugam, 1980). Such sediments are probably more common in fine grained basinal sediments than has been previously supposed.

If the permanent pycnocline results in meromixis and the bottom

Table 2.5 Hydrographic stability zones and sedimentary regimes

Hydrographic regime	Benthic boundary	Pelagic boundary		Zone	Sand bodies
Continuously mixed			NERITIC	Upper mixed layer	Shoreface, deltas, bars
	Fairweather wave base	Shallow sea front		Seasonal pycnocline	
Seasonally stratified				Lower mixed layer	tidal bars, storm layers, deltas
	Storm wave base	Ocean-shelf front		Permanent pycnocline	
Permanently stratified			OCEANIC		turbidites, mass flows, contourites
				Bottom water	submarine fan channels

Note: the ichnofacies zonation scheme of Seilacher (1967) may be related to this classification of environments as follows: *Skolithos* ichnofacies - upper mixed layer; *Cruziana* ichnofacies - lower mixed layer; *Zoophycus* and *Nereites* ichnofacies - pycnocline and bottom water.

water becoming anaerobic, all the sediments deposited below the pycnocline (other than resedimented units) will be laminated. Sediments deposited within the pycnocline (as it exists at any one time) will show a vertical and hence lateral gradation of increasing bioturbation as the mixed layer is approached, and those deposited above the pycnocline will be rippled, cross-bedded and variously bioturbated depending upon substrate and sedimentation rate controls. These features are discussed more fully in the following section. Since the development of a pycnocline can be brought about without there necessarily being any deepening of the water column, the presence or absence of current structures in sediment sequences containing black shales should not be taken as an infallible indicator of relative water depth. Laminated black shales in themselves are no proof of actual increase in water depth and their transgressive nature must be determined by examining their temporal and spatial distribution.

THE PALAEOECOLOGICAL IMPLICATIONS OF WATERMASS STRATIFICATION AND MEROMIXIS

1. Plankton

Tappan (1968) has proposed that the early phase of a transgression is characterised by high primary productivity and that the later stages are characterised by low primary productivity. She believes that "phytoplankton abundance may have been controlled by contemporaneous continental physiography, through its effect on climate, atmospheric circulation and oceanic upwelling" (p.187). However, there is little reason to suppose that the environment associated with the early part of a transgression should be especially favourable to the phytoplankton. Although transgressions may be rapid in the geological sense they are infinitesimally slow when compared with the life span of planktonic organisms and the rapidity of nutrient recycling processes. Any changes brought about by the transgression itself will be so slow that they will be imperceptible to the living ecosystem. However, as the sea level rises the coastal zone (the most productive part of the marine habitat) will migrate laterally and so a phase of apparent increased productivity may be reflected in the basal transgressive sediments (if they are of suitable facies). If the later phases of the transgression are characterised by widespread stratification of the epeiric shelf seas and oceans then this would result in low primary production due to the lack of biolimiting nutrients (especially P and N). The stratification model thus provides a more eloquent explanation of variations in palaeo-productivity than the model described by Tappan (1968).

Schlanger and Jenkyns (1976) and Jenkyns (1980) have suggested that the productivity of shelf seas increases during transgressive episodes but this is largely incompatible with a model which assumes widespread watermass stratification. An increase in the net area of the shelf seas does not necessarily imply an increase in productivity per unit area. Some confusion undoubtedly arises from the tendency to place too much emphasis on productivity as a control on the organic content of sediments. In the majority of cases the organic-rich nature of black shales may be attributed to a decrease in the efficiency with which organic matter is normally recycled within the environment rather than a large increase in the rate of primary productivity. Having said this, transgressions may result in an increase in the area of coastal waters whose higher productivity may to an unknown degree compensate for the general decline in the rest of the sea. Should the shelf seas become meromictic then the accumulation of nutrients in the bottom waters may increase the productivity of frontal zones and result in various degrees of eutrophication.

If epeiric black shales <u>are</u> indicative of regional watermass stratification and decreased efficiency of nutrient recycling,we might expect an overall decline in primary productivity and a relative expansion in those groups of phytoplankton adapted to stable conditions and low nutrient concentrations in the mixed layer. We have already seen that the two major groups are the dinoflagellates and blue-green algae; another is the coccolithophores which presently reach their maxima in oligotrophic tropical waters (Riley & Chester, 1971; Berger, 1976), but are widespread as a group. Under stable meromictic situations the dinoflagellates may cease to produce benthic cysts (see Chapter 3), while the blue-green algae often do not produce distinct remains and it is therefore difficult to assess their specific contributions to the organic content of black shales (except perhaps by geochemical means). Coccoliths on the other hand are often conspicuous in Mesozoic to Quaternary black shales (e.g. Müller & Blaschke, 1971; Busson & Noel, 1972; Ross & Degens, 1974; Kemper & Zimmerle, 1978, Noel & Manivit, 1978). Coccolithic pelletal laminae in these deposits probably record seasonal mixing events which supplied nutrients from below the pycnocline into the mixed layer and made them accessible to plankton other than the dinoflagellates and blue-green algae. Their presence suggests that the nutricline was probably present at a level within the reach of seasonal mixing processes. An abundance of blue-green algal organic matter in the sediments probably suggests eutrophication, with the nutricline and main pycnocline occurring within the euphotic zone. Dinocysts will only be abundant in areas of seasonal stratification (organic-poor or black mud sediments) or seasonal up-welling and therefore usually only in marginal areas.

It should be noted that the presence of conspicuous coccolithic faecal pellets in black shales is at least partly a consequence of their contrast in colour with the sediment matrix. Supply of the faecal pellets may be constant but only become visible during black shale deposition. Anaerobic bottom conditions will also enhance the preservation of the coccolith calcite (see later) and also reduce the disruption of the pellets by bioturbation. More significant is perhaps the variation of abundance of coccolith pellets <u>within</u> black shales; I have observed that they often increase upwards, which suggests a progressive modification of plankton dynamics in the basin of deposition.

2. Benthos

Black shale biofacies

In stratified anaerobic watermasses three main layers of differing dissolved oxygen concentrations can be observed:

1. The mixed surface layer with 8-5 ml/l of dissolved oxygen,

2. the pycnocline through which the level of oxygenation progressively declines as turbulent mixing decreases with depth, and

3. the bottom water in which oxygen is absent and free H_2S may be present.

That part of the sea floor corresponding to the mixed layer will contain a normal marine fauna subject to the usual ecological controls of substrate, etc., but dissolved oxygen rapidly becomes the most important control in the deeper parts of the basin. Theede et al. (1969) have shown that the invertebrates most resistant to low oxygen concentrations and H_2S are those which normally inhabit soft mud substrates (particularly the infaunal forms). In fine grained muds with poor interstitial water circulation, conditions rapidly become reducing below the sediment-water interface (within 1-2 mm of the surface in the absence of bioturbation; Rhoads, 1974), and so any inhabitants of this type of substrate must be able to tolerate anaerobic conditions (if only temporarily) or be able to constantly irrigate their burrows. Burrowing pelecypods of tidal flats close their shells and carry out anaerobic respiration during the short interval between tides (Braefield, 1972; Newell, 1973). The major adaptions to low oxygen levels and H_2S which will be reflected in trace and body fossils are:

(a) small body sizes (providing a more favourable surface area to volume ratio),

(b) efficient burrow ventilation,

(c) the ability to carry out temporary anaerobic respiration (the effects of which may be preserved in the shell microstructure; Lutz & Rhoads, 1977).

Pelecypod faunas of fine grained muds often consist of small, thin-shelled individuals. Small sizes are an adaptation to flotation in soft soupy substrates as well as to low oxygen concentrations (Stanley, 1970; Rhoads, 1974; Thayer, 1975). The thin shells serve to reduce the bulk density of the organisms (again favouring bouyancy within the substrate) and may result from impaired calcite precipitation due to high pCO_2 levels produced by organic decay and the use of the shell in buffering the acid products of anaerobic respiration (Rhoads & Morse, 1971).

The profound effects that the differing levels of oxygenation have on the benthic life has allowed Rhoads and Morse (1971) to define three biofacies (see also Byers, 1977). The three biofacies are here related to watermass structure as observed in anoxic basins:

(a) The aerobic biofacies: this occurs in that part of the basin corresponding to the mixed layer and top part of the pycnocline where dissolved oxygen concentrations are greater than 1.0 ml/l. It contains an abundant and diverse benthonic fauna (including heavily calcified forms) and the sediments are intensively bioturbated unless sedimentation rate is limiting.

(b) The dysaerobic biofacies: this occurs in that part of the basin corresponding to the lower part of the pycnocline where dissolved oxygen concentrations range between 1.0 and 0.1 ml/l. It contains a low diversity (but often high density) fauna which is characterised by an almost total absence of organisms with calcareous hard parts (except for small, thin-shelled, infaunal protobranch pelecypods and echinoderms such as holothurians). The fauna is dominated by soft bodied metazoans (i.e. polychaetes, crustaceans, etc.); bioturbation is present but variable. The dominance of infauna does not reflect a selection for the infaunal habitat but reflects the fact that those invertebrates most tolerant of low oxygen concentration are infaunal forms. Deep burrowing will be precluded by highly toxic interstitial waters and low metabolic activity.

(c) The anaerobic biofacies: this occurs in that part of the basin corresponding to the very base of the pycnocline and all of the bottom water where oxygen concentrations are less than 0.1 ml/l. There are no metazoan organisms and primary laminations are preserved. The term anoxic is used in this paper to describe that part of the anaerobic zone which contains free H_2S.

Much of the information used by Rhoads and Morse to construct this biofacies scheme came from the oxygen minimum zone of the Gulf of California (and the Black Sea). It should be appreciated that such oxygen minima represent an unusually stable and thick layer of dysaerobic conditions in comparison with the relatively thin dysaerobic layer in anoxic basin pycnoclines. Byers (1977) has already emphasised that the pycnocline represents only a small proportion of the water column and so in anoxic basins we should expect that,at any one time,the dysaerobic biofacies will be very limited in comparison with the anaerobic biofacies even when bottom slope effects are taken into account. This observation suggests that the maximum spread of dysaerobic conditions occurs only during the initial de-oxygenation of the basin or during transient (seasonal?) hypoxic episodes in basins where anaerobic conditions never fully (or only locally) develop.

There are relatively few modern accounts of the effects of oxygen stress on recent faunas which are of use to the geologist. Rhoads and Morse (1971) have synthesised much relevant data on both stable anoxic stratified basins and oxygen minimum layers (but see also Rosenberg, 1977 and Thiel, 1978). The faunal effects of long term, periodic anoxic episodes in the Baltic Sea (Bornholm Basin) have been described by Tulkki (1965) and Leppakoski (1969, 1971) and seasonal anaerobic to anoxic events are discussed for various areas by Hoos (in Davis, 1975, p.2317), Kitching et al. (1976), Pearson and Rosenberg (1978), Kolmel (1979) and Jørgensen (1980). Experimental studies on the effects of hypoxia and H_2S in marine invertebrates have been reported by Theede et al. (1969), Theede (1973) and Oertzen and Schlungbaum (1972); the latter indicate that the pelecypod genera Arctica, Astarte, Corbula, Macoma, Mya, Mytilus and Scrobicularia are among the most oxy-tolerant at the present day.

The descriptions of seasonal anoxic events cited above indicate that in relatively small, nearshore basins (bays, estuaries, fjords) elimination and recolonisation of benthic faunas can be a regular occurence. These changes are brought about by seasonal trends in stratification, productivity and sediment B.O.D., with oxygen depletion taking place in the summer, followed by rapid re-colonisation in the Autumn. This time scale contrasts with the longer periods of anoxia and the slower, transitory return of benthos observed in the Baltic Sea. Leppakoski (1969) notes that it may take some time before the R.P.D. retreats below the sediment surface even after aerobic conditions have been established in the watermass. In common with other authors he notes that polychaetes are the most resistant to hypoxia and are the first to re-colonise the sediment (e.g. Scolopos and Capitella in the Baltic). In the geologic context these observations are of some significance since they indicate that:

(i) If the interval between transient anaerobic or anoxic episodes is too brief to allow benthic recolonisation, the sedimentary evidence may appear to indicate stable, continuous anaerobic conditions. This is particularly so in the case of transient oxygen depletion in a permanent bottom water zone; where the changing conditions represent bottom water - mixed layer alternations, sedimentological evidence should indicate the real situation.

(ii) If the interval between transient anaerobic or anoxic episodes is very short, polychaetes (i.e. soft-bodied organisms) may be the only forms to recolonise the sea floor. If the 'improved conditions' are still only dysaerobic (or more specifically <0.5 ml/l O_2 according to Calvert, 1964, p.319-320) bioturbation will be absent, but above this level will be more extensive (but probably limited by a shallow R.P.D.). Such a condition may explain occasional intervals of slightly or moderately bioturbated sediment devoid of macrofossils within an otherwise apparently anaerobic or anoxic facies.

(iii) The larger the area that has been subjected to anaerobic conditions the longer the interval required for recolonisation. The first recolonisers will probably be eurytopic forms with pelagic larvae; repeated oxygen stress may prevent the subsequent return of more stenotopic species that would normally characterise that environment under aerobic conditions.

(iv) During brief (e.g. seasonal) episodes of generally dysaerobic/aerobic bottom water conditions sediment B.O.D. can result in the $O_2:H_2S$ interface (i.e. the R.P.D.) rising to a few cm above the sediment water interface. Under these conditions only benthos on slightly raised areas of the sea floor or those forms with long siphons will be able to survive (see Jørgensen, 1980 for analogue). This may account for occasional occurences of isolated patches of eurytopic benthic organisms or the presence of epifauna-biased benthic assemblages observed in some 'black shales', and is a dysaerobic biofacies indicator. This interpretation conforms with the comments made by Kauffman (1979, p.428-9); see also Kauffman (1978) and Seilacher (1982).

Modern examples of temporary oxygen depletion in the marine environment are not just limited to coastal embayments. Unusually prolonged thermal stratification of the shelf watermass coupled with atypical development of phytoplankton can produce widespread dysaerobic, anaerobic, and even anoxic conditions on the open shelf (e.g. New York Bight 1976; see Mahoney & Steimle, 1979; Steimle & Sindermann, 1978 and

Falkowski et al. 1980).

Rhoads and Morse's biofacies scheme can be combined with the biofacies classification of Schafer (1972) to give a more complete picture of the biological environment in stratified anaerobic basins (see Fig. 2.1). Schafer's scheme is a useful one which utilises two main criteria, bottom turbulence and oxygenation to describe various environmental regimes, its only draw-back being the rather unwieldy terminology. The four relevant biofacies of Schafer's scheme are:

(1) Lethal-lipostrate biofacies: bottom conditions are very turbulent and well aerated. Erosive structures and current bedding are common; high rates of sedimentation and reworking prevent colonisation by benthos. Shell lag deposits are common. Coarse, sandy substrates. Also termed lethal-heterostrate.

(2) Vital-lipostrate biofacies: bottom conditions are turbulent and well aerated. The sediments are sandy and contain common current and erosive structures. High sedimentation rates are typical and mobile deep-infaunal forms dominate the benthos. Escape structures (see Kranz, 1974) are common. Also termed vital-heterostrate.

(3) Vital-pantostrate biofacies: bottom conditions are calm and aerated. The sediments are muddy and without current structures. Bioturbation is usually complete,but lamination may occasionally be preserved by higher rates of sedimentation. Transient anaerobic episodes may occur due to organic decay during periods of low circulation. Also termed vital-isostrate.

(4) Lethal-pantostrate biofacies: bottom conditions are calm and anoxic. The sediments are laminated due to the absence of bioturbation and benthos. Nektonic and planktonic remains are common. Also termed lethal-isostrate.

<u>Black shale benthic foraminiferal biofacies</u>

Greiner (1969, 1970) has shown convincing evidence to suggest that calcium carbonate availability is the major control on the distributions of the three main types of benthonic foraminifera (i.e. porcellanous, and hyaline calcareous foraminifera and arenaceous agglutinating foraminifera). Low temperatures and low salinities both result in low levels of carbonate saturation making it increasingly difficult for organisms to secrete calcareous hard parts. The distribution patterns of the benthic foraminifera will therefore depend upon the relative efficiency with which they can secrete $CaCO_3$ and on the distribution of temperature and salinity within the watermass. Low salinity nearshore and cold bathyal environments are dominated by simple agglutinating foraminifera (e.g. see reviews by Ksiazkiewics, 1975; Boltovskoy & Wright, 1976) while normal marine environments with higher $CaCO_3$ levels are characterised by porcellanous calcareous forms, and environments of intermediate saturation are dominated by hyaline calcareous foraminifera. This trend reflects the fact that many agglutinating foraminifera do not secrete $CaCO_3$ to form their tests (Hansen, 1979) and that hyaline calcareous foraminifera, by using an organic nucleating surface for calcite growth (see also Govean, 1980), can secrete $CaCO_3$ more efficiently than the porcellanous forms (Greiner, 1969, 1970). Although Greiner confined his investigations to the effects of salinity and temperature, it is possible to extend this model to include low oxygen conditions and thereby model foraminiferal biofacies in black shale basins.

In low oxygen (dysaerobic) bottom conditions high rates of organic decay lead to high levels of P CO_2, resulting in a lowered pH and the tendency for solid carbonate exposed at the sediment-water interface to dissolve. The physio-chemical conditions therefore tend to make it rather difficult for organisms to secrete calcium carbonate, which will result in dysaerobic environments being dominated by small, thin-shelled, poorly ornamented hyaline calcareous microfaunas (Harman, 1964; Phleger & Soutar, 1973; Boltvoskoy & Wright, 1976). At the present day the hyaline Buliminids (e.g. Bolivina) dominate the dysaerobic biofacies of the oxygen minima (Harman, 1964; Phleger & Soutar, 1973; Douglas & Heitman, 1979; Ingle et al. 1980; Sen Gupta et al. 1981). Agglutinating foraminifera are apparently less tolerant than the hyaline ones (at least to oxygen and redox stress) and are mainly restricted to areas with greater than 0.3 ml/l of dissolved oxygen (Saidova, 1971; Boltovskoy, 1972). The nodosariids, another group of hyaline foraminifera were important members of the dysaerobic biofacies in the Mesozoic black shale basins (pers. comm. J. Exton). Ancient organic-rich sediments with microfaunas dominated by agglutinating foraminifera may be produced by differential preservation; gyttjas often show poor carbonate preservation when deposited in dysaerobic conditions (especially >0.4 ml/l O_2; Sholkovitz, 1973), but excellent preservation under anaerobic conditions due to differing bacterial diagenesis and poor water chemistry. It should be noted that hyaline foraminifera can survive in the 'anaerobic' biofacies since they require much less oxygen than metazoans (Phleger & Soutar, 1973; Boltovskoy & Wright, 1976).

The higher stress dysaerobic environment is characterised by a low diversity, but often high density microfaunal biofacies. Since much of the epeiric sea may have been anaerobic or dysaerobic during black shale sedimentation, under such conditions the maximum diversity would have occurred in relatively coastal areas and not increased offshore (except on the ocean margin) as it tends to do at the present day. The main microfaunal trends are summarised in Fig. 2.12.

WATERMASS STRATIFICATION AND ENVIRONMENTAL STABILITY

The physico-chemical gradients produced by watermass stratification provide an excellent basis for the interpretation of biological and faunal gradients in both normal (oxygenated) and black shale basins. Earlier we saw that the initial distributions of temperature, salinity, dissolved gases and nutrients in the water column are controlled by the distribution of turbulent mixing processes and that the degree of

	Environment	Benthic Foraminifera
Mixing of assemblages due to fluctuating environments and redeposition (varying with size and morphology).	Aerobic with hyposaline influence	Low diversity assemblage dominated by simple agglutinating foraminifera. Some small hyaline sp. Preservation variable.
	Normal marine aerobic	Moderate to high diversity assemblage composed of large, thick-walled, well ornamented calcareous forms (hyaline and porcellaneous) and more complex agglutinating sp. Reduced preservation associated with carbonate-poor sediments and bioturbation.
	Dysaerobic (especially $<0.4\,ml/l\,O_2$)	Low diversity (decreasing). Thin-walled, poorly ornamented, hyaline foraminifera dominate calcareous forms. Eurytopic agglutinating sp.. Foraminifera frequently small, but larger, flattened, forms with high surface : volume ratios also occur. Deteriorating preservation of $CaCO_3$. Densities increase below $0.2\,ml/l\,O_2$ due to absence of deposit feeders. Typical genera *Bolivina*, *Bulimina* and *Uvigerina*. Also partly meroplanktonic (?) forms such as some *Brizalina*, *Globulimina* and *Chilostomella* sp.
	Anaerobic	Lower diversity. Good preservation of $CaCO_3$. Otherwise as dysaerobic.
	Anoxic	No in situ foraminifera.

Fig. 2.12 Benthic microfaunal trends in stratified oxygen-deficient basins.

turbulent mixing decreased with depth. In a stratified watermass the strong turbulent mixing in the surface mixed layer diminishes through the pycnocline until diffusion appears to become the dominant process in the bottom water layer (assuming there is no benthic mixed layer). This means that the processes which control and bring about changes in the physico-chemical nature of the environment become less efficient and slower acting as depth increases. We have also seen that the development of a pycnocline stabilises the water column (decreasing temporal variation in the depth of the mixed layer). The combined result is that the degree of environmental variability and the rate at which changes in the environment take place,both decrease from the surface mixed layer to the bottom water; the environment is thus said to become more stable with depth. This kind of approach to the marine environment readily allows the modern ecological concepts of environmental stress (Sanders, 1969), environmental predictability (Slobodkin & Sanders, 1969), environmental stability (Bretsky & Lorenz, 1970), trophic resource stability (see Valentine, 1971, 1973, 1976; Ayala et al. 1975a) and spatial environmental heterogeneity (e.g. see discussions in Levins, 1968; Valentine, 1973) to be applied to the palaeoecological modelling of stratified basins. In the following section the term stability is used with reference to the physico-chemical environment rather than the biological systems or communities which it contains.

The concept of environmental stability embraces several meanings, the principal ones being what may be termed 'constancy' and 'persistence'. As used here "constancy" describes the amount of variability within the environment over relatively short time scales, while "persistence" relates to environmental changes over geological (and hence evolutionarily significant) periods of time. In order to characterise the constancy of the environment the following values must be considered for each variable comprising the physico-chemical environment (see Valentine, 1973; Jackson, 1977):

(i) the mean value

(ii) the variance

(iii) the pattern of predictability of the variance

(a) amplitude of perturbations

(b) frequency of perturbations

(c) regularity of perturbations, (i.e. are they predictable?)

(d) rate at which perturbations take place

Since many, if not most, of these variables are genetically linked, an examination of the parameters listed above will provide a general 'value' for the stability (constancy) of the environment as a whole. This value increases with depth and distance from shore, but is not a uniform trend since it must show discontinuity (of variable intensity) in the region of the pycnocline as do all the other environmental variables.

The mixed layer (especially its uppermost and coastal parts) is the least stable part of the marine environment and experiences variations on diurnal, seasonal and secular time scales. Below the mixed layer (i.e. in the vicinity of the pycnocline) variations are dominantly seasonal in nature, while hardly any variation is discernible in the stable bottom water environment. This is the situation which prevails in permanently stratified watermasses (such as the present day oceans).

Superimposed upon this pattern of stability distribution with depth we have lateral, latitudinal effects where seasonality in the physical environment is greatest at mid-latitudes and least at the poles and equator. Mid-latitude temporal stability changes (due to seasonal variations in insolation, etc.) include the establishment of seasonal thermocline-pycnoclines on the shelf (see earlier) and migration of the main thermocline-pycnocline in the oceans (see for example Pingree, 1975; Glemarec, 1973). These temporal and spatial migrations of the stability fields will not alter the general trend of increasing stability with depth but will mean that the benthic stability zones will reflect the vertical (and hence lateral) shifts of the stratification within the watermass; since the actual depths of these stability zones will vary widely in time and space (within the absolute limits determined by latitude) this emphasises the folly of the unwarranted pre-occupation with palaeobathymetric analysis that has characterised many palaeoenvironmental studies in recent years. There are no absolute depth markers which can be applied universally in time and space.

The biological effects of environmental stability (constancy) may be termed stress. A variable, unstable environment exerts a higher level of physiological stress and makes more demands on the homeostatic capabilities of organisms than does a stable environment. However, an unstable, high stress environment does not necessarily contain a highly stressed fauna since many organisms are acclimatised and/or possess behaviourial responses which act to reduce the level of stress. These stress responses are most efficient when the fluctuations in environmental conditions are of predictable intensity, duration and periodicity (see discussion in Slobodkin & Sanders, 1969). They allow many organisms to colonise otherwise inhospitable environments, such as high stress intertidal regions, where the range in conditions is generally greater than can be buffered by homeostatic regulation alone.

Since the homeostatic capabilities of an organism relate to the normal expected range in environmental conditions to be encountered during the life of an individual of any particular species (and although they may be mediated as above), unstable environments usually have the most tolerant faunas. Organisms which can tolerate (by one means or another) variable conditions are said to be eurytopic and those which are limited to more restricted conditions are said to be stenotopic (see Valentine, 1973; Menzies et al. 1973). Eurytopes dominate high stress, unstable environments and stenotopes dominate the lower stress, usually deeper, stable environments.

As well as differing in their tolerance to stress, eurytopes and stenotopes are characterised by different ecological and reproductive strategies. Eurytopes are opportunists that are adapted to exploit periodic (e.g. seasonal) peaks in resource availability and can thus reproduce quickly and in large numbers when such conditions arise. Eurytopes are also characterised by their wide geographic dispersal, both latitudinally and longitudinally, which is associated with long-lived, planktonic larvae with high tolerance to variable conditions and substrates (Jackson, 1974). Although they can occur in a variety of environments, eurytopes only dominate in high stress environments. This is because they are competitively inferior to the specialist stenotopes which occur in lower stress environments (Connell & Slatyer, 1977). The distinction between eurytopic and stenotopic faunas is maintained by the stability-stress gradient in the environment; stenotopes cannot cope with the environmental variability of high stress regions, while eurytopes are quantitatively limited to the latter regions by competition. Because of their more efficient dispersal eurytopes are the first to encounter new ecospace (e.g. new sea floor during transgressions or areas defaunated by temporary de-oxygenation) and are the only organisms that have sufficient tolerance to colonise physically or chemically stressed areas. This tolerance and their wide distribution results in eurytopic organisms being less likely to speciate or become extinct than are stenotopic organisms (Jackson, 1977; Kauffman, 1977).

Stenotopes are specialists which are adapted to,and have evolved under,stable (i.e. constant and persistent) environmental conditions. Because they live in a constant environment the reproductive strategy of stenotopes is not required to produce large numbers of young during short lived periods of optimum resource supply. Characteristically they produce a small number of young (most of which will survive to become adults), have slow growth rates and high longevity (e.g. see Wolff, 1977). Stenotopic faunas show well developed niche specialisation because of the more efficient resource partitioning which is possible in stable environments. Since they are adopted to a constant, invariable set of environmental conditions stenotopes can only be as widely distributed as is the environment in which they live. The high genetic variability of stenotopes (Ayala & Valentine, 1979) is employed in promoting specialisation rather than adaptive flexibility and hence makes them more dependent on the environment and, unlike as was once thought by Bretsky & Lorenz (1971), <u>less</u> resistant to environmental changes. In shelf seas 'environmental persistence' favouring stenotopes is probably a function of sea level, climate and the corresponding hydrographic patterns, with stable watermass stratification tending to introduce inertia into the system in such a way that changes are only produced by major events. In epeiric seas these phenomena are likely to be widespread and synchronous (i.e. within geological resolution), allowing an 'event stratigraphy' in both a sedimentologic <u>and</u> biostratigraphic sense.

Present day faunal influences of watermass stratification

Glemarec (1973), Menzies et al. (1973), Iverson et al. (1979) and Rosenberg & Moller (1979) have all documented evidence which indicates that the ecological zonation of benthic faunas is strongly related to environmental stability, which is in turn controlled by the structure of the water column. Glemarec (1973) divides the shelf into three faunual zones (etages) of differing degrees of seasonality, one in which the overlying waters are seasonally stratified, one in which they are permanently mixed, and one in which the overlying waters are always stratified (i.e. oceanic). Very similar hydrographic-faunal correlations have also been observed in shelf waters by Iverson et al. (1979). The prime significance of the pycnocline in maintaining the observed faunal patterns has been demonstrated by Rosenberg & Moller (1979). The work by Menzies et al. (1973) is mainly concerned with the shelf/ocean benthic faunal boundary. They conclude that the shelf faunal province cannot be defined by bathymetric limits or submarine topography but noted that the lower boundary of the province correlates with the pycnocline associated with the main ocean thermocline. They also observed that in latitudes where the thermocline (and therefore stress) gradient is high the shelf, transitional and oceanic faunal provinces were characterised by higher endemism and less mixing than in latitudes where the thermocline is weak. Thus high endemism occurs at the equator where the main thermocline is most intense, intermediate endemism occurs in mid-latitudes and polar faunas are cosmopolitan with depth.

Both Glemarec (1971) and Menzies et al. (1973) show their ecological depth-ordered zones deepening towards the equator. This should only be so if temperature (stenothermy) is an important factor and is not predicted by the environmental stability model which would otherwise require that the zones shallow toward the equator (along with the thermocline) and increasing hydrographic stability. It should be noted that the boundaries between the benthic stability zones will correspond to hydrological fronts within the watermass and that this provides a fundamental link between benthic and pelagic ecology (see also Table 2.5 and Chapter 3 for the possible use of dinocysts as palaeo-watermass stability indicators).

THE GEOCHEMICAL SIGNIFICANCE OF WATERMASS STRATIFICATION AND MEROMIXIS

Introduction

It is very apparent that the hydrodynamic conditions which result in watermass stratification have a profound effect on the distribution of physical and biological processes within the watermass. Where watermass stratification combines with restricted lateral circulation (and a sufficient supply of organic matter) bottom water deoxygenation will also result in a major shift in the chemical nature of the depositional environment. In shelf waters this deoxygenation will occur from the seafloor upwards, while in open ocean conditions oxygen minimum zones may be formed as described earlier. Richards (1965) and Grasshoff (1975) provide comprehensive summaries of the hydrochemistry of modern landlocked basins and fjords, and Hutchinson (1957) and Beadle (1974) discuss chemical aspects of stratified lakes. From the geological viewpoint the most relevant general papers are Krumbein and Garrels (1952), Krejci-Graf (1964) and Calvert (1976). The most significant papers in the author's view are, however, those by Borchert (1960, and slightly revised 1965).

Borchert's infrequently quoted papers on marine sedimentary iron ores already contain most of the elements of the watermass stratification model presented here even though they were written in the early 1960's. The similarity of the present model and that developed by Borchert specifically for sedimentary iron deposits may be appreciated from the following quote from his 1965 paper:

"Under more-or-less stagnant conditions a stratified system of zones may develop. The deepest zone of the watermasses will be richest in H_2S and will be in slow exchange by diffusion with the bottom sapropel enriched in organic matter. This is termed the H_2S-zone. The surface water, which is the main zone of plant life, is rich in oxygen and is in free exchange with the atmosphere; this is termed the O_2-zone. However, in the middle there is a gradual transition from the well-oxygenated surface zone down to regions of decreasing oxygen and increasing CO_2 content; this is termed the CO_2-zone" (Borchert, 1965 p.182-183).

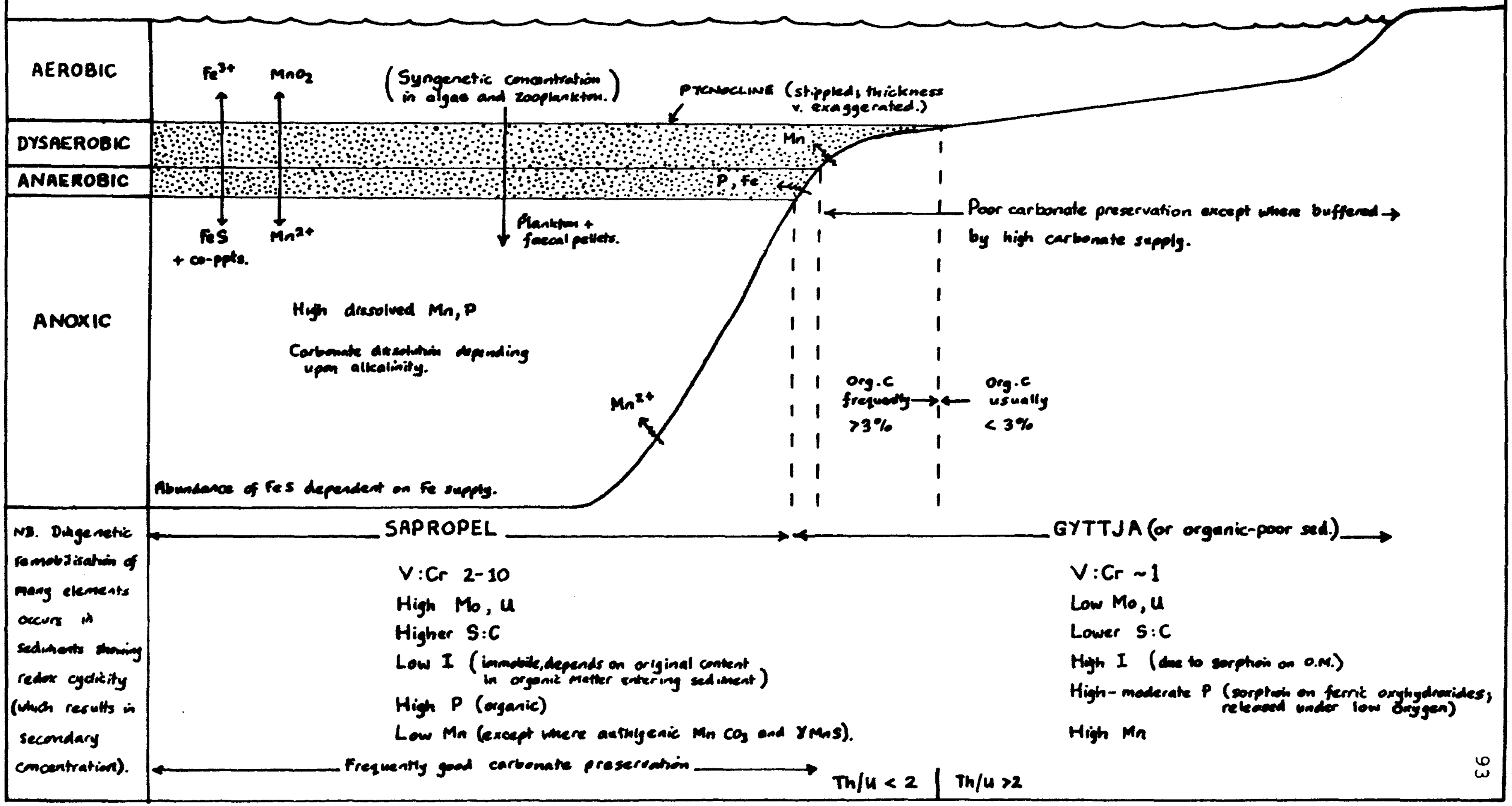
Although the terminology invented by Borchert is not used here, his three zones can be recognised on Fig. 2.13 which summarises the geochemical character of stratified anoxic basins. It should be noted that Borchert's CO_2-zone only really correlates with the dysaerobic and anaerobic parts of the pycnocline.

The essential chemical parameters of the mixed layer, pycnocline and bottom water layers of an idealised meromictic basin are summarised below. The generalised Eh and pH values given are based on Richards (1965) and Grasshoff (1975) and are also shown on the Eh-pH-O_2 diagram in Fig. 2.14 along with the stability fields of various minerals.

(a) The mixed layer; biological processes and gas exchange at the sea surface result in this layer being characterised by oxidising conditions (Eh = +400mV to +200mV) and high pH values (7.9 to 8.5). Organic matter is normally oxidised but may be preserved under the depositional conditions of the black mud megafacies, where in the pH and Eh of the water immediately above the bottom may be depressed. Oxygen content is high, normally between 6.0 and 8.0 ml/l. Sediment type will be influenced by hydrodynamic factors, clastic deposition rates and the abundance of calcareous benthic organisms.

(b) The pycnocline layer; almost by definition the chemical characteristics of this layer represent a transition between the surface and bottom waters and are more variable than either of the other two zones. The oxygen content declines from around 5.0 or 6.0 ml/l to less than 1.0 ml/l and is totally absent at the base of the layer. The total range of Eh values encountered is +400mV to -200mV. The large amounts of organic matter preserved by the low oxygen and Eh values near the bottom of the pycnocline decay and produce CO_2 and therefore lower the pH to 7.0 to 7.6. Values as low as pH 6.0 suggested by Borchert (1960,

Fig. 2.13 Some geochemical features of stratified, anoxic basins.



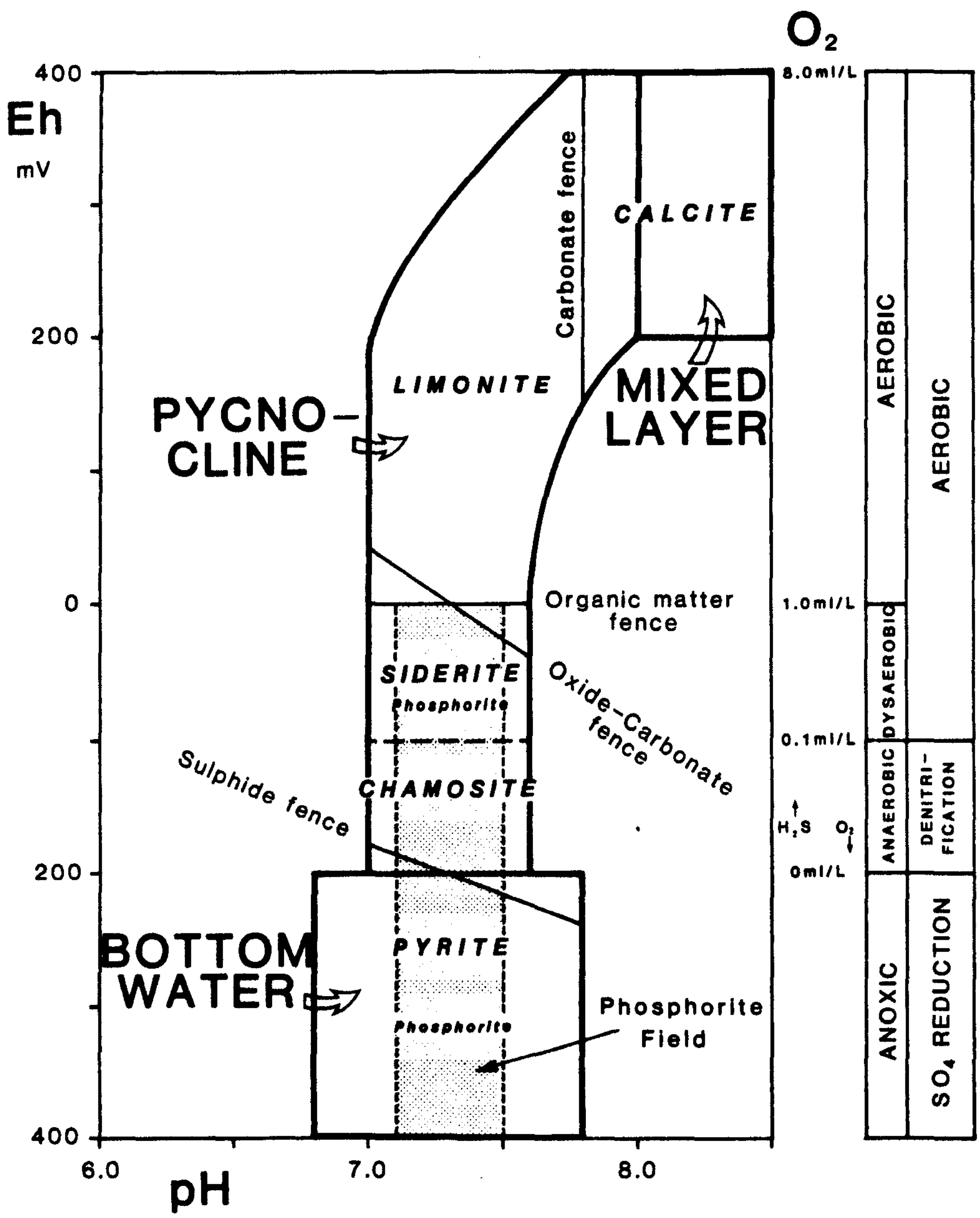


Fig. 2.14 Eh, pH and dissolved oxygen characteristics of the main watermass layers in stratified anoxic basins. Data from Krumbein & Garrels (1952), Richards (1965), Grasshoff (1975), Fenchel & Jørgensen (1977) and Sorokin (1978).

1965) are not supported by modern studies. Bacterial denitrification occurs at the base of the pycnocline where Eh values are below -100mV ($O_2 \leq 0.1$ ml/l). Contamination with H_2S from the bottom water layer may occur immediately above the $O_2:H_2S$ interface.

(c) The bottom water layer; oxygen is absent and free H_2S is present due to the action of sulphate reducing bacteria. Conditions are strongly reducing (Eh -200mV to -400mV). The production of ammonia and consequences of sulphate reduction often lead to somewhat higher pH values in the bottom water than in the pycnocline (pH 6.8 to 7.8). Again, pH values as high as 9.0 as suggested by Borchert (Ibid) are not supported by recent studies. The high concentration of solutes characteristic of bottom waters may in fact so increase their density that it can be a significant factor in the stability of the water column (Hutchinson, 1957).

Since the pycnocline generally corresponds with the location of the redox boundary in the water column it is the site of intense biological and chemical activity (e.g. Sorokin, 1978). Mixing through the redox boundary (Eh interface, $O_2:H_2S$ interface or chemocline) and vertical migrations of the same,may lead to the precipitation and/or dissolution of various mineral species, especially carbonate, sulphide and oxide phases (e.g. Degens & Stoffers, 1976; Degens et al. 1978; Stoffers & Muller, 1978; Spencer et al. 1972). It should be remembered that the $O_2:H_2S$ interface will only correspond with the base of the pycnocline under 'optimum conditions' as de-oxygenation has to proceed from the sea floor upwards through the bottom water. In shallow basins (such as lakes) this de-oxygenation may be rapid (e.g. anoxic conditions may be developed on a seasonal basis) but in deeper basins the time delay between the onset of a stagnation and the arrival of the $O_2:H_2S$ interface at the base of the pycnocline may be significant (see Deuser, 1974). The geochemical significance of the various conditions in each of the watermass layers is discussed below with particular reference to carbonates, phosphates and uranium and other trace elements.

Carbonates

Since temperature, oxygenation, pH and Eh are the main controls on the production, precipitation and preservation of biogenic and inorganic carbonates, the different watermass zones are characterised by varying amounts and types of carbonate materials. In terms of the production of calcite by benthic organisms there is a clear trend from the mixed layer to the bottom water. The mixed layer normally contains abundant heavily calcified benthic organisms which contribute significant proportions of carbonate to the sediment,but with increasing depth through the pycnocline falling oxygenation and pH conditions result in progressively poor and less calcified faunas. The production of biogenic calcite by benthic organisms in a stratified, anaerobic basin therefore shows a marked decrease in depth.

In surficial sediments deposited under aerobic and dysaerobic conditions the decay of organic matter and the production of CO_2 produces low pH values with promote the dissolution of carbonate (which is released into the bottom waters) and results in chalky, etched shells and tests. However, in the more restricted pore water environment the activities of the sulphate reducing bacteria result in increased alkalinity and shift the chemical equilibria in favour of solid carbonates. The principle problem for good carbonate preservation is therefore to get particulate carbonates through the R.P.D. as quickly as possible following sedimentation and/or death of carbonate secreting organisms. The most effective ways of doing this are to increase the sedimentation accumulation rate or to elevate the position of the R.P.D. within the sedqment by decreasing the bottom water oxygenation. Sholkovitz (1973) has, for example demonstrated poor carbonate preservation under dysaerobic bottom waters but excellent preservation under anaerobic bottom waters where the R.P.D. is presumably situated almost at the sediment-water interface (see also sections on the preservation of organic matter, the pericontinental megafacies and black shale biofacies). The physiochemical effects of bioturbation have been shown to lead to increased rates of carbonate dissolution (Aller, 1982) and, therefore, reduced bottom water oxygenation to dysaerobic and anaerobic levels presumably results in better preservation than under aerobic conditions because of the reduction in biological reworking. Exon (1973) and Warme (1971) have both noted the poor preservation of calcareous fossils in modern gyttja sediments and one must conclude that unless the accumulation of organic matter and hydrographic factors result in anaerobic conditions at the sediment-water interface, gyttjas will show worse carbonate preservation than ordinary organic-poor terrigenous clays and muds.

In the anoxic bottom water zone the dynamics of preciptation and dissolution of carbonate materials will be determined by the alkalinity of the anoxic waters. The modern Black Sea has unusually high bottom water alkalinity (Grasshoff, 1975) presumably reflecting the high input of calcareous detritus from its margins (Trimonis, 1974; Müller & Stoffers, 1974), and explaining the relatively high carbonate contents of the recent bottom sediments. The distribution of alkalinity in the oxygen deficient bottom waters of the Baltic is also primarily influenced by the amount of calcareous detritus supplied to the various sub-basins (Manheim, 1961).

It should be remembered that the hydrodynamic conditions prevailing in a persistently stratified basin will prevent the transport of shell debris from the mixed layer into deeper water by all means except turbidity currents and mass flows. The only input of carbonate into the bottom water independent of hydrodynamic and benthic ecology controls will be from the surface waters via calcareous plankton and inorganic precipitates. If the anoxic bottom waters are rich in bicarbonate ions lowering of the chemocline may lead to the precipitation of carbonates due to the change in chemical equilibria (see Degens & Stoffers, 1976; Degens et al. 1978; Stoffers & Müller, 1978). The composition of the carbonates will be determined by the cation balance of the watermass and influenced by evaporation, the ionic balance of the rivers entering the basin and the distribution of Eh and pH within the watermass. If the inorganic carbonates precipitated are not dissolved under low pH conditions they may be preserved as single laminae within sapropels (as described by Stoffers & Müller, 1978). The precipitation of carbonates in the water column produces 'milky water' or 'whitings' as described in detail from Lake Zürich (Kelts & Hsü, 1978).

Hsü and Kelts (1978) have described calcitic rhythmites (chalks and seekreides) from the Pliocene of the Black Sea. Chemogenic calcite is common in the modern sediments of the Black Sea but is apparently absent or rate in the 'halistatic' (oxygen-deficient) areas where carbonate is predominatly coccolithic (Trimonis, 1974). Coccolith-rich laminae are common in Mesozoic to Quaternary black shales and calcareous nannoplankton must be considered one of the dominant sources of calcite in stratified basins. Coccoliths sedimented via faecal pellets may be protected from dissolution prior to their burial by pellet 'peritrophic' membranes. Widespread coccolithic rhythmites were deposited under anaerobic (but fluctuating) conditions in the Blake-Bahama Formation of the central Atlantic during the Neocomian (Lancelot et al. 1972; Jansa et al. 1977; Tucholke & Vogt, 1979).

When studying kerogenous sedimentary rocks one must bear in mind the considerable information loss that may have occurred due to the dissolution of calcareous fossils. This is especially important when determining whether anaerobic conditions were present or not and emphasises the importance of trace fossils (see Byers, 1979). In contrast to the normal trends it appears that aragonite (which is metastable) is better preserved in organic-rich sediments (e.g. Callovian clays, Hudson & Palframan, 1969). Kennedy and Hall (1967) suggest that this may be due to the formation of a protective layer of hydrophobic organic molecules around the shell material isolating it from the pore fluids. The characteristic minimal carbonate contents of most black shales indicates that dissolution combined with originally low calcite production are major determinants of the sediment type.

Phosphates

Phosphorus is one of the most essential components of all living organisms and is therefore present in all living organic matter (phytoplankton for example have a carbon to phosphorus weight ratio varying between 24 and 41; Calvert, 1976). When organisms die and decompose the phosphorus contained in their organic matter is re-mineralised and recycled through the living ecosystem. In stratified watermasses (such as temperate seas in summer) dead phytoplankton may settle through the water column and decompose and be re-mineralised below the pycnocline in such a way that the phosphorus liberated is trapped and accumulated in the bottom water. The stability of the water column resulting from a pycnocline results in low mixing and the regenerated nutrients accumulate beyond the reach of most of the phytoplankton living in the surface waters. Generally this situation will lead to decreasing primary productivity,but in quasi-estuarine circulation cells associated with coastal upwelling,high productivity and nutrient-trapping may occur hand-in-hand (see Brongerma-Sanders, 1965 et seq.). The bottom waters of stratified basins are, therefore, usually significantly enriched in dissolved phosphorus. Anoxic to dysaerobic bottom conditions will tend to favour the accumulation of phosphorus by the preservation of phosphorus-containing organic matter, even though such conditions will lead to the release of inorganic phosphorus by the reduction of ferric oxyhydroxides (e.g. see Krom & Berner, 1981). It is not surprising that there is a common association between black shales and phosphorites.

According to Youssef (1965) and Pettijohn (1975) two general types of ancient phosphorite deposits can be recognised:

(a) Basinal ("geosynclinal") bedded phosphorites associated with black shales and cherts. These phosphorites are generally dark coloured, fine grained with elongate, angular grains of fluorapatite (francolite) and contain organic matter, pyrite, clay, micrite and rate silt.

(b) Platform nodular phosphorites with little or no organic matter or pyrite and containing coarse, rounded grains of fluorapatite (francolite).

It is apparent from the literature that the second type is largely considered to be derived from reworking and modification of the former (see for example El-Tarabili, 1969; Baturin, 1971; Birch, 1979; Bremner, 1980; Burnett, 1980) and so only the genesis of primary, type (a) phosphorites is considered below.

Gulbrandsen (1969) notes that the accumulation of abnormal concentrations of dissolved phosphorus, either in bottom waters or interstitial solutions, must be an essential pre-requisite of phosphorite formation. This can only happen when and where the supply of phosphorus (via organic matter) exceeds the rate at which it is recycled; a condition which is effectively only fulfilled in stratified, oxygen-deficient environments. Since this equation must be most favourable in quasi-estuarine cells at the margins of stratified basins,it is not surprising that the contemporary phosphorites are largely confined to the oxygen minimum zones of upwelling areas (on the coasts of Peru, Chile, Namibia and South Africa; Veeh et al. 1973; see also O'Brien & Veeh, 1980). These contemporaneous phosphorite nodules occur in laminated, diatomaceous oozes rich in organic matter (incipient diatomaceous black shales). Most modern theories of phosphorite formation have used these recent deposits as actualistic or physiochemical analogues.

Manheim et al. (1975) have strongly argued for the formation of phosphorite by the diagenetic replacement of calcite. They consider that the high phosphate content of the bottom water is essential only for maintaining high productivity and the input of P-rich organic matter into the sediment. Similarly, and more to the point, Calvert (1976) considers that low Eh conditions are essential to prevent the organic matter being totally re-mineralised (and its phosphous returned to the bottom water) before it enters the sediment, so that decomposition within the sediment will provide P-rich, diagenetically active pore waters. One problem with the replacive diagenetic model is that organic-rich sediments often contain very little carbonate (see above) which is available to be replaced, although Birch (1979) emphasises that lateral migration of oxygen minimum and upwelling zone may occur, and that the calcareous material being replaced may have formed under a previous set of environmental conditions unlike those during phosphorite genesis. Birch explains the occurrence of unaltered shell material in phosphorites as a surface-area effect during diagenesis, where smaller particles such as micrites are much more easily altered than larger shell debris. Other workers (e.g. Burnett, 1977, 1980) consider that the fluorapatite precipitates directly from anoxic interstitial waters rich in phosphorus by nucleating on detrital grains. Presumably both mechanisms occur but to varying extents.

Veeh et al. (1973) have made an important observation which may resolve some of the problems surrounding phosphate formation. They observed that contemporaneous phosphorite nodules forming off Peru were "confined to two narrow bands roughly coinciding with the upper and lower boundaries of the oxygen minimum layer" (p.844). This is also indicated by Fig. 4 in Burnett (1980). If this observation is of a general applicability it suggests that phosphorite formation is localised in the vicinity of the anaerobic-dysaerobic boundary, and as can be seen from Fig. 2.14, such conditions do in fact encompass the phosphorite window (pH 7.1-7.5). It would appear that the optimum geochemical conditions for apatite precipitation occur in these areas at the margin of the oxygen minimum, but a rationalisation of this fact has yet to be forthcoming.

In summary, the conditions required for phosphorite formation are:-

(a) Persistent quasi-estuarine circulation on the margins of stably stratified basins.

(b) High productivity resulting from upwelling (corollary of (a)).

(c) Oxygen deficient bottomwaters (dysaerobic-anaerobic conditions).

(d) The presence of pH conditions between 7.1 and 7.5.

(e) Minimal clastic sediment accumulation rates.

Trace elements other than Uranium

One of the major characteristics of black shale type sediments is their enrichment in various trace elements, particularly Ag, Mo, Zn, Ni, Co, Cr, V, Pb and U (Vine & Tourtelot, 1970; Pettijohn, 1975; Calvert, 1976; Brumsach, 1980). Their trace element composition is strongly influenced by the nature of the mineralic components and the amount of organic matter which they contain and may result from syn-depositional and/or diagenetic processes or mineralisation due to external agencies. Vine and Tourtelot (1970) considered that if trace element enrichment resulted from a syngenetic process (where the trace element composition of black shales would be strongly influenced by the depositional environment) it would be necessary for abnormally high trace element compositions to have occurred in the seawater. Holland (1979) has reappraised the situation with the assistance of more accurate trace element data than was available to Vine and Tourtelot,and has concluded "that the removal of trace elements from seawater in anoxic basins can account for the observed enrichment of many metals in black shales" (p.1679).

Since the presence or absence of reducing conditions and free H_2S in the environment will undoubtedly have a considerable influence on the geochemical processes determining the distribution of dissolved, complexed and particulate trace elements,it is not surprising that gyttjas and sapropels differ in their trace element compositions.

Krecji-Graf (1964, 1972) has discussed these differences. Chromium is apparently characteristic of gyttjas, Pb and Zn are concentrated under negative Eh conditions and Cu, Ag, V, Mo and Ni characterise more strongly reducing conditions. These differences are conveniently expressed by the ratios of various trace elements, amongst which the chrome/vanadium ratio is most useful (see also Ernst, 1970). According to Krejci-Graf (1972) Cr/V ratios of gyttjas are around one, while for sapropels this value is said to be grater than unity; trends are doubtless more informative than are specific ratios. Piper (1971) studied the distribution of various elements including chromium in a stratified, anoxic fjord and found that while chromium was abundant as a suspended phase above the pycnocline (and therefore in the gyttja sediments) it decreased sharply through the pycnocline and dissolved in the anoxic bottom waters where the pH was ≤ 6.9.

A variety of elements precipitate at the redox boundary associated with the $O_2:H_2S$ interface at the base of the pycnocline. Iron and zinc descending through the pycnocline precipitate as insoluble sulphides and probably lead to a flux of various other trace elements (e.g. Mo) to the bottom because of co-precipitation (e.g. Piper, 1971; Spencer et al. 1972). It is not surprising that many trace elements are positively correlated with pyrite and with organic matter (which is responsible for the reducing conditions). Upward movement of water from below the pycnocline also results in precipitation (e.g. the oxidation of Mn (II) to MnO_2 seen in the Black Sea, Spencer et al. 1972) and the co-precipitation of various elements and trace elements. Because the pycnocline-chemocline is such a chemically active part of the watermass, the content of certain trace elements may be higher in sediment deposited in this region than either the mixed layer or bottom water (e.g. Manheim, 1961).

The question of whether the association between organic matter and high trace element contents is due to biochemical processes (i.e.

occurring while the organic matter is part of a living organism) or chemical processes (i.e. inorganic chemical reactions involving inert organic matter) is a fairly controversial one. Holland (1979) concluded that the concentration of metals in organic-rich sediments owed more to chemical precipitation and to reactions with dead organic remains than to their incorporation in living organism,but did not altogether dismiss the alternative possibility. Brongersma-Sanders (1965- 1966, 1969 and 1971) has been a great champion of the syngenetic theory of trace element enrichment in black shales. She believes that the nutrient trap conditions produced by quasi-estuarine circulation in areas of upwelling (which result in high productivity),also lead to a significant flux of biochemically combined trace elements to the sea floor which are liberated by the decay of the organisms involved and fixed in the sediment by precipitation with H_2S. In their work on organic-rich diatomaceous oozes deposited under anaerobic conditions in the upwelling area of S.W. Africa, Calvert and Price (1970, 1971) did find a clear association between organic matter and certain trace elements (Cu, Pb and Zn) but much lower enrichment factors than those that occur in Kupferschiefer, the specific example which Brongersma-Sanders was trying to explain. Holland (1979) considers this particular deposit unexplainable in normal environmental terms and it would appear that Brongersma-Sanders' (1969) statement that the "syngenetic origin of the metals of the Kupferschiefer is now well established" was somewhat premature. This has now been confirmed by Brongersma-Sanders et al. (1980).

Chester et al. (1978) report that the ratio between Pb, Zn, Cu and organic carbon in surface water particulates from the highly productive waters off W. Africa was very similar to that in the most metal enriched sediments described by Calvert and Price (1971). This clearly suggests that the metal enrichment in the organic-rich diatomaceous oozes was due almost entirely to the content of metals in the particulate matter (dominantly plankton and faecal pellets), and that specific values were influenced by the degree of sediment dilution. Martin and Knauer (1973) have reported on the trace element contents of microplankton and have shown that although there is significant variation some enrichment is at least evident. Fowler (1977) has more recently shown that zooplankton (copepods in particular) also effectively concentrate a number of trace elements. Fowler notes significant trace elements concentration factors for copepod faecal pellets, which because they are produced in abundance (about 200 per individual per day, Honjo & Roman, 1978) and decompose relatively slowly, must be considered significant sources of trace element input to sediments. Thus not only do microplankton concentrate trace elements in comparison to their levels in sea water, but zooplankton lead to an even greater concentration. Bostrom et al. (1974) believe that elements once incorporated into the pelagic food chain may tend to become concentrated despite moulting, etc.

Landry (1977) notes that a consequence of the highly complex organisation of pelagic food webs (particularly the absence of well defined trophic levels) is that 'toxic pollutants' become highly and uniformly concentrated in the zooplankton. If this effect is also evident with trace elements the hypothesis of Bostrom et al. (1974) may be correct. The latter authors also speculated that under low nutrient concentrations the micro-plankton may have to 'filter' more sea water to obtain their nutrients and that as a consequence they might also concentrate other metabolically inactive trace elements. In retrospect it would appear that syngenetic concentration of trace elements is a realistic process but that it cannot be used to explain the very high metal contents observed in some shales.

Uranium

The concentration of significant amounts of uranium in modern organic-rich sediments (black muds, dysaerobic to anaerobic gyttjas and sapropels) and ancient black shale deposits is a well known phenomenon of considerable academic and commercial interest (Veeh, 1967; Baturin, 1973; Swanson, 1960). Whereas normal argillaceous sediments contain between 2-4 ppm, recent organic-rich gyttjas and sapropels may contain up to 80 ppm or more (Table 2.6) and phosphatic sediments have been reported to contain as much as 158 ppm of uranium (Veeh et al. 1971).

The majority of the uranium present in sea water occurs as the soluble stable uranyl carbonate complex $(UO_2(CO_3)_3)^{4-}$ in which the uranium is in its 6^+ valency state (Baturin, 1973). There are several possible mechanisms by which this uranium may become incorporated in sediments:

(i) Reduction to the 4^+ valency state and precipitation as insoluble $U(OH)_4$.

(ii) Adsorption on clays.

(iii) Adsorption on and incorporation into organic matter.

(iv) Substitution for Ca^{2+} in the crystal lattice during growth of flourapatite.

While the reduction of uranium from U(VI) to U(IV) is generally reported to occur at Eh values below -100 mV (e.g. see Calvert, 1976), Nikolayev et al. (1977) claim that this reduction only occurs at redox potentials between -400 and -500 mV. From Fig. 2.14 it can be seen that this process will not occur in the water column of anoxic basins and will be limited to the interstitial waters of highly reducing sediments. This is in keeping with the fact that no tetravalent uranium has been reported from the waters of the Black Sea (Kolyadin et al. cited in Kolodny & Kaplan, 1970). Kolodny and Kaplan (1970) believe that at pH values of $\geq$ 7, uranium should be reduced and precipitated by co-existing pyrite to its tetravalent state until the concentration of uranium in the interstitial waters is less than 1 ppb.

The uranium content of sedimentary rocks is clearly correlated with high organic contents, and is therefore also correlated with fine grained sediments and dysaerobic-anoxic bottom conditions (Baturin, 1973; Calvert, 1976). Baturin (1973) considers that "uranium is abiogenic and is not concentrated in the living tissue, or in oceanic suspensions containing up to 20% organic carbon". If, however, uranium could be syngenetically concentrated as in the case of certain other trace elements, this would help to explain the high uranium contents in some sapropelic black shales. Some evidence for syngenetic enrichment has been found by Degens et al. (1977) when studying Holocene sediments from the Black Sea. The latter authors found that the bulk of the uranium in the uppermost coccolithic sapropel sediments seemed to be bound to planktonic organic matter and that the land-derived (humic) organic debris contained comparatively little uranium. They concluded that (in this case) coccoliths were the prime host for the uranium but that other planktonic organisms showed the same affinity. This is even more surprising when one considers that the underlying sediment, a non-coccolithic sapropel richer in organic matter, has a lower uranium content (30 compared with 39 ppm). Degens et al. (loc. cit.) also described a 10,000-fold uranium enrichment in diatoms in a Holocene lake in Ontario polluted by waters from a uranium mine; the diatoms contained 210 ppm in comparison with 20 ppb of uranium in the lake water. These observations suggest that where dissolved uranium is available in slightly greater amounts than normal it can be considerably concentrated by plankton and fixed in the sediment. Degens et al. (1977) note that in the Black Sea the uranium content is higher in the pycnocline than above or below it (5.9 ppb as opposed to 2.4 ppb above and 3.5 ppb below), which may also enhance the potential for syngenetic enrichment, since in the basin centre the plankton is concentrated at the base of the mixed layer (Caspers, 1957). Note that during the deposition of the non-coccolithic sapropel (Unit Two) this dissolved uranium peak would have been deeper and 'unavailable' to the plankton.

In his study of the geochemistry of recent sediments from the Baltic Sea, Manheim (1961) found higher concentrations of uranium in gyttjas deposited within the pycnocline than in either sapropels or other sediments (see Table 2.6). While this might be an expression of higher dissolved uranium concentrations in this vicinity (as in the Black Sea) and a number of geochemical parameters that are unique to the pycnocline, Baturin (1968) considered that this observation was invalid and that there was a straightforward correlation between sediment organic content (and hence textural paratemers) and uranium. Krejci-Graf (1964, 1972) believes that uranium is concentrated in the transitional stages between sapropels and gyttjas which would be in agreement with Manheim's findings in the Baltic. Under oxidising conditions uranium is relatively mobile and hence in a sequence of alternating oxidised and reduced sediments it will tend to migrate and become concentrated (immobilised) at the contact between the two (Baturin, 1973). This may in part explain Krejci-Graf's conviction that uranium concentrates near the zero potential and "straddles the limit to anaerobic conditions" (Krejci-Graf, 1972). Baturin (1973) notes that under very reducing (anoxic) conditions uranium is intensively extracted from the sea water (on the sea floor) and that this is inhibited in the presence of a surface oxidised layer (i.e. where the RPD is below the sediment surface). This process will, therefore, be favoured only in anaerobic gyttjas and sapropels. In phosphate-rich sediments uranium is effectively scavenged during the growth of fluroapatite and the strong association between phosphorite and uranium can often obscure correlations between uranium and organic matter (Veeh et al. 1974; Calvert, 1976). In very strongly stratified basins where no syngenetic mechanism is operating, low supply of dissolved uranium to the bottom water (due to the inhibition of mixing by the pycnocline) may lead to rather low uranium contents determined by the level of enrichment in particulate phases entering the basin (Weber & Sackett, 1981).

TABLE 2.6 Uranium in recent organic-rich sediments

Location	Sediment	Uranium content	Reference
Baltic Sea	nearshore (organic-poor)	~5 ppm	Manheim, 1961
"	gyttjas within pycnocline	18 - 32 ppm	"
"	sapropel	6 - 7 ppm	"
"	sands and sandy silts	1 - 4 ppm	Baturin, 1968
"	clays and silty clays	4 - 16 ppm	"
Black Sea	coccolithic sapropel	15 - 60 ppm	Degens et al. 1977
Norwegian fjords	sapropel	13 - 60 ppm	Strøm, 1948
Pettaquamscutt River	sapropel	7 - 30 ppm	Mo et al. 1973
Saanich Inlet	sapropel	2 - 9 ppm	Kolodny & Kaplan, 1970
Orca Basin	sapropel	2 ppm	Weber et al. 1981
Cariaco Trench	sapropel	<30 ppm	Dorta & Rona, 1971
S.W. African margin	gyttjas	79 - 158 ppm	Veeh et al. 1971
Californian margin	gyttjas	11 - 28 ppm	Veeh, 1967
Peruvian margin	gyttjas	9 - 20 ppm	"
Average argillaceous sediment		2 - 4 ppm	Baturin, 1973

SOME ASPECTS OF BLACK SHALE SEQUENCES

The temporal and sequential evolution of black shale basins can be thought of in terms of three phases of development: the initial stagnation phase, a mature 'sapropel' phase, and a termination or breakdown phase. This sequence reflects the climatically controlled waxing and waning of watermass stability, or in oceanic sedimentation, variations in the intensity and nature of the thermohaline circulation. The transition to the 'black shale mode' can be very rapid in which case the initial stagnation phase may hardly be represented, sapropels lying abruptly above normal sediments (e.g. Quaternary sequence of the Black Sea). In most cases, however, the transition is gradual and suggests an overall increasing frequency, intensity and/or duration of periodic episodes of oxygen deficiency. Rapid stagnation and virtually immediate sapropel deposition (in an on/off manner) is most likely in barred basins (and times of high amplitude climatic cyclicity), while in epeiric basins gyttjas may remain dominant and sapropels may occur only locally or not at all. Most black shales occur as relatively thin interbeds (generally <1m thick) within cyclic redox sequences with estimated periodicities usually in the range of 20-50,000 years. Such cyclicity is clearly climatic but is apparently only expressed in preserved redox couplets (of alternating organic-rich laminated and organic-poor bioturbated beds) when the longer term climatic trends influenced by sea level variations, or unusual palaeogeographic settings, result in the global ocean (or parts of it) being in a generally more delicate balance by lowering the initial bottom water oxity. Thick and truly uniform black shale sequences are a rarity and appear to only occur in unusual palaeogeographic situations and rapidly subsiding 'deep' water depocentres with sufficient environmental inertia to defeat the 'oxidising' phase of the 20-50,000 year climate cycles.

The exact expression of the mature or sapropel phase of the black shale basin is determined by the dynamic relationships of the pycnocline, compensation depth and hence the general water depth (see earlier and Chapter 5). Perhaps more interesting is the nature and expression of the event(s) which terminate black shale sedimentation. In black shale sequences the sedimentological expression of the termination phase will be determined by a number of factors including:-

(a) the rate and magnitude of the environmental change

(b) the stability of the basin before and after the termination event

(c) any changes in clastic supply or the dynamics of redeposition

(d) the mean oxygen status of the new environmental regime

(e) any changes in the sediment accumulation rate (note that differential compaction may occur)

and (f) the dynamics of benthic re-colonisation.

There is always a degree of assymetry between the basal and upper margins of black shale units purely because the first aerobic event leaves a much more tangible record than the first oxygen deficient event (i.e. a bioturbated horizon compared with one or a few organic-rich laminae). This bias means that aerobic conditions will tend to be over-estimated since the thickness of aerobic facies expands at the expense of dysaerobic to anoxic sediments. Although the changes in depositional conditions may have been symmetrical in time the corresponding sedimentary record is frequently not. Sediments of the "dysaerobic facies" probably often represent a temporal and physical mixture of other facies, recording the partial and incomplete destruction and incorporation of anaerobic and anoxic sediment layers back into the aerobic facies, rather than sustained, uniform, strictly dysaerobic conditions (1.0-0.1 ml/l dissolved oxygen). The preservation of anaerobic-anoxic sediment layers will depend partly on their thickness and their toxicity to infauna, but also on the accumulation rate of the overlying sediment, the character of the infauna, the level of oxygenation and the time taken for benthic recolonisation (a function of the size of the basin, the size of the defaunated area, and the dispersal patterns of the benthos).

The termination of thick uniform black shale sequences, or of any black shales deposited in shallow shelf settings, is particularly interesting because of the greater changes in the environmental regime which are implied (in the former case an event sufficiently extreme to overcome the high environmental inertia which had maintained the black shale basin, and in the latter the highly significant change from permanently to seasonally stratified watermass conditions). The scale of these environmental changes implies major episodes of re-equilibration in the sedimentary regime which are frequently represented by current scouring, omission surfaces and/or reworking, and condensation as well as marked changes in lithofacies. Such features, which reflect the importance of watermass stability in the dynamics of sedimentation, may be detectable by seismic stratigraphy but have little to do with eustatic cycles. What kind of event could initiate such a catastrophic change in the marine environment? The basin-wide synchroneity and abruptness of the events which terminate black shale sedimentation tend to suggest tectonically initiated changes in the hydrography or oceanography, e.g.

(1) The establishment of deep-water connections with normal oxygenated basins by the tectonic breaching of sills.

(2) Growth of the basin (e.g. an opening ocean) to a size where thermohaline circulation becomes possible (q.v. Andrews, 1977).

(3) An increase in cross-shelf and/or tidal mixing due to the tectonic foundering of tide and current-damping shallow, ocean-margin carbonate platforms, or increases in the ratio of oceanic seaboard to shelf sea area.

(4) Cessation of subsidence and infilling of bathymetric 'stagnant' depressions resulting in shallowing and decreased watermass stability.

(5) Tectonic events resulting in shallow water connections between basins of differing watermass (salinity/temperature) characteristics, leading to watermass exchange and density-driven 'flushing' of deeper parts of one (or both) basins.

It is also possible that a 'castrophic' overturn event could occur for purely climatic reasons if along a climatic gradient of decreasing temperature or precipitation there occurs a critical point at which the stability of the watermass suddenly and irrevocably breaks down (particularly if this is self-catalysing). With regard to tectonically initiated changes (particularly 1 and 5 above) it should be noted that oceanographically significant connections between basins can be very limited in size in relation to their importance (e.g. Straits of Gibraltar) and may be unresolvable by palaeogeographic techniques.

The lithofacies response of black shale termiation is frequently not just a return to normal 'background' oxic sediments. The dispersal of large volume of oxygen deficient watermass must have profound geochemical and biological consequences, particularly in relation to 'biomineral' substances such as carbonates, phosphates and silica, and redox sensitive elements such as iron. Some of these consequences are discussed below.

Degens and Stoffers (1976) have noted that the lowering of the pycnocline in a stratified anoxic system may lead to massive precipitation of limestones. This process is an extremely rapid one in geological terms and it seems unlikely that the total overturn of an anoxic bottom water layer could occur quickly enough to precipitate all the dissolved carbonate inorganically. A more likely process is that the high nutrient supply from a dispersing anoxic watermass leads to higher productivity, which is reflected in foraminiferid and coccolith-rich limestones or calcareous clays. The aerobic conditions and, in ocean basins, the favourable shift in the calcite compensation depth resulting from the higher productivity, will also favour greater carbonate preservation. The supply of dissolved carbonate (as well as nutrients) may particularly favour the calcareous plankton and the reappearance of in situ benthos (and increased redeposition?) will also favour a transition form black shales to calcareous clays or limestones. The overturn of large, oceanic, carbonate reservoirs may also mean that black shale termination is also associated with a carbon isotope signal (Scholle & Arthur, 1980).

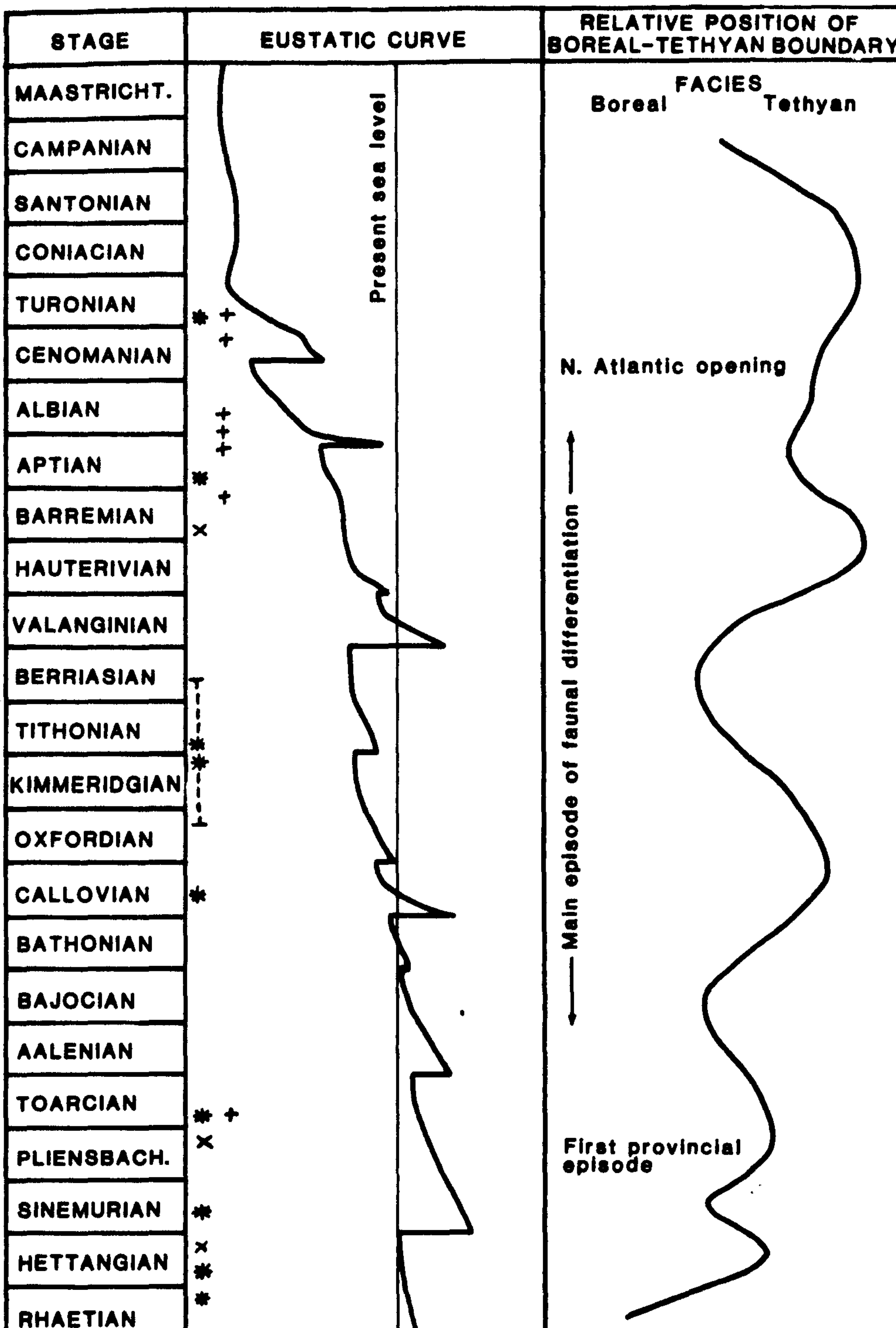
Increased productivity due to enhanced nutrient supply from overturning oxygen deficient watermasses may also promote greater formation of phosphorite deposits (see earlier). As noted previously black shales/ stable watermass stratification and high productivity are mutually exclusive except in the peri continental black shale facies and it is not, therefore, surprising that Arthur and Jenkyns (1980) found a lack of synchroneity between global 'oceanic anoxic events' and the major phases of phosphorite deposition. Periods of overturn and dispersal of oxygen depleted watermasses must, however, represent in areal terms at least, the prime conditions for phosphorite formation. The phosphorites may form mainly during the period of low sedimentation and condensing associated with the re-equilibration of the sedimentary regime following the initiation of the overturn, and then become subsequently reworked. Phosphatic horizons above black shales in epeiric sequences are frequently followed by sandy, glauconitic, fully oxic sediments. The grains of glauconite are probably derived from the winnowing of argillaceous sediments, wherein glauconite is believed to form in reducing micro-environments (Calvert, 1976); the reworking and progradation of coarser sediments being a consequence of the change in watermass stability (e.g. the cessation of the 'false bottom effect' of stable stratification - see earlier). Primary chamositic ooliths, as in Jurassic ironstones of north west Europe, are probably also derived in a similar fashion from the winnowing of organic rich muds after the destabilisation of the shelf watermass. Hallam and Bradshaw (1979) note that most British Jurassic ironstones are associated with minor or major phases of reworking which they attribute to 'regression' but are more likely episodes of watermass instability in the Jurassic epeiric seas. The decline of stratification would have allowed clastic material previously trapped in the nearshore zone to have spread outwards into the basin; only the finer clastic material would have reached the basin centre (after an uncertain time lag) and hence relatively pure ironstones would have been deposited centrally, and sandier impure ironstones towards the margins. The 'clastic traps' of traditional ironstone models would have been built into the mechanism.

FACIES ZONATION IN EPEIRIC SEAS AND ORIGINS OF THE BOREAL AND TETHYAN FAUNAL REALMS

The regional dynamics of watermass stability can be used to explain the facies belts observed in the north west European Jurassic epeiric sea and first commented upon by Hallam (1969, 1971). Hallam described three facies belts; an internal 'terrigenous clastic facies association', a central 'intermediate facies association' and an external 'calcareous facies association', whose origins he interpreted mainly in terms of a clastic supply-water turbidity gradient. This explanation is essentially empirical, and undeniable as far as it goes, but is inadequate to explain all the features of the facies patterns by itself. The variable lithofacies character of the intermediate facies assocation in particular can only be adequately explained by invoking variations in watermass stability.

For most of the Jurassic and Cretaceous the worldwas divided into two faunal realms; the northern part of the Northern Hemisphere containing the Boreal Realm and the rest of the world comprising the Tethyan Realm (Hallam, 1969). Despite the fact that Hallam (1969, 1971) has pointed out the quite consistent relationship between sedimentary facies and these faunal realms, other workers do not appear to have given this point sufficient attention, although the association between carbonate facies and the Tethyan Realm has often been cited as evidence of climatic control on Mesozoic faunal provinciality.

Fig. 2.15 Jurassic-Cretaceous sea level curve (from Vail et al. 1977) and fluctuations of the Tethyan/Boreal realm facies boundary (after Khudoley, 1974).



✱ Principal developments of organic-rich mudrocks on the European shelf.

× Lesser or more localised developments.

I Full range of Late Jurassic 'event' in North Sea basin.

+ Organic-rich intercalations in the pelagic facies of the N.W. Tethys.

Not only does the boundary between the realms generally reflect the position of the change-over from the shallow calcareous facies association to the intermediate facies association, the two boundaries also show synchronous shifts through time (see Hallam, 1969, 1971; Khudoley, 1974- and Figure 2.15). Another factor whose importance has undoubtedly been underestimated is the great extent of the Boreal Sea in relation to the length of its oceanic margin; prior to the opening of the North Atlantic and the Cenomanian transgression, the Boreal realm was one of the widest epeiric seas in the Mesozoic.

Various mechanisms have been proposed for the origins of the faunal differentiation which resulted in the Boreal and Tethyan realms. Hallam (1969, 1971) once believed that slightly lowered salinity was responsible for the distinction of the Boreal Realm but has since abandoned this hypothesis (Hallam, 1975, 1973), which never found favour with the majority of workers (e.g. Reid, 1973; Stevens, 1973; Gordon, 1974). The general consensus of opinion is that the Boreal-Tethyan differentiation is due to a latitudinal climatic control, modified in certain areas by circulation pattern and physical barriers (e.g. Gordon, 1974, 1975, 1976; Stevens, 1973). Arguments given in favour of this hypothesis include:

(i) Temperature is the dominant control on marine zoogeography at the present day.

(ii) Longitudinal faunal similarities are consistently higher than latitudinal faunal similarities (see Gordon, 1976).

(iii) The positive correlation between the Tethyan realm and carbonate facies.

(iv) The recognition of apparent (but poorly developed) anti-Boreal faunas at various times in the Mesozoic (see Gordon, 1975).

(v) The supposed lack of a viable alternative hypothesis.

The importance of temperature in marine zoogeography cannot be denied, however, geological evidence strongly suggests that the Mesozoic climate was very equable with rather low latitudinal temperature gradients (Hallam, 1969; Frakes, 1979). Such a situation is hardly conducive for the formation of faunal realms based on the stenographic distribution of stenothermal organisms (see Menzies et al. 1973). It should be noted that the Mesozoic Boreal Realm is only 'Boreal' in the geographic sense and not in the climatic or modern zoogeographic meaning of the term. The Tethyan-Boreal boundary is generally believed to mark the transition between tropical and warm temperate conditions. The absence or weak development of true anti-Boreal faunas in the Mesozoic would suggest that the significance of temperature in faunal provinciality was slight though not altogether absent. If temperature is not the answer what other latitudinal factors may be important? Valentine (1976) has shown that trophic resource stability is a potent latitudinal control on faunal diversity resulting from the extent of seasonality in primary production (see also Reid, 1973). However, this factor is an independent variable which will have shown little if any variation through time. With the rejection of the classic climatic control hypothesis it is perhaps in order to examine some of the basic aspects of Boreal-Tethyan provinciality.

The distinction between the two faunal realms is based fundamentally on the distribution of certain ammonites (Arkell, 1956; Sachs et al. 1973; Khudoley, 1974) and belemnites (e.g. Stevens, 1973) i.e. nektonic organisms. It has also been pointed out that certain groups of benthic organisms apparently act as provincial indicators as well but this is misleading since in reality they indicate facies which correlate with the faunal realms. In this context it is worth repeating a comment by Laubscher and Bernoulli (1977) - "It seems that the terms 'Tethyan', 'Alpine' or 'Mediterranean' facies in the Jurassic often merely imply pelagic deposits as they are found along the opening ocean".

In addition the selectivity of some palaeontologists has propagated certain myths about the provinciality of benthic faunas; larger foraminifera are often considered indicative of the Tethyan province, but if one looks at the adjacent basinal facies as well as the carbonate platform environments,one can find nodosariid faunas considered typical of the Boreal Realm (pers. comm. J. Exton). Alas much palaeontological information is often considered outside of its facies context.

The first real differentiation of the Boreal and Tethyan realms in the Mesozoic occurred in late-Aalenian-Bajocian after a brief provincial episode in the Pliensbachian (Gordon, 1976; Sachs et al. 1973; Stevens, 1973). From this time onward two distinct realms existed whose boundaries showed minor fluctations (see Figure 2.16) until the Albian, after which the faunal distinction decreased rapidly (Gordon, 1976). During this interval,Hallam (1978) has noted that regressions enhanced faunal differentiation (leading to the recognition of separate provinces within each of the realms) and that transgression led to faunal expansions. Geographic separation was evidently an effective process in intra-realm differentiation, but what of the realms themselves?

One of the most detailed analyses of the Boreal-Tethyan problem is that by Fursich and Sykes (1977) for the Oxfordian. The conclusions of this paper include:-

(a) Faunal diversity is clearly influenced by facies.

(b) Overall benthic diversity decreased northwards into the Boreal Sea.

(c) The absence of certain Tethyan groups from northern Boreal areas was probably due to limited dispersal from Tethys.

(d) Physical and physiological barriers probably operated between the various Boreal basins.

Fursich and Sykes interpreted the low diversity of the Boreal sea as a reflection of the low habitat diversity characterising what was essentially a clay basin, but note that benthic diversity was in fact relatively high in the Boreal Oxfordian clays. They located the Tethyan-Boreal boundary in the general region of the Hercynian basement islands of the London-Brabant Massif, the Rhenish, Bohemian, and Sudetic Mountains and Pompeckji swell (the so-called European Archipelago), which corresponds with the terrigenous clastic - intermediate facies association boundary. In this region complex facies variations and physio-geography were believed to be responsible for a rather complex pattern of benthic diversity because they produced conditions of low environmental stability (see earlier).

I consider the Boreal-Tethyan division to be what is little more than a contrast between the oceanic and neritic environments. At the present day (and virtually since the opening of the North Atlantic) shelf seas have generally been of the rather narrow marginal type; the Boreal sea was an epeiric sea of much larger dimensions (large by even epeiric standards), wherein the contrast between the oceanic and neritic environment could attain its maximum expression. Passing from the ocean margin to the internal epeiric sea (from the calcareous to terrigenous clastic facies association) we pass from a normal stratified tropical ocean across a shallow, turbulent margin to a broad carbonate platform (containing rather small land areas) into somewhat deeper epeiric seas. In the epeiric sea water depths, bottom water physico-chemistry and substrate conditions were not as suitable to heavily calcified benthic organisms as those conditions which occurred further south and sediments were dominantly argillaceous. This internal epeiric sea varied in character; when it was shallow (presumably due to low subsidence rates) it corresponded only to the mixed layer and limestones were widespread, reflecting an extension of the margin carbonate platform environment. When subsidence was higher (and when perhaps, denudation of internal land areas was greater) the epeiric seas were generally somewhat deeper clay basins which may have been thoroughly mixed, seasonally stratified or occasionally persistently stratified.

Offshore transport of coarse clastics would have been restricted by storm wave base and possibly by shallow water carbonate facies fringing the land areas. During episodes of transgressive deepening (sometimes associated with tectonic subsidence as well) watermass stratification would have intensified, resulting in meromixis, and dysaerobic to anaerobic kerogenous clays and shales become widespread.

Essentially we have two 'deep' basins divided by a shallow platform area, one showing permanent oceanic stratification, the other varying from fully mixed to persistently stratified. The shallow area between the two will undoubtedly represent a region of low environmental stability and high stress. This high stress barrier would, with time, result in allopatric speciation and faunal differentiation. One should note, however, that many benthic organisms managed to cross this barrier and that others did not (or more accurately did not, or did but could not survive) merely because their environmental preferences were not met on the other side. Furthermore, it seems likely that the circulation patterns in the epeiric sea proper would have been largely decoupled from the oceanic-marginal shelf circulation on the Tethyan platform by hydrodynamic frontal systems, etc. (particularly in the region of the 'European Archipelago'). We must anticipate that the patterns of primary productivity would have shown considerable differences between the oceanic and epeiric seas and this would have resulted in behaviourial and physiological differences that would have promoted speciation and differentiation in the nekton (see also Reid, 1973). It should be noted that since the Boreal-Tethyan boundary is determined by watermass conditions, relatively uniform sediments may occur in the transition zone which show alternations of Tethyan and Boreal nekton but a constant benthic fauna. In the absence of an intermediate high stress barrier and shallow water zone which was the site of circulation decoupling, it is apparent that the Boreal mixed layer environment was not that different from the Tethyan mixed layer (apart from absolute depth which is irrelevant) and exchange of nektonic faunas would be possible. In the Toarcian, a time of platform collapse on the Tethyan margin, Tethyan ammonites penetrated into Yorkshire testifying to the similarity of the mixed layer environments.

The differentiation of the Boreal and Tethyan realms is thus believed to be a watermass effect largely resulting from the Mesozoic (pre-North Atlantic) palaeogeography. Faunas expanded into the newly created Boreal epeiric sea created during the Early Jurassic and became genetically isolated by stress barriers and developed separately into epeiric and oceanic - marginal sea faunas. Latitudinal temperature gradients were relatively unimportant but may have intensified later in the Cretaceous. Geographic isolation by physical barriers was probably important at provincial level. With the opening of the North Atlantic the main cause for the faunal realms was removed by a fall in sea level, improved circulation and increased ratios of ocean margin length to shelf sea area.

CHAPTER THREE

A review of palynofacies studies in the palaeoenvironmental analysis of marine sediments.

"The data on the distribution of pollen and other microfossil groups can also be utilised in studies of depositional environment, and therefore are valuable in facies recognition and in the palaeogeographic reconstruction of ancient sedimentary basins. The recognition of palynological provinces makes it possible to distinguish rather sharply the characteristics of each of several major types of depositional environment"

Jan Muller (1959, p.29)

INTRODUCTION

The term 'palynofacies' was first introduced into the literature by Combaz (1964) who defined the palynofacies of a rock as including all the microscopic organic constituents present after maceration, concentration and mounting using normal palynological preparation procedures. It is not precisely synonymous with the total kerogen content of the rock as this is generally defined as that part of the particulate sedimentary organic matter which is insoluble in ordinary organic solvents (for a discussion of the historical development of the term kerogen see Durand, 1980). In the broadest sense palynofacies studies include both the geochemical characterisation of particulate organic matter and the palynological classification and quantification of palynomorphs and phytoclasts etc. combining the expertise of palynologists, organic geochemists and sedimentologists. Organic petrography is a much longer established, closely allied field that historically involves the examination of coals in reflected light, rather than the study of dispersed organic matter in transmitted light which is the norm in palynofacies investigations. The division between the two is artificial but has resulted in differing terminologies which have still not been fully reconciled (see Table 3.1).

In this chapter the 'palynofacies literature' has been reviewed to provide the basis for the interpretation of the data presented in chapters 4, 5, 6 and 7. This was found to be essential because no comprehensive review exists in the literature and previous reviews have tended to show too strong a geochemical or palynological bias.

GENERAL COMMENTS

The objectives of palynofacies investigations are several:-

1) To provide information on the depositional environment (see also Cross et al. 1966, p.470).

(a) to determine open marine sediments from paralic and from non-marine deposits

(b) to estimate the relative proximity to shore or sediment source in marine sediments

(c) to provide an insight into the watermass conditions occurring at the time of deposition, including the productivity of the environment and nature of the bottom conditions

(d) to locate the sources of sediment input and, indirectly palaeocurrent directions

(e) to obtain palaeo-botanical information relevant to palaeo-climates etc .

2) To provide information on the diagenetic environment

(a) Eh conditions

(b) Depth of burial.

3) To provide information on the source rock potential of sediments

(a) presence of sapropelic (oil-prone) or humic (gas-prone) organic materials and their relative proportions

(b) the degree of maturity of the kerogen.

The first of these three objectives represents the classical field of palynofacies while the latter two are based on kerogen studies which are traditionally the province of organic geochemists. In order to accomplish the first objective it is necessary to appreciate the nature of particulate organic materials, their behaviour in various transporting media (air and water) and the broad controls on their distribution. In this respect the key to palynofacies investigations has been summarised by Cross et al. (1966, p.480) in the following statement: "The distribution and accumulation of pollen and spores together with other degraded or dissociated plant detritus and a whole variety of plankton is remarkably similar to some patterns of sedimentation of terrigenous clastics and biogenous marine sediment. The fact is inescapable that these organic entities behave as sedimentary particles, particularly in a water medium". Since organic matter is concentrated in clay and silt grade sediments (see Chapter Two) it follows that the majority of naturally occurring organic particles are hydrodynamically equivalent to this size of material and this has certainly been proved for sporomorphs (Stanley, 1969; Cross et al. 1966). Many organic particles including spores and pollen are, however, non-Stokesian in their hydrodynamic behaviour (e.g. Brush and Brush Jr., 1972; Davis & Brubaker, 1973), which tends to make more rigid interpretations invalid. This is due to a variety of reasons including adsorption of water, swelling and alteration of shape, changes in chemical composition with time, and aggregation. Adhesion to, and inclusion within, amorphous aggregate particles (see later) will also clearly alter the sinking properties of many small particles (see Riley, 1970; Wiebe & Pomeroy, 1972; Johnson, 1974). In addition large amounts of particles are probably sedimented via the faecal pellets of planktic and nektic organisms and so detailed estimates of settling behaviour of different organic particles may be highly misleading.

The distribution of organic particulate material (palynomorphs and phytoclasts) in sediments is determined by a complex combination of factors (Muller, 1959; Bottema & Van Straaten, 1966; Williams & Sargeant, 1967; Heusser & Balsam, 1977) including:-

(i) The location in which the various particles are produced in relation to their final place of rest.

(ii) The rate at which the various particles are produced.

(iii) The relative rates at which the particles are supplied to the depositional environment.

(iv) The extent and efficiency with which the various particles are transported within the depositional environment.

(v) The location, number and importance of river inputs into the basin (and efficiency of the drainage system in adjacent land areas).

(vi) The intensity, duration and direction of bottom and surface currents within the depositional basin (including overall circulation pattern e.g. estuarine or anti-estuarine etc).

(vii) The presence or absence and distribution of watermass stratification in the depositional basin.

(viii) The distribution of grain sizes in bottom sediments of the depositional environment.

(ix) The preservation potential of the various particle species.

(x) The bottom water oxygenation of the depositional environment.

(xi) The accumulation rate of sediments in the depositional basin.

(xii) The pattern of mineralic sedimentation compared to that of the organic particles.

There are a great number of secondary factors and allegories that may be drawn from this list, e.g. differences in transport time (and hence likelihood of preservation) are implicit in (i) and climatic factors are clearly of major importance in (ii), (v), (vi) and (vii). The overall intention here is to indicate the complexity of the problem and the inextricable way in which palynofacies is linked to the environment. Despite this complexity there are still a good number of useful generalisations that can be made. In particular, since spores and pollen are produced onshore and transported offshore (along with phytoclastic material), and plankton is produced within the depositional basin, a gradient will exist between the two which has a roughly onshore-offshore orientation. All but major anomalies in the production and supply of terrigenous and marine particulate organic materials will be evened out by the redistributing effects of marine currents. Thus Scott and Kidson (1977, p.176) note that "spores and pollen tend to concentrate in the nearshore sediments where phytoplankton numbers are typically lowest and they become decreasingly important in a more marine direction as marine phytoplankton become more diverse".

DISTRIBUTION OF ALLOCHTHONOUS ORGANIC MATTER

(a) Macrophyte debris (wood etc)

According to Degens and Mopper (1976,p.64) the contribution of terrestrial organic materials is apparently small compared to the marine autochthonous component regardless of whether the sediments are deposited during transgressive or regressive phases. However, woody debris is usually abundant in the vicinity of river mouths and is clearly influenced by palaeo-geography and climate. Pocklington and Leonard (1979) found that terrigenous organic matter was abundant in the upper part of the St. Lawrence Estuary but decreased rapidly in a seaward direction. Reid (1972) also reports large amounts of plant debris in fine-grained estuarine sediments from around the U.K. As a general rule it would appear that high contents of woody material tend to be restricted to the immediate vicinity of fluvial inputs (except where the land areas are poorly vegetated) but smaller amounts are probably widely distributed (e.g. Cross et al. 1966). Although the normal textural controls apply (i.e. woody debris is usually most abundant in clays and silts) coarse fragments (drift wood etc.) may be found in sandy sediments. Like any other sedimentary particles, phytoclasts may be carried across shallow shelves, channelled through submarine canyons and redeposited in deeper waters (Cross et al. 1966; Drake et al. 1978; Scott & Birdsall, 1978; Tissot & Pelet, 1981) and are apparently common in recent turbiditic sediments (Heezen et al. 1955; Rupke & Stanley, 1974). Menzies et al. (1973) consider this process an important source of food for the organisms in the abyssal zone (see also Wolff, 1979).

Before discussing the pattern of distribution of macrophyte particle organic matter that has been observed in some ancient sediment sequences, it is necessary to comment on the relationship between modern organic debris and the humic kerogens observed in ancient rocks. The most resistant plant tissues are those that contain lignin which is deposited in the cell walls of sclerenchyma (mechanical support tissue) and the tracheids of xylem tissue. Lignin normally comprises about 25-30% of the wood of trees and tends to be more abundant in gymnosperms than angiosperms (Correia, 1971). Collenchyma and parenchyma probably have poor preservation potential and the bulk of phytoclasts which survive degradation within the depositional environment are probably composed of tracheids and sclerenchyma. Masran and Pocock (1980) note that "lower vascular plants and bryophytes lack the strengthening structures of woody plants" and "after both mechanical and chemical breakdown these plants yield an abundance of thin, structureless tissue". The "cell walls and stomatal structures are gradually obliterated leaving sheets of transulcent amorphous material" but "vascular and conducting tissues are frequently preserved". Aquatic macrophytes require less support than land plants and are probably less lignified and hence more easily degraded, but marine forms are more resistant than freshwater equivalents (Godshalk & Wetzel, 1977, 1978; see also Josselyn & Mathieson, 1980). Masran and Pocock (op. cit.) state that "seaweeds break up into cell masses, individual cells and amorphous material" and that "under anaerobic conditions green-yellow amorphous material is formed .... whereas marine grasses are converted into green-yellow amorphous material and orange sheet-like masses". The less lignified nature of marine macrophytes probably explains why, although they are such an ecologically significant component of modern marine ecosystems, they have never been conclusively identified in ancient sediments. Lignin is only very slowly decomposed by the attack of specific lignolytic fungi and bacteria (Stach et al. 1975; Huc, 1980) although the rate does depend on particle sizes (surface area/volume relationships). Mann (1972) has shown that many grazing benthic organisms feed on the bacterial microfloras associated with macrophyte detritus rather than this material itself, but their activity breaks the particles down into smaller sizes and hence makes them more susceptible to decay. Although preservation of ligno-cellulosic materials is relatively good under aerobic conditions it is even better under anaerobic conditions which exclude lignolytic fungi (Huc, 1980) as well as benthos. The resistance of ligno-cellulosic phytoclasts allows them to be recycled several times and to survive sporadic,intermittent, transport.

During early diagenesis phytoclastic materials are converted into vitrinites by a two stage process (Stach et al. 1975). The first stage, humification, is effectively a slow biochemical oxidation process brought about by fungi and bacteria which converts the ligno-cellulosic material into humic acids and subsequently to humins. The second stage, gelification, leads to conversion of the humins into vitrinites. Vitrinites have two principal components: telinite which is typical, structured phytoclast material and collinite which is an amorphous secondary structureless cell filling (Stach et al. 1975; Correia & Peniguel, 1975).

Vitrinites show much the same pattern of distribution in ancient rocks as does woody debris in recent sediments but there are few published palynofacies case studies to prove the point. Work carried out by Robertson Research on Middle Jurassic deltaic sequences in Yorkshire and the North Sea has, however, documented the abundance of vitrinites in these sediments (Fisher, 1980; Denison & Fowler, 1980; Hancock and Fisher, 1981), and similar results have been reported by Batten (1973, 1974) and Parry et al. (1981). Habib (1979b) has made some brief comments on the distribution of woody materials in Cretaceous oceanic sediments which are clearly related to the nature and distribution of redeposition processes in space and time, and Weidmann (1967) and Fisher (1980) report significant vitrinite contents in turbiditic silts and muds as observed in recent equivalents.

(b) Cuticle

Cuticle is the cellular, waxy, thin epidermal tissue of higher plants of which much is probably derived from leaf debris. In his classic study of the Orinoco Delta Muller (1959) observed that fragments of cuticle were concentrated in belts opposite the large delta distributaries and that the size and abundance of these particles rapidly diminished offshore. Cross et al (1966) have observed the same pattern of distribution in the sediments of the Gulf of California and report that cuticular debris is hydrodynamically equivalent to sporomorphs, but because of its thin-walled and delicate nature is less capable of surviving long periods of transport. There is very little documentation of the palynological distribution of fossil cuticle debris in ancient sediments but it does appear to conform with modern trends. Batten (1973, 1974) reports that cuticles are quite common in the Wealden Beds of southern England where they are most abundant in clays and medium silts, and Parry et al. (1980) have shown that they are also common in Middle Jurassic deltaic sequences where they reach their peak relative abundance in pro-delta clays and interdistributary marginal marine embayments.

(c) Inertinite (fusinites)

These particles represent carbonised wood material and are of two genetic types (Stack et al. 1975):-

(i) Pyrofusinite formed by the imcomplete combustion of wood or peat during natural fires (see also Harris, 1958)

and (ii) Degradofusinite formed by subaerial, subaqueous or groundwater oxidation of woody materials.

Little is known about the distribution of inertinite in recent sediments but it is common, if not particularly abundant, in ancient rocks. Both Batten (1973) and Parry et al. (1980) report significant amounts of 'black wood' in deltaic sediments and Habib (1979a, 1979b) has shown that fine fusinised debris is abundant in Cretaceous oceanic sediments which were deposited furthest from active terrestial inputs (see his "micrinitic palynofacies" described later). Much of the wood content of sandy rocks is often inertinite or semifusinite and it is probable that part at least was formed by in situ post-depositional oxidation processes. Such processes are probably very important in deltaic sediments where they will be strongly controlled by the position of the water table.

(d) Pollen and spores (sporomorphs)

Stanley (1969) has reviewed the pattern of distribution of sporomorphs in marine sediments. He concluded that while the wind only transports most sporomorphs ≤100 km and rivers may transport them ≤2,500 km, most are probably derived from between 30-100 km of the coast. Field surveys in marine sediments strongly suggest that aeolian transport is of little quantitative significance (Muller, 1959; Cross et al. 1966; Heuser & Balsam, 1977). The pattern of distribution of most sporomorphs is directly related to fluvial inputs, with the highest concentrations occurring in the vicinity of deltas and estuaries (Muller, 1959; Cross et al. 1966; Darrell & Hart, 1970; Reid, 1972; Heusser & Balsam, 1977). Cross (1975) notes that "low concentrations occur along some shores and shelves due to dilution by terrigenous sediments, winnowing and distance from active delta. Although rate of deposition is greater in and in front of deltas and fans, land derived palynomorphs generally occur in even greater relative and absolute frequencies".

Heusser and Balsam (1977) report a bimodal sporomorph distribution in the sediments of the N.E. Pacific off of Oregon. They found that sporomorphs were abundant in nearshore sediments (particularly adjacent to rivers) but generally low on the shelf, increasing again on the continental slope and rise before susbsequently diminishing oceanward. This may reflect textural parameters, the likely preservational effect of the oxygen minimum on the outershelf-slope (see Gross et al. 1972) or the effect of stratification of the shelf watermass on the sedimentation of organic particles (as hypothesised by Kulm et al. 1975) or (more probably) a combination of the three. The effect of watermass stratification on sporomorph distribution has been demonstrated for lacustrine sediments by Davis and Brubaker (1973). These authors found that relatively buoyant pollen were transported across the pycnocline (metalimnion) and hence across the lake, while denser pollen soon settled out and were deposited in the bottom sediments. Although the major conclusions of the authors result from the geometry and size of the lake it is also clear that, given offshore transport in a stratified sea, appreciable sporomorph sorting is likely to result. At the present day bisaccate pine pollen are among the most buoyant forms and are often the most characteristic sporomorphs of distal sediments.

Brush and Brush (1972) have carried out some laboratory experiments on sporomorph transport in moving water and made the following observations:-

(i) The size, shape, density and fall velocity of pollen grains can vary with the length of time they are immersed in water (hence their non-Stokesian behaviour)

(ii) There is a selectivity for certain pollen types to remain in suspension but some percentage of all pollen types become included in the sediment (therefore palynofacies boundaries will be gradational rather than absolute)

(iii) Hydrodynamic effects are more important than supply in determining the concentration of pollen in sediments (this emphasises the need for sedimentological inputs in palynofacies investigations).

The observations above emphasise the need for caution in using palynofacies in any palaeobotanical sense and point out that palynofacies investigations must be conducted at the level of general trends rather than specific, localised details. Heusser and Balsam (1977) have noted that although the character of the vegetation is reflected in the relative frequency of the pollen, abundance is controlled by hydrodynamic factors and this supports (iii) above. Muller (1959) has described how the pollen of the local swamp vegetation is over-represented in the Orinoco Delta because of the limited hydrodynamic sorting, and how the diversity then subsequently increases offshore. Similarly, Darrell and Hart (1970) have noted that the lack of sorting in marsh and bay sediments of the Mississippi Delta leads to reduced diversity and more irregular distributions of palynomorphs. Thus the general rule is that near source the diversity will be low, then increase offshore for a while before decreasing again as all but the most bouyant morphotypes are left behind (see Fig. 3.1).

There are very few palynofacies orientated accounts of the distribution of pollen and spores in ancient sediments and it is often very difficult to draw any meaningful conclusions from biostratigraphical works. Hughes and Moody-Stuart (1967) have shown that fern spores tend to become smaller and less diverse in more offshore Wealden sediments, and along with many other authors (including Riley, 1974; Batten, 1973, 1974; Habib, 1979a, 1979b; Reneville & Raynaud, 1981) have noted that Mesozoic distal palynomorph assemblages are often dominated by bisaccates and Classopollis pollen. Although it is clear that in the bisaccates the air sacs are responsible for their increased bouyancy and transportation, in the case of Classopollis the explanation is a less obvious one which partly relates to its porous wall microstructure (M. Partington, pers. comm. 1981). Batten (1973, 1974) notes that most sporomorphs are equivalent to clay and silt grade clastics but that relatively heavy (thick walled and or strongly ornamented) fern spores were most abundant in medium to coarse silt and fine sandstones. Chaloner and Muir (1968) have interpreted changes in Carboniferous spore assemblages as reflecting changes in sea level and hence the proximity of the various floras to the site of final deposition. This early paper signals the recognition of the significance of differential transport as a factor which at least equals or outweighs palaeobotanical controls on sporomorph assemblages, but suffers from too great an emphasis on sea-level rather than more general 'proximality and distality'.

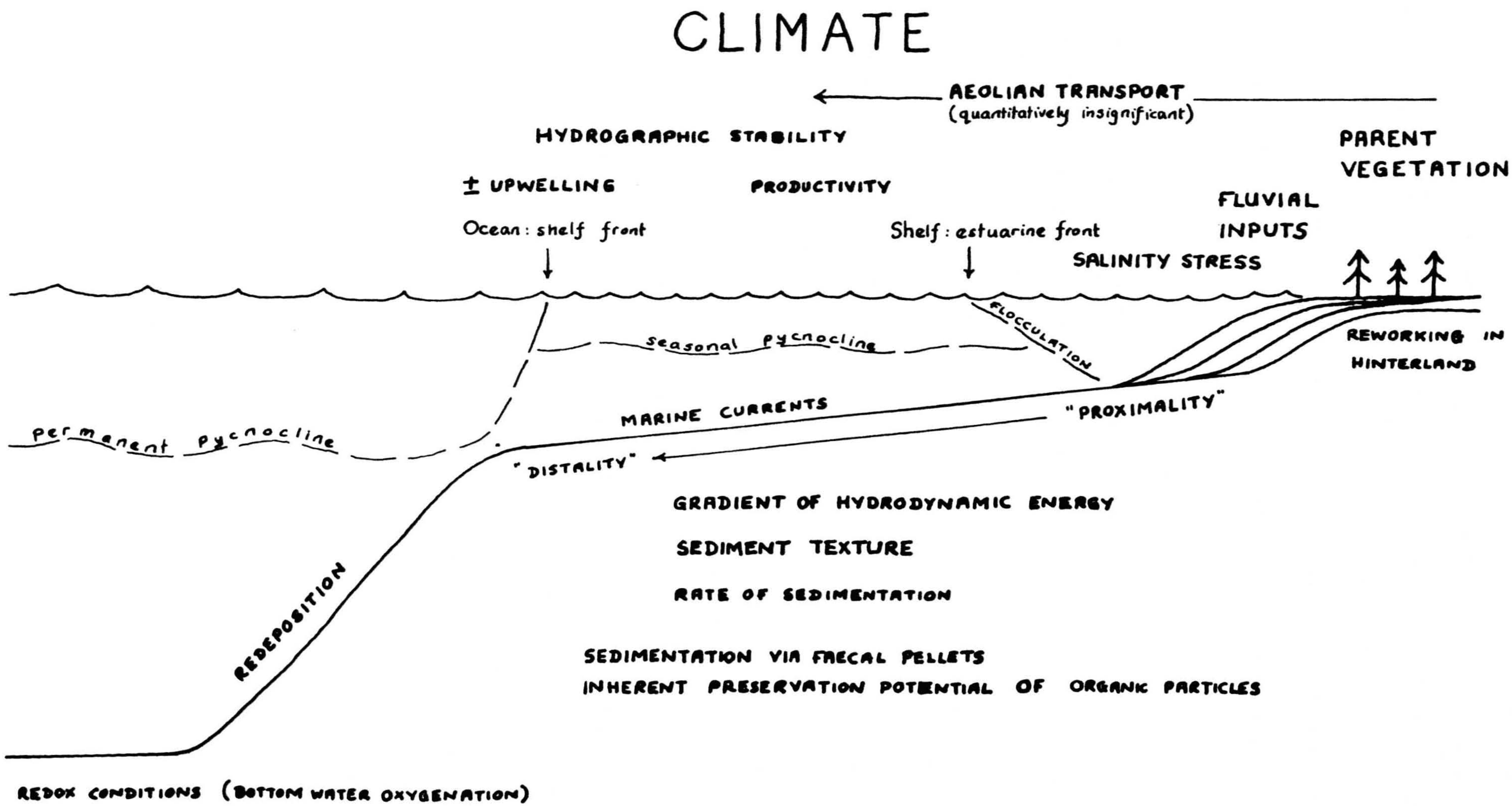


Fig. 3.1a Palynofacies summary diagram: principal controls on the palynofacies of marine sediments

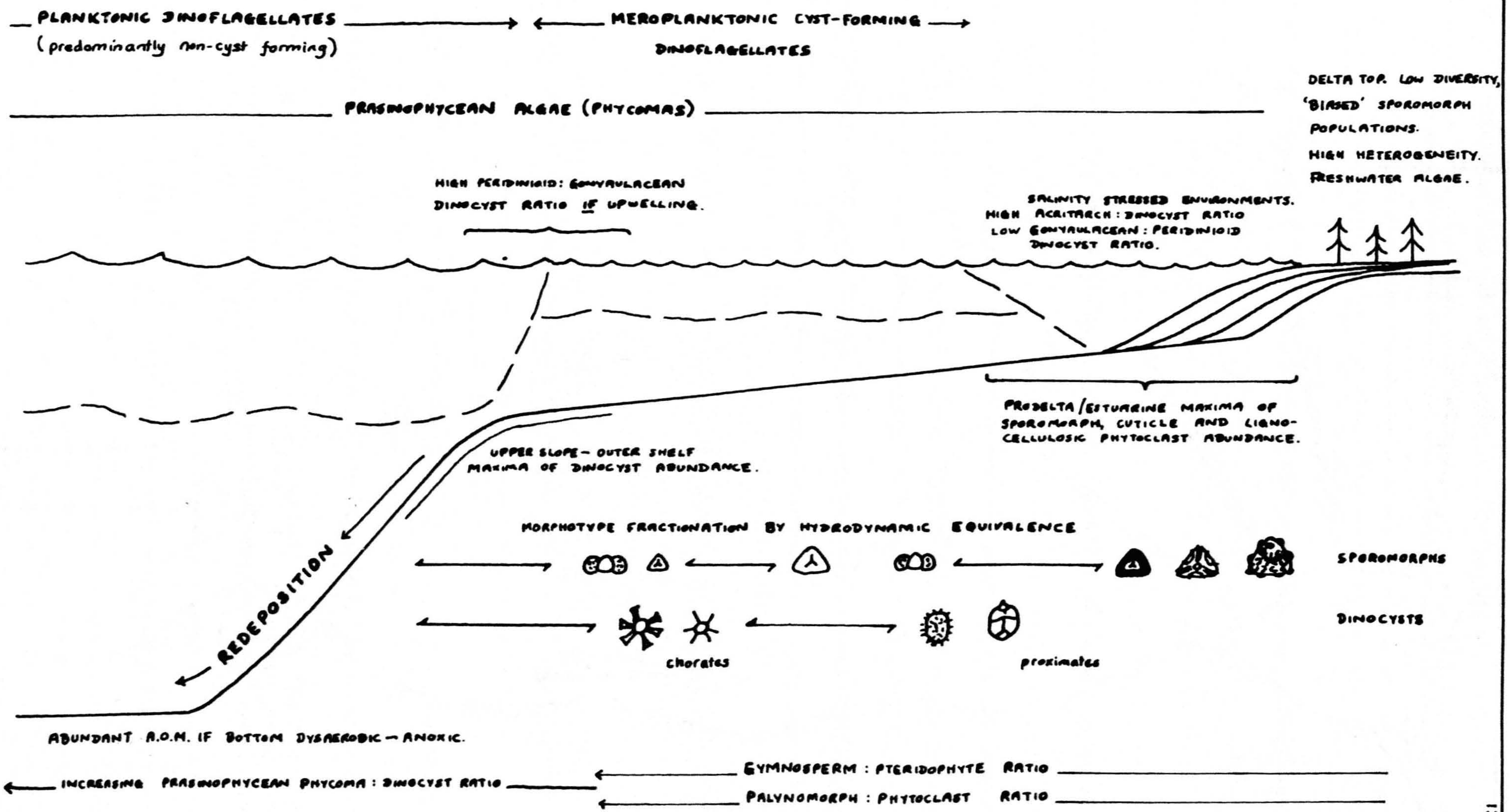


Fig. 3.1b Palynofacies trends: cross-section

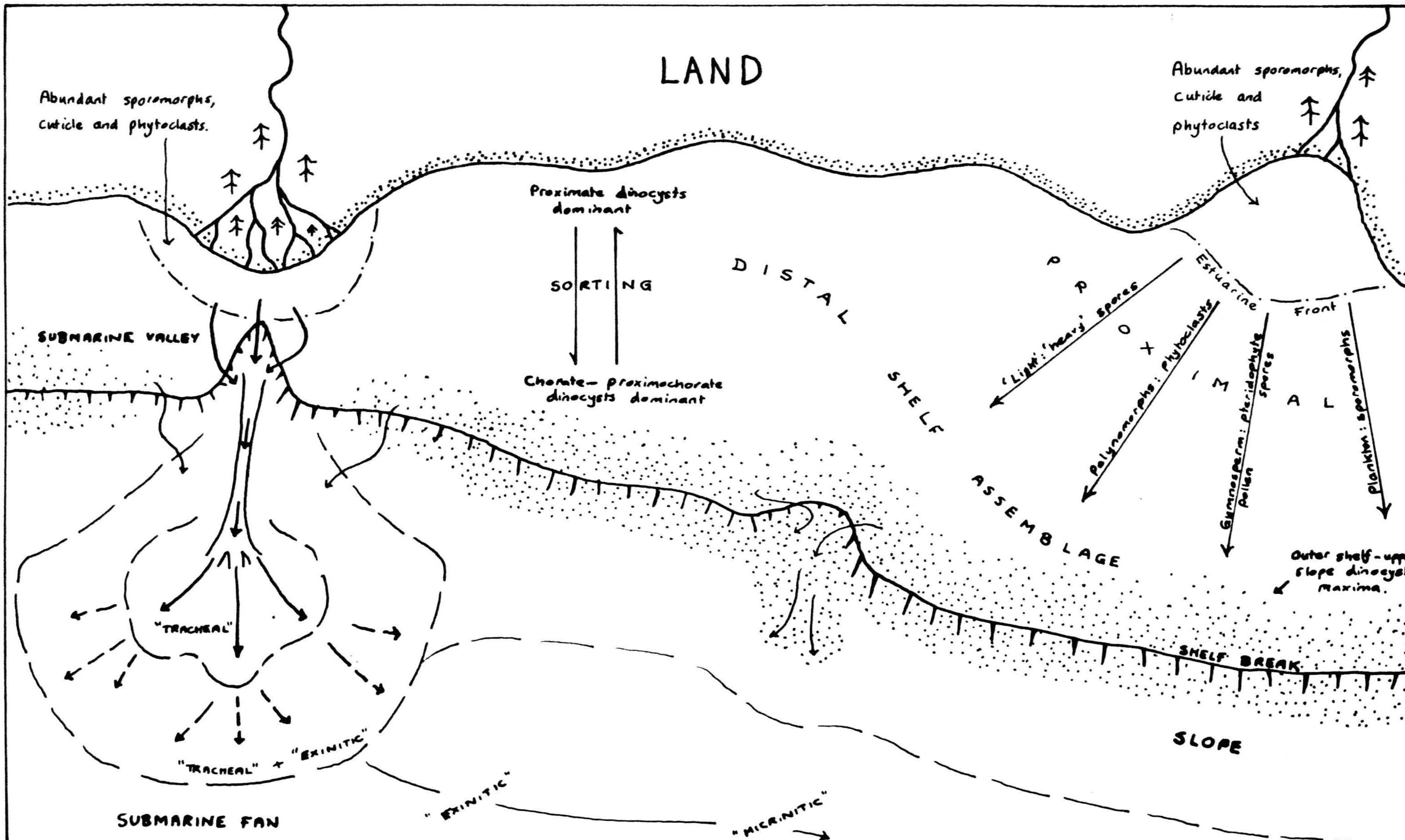


Fig. 3.1c Palynofacies trends: plan view. Terms in inverted commas from Habib, 1979 et seq.

THE DISTRIBUTION OF AUTOCHTHONOUS ORGANIC MATTER

(a) PLANKTON (DINOCYSTS, ACRITACHS AND PRASINOPHYCEAN PHYCOMAS)

The principal controls on marine primary productivity were reviewed in Chapter 2. The key factor was identified as water column stability since this determines the distribution and availability of nutrients within the watermass and the extent to which the phytoplankton can utilise the available light. Where the mixed layer extends to the bottom, nutrient recycling processes are efficient and good phytoplankton populations are developed (e.g. coastal waters) but where the water column is stratified a nutrient trap situation develops, the surface waters are oligotrophic and poor phytoplankton populations are present. Well mixed conditions tend to be characterised by diatoms and stratified conditions by dinoflagellates (Landry, 1977; Margalef, 1978) but the two states usually alternate seasonally. Eppley and Peterson (1979, p.679) note that "over geological time climatic changes that reduced vertical mixing must have greatly affected the planktonic production process" and that "both too little and too much mixing will reduce production". These fundamental relationships suggest that watermass controls and their effects on productivity and nature of planktic floras are likely to be recorded in the geological record.

(i) Dinoflagellate cysts (dinocysts)

All the dinoflagellate cysts encountered in ancient sediments represent the zygotic resting stage (hypnozygote) of the life cycle of meroplanktic dinoflagellates which is normally formed in the late summer and autumn as a response to decreasing light, temperature and stability or under conditions of crowding (as in red tides). Although it is clear that encystment occurs as a response to stress and that this occurs in unstable hydrographic regimes (Wall et al. 1977; Margalef et al. 1979), the relationships between particular motile phases and cysts are still rather obscure (see Wall, 1971; Tappan, 1980). It is unclear to what extent the cyst morphology is controlled by phenotypic as opposed to genotypic factors and therefore palynofacies workers should regard dinocysts only as morphologically variable sedimentary particles produced by motile dinoflagellates and not necessarily expect one-to-one relationships between cyst and dinoflagellate taxa.

Although dinocysts show a great deal of morphological complexity, three major morphotypes can be distinguished (see Sargeant, 1974; Evitt et al. 1976; Williams 1978; Tappan, 1980) as illustrated in Fig. 3.2.

(a) Proximate cysts: these consist of an inner (endophragm) and outer wall (periphragm) which are in close contact. Their morphology strongly reflects that of the motile thecate stage in the life cycle and shows an equatorial indentation (the paracingulum) corresponding to the cingulum of the motile phase. The wall structure is strongly tabulate and although large processes are not present the margins of the paraplates often show simple or ornamented crests or ridges and small spinelets may occur.

(b) Cavate cysts: in these cysts the endophragm (endocyst) and periphragm (pericyst) show a marked degree of separation; the cavity between the two walls (the pericoel) may be continuous or occur as separate apical and antapical cavities (bicavate cysts) or lateral, equatorial cavities (pterocavate cysts). Traces of paratabulation are common on the periphragm but rather rare on the endophragm; the former may show parasutaral ornament, but ornament is otherwise lacking.

(c) Chorate cysts: these cysts are characterised by processes of comparatively great length which show a great deal of morphological variation (e.g. see Fig. 32 in Sarjeant, 1974). The processes arise from the periphragm which is in close contact to the endophragm, and although the arrangement of the processes generally reflects the thecate tabulation, the paratabulation itself is not directly indicated. The length of the processes does not, as was previously thought, reflect the contraction of the chloroplast within the cyst (Tappan, 1980). The processes may have their distal

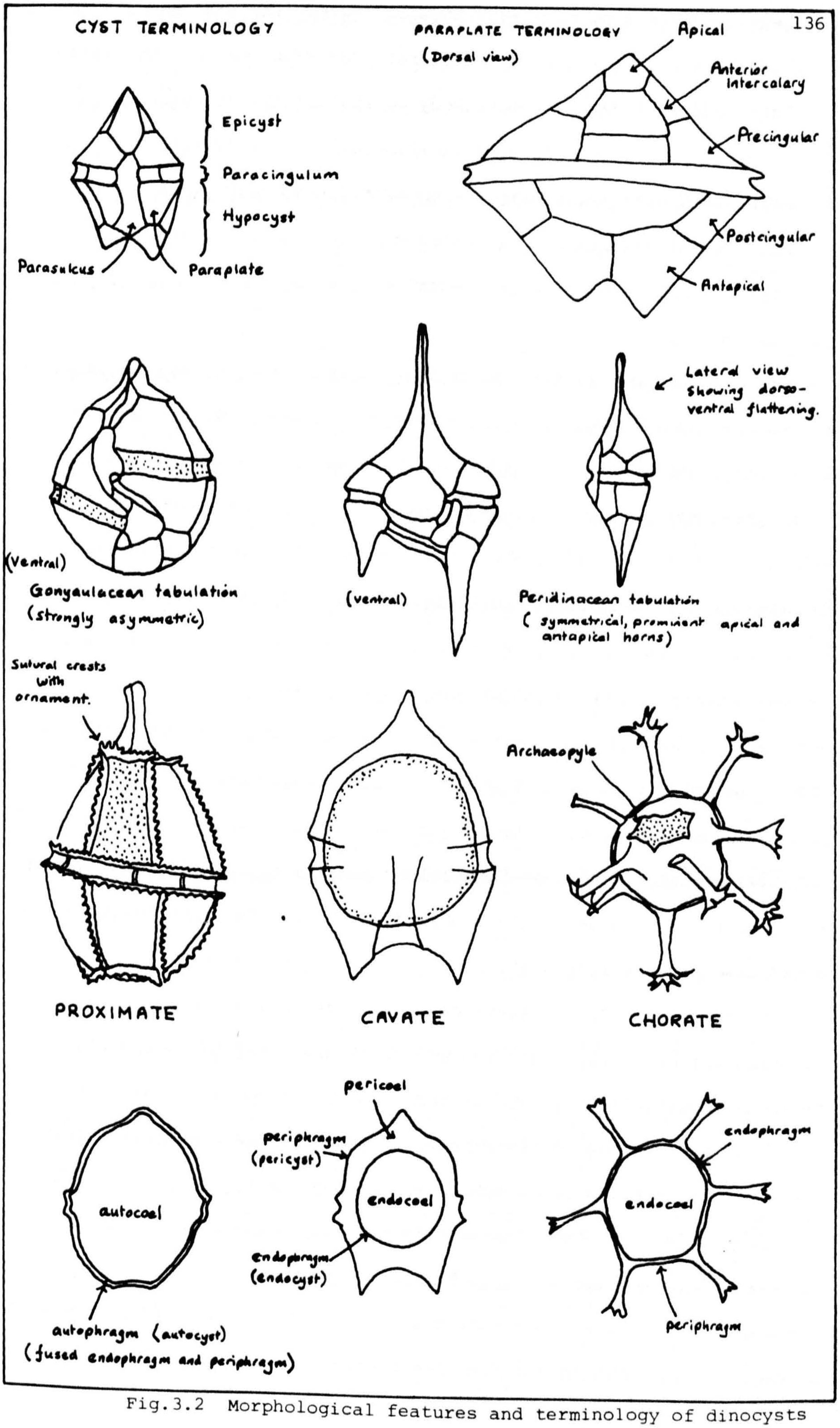


Fig.3.2 Morphological features and terminology of dinocysts

ends connected by thin membranes or narrow 'threads' and are then respectively described as membranate and trabeculate. The ratio of the radius of the main cyst body to that of the overall radius including processes is by definition 0.5 to 0.6 or less. Chorate dinocysts were previously termed hystrichospheres prior to their biological affinities being proved by culture experiments. The typical hystrichosphaerid chorate morphology is sometimes called skolochorate.

Many intermediate types occur between these end members; particularly noteworthy are the proximo-chorates which bear short spines and have an inner:outer radius ratio of 0.6 to 0.8 (above 0.8 the cyst is considered to be proximate). Two types of paratabulation are observed in dinocysts and these are thought to reflect biological differences within the dinophyceae. Peridinacean tabulation is nearly bilaterally symmetrical; a prominent group of anterior intercalary paraplates in a median position is characteristic and two antapical horns and dorso-ventral flattening are common. Some peridinacean forms show pandasutural bands (Evitt et al. 1974). Gonyaulacean tabulation is strongly asymmetrical with a 'spiral' paracingulum, one large posterior intercalary paraplate and (if present) small assymetrical anterior intercalary paraplates. Prominent horns are rare. Further discussion may be found in Evitt (1969) and references cited in Evitt et al. (1974).

Wall (1971, p.401) states that the "factors influencing the distribution of modern cysts include first the environmental and biological controls which affect the primary biogeographic distribution of the living species in the planktic (motile) phase and its encystment and secondly, the factors which determine the behaviour of both viable and non-viable cysts hydrodynamically as small organic particles in transportation and sedimentation". In Chapter Two it was noted that dinoflagellates are most characteristic of at least seasonally stratified neritic waters and Margalef (1978) and Margalef et al. (1979) consider that cyst forming dinoflagellates are adapted to environments with at least temporarily strong vertical gradients. Further comments on the ecology of the motile phase may be found in Williams (1971); the remainder of this section is devoted to cyst distributions.

Davey (1970) and Davey and Rogers (1975) have studied the palynofacies of sediments on the shelves of South and South West Africa (a classic area of upwelling - see Chapter Two). Dinocyst densities were high (>1000/gm.) presumably reflecting the productivity of the upwelling ecosystem, but diversities were low and showed high dominance (assemblages dominated by two species). The dinocysts showed a bimodal distribution (like that reported for spores in the N.E. Pacific by Heusser and Balsam, 1977), being most abundant on the inner shelf, less abundant on the outer shelf (which is characterised by coarser and probably winnowed sediments) and increasing again in continental slope sediments. The ratio of spormorphs to dinocysts was found to decrease offshore with a paucity of dinocysts in the vicinity of deltas. The antipathetic distribution of dinocysts and sporomorphs (particularly in deltaic areas) has previously been noted by Muller (1959), Correia (1971) and Heusser and Balsam (1977), and appears to reflect both dilution of plankton by sporomorphs and the relatively stenohaline nature of most fossilising dinoflagellates. Davey and Rogers (1975) also observed a tendency for oceanic and warmer water dinocysts to have longer and more delicate processes than more nearshore forms (see later).

Wall et al. (1977) have carried out an extensive investigation of dinocyst distribution in recent sediments of the North and South Atlantic margins. Their studies demonstrated the presence of two concurrent distributional trends, one in the inshore to offshore direction and the other latitudinal. Dinocyst diversities and species equitability were found to increase offshore, while dinocyst densities demonstrated a similar pattern and reached their maximum in continental slope and rise sediments. Latitudinal trends were revealed as a peak in diversity in the low latitudes and a peak in density ("by and large") in temperate latitudes. The latitudinal distribution of offshore, oceanic cyst assemblages was found to correlate with the extent of the 'warm water sphere' (i.e. between the northerly and southerly surface outcrops of the main ocean thermocline), but it was also noted that almost all oceanic dinoflagellates find no representation as cysts in bottom sediments. Within individual samples Wall et al. (loc. cit) found that dinocyst density was strongly controlled by textural parameters; values increase semi-logarithmically with the percentages of silt and clay until these comprise 50-60%, after which no further increase occurs. Their study also demonstrated that all areas of upwelling are characterised by an enrichment in the proportion of peridinoid dinocysts (see also Davey, 1970, and Davey and Rogers, 1975). Elsewhere, Dale (1976) has noted that peridinoid cysts are poorly preserved and are therefore probably 'grossly under-represented' in most marine sediments.

Wall et al. (1977, p.165) consider that for environmental correlations to made "two inferences must be drawn regarding the inter-relationship between the dinoflagellate biocoenosis and thanatocoenosis. One is that the distribution of cyst-based species in bottom sediments corresponds reasonably well with the distribution of cysts and their parental dinoflagellates as they occur in surface water environments (i.e. as members of phytoplankton communities). A second is that the primary distribution of cyst producing dinoflagellate species in the plankton is related in like manner to contrasting surface water types". Their observations did in fact demonstrate that "locations where important changes occur in bottom sediments are closely coincidental with hydrodynamic boundaries in the surface water circulation which are usually marked by temperature-salinity discontinuities" (p.166). These hydrodynamic boundaries are the frontal zones discussed in Chapter Two. Wall et al. (op. cit.) found that the most significant boundaries were the estuarine-shelf and shelf-ocean frontal systems, and were able to demonstrate the relationship of dinocyst distributions to thermal stability regimes in the marine environment. They concluded that modern fossilising dinoflagellate species are adapted to unstable and unpredictable hydrographic regimes along continental margins and around oceanic islands. Some degree of relationship between dinocyst distributions in the plankton and hydrographic boundaries is also apparent from other works (compare Reid, 1978; Pingree, 1978; Holligan et al. 1980). Diatom assemblages in shelf sediments have also been related to cross-shelf variations in hydrological structure, with major changes in assemblages corresponding to frontal zones (see Sancetta, 1981; Iverson et al. 1979).

As with sporomorphs, there are rather few palynofacies oriented accounts of the distribution of dinocysts in ancient marine sediments. In an early study of lower Jurassic dinocysts Wall (1965) was able to identify 'inshore' plankton assemblages of low diversity and high dominance and 'offshore' assemblages showing higher diversity and equitability. Scull et al. (1966) found that Oligocene shallow water sediments were characterised by thick-walled, short-spined dinocysts (i.e. proximates to proximo-chorates) and deeper water sediments by thin-walled dinocysts with longer, more delicate processes (i.e. chorate morphotypes). Identical trends have also been described in various Mesozoic sequences by Riley (1974), Scott and Kidson (1977), Harker (1978) and Aurisano (1980). In an investigation of Cretaceous sediments, Harland (1973) found changes in the ratio of gonyaulacean to peridinacean dinocysts which he interpreted as a palaeosalinity indicator. The tendency for peridinacean cysts to be more abundant in more marginal areas is a common observation and has also been documented by Riley (1974), Scott and Kidson (1977), Harker (1978) and Aurisano (1980). It should be noted that peridinacean dinocysts are also enriched in areas of coastal upwelling (see above) as well as in areas of marginal marine conditions and the intepretation of these trends should not be thought of wholly in terms of salinity. Scott and Kidson (1977, p.177) set out four main principles or assumptions which form the basis for the palynofacies interpretation of dinocyst assemblages; these are:-

(1) Specific palynomorph morphotypes are adapted to particular environments (see below).

(2) The most consistent environmental parameter that controls dinoflagellate distribution in the euphotic zone is wave action or water energy.

(3) Morphologic end members that have adapted to very high energy or very low energy environments can be readily identified (see below).

(4) Displacement of dinocysts from their adapted environments to a different environment before they are incorporated into the sediment is an important factor but one that does not typically blend the resulting associations beyond our ability to resolve environmental boundaries (q.v.Wall et al. 1977).

Comments on the possible functional morphology of dinoflagellate cysts.

Vozzhennikova (in Williams & Sarjeant, 1967) suggested that thin-walled delicate dinocysts with elaborate processes (i.e. chorates) were adapted to reduced settling rates and characterised open sea environments, while thick walled simple cysts were more typical of nearshore conditions. Although this is to a certain extent an empirical observation, the main question is whether the distribution of different dinocyst morphotypes results from their differing hydrodynamic behaviour alone, or at least partly reflects a biological adaptation to the hydrographic conditions more-or-less in the area in which they were sedimented. Scott and Kidson (1977) clearly believe the latter. They state that since "all fossil dinoflagellates are considered to be photosynthetic .... it is necessary that the cyst between encystment and excystment remain within the light-penetrating zone of the water column.

If the water is shallow the excystment will be within the euphotic zone, but if the water is deep then the cyst must evolve an elaborate flotation capacity to reduce its settling velocity" (p.177). They also say that the 'elaborately evolved' morphology of chorate dinocysts may be explained as an adaption to increased frictional resistance to sinking by increasing cross-sectional surface area. However, it is clear that light is far from being the only control on excystment,and Smayda (1970) concluded from his review of the suspension and sinking of marine phytoplankton that the extra skeletal material used in processes is likely to nullify the small advantageous decreases in settling rate resulting from increased form resistance. In addition, it is clear that dinocysts capable of being fossilised are only produced by meroplanktic dinoflagellates and that the cysts are benthic and not pelagic.

Since cyst forming dinoflagellates are adapted to unstable euphotic environments they should reach their acme in temperate latitudes in those areas above or within the upper part of the permanent ocean thermocline,and this is exactly what is observed (Wall et al. 1977). Since the motile phase is adapted to stratified conditions, the combination of the two requirements implies that dinocysts are only formed in the seasonally stratified parts of the neritic watermass. Encystment is likely to be the result of the breakdown of stratification (usually in late summer and autumn coincidental with reduced light and temperatures) while excystment is possibly at least partly triggered by changes in bottom water chemistry following restratification (e.g. see Ilsuka, 1972). The dinocyst must clearly remain in the seasonal part of the water column (or corresponding part of the sea floor) and this would imply that they remain on the shelf or upper slope as hypnozygotes or that the dinoflagellate would have to be holoplanktic and possess a pelagic resting stage which would remain in suspension above the main thermocline.

Smayda (1970) believes that the major role of protuberances and processes may be to increase the chances of entrainment in micro-convection currents (e.g. Langmuir cells) and hence effectively reduce the chances of the cyst being sedimented out of the mixed layer. The main advantage in this may be that it enhances dispersal and possibly represents a mechanism for segregating natural populations of dinoflagellates. It is not clear whether there is any tendency for different cyst morphotypes to be preferrentially produced in different parts of the seasonally-stratified shelf watermass. It may be that such differences do exist between areas of differing hydrological structure and stability, e.g. between areas with and without a core flow layer (see Chapter 2). Although changes in cyst morphotype ratios may correspond with various intra-shelf and shelf-ocean frontal systems the absence of any firm modern evidence for such a phenomenum makes it difficult to comment on this possibility.

Proximate dinocysts are characteristically more robust than other cyst types and if they possess processes they are generally short and sturdy (transitional to proximo-chorates). Chorate cysts are the most delicate and thin-walled, while the pericoels of cavate dinocysts may be gas filled (Sarjeant, 1974) improving their flotation characteristics. Hydrodynamic sorting should be expected between these cyst morphotypes and also within each group depending on size, thickness of cyst wall and possibly degree of ornamentation. If we were to imagine that proximate and chorate dinocysts were formed uniformally during the break down of seasonal watermass stratification or due to seasonal changes in light intensity etc., then:

(a) in areas of bottom currents we would expect the winnowing and offshore movement of the chorate dinocysts (and, in part, lighter proximates), while the heavier proximates would remain more-or-less where they were deposited or would be moved preferrentially onshore. Sorting according to hydrodynamic equivalence along an offshore gradient of decreasing turbulent energy would therefore tend to create two end-member thanatocoenoses, one characterised by robust proximates and the other by delicate chorate dinocysts.

(b) in areas of no bottom movement the ratio of proximate to chorate dinocysts would depend on their relative production and the amount of chorate dinocysts being exported in suspension from shallower areas as in (a), i.e. essentially the dilution of one by the other.

If proximate cysts are only produced in the inner most part of the seasonally stratified neritic watermass and chorates only in the outer part, then we would expect the proximate dinocysts to remain more-or-less in the region where they are produced while the chorate dinocysts would (because of their better dispersal) occur in all areas but in varying proportions. Under these conditions proximate dinocysts would be largely absent in quiet offshore areas unless there was appreciable tidal circulation, storm resuspension or turbidity current activity.

If, in the geological past, areas of the neritic watermass were to become stably and persistently stratified (essentially 'oceanic') then we might anticipate that the sediments of such areas would have hardly any dinocysts except for small amounts of chorates that may have drifted offshore from surrounding areas with less stable (seasonally stratified?) hydrographic conditions. This trend might be obscured by offshore changes in the relative rates of deposition of different palynomorph and/or organic matter components. If for example, stable stratification results in bottom water deoxygenation, the sediments may contain large amounts of A.O.M. which dilutes the dinocysts and makes them appear more rare than they really are. The sparsity of dinocysts in persistantly stratified regimes, does not of course imply a sparsity of dinoflagellates! The fact that many dinoflagellates do not produce cysts at all, or cysts which are likely to be preserved, should be a constant source of caution, although it does appear that where productivity is high in shelf-slope waters so are dinocyst densities

(Davey, 1970; Davey & Rogers, 1975). Some massive dinoflagellate blooms leave no dinocyst record at all (q.v. Mahoney & Steimle, 1979; Smayda, 1976).

(ii) Acritarchs

Little is known about the distribution of acritarch plankton in recent sediments; the majority of the taxa are extinct forms whose biological affinities and evolutionary history remains rather obscure (see general reviews by Williams, 1978 and Tappan, 1980). The main varieties of acritarch morphotypes are illustrated in Fig. 3.3. Wall (1965), Williams and Sarjeant (1967), Riley (1974) and Jacobson (1979) have shown that nearshore acritarch assemblages are typically of lower diversity and consist of sphaeromorphs and small, short spined acanthomorphs; passing offshore there is an increase in diversity, in the size and spine length of the acanthomorphs and the abundance of polygonomorphs.Beyond these few generalisations little consistent pattern has emerged (see Tappan,1980 for discussion) but it does appear that the optimum ecological requirements of the acritarchs differed from those of the prasinophyceae (and dinophyceae in the Mesozoic). Acritarchs were apparently more euryhaline than dinoflagellates. Dale (1976) describes how a common small acanthomorph acritarch in the recent sediments of Trondheimsfjord (whose surface waters are brackish) is in fact the cyst of a peridinoid dinoflagellate,and it may be that at least some of those acritarchs found in nearshore,high-stress palaeoenvironments may have been the cysts of euryhaline dinoflagellates. This is possibly supported by the fact that acritarch:dinocyst ratios normally decrease offshore.

(iii) Prasinophycean algal phycomas

Our present knowledge of the character and biology of prasinophycean algae is based heavily on the work of Dr. Mary Parke (now retired) and

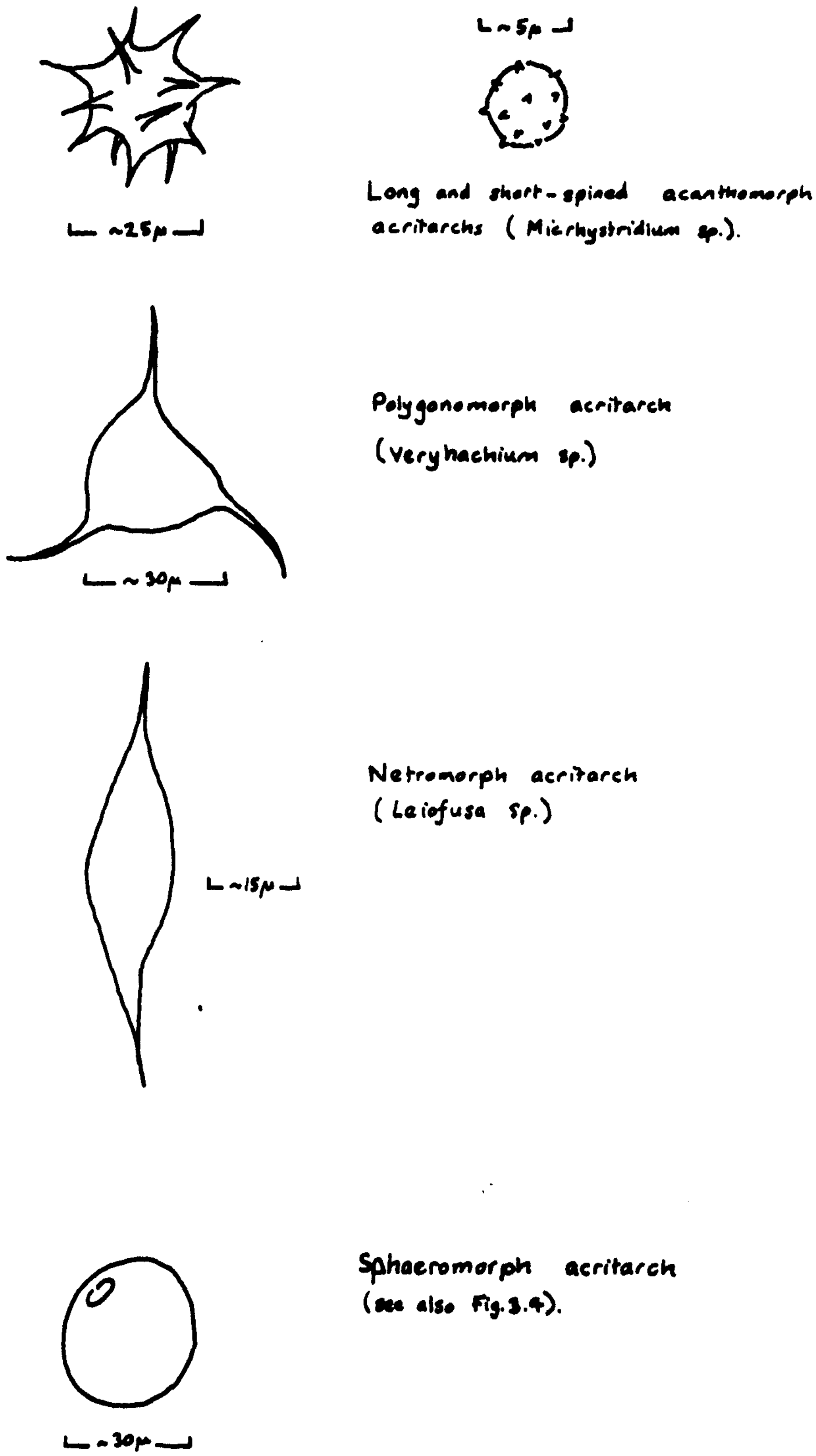
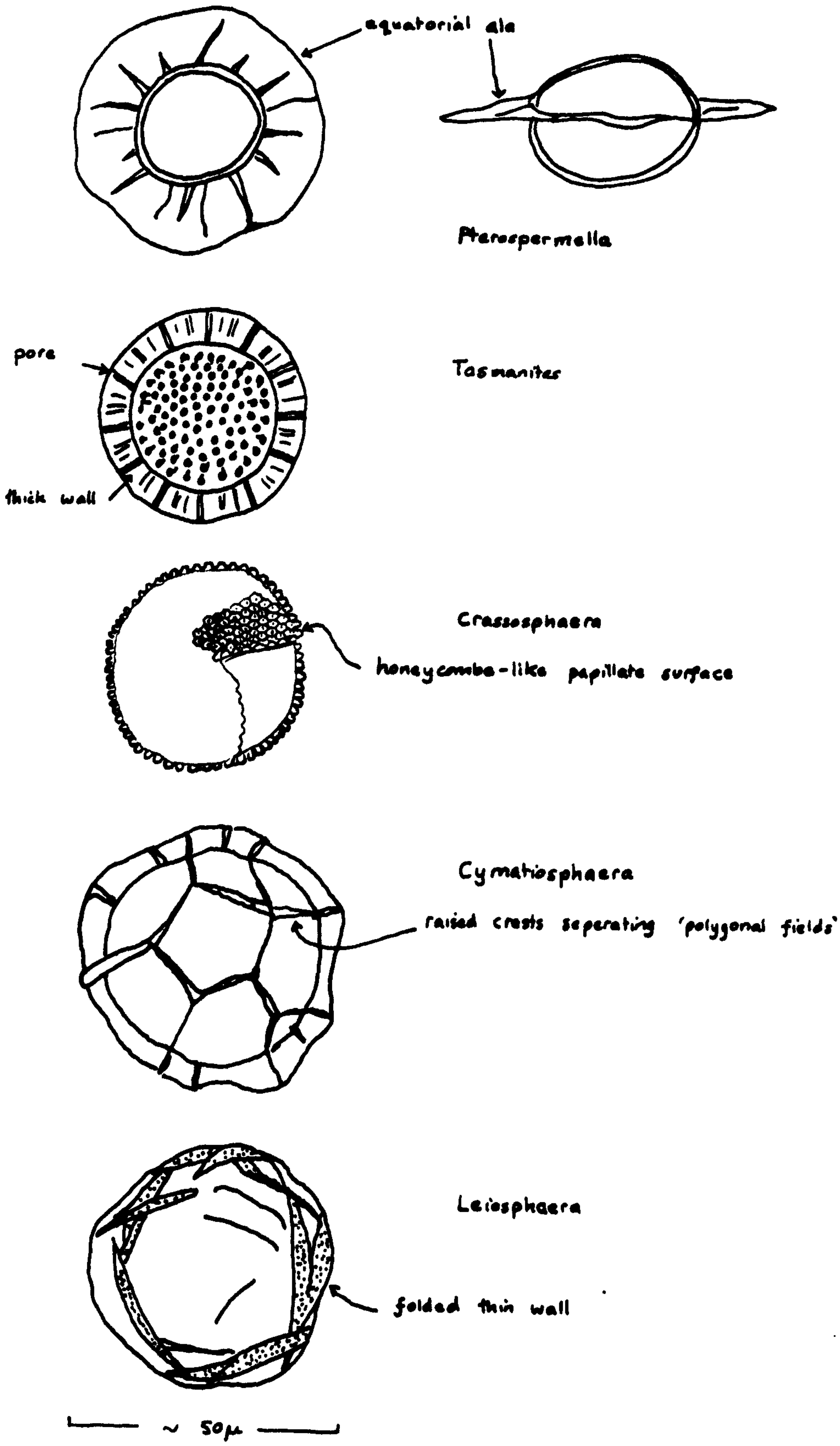


Fig. 3.3 Principal varieties of acritarchs relevant to this study

her colleagues at the laboratory of the Marine Biological Association at Plymouth. A series of papers including Parke & Hartog-Adams (1965), Parke (1966), Boalch & Parke (1970), Parke & Dixon (1976), and Parke et al. (1978) have been published on the three major genera Halosphaera, Pachysphaera and Pterosperma. Parke and Dixon (1976) list the prasinophycean algae as a class of the Chlorophyta but other workers (e.g. Round, 1971) have separated them under the division Prasinophyta (as followed by Tappan, 1980). The life cycle of these algae consists of two principle stages, a motile flagellate stage and a pelagic phycoma phase (Boalch & Parke, 1970; see also Tappan, 1980 for a general review). The phycoma is not a cyst as such, since it is not dormant and its contents are undergoing active cell division prior to their final release. It has a double wall structure consisting of an inner pectic wall and an outer very tough sporopollenin-like wall which is variously punctate, smooth, papillate or alae bearing in the various genera (see Fig. 3.4). When new, the phycoma contains a large lipid globule which acts as a food source and because of its low density provides bouyancy. Under extremely calm conditions halosphaerids and pterospermids have been observed to accumulate as a green film on the sea surface but these are probably not 'blooms' as such (see Parke et al. 1978). Sedimentation of the phycoma may occur after the release of the contents, via zooplankton fecal pellets or possibly after exhaustion of the lipids and loss of viability following a long period of residence in a subsurface watermass (G.T. Boalch pers. comm. 1980). Phycomas are most common in the sea during the winter and early spring, the flagellate phase predominating for the remainder of the year. The prasinophyceae appear to have a cosmopolitan distribution (perhaps because of the pelagic character of the phycoma and its entrainment in circulation patterns) and are dominantly marine although representatives of the class occur in all aquatic regimes. Little as yet is known about their ecology.

Over the last twenty years a series of papers including Wall (1962),

Fig. 3.4 Principal types of prasinophycean algal phycomas



Boalch & Parke (1970), Eisenack (1972) and Taugourdeau-Lantz (1979) have shown that certain palynomorphs previously classed as acritarchs are probably fossil relatives of the prasinophycean algae. The suspected relationships are:-

Tasmanites, *Monogemmites*, *Tytthodiscus*	≃	Pachysphaerid or halosphaerid
Leiosphaeridia, *Crassosphaera*, *Pleurozonaria*	≃	Halosphaerid phycoma
Cymatiosphaera, *Pterosphaeridia*	≃	*Pterosperma* phycoma with polygonal fields (q.v. Parke et al. 1978)
Pterospermopsis, *Pterospermella*	≃	*Pterosperma* phycoma with equatorial alae (q.v. Parke et al. 1978).

Rather little is known about the distribution of the fossil forms other than they occur in a wide range of salinities and occasionally in great abundance (e.g. Tasmanite rock, q.v. Cane, 1976). In the late Quaternary sequence in the Black Sea, *Cymatiosphaera* shows a peak abundance (up to 3 to more than 6 times its normal occurrence) in the Unit Two sapropel (Wall & Dale, 1973, 1974). Its subsequent reduction in the overlying coccolithic sapropel (Unit One) may indicate that the organism was adapted to stable low nutrient conditions. Prasinophycean algae are often conspicuously characteristic of ancient sapropel-type black shale sediments (e.g. see Chapter Four and Mädler, 1963; Singh, 1971; McLachlan & Pieterse, 1978; Davey, 1978; Hochuli & Kelts, 1979 and Cluff, 1980) but this may largely be due to lowered sediment accumulation rates and the cessation of dinocyst production. LIttle is known about their actual palaeoecological significance.

(b) AMORPHOUS ORGANIC MATTER (A.O.M.)

Particulate amorphous organic materials are very common in the modern marine environment (see general reviews by Riley, 1970; Parsons, 1975 and Whittle, 1977). These particles are normally called organic aggregates, organic-mineral aggregates or flocculent aggregates and are distinct from detritus which consists of recognisable fragments or debris (see Riley, 1970). Many workers, however, are guilty of calling this material 'detritus'. Parsons (1975) has estimated that organic aggregate material exceeds phytoplankton carbon by a factor of ten or more except during algal blooms. With the same exception, Finenko and Zaika (1970) have estimated the non-living component of the particulate organic carbon (P.O.C.) in the water column to be 50-90%. Wiebe and Pomeroy (1972, p.327) state that "it is clear that in most regions below the euphotic zone 'detritus' constitutes a source of particulate reduced organic carbon much greater than the biomass of living organisms. It is also clear that 'detritus' represents but one portion of a reversible continuum which includes dissolved organic matter and organisms." Riley (1963) believes that organic aggregates originate by a physico -chemical process that involves bubbling and adsorption and Parsons (1975) notes that their visual appearance suggests they have formed in situ rather than from simple decomposition of planktic organisms.

The following are composite descriptions of the two major types of suspended particulate amorphous organic materials found in modern marine waters:

(i) Amorphous aggregates (senso stricto).

Gordon (1970, p.178) has described amorphous aggregates from the Atlantic Ocean as "amorphous particles composed of a variety of aggregated yet distinctive organic and inorganic sub-units of variable composition. Some may contain phytoplankton and bacteria plus recognisable mineral crystals while others are mere blotches which are extremely difficult to quantify and are visible only after staining. Their outline is roughly circular and poorly defined. When suspended in seawater, they are almost spherical." These aggregates were principally of a carbohydrate composition, 7.25μ in diameter and most abundant in surface and coastal waters. Mel'nikov (1976, p.401) has described aggregates from the eastern Pacific as "heterogeneous amorphous masses apparently consisting of separate particles bound together in a loose gelatin-like matrix or tightly packed complex. Phytoplankton cells, plant fibres and mineral inclusions could frequently be seen in some of them". They ranged in size from a few to several hundred microns. Riley (1963, p.373) found that "naturally occuring organic aggregates in Long Island Sound commonly consist of pale yellowish or brownish amorphous matrices with inclusions of bacteria, silt particles and sometimes phytoplankton". These particles were delicate, plate-like and $\sim 5\mu$ to several millimetres in diameter. Wiebe and Pomeroy (1972) found aggregates of identical character to those described above which were chemically heterogeneous and contained increasing numbers of bacteria as the coast was approached. Riley (1970) comments that the organic content of the aggregates is rarely >30% (clay and calcite crystals are common) and that aggregates may show a maximum in the oxygen minimum layer ($\simeq$5 times their normal concentration).

(ii) 'Flakes'

Gordon (1970, p.179) describes flakes from the Atlantic Ocean as "thin, scale-like particles having a distinct margin and circular to elongate outline with a varying number of small inclusions which appear to be mostly bacteria". Gordon (loc. cit.) believed these particles to be dominantly proteinaceous. Equivalent particles have been described from the eastern Pacific ocean by Mel'nikov (1976, p.401) who called them 'flocs'. These were "fine, transparent platey particles with irregular angular edges. They appear to be fairly homogenous matrices often with organic and mineral inclusions. They are mostly large particles (50-200μ) and are distributed mainly in the top layers of the euphotic zone". Wiebe and Pomeroy (1972, p.332) have described them as having a "rather homogeneous looking matrix with irregular angular edges and organic and inorganic materials embedded in or attached to the particle. They appear evenly distributed through the water column and their morphology is more constant with depth than other materials".

The chemical composition of this particulate material appears to depend upon the extent to which it is degraded (see Chapter 2). The proportion of relatively labile components in the particles is likely to decrease with depth and distance from shore. Eppley et al. (1977) concluded that 'detritus' (aggregates) in surface waters off California has a composition little altered from its original chemical composition in organisms from which the particles were considered to have been derived. My contention in Chapter Two that "in an aerobic environment the organic aggregate material will be destroyed or degraded to refractory material with a more 'humic' composition" is supported by a recent reference in which Huc (1980, p.464) states that organic matter in surface waters is "rapidly transformed into material with a high molecular weight of the humic compound type". The lipid component of the aggregates will be rapidly destroyed in the mixed layer.

Davey and Rogers (1975) have reported A.O.M.-rich sediments in the highly productive upwelling area off S.W. Africa. Hobson (1971) found that 30-60% of the suspended P.O.C. in waters of this region consisted of 'detritus' (see also Emery & Honjo, 1979) and so the A.O.M. content of the (dinocyst-rich) sediments may represent the settling and preservation of this material. Mahoney and Steimle (1979) describe how following a massive, non-toxic dinoflagellate bloom and bottom deoxygenation in the New York Bight the sea floor was covered in a 1cm layer of yellowish-coloured flocculent material containing dinoflagellate debris and bacteria. As bottom oxygenation increased and this material was oxidised the recognisable content progressively decreased and it became fine textured and blackened in appearance. This event was atypical for this area but it is clear that had the bottom water been permanently oxygen depleted (as off S.W. Africa) a large amount of A.O.M. may have become preserved in the sediment. The normal hydrographic regime and sediment textural parameters were also unfavourable in the New York Bight. However, given the correct conditions it is possible that large dinoflagellate blooms and their decay products may also be a significant contributor to A.O.M. in fine grained sediments. See Chapter 5 for further discussion.

The other major source of amorphous organic matter in recent sediments are the faecal pellets of benthic organisms and zooplankton which are described below.

(i) Zooplankton (copepod) faecal pellets

Honjo and Roman (1978) have provided an excellent review of this topic (see also Turner & Ferrante, 1979; Porter & Robbins, 1981). They note that copepods produce faecal pellets at a considerable rate ($\leq$200 per individual per day) and that they are an important food source for detritivores. Since they are larger than most particles in the open sea (150-700µ x 30-80µ) and because they are held together by a resistant ('peritrophic') membrane, they sink faster (80-200m/day) and tend to maintain their integrity longer. Honjo and Roman (ibid) found that the combustible (organic) content of the pellets they analysed was 25-45% but report that some pellets that have been studied "contained primarily amorphous organic material" (p.53). As they settle the outer binding membrane of the pellets decays (at a rate depending primarily on temperature) and the pellets may disintegrate before reaching the bottom, breaking up into "small amorphous aggregates" (p.45). It would appear that faecal pellets are of considerable importance in the transfer of materials from the surface waters to the sea floor (see also Fowler, 1977). In Chapter Two it was concluded that the process responsible for the preservation of organic matter occurs primarily at the sea floor. Müller and Suess (1979, p.1356) also consider that "the finding that the sedimentation rate largely determines the fraction of primarily produced organic carbon that eventually becomes fossilised in (aerobic) sediments also suggests that most of the organic material escaping the euphotic zone is remineralised at the sea bottom rather than during settling through the water column." When one considers that about 80-90% of the primarily produced organic matter is recycled within the mixed zone (Whittle, 1977) this suggests that zooplankton are very important in transferring organic matter from the euphotic zone to the bottom, particularly in deep stratified waters.

(ii) Faecal pellets of macrobenthic organisms

Macrobenthic organisms are prolific producers of faecal pellets (see Tables I and II, pages 267 and 275 in Rhoads, 1974). Although deposit feeders produce much faecal material their main effects take place "in the direction of oxidation" (Rhoads, 1974, p.278) and hence ultimately tend to reduce the organic content of the sediments, partly because bioturbation tends to lower the R.P.D. and partly because faecal pellet formation results in an increase in the surface area available for colonisation by micro-organisms (Driscoll, 1975; see also Mann, 1972). According to Rhoads and Young (1970) suspension feeding organisms may, however, produce an enrichment of organic matter in the sediment due to 'biodeposition' of suspended materials including organic matter (which is only partly utilised prior to egestion). Johnson (1974) has examined sedimentary 'organic-mineral aggregates' which probably originated as macrobenthic faecal pellets, from coastal sediments near Woods Hole. These aggregates constituted 13-71% of the sediment, were yellowish-brown in transmitted light and consisted of an amorphous, predominantly organic (largely carbohydrate) matrix in which a wide size range of mineral particles from a few to several hundred microns were embedded. Johnson noted that some of the quartz particles which they contained were much larger than the settling rate of the aggregates suggested, indicating the matrix had a buoyant effect. He considers these aggregates to be genetically different from the pelagic amorphous aggregates described earlier. Webster et al. (1975) also record significant amounts of A.O.M. in coastal sediments.

Modern amorphous organic materials are clearly polygenetic and heterogeneous and result from a complex combination of biochemical and physico-chemical processes involving dissolved and particulate precursors. Very little is apparently known about the distribution of amorphous organic matter in ancient sediments other than it tends to correlate with T.O.C. in marine rocks, is abundant in black shales and is apparently only preserved under strongly reducing conditions (e.g. Kendrick, 1979). It has traditionally provoked more interest from organic geochemists than palynologists since the majority of the latter regard it solely as a nuisance because it tends to obscure palynomorphs in palynological preparations. It is often called amorphous sapropel but this implies geochemical affinities and the general label 'A.O.M.' is far more desireable as it is often very difficult to distinguish from amorphous humic materials (e.g. collinite), particularly in ordinary light (e.g. Staplin, 1969). The following descriptions of this material (and the terms used to describe it) occur in the literature:-

Staplin (1969): sapropel: unorganised fluffy to semi-coherent masses. "Un-organised, amorphous, light-coloured, translucent debris, usually in semi-coherent fluffy or somewhat platey masses, indicates relatively anaerobic sapropelic deposition" (p.52).

Correia (1971): MOC (matiere organique colloidale): granular in mass and without true structure, of variable colour and with or without various inclusions. 'Sapropelic material'.

Batten (1973): Mush: "Variously homogenised, frequently coagulated, combination of .... minute fragments of poorly preserved miospores; entire but pale, crumpled 'ghosts' of miospores; small fragments of wood; very pale flimsy structureless tissue, partly cuticular-epidermal; finely divided organic (soft tissues of plants and animals) and inorganic (pyritous?) material; and mineral debris" (p.28). This was considered in part equivalent to Correia's 'sapropelic material'.

Burgess (1974): Amorphous debris: "relatively structureless masses of organic matter, often with granular texture, but usually without other distinctive structure and in the thermally unaltered state bright yellow to orange in colour". "Only rarely is distinctive algal structure preserved following diagenesis and compaction. This amorphous debris is rich in the lipid fraction" (p.24).

Tissot (1977): Amorphous material (Type II kerogen in part): vaporous or cloudy shape and no identifiable structure.

Timofeev & Bogolyubova (1979): Sapropel-humic and humic-sapropelic matter: "composed of sapro-humo-collinite and humo-sapro -collinite microcomponents which form a mechanical admixture of gelification products (of varying quantitative ratios) of allochthonous remains of higher plants, autochthonous algae and decomposition products of animal origin. Like sapro-collinite they are finely dispersed with clay minerals. The petrographic aspect of humo-sapro-collinite abounds in beige-olive sapropelic structureless brownish flakes (>50%) with admixed clots and fine grains (<1μm) of humic material. The mixture produces a nonuniform colour with a granular, sometimes dotted granular or flocculent structure. With increase in humic components of this nature (>50%) the groundmass becomes sapropelic-humic, distinctly brown in colour, with a more pronounced lumpy and granular structure" (p.831-832). Collinite appears to have been used to describe any amorphous organic material regardless of composition.

Timofeev and Bogolyubova (1979) report that the humic content of the amorphous ("collinitic") organic material is greatest in the nearshore facies of Cretaceous black shales in the Atlantic. Correia (1971) states that A.O.M. is usually encountered when micro-organisms are rare and plant debris less abundant (q.v. Chapter 2). It should be noted that the descriptions of ancient A.O.M. reproduced above are very similar to those given for modern amorphous materials. There is however, a fair range in the morphology of amorphous kerogens and Combaz (1980) has recently proposed the following subdivision:-

(i) Grumeleuse organic matter: apparently the 'normal' A.O.M. particles.

(ii) Granular organic matter: distinguished from (i) by the presence of 'innumerable' granules of organic material, pyrite or bacteria of about 1μ in size.

(iii) Sub-colloidal organic matter: an artifact of palynological preparation techniques.

(iv) Pellicular organic matter: typical of carbonate rocks; dull in colour and with common imprints of carbonate crystals.

(v) Gelified organic matter: coherent, compact, consolidated particles in the form of angular or rounded grains; from my understanding of his paper this includes amorphous vitrinite and collinite s.s.

Ancient amorphous kerogens (collinite excluded) are herein considered to be directly related to modern A.O.M., allowing for the complex reactions that occur during the transformation of organic materials into kerogen (cf. Brooks, 1978: Huc, 1980). Many of the morphological features of amorphous kerogens are, however, strongly influenced by compactional effects and the nature of the mineralic matrix of the host rock. It should be appreciated that thermal maturation in itself affects the visual appearance of the kerogens, including more subtle changes than those of spore colour etc. Peters et al. (1981, p.701) report that during experimental maturation of sediments rich in filamentous, stringy, cyanobacterial debris all structure is lost at a stage corresponding to H/C = 1.1 with the algal components being converted to amorphous material resembling "bituminite".

KEROGEN ASSEMBLAGES ("PALYNOFACIES")

In two recent papers Habib (1979a, 1979b) has recognised and defined four "palynofacies" (kerogen assemblages) present in Cretaceous sediments cored by D.S.D.P. in the Atlantic Ocean. The four assemblages are:-

(a) The exinitic palynofacies (Habib, 1979a, 1979b)

This assemblage consists of a large number of spormorphs and abundant structured palynodebris (vis. phytoclasts) with phytoplankton and amorphous material rare or absent. The sporomorph assemblage is diverse and characterised by heavy, ornamented trilete fern spores (especially striate, schizaeaceous forms). The palynofacies occurs in sediments with high organic carbon values (0.5-2.5%) and is interpreted as having been deposited during times of high rates of sedimentation of land plant materials contributed from actively prograding deltas.

(b) The tracheal palynofacies (Habib, 1979b)

This assemblage is similar to the exinitic palynofacies in that it contains numerous sporomorphs but it has more abundant phytoclastic material and phytoplankton is well represented (16-60%). The sporomorph assemblage is of low diversity and is dominated by bouyant forms.

(c) The micrinitic palynofacies (habib, 1979a, 1979b)

This palynofacies is characterised by an overwhelming abundance of unstructured, small (>20μ), even-sized, more-or-less equidimensional carbonaceous particles that are generally opaque to semi-opaque and black to dark brown. More phytoplankton are present than in the exinitic palynofacies and the sporomorphs are correspondingly less abundant and characterised by relatively bouyant forms including bisaccates and smooth deltoid trilete fern spores. Organic carbon values are relatively low (<1%) and the palynofacies is interpreted as representing a much diminished sedimentation of terrigenous organic materials and deposition by marine currents which concentrated those sporomorph types morphologically suitable for selective transportation.

(d) The xenomorphic palynofacies (Habib, 1979a, 1979b)

This consists of abundant "amorphous palynodebris" which has irregular outlines and is translucent and pale yellow-orange in colour. Two types are present: large (>50μ) globular particles (probably equivalent to normal A.O.M.) and smaller (<15μ) particles of irregular and shredded appearance (probably highly degraded plant material). Palynomorphs are not usually abundant but phytoplankton are generally more numerous than sporomorphs which consist of bisaccates and the more bouyant fern spore morphotypes. There is little correlation between organic carbon values and palynomorph abundance in this palynofacies and at least part of the amorphous material may represent autochthonous marine organic matter. Where "amorphous palynodebris" is present it is usually the most abundant component.

Classopollis pollen were found in high relative abundance in all these assemblages. It is clear that the sequence (a) to (d) represents a proximal-distal trend, although to a certain extent the difference between the micrinitic and xenomorphic palynofacies may be essentially one of redox conditions. When comparing Habib's observations with other works it is important to note that his palynological preparations have been subjected to partial oxidation with nitric acid and that this will undoubtedly have altered the abundance and appearance of amorphous materials present in the kerogens. In geochemical terms the exinitic, tracheal and micrinitic assemblages represent Type III kerogens (see below) and the xenomorphic palynofacies probably varies from Type III to II depending on the type of amorphous debris present (theoretically it could also correspond to a Type I kerogen but this is unlikely for the sediments studied by Habib).

Habib's palynofacies terminology has also been used by Aurisano

(1980) in a palynological study of Upper Cretaceous sediments from the coastal plain of New Jersey. Unfortunately Aurisano's work does not include any detailed (or even general) description of the sediments studied and this makes it very hard to assess the significance of the palynofacies distributions that were recorded. The sediments appear to be predominantly sandy with argillaceous interbeds, are apparently of shallow water shelf facies, and thus not surprisingly consist mainly of the terrigenous dominated palynofacies with the xenomorphic palynofacies only well developed in Santonian-Campanian clays in the most offshore well studied (Cost B-2, Baltimore Canyon Trough). It is apparent that Aurisano (1980) found the kerogen character to represent more of a continuum than the four clear cut assemblages described by Habib, and although this may to a certain extent reflect a genuine difference between oceanic and shelf sediments it does high-light the weakness of Habib's classification. Such a classification depends on a general visual estimation of the kerogen character and the subsequent placing of the assemblage into only one of four possible categories; clearly only the coarsest trends can be revealed by such an analysis. The other major problem with Habib's scheme is the inherent ambiguity in the term "xenomorphic". It is absolutely essential that a clear differentiation be made between what is amorphic degraded plant debris and what is amorphous organic matter in the accepted sense of the term. This distinction is of prime importance to the palynological assessment of source rock character and potential.

GENERAL KEROGEN CHARACTERISTICS: GEOCHEMICAL ASPECTS OF PALYNOFACIES

It is beyond the scope of the present work to provide a comprehensive review of the geochemical analysis of organic matter in acient sediments and hence only particular aspects directly related to this study will be discussed. Reviews of the geochemical parameters and techniques used in the assessment of ancient depositional environments may be found in

Correia & Peniguel (1975), Tissot (1977), Didyk et al. (1978), Tissot & Welte (1978), Hunt (1979) and Durand & Monin (1980).

Excluding inert carbon (inertinite), the two most important, geologically stable, naturally produced organic substances are lignin and lipid (Huc, 1980). This division forms the basis for the recognition of the two principle types of kerogens in sedimentary rocks, namely humic material (vitrinites- ligno-cellulosic tissues of higher plants) and sapropelic material (exinites or liptinites - the lipid-containing substances of spores, plankton, cuticle material and some amorphous materials). The geochemical character of humic and sapropelic materials is usually expressed in terms of their carbon, hydrogen and oxygen ratios (e.g. see Durand & Monin, 1980). The lipid-rich sapropelic materials are characterised by relatively high hydrogen contents (10% or more) and prior to maturation show high H/C ratios and low O/C ratios. The lignin-rich humic material has originally lower H/C and higher O/C ratios. Although the character and maturation paths of these end member materials have been well established it is clear that the kerogen content of marine sediments will represent a complex and varying mixture of these components. However, since humic materials are essentially land derived and sapropelic materials chiefly of marine origin (excepting sporomorphs and cuticle), it has long been appreciated that there will be an offshore increase in the ratio of sapropelic to humic materials (e.g. Uspenskiy, 1978, cited in Swanson, 1960 and data in Breger & Brown, 1963 and McIver, 1967).

The principle technique used to determine the C/H/O ratios of kerogens is pyrolysis (see Tissot & Welte, 1978) and the results are generally plotted on a "Van Krevelen diagram", whose vertical axis is the atomic ratio of hydrogen to carbon (or hydrogen index) and horizontal axis the atomic ratio of oxygen to carbon (or oxygen index) as shown in Fig. 3.5. Tissot and other members of his group at the Institut Francais due Petrole have used this pyrolysis technique as a

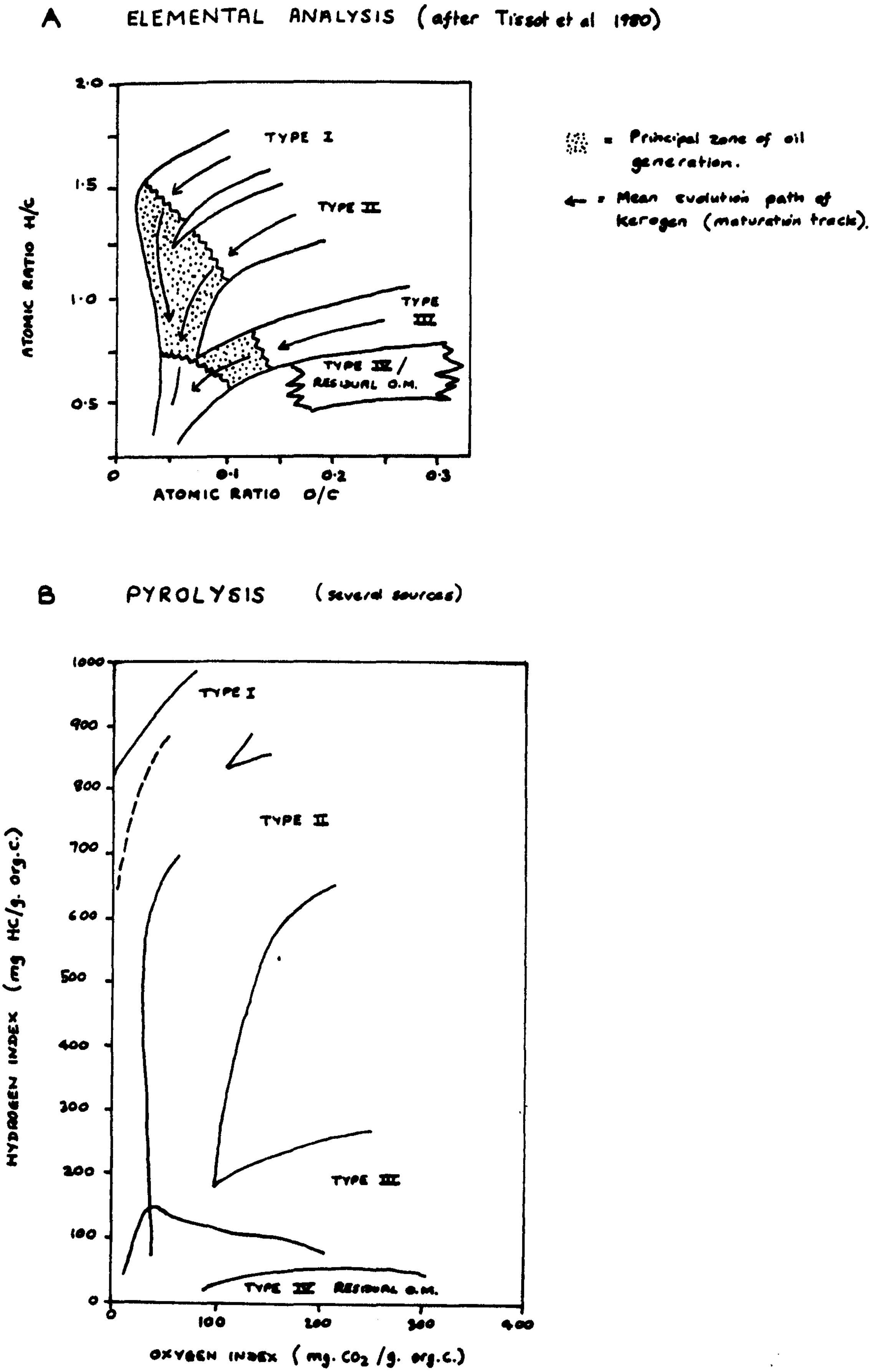


Fig. 3.5 Geochemical characterisation of organic matter Type

basis of a three-fold kerogen classification (see Tissot, 1977; Durand & Monin, 1980) which is outlined below.

Type I Kerogen: "Kerogens having an originally high hydrogen content This type of kerogen is very rich in lipids, and particularly aliphatic chains. It contains only a small proportion of polyaromatic nuclei and oxygenated functional groups. The organic matter may be derived from an accumulation of algae, like oil shales containing Botryococcus and the recent coorongite. It may also result from an intense reworking of various kinds of microbes, leaving mostly the lipid fraction of the original material and microbial lipids" (Tissot, 1977, p.56). This type of kerogen was defined on the Green River Oil Shales of Colorado and Utah which are rather unique and hence Durand and Monin (1980) have extended it to cover most of the bogheads consisting of algae of the Botryococcus family. General aspects of bogheads and associated sapropelic coals and other analagous sediments may be found in Moore (1968) and Cane (1969, 1976). H/C values >1.2.

Type II Kerogen: "Kerogen having a fairly high original hydrogen content, although it is somewhat lower than Type I. Kerogen contains abundant aliphatic chains of various lengths and saturated rings. Also, it comprises polyaromatic nuclei and hetero-atomic functional groups, like carbonyl or carboxyl groups which contain oxygen. The organic matter is usually derived from marine phytoplankton and zooplankton, laid down in a confined environment"

(Tissot, 1977, p.56). "Under the microscope it consists essentially of amorphous material probably derived from planktonic (algal) biomass. The identifiable remains of organisms (20% to 30% of the organic content) are <u>Tasmanaceae</u> and <u>Nostocopsis</u> algae and ligneous debris in variable proportions indicating a small terrigenous contribution" (Durand & Monin, 1980, p.128). H/C values 1.0 to ∿1.2.

Type III Kerogen: "Kerogen having a low hydrogen content and a high oxygen content. Its chemical structure comprises mainly polyaromatic nuclei and oxygen containing functional groups. In addition to these some aliphatic chains, including long straight chains from natural waxes, are attached to the polycyclic network. The organic matter is mostly derived from higher land plants. This terrestial input may be transported by rivers and currents and laid down either in non-marine or in deltaic or continental margin environments" (Tissot, 1977, p.56). "Under the microscope it appears to consist of amorphous cement containing recognisable debris of evolved plants and vitrinite splinters" (Durand & Monin, 1980, p.128). Humic coals are essentially equivalent to this type of kerogen. H/C values 0.5-0.8.

In addition to these three types, other workers have recognised a type IV kerogen (Harwood, 1977) which has H/C values between 0.5 and 0.8 and "generally appears as wood or opaque, coaly fragments under a transmitted light microscope" (Powell & Snowdon, 1980, p.424). This appears to correspond to the "residual organic matter" category recently introduced by Tissot et al. (1980). p.2053) which is described as having been "either oxidised in subaerial environments and/or sediment transit, or recycled from older sediments". Tissot et al. (1980) state it has no hydrocarbon potential while Powell and Snowden (1980) report it to generate primarily dry gas on maturation.

The idealised nature of the Tissot classification and its basis on bulk rock, or more precisely bulk organic matter analysis, should be appreciated. In reality pyrolysis data often shows a high degree of scatter when plotted on the Van Krevelen diagram and the classification scheme is essentially a rationalisation based on the known compositions (and positions) of relatively pure samples of end-member materials (e.g. vitrinites, exinites, alginites). Naturally occurring kerogen mixtures not suprisingly often fall outside the idealised fields since their composition potentially reflects a whole spectrum of constituent materials.

Despite the somewhat artificial basis of the Tissot scheme there is a fairly logical argument that can be made for there being basically three main types of kerogen mixtures. "Type I" kerogens are very distinct and rather rare, their unique character largely the result of abundant cyanobacterial (blue-green algal) organic matter as in sapropelic bogheads and algal oil shales (e.g. Cane, 1976). This essentially leaves only "Types II" and "III" to be accounted for. Modern studies of organic matter decomposition have demonstrated that most naturally occurring particulate organic materials consist essentially of two components, a labile component and a refractory component (see Godshalk & Wetzel, 1978 and Parsons, 1975). The labile component includes most of the lipid fraction and is rapidly degraded in aerobic environments, being largely converted to D.O.M. or consumed by bacteria within a few days of the beginning of decomposition. The remaining refractory material consists of less digestable structural compounds such as lignin which decompose much more slowly (months to years). In terms of the main species of palynologically recognisable particles, the material with the potentially largest labile component is A.O.M.; both vitrinites and exinites are predominantly refractory. As has been discussed previously, the labile lipid component of A.O.M. can only be preserved under highly reducing conditions (e.g. Kendrick, 1979).

From the considerations of preservation and palynofacies developed here and in Chapter 2, the following interpretations for "Type II and III" kerogens becomes evident. Under stable dysaerobic and especially anaerobic or anoxic conditions, both autochthonous and allochthonous labile lipids are preserved, the autochthonous component being largely A.O.M. In the centre of the basin the effects of watermass stratification and the swamping effects of the preserved A.O.M. will tend to result in a low humic content and hence barred basin, pericontinental and epeiric black shale megafacies sediments will be largely characterised by "Type II" kerogens. Under aerobic conditions (either basin margins or in entirely oxygenated basins) the labile lipids (and A.O.M.) will be rapidly destroyed leaving only resistant ligno-cellulosic and sporopollenin materials (see also Merewether & Claypool, 1980, p.495; Demaison & Moore, 1980b, p.1194; Pratt, 1981). Under these conditions the A.O.M. will be destroyed or degraded to a more refractory 'humic' composition, leaving a "Type III" kerogen. This will be characteristic of the black mud megafacies and organic-poor sediments which often show a good correlation between the humic content and T.O.C. values. The fact that degraded and oxidised A.O.M. has a relatively 'humic' composition (Huc, 1980) should be taken as a warning that not all "humic" materials are of terrigenous origins. Since it is clear that the nature of the organic matter supplied, the redox conditions and the rate and pathways of decomposition (as also influenced by sedimentation rate, temperature, environmental stability, etc.) all will influence the composition of the resulting kerogen, it is not surprising that Van Krevelen plots tend to show a good degree of scatter. Too many factors are in reality involved to allow a simple, straightforward three or four-fold geochemical classification of kerogens, although such a rationalisation may be a useful one for the purposes of general discussion.

During decomposition it is the decay of the labile component of particulate organic matter (and of the D.O.M. released during this process) which results in the greater part of the B.O.D. (see Godshalk & Wetzel, 1978). The slow decay of refractory materials uses rather little dissolved oxygen per unit time. Given that the labile component only survives a few days in an aerobic environment, this suggests that bottom water deoxygenation is most likely where there is reduced circulation and a supply of relatively fresh organic matter which has an initially significant component of labile material. In general, refractory terrigenous organic matter, which in most cases will have probably been present in an aerobic environment for at least several days prior to final deposition, is unlikely to create oxygen-depleted bottom water in offshore marine settings. It is likely that organic matter sedimentation via zooplankton faecal pellets may play a key role in the transfer of relatively undegraded labile material into the bottom water (e.g. see Seki et al. 1974). Given the above reasoning, the abundance of terrigenous organic matter occasionally observed in marine black shales may reflect purely sedimentational factors; it cannot be inferred that this material is only present because of high preservation rates due to oxygen depletion, or that the abundance of this material would have necessarily resulted in such oxygen depletion. I believe that this relationship is supported by the nature and distribution of organic matter types in the Cretaceous Hatteras Formation of the Atlantic Basin that have been reported by Tissot et al. (1979, 1980).

During burial and maturation (e.g. see review by Staplin, 1975) all kerogen types show a decrease in their H/C and O/C ratios as they approach graphite. Reworking and oxidation within the depositional environment can, however, lead to anomalous O/C ratios when compared to the general level of maturation (e.g. Kendrick, 1979). The hydrocarbons which are generated during organic metamorphism depend upon the nature of the original kerogens. "Type I" kerogens generate oil, "Type II" oil and wet gas, and "Type III" dry gas; dry gas generation is characteristic of all kerogens at late stages of maturation (Staplin, 1969; Correia, 1971; Burgess, 1974; Lijmbach, 1975; Dow, 1977).

PART TWO : METHOD AND PRACTICE

INTRODUCTION:

All the palynological slides examined during this research project were prepared by the laboratories of Robertson Research International according to standard palynological techniques (see Barss & Williams, 1973). A small batch of samples was, however, processed by me in order to familiarise myself with the technique. The mineralic matrix of the samples was digested by the use of concentrated HCL and HF and part of the resulting kerogen concentrate used to produce a kerogen slide, while the remainder was in some instances oxidised with HNO_3 to provide a 'palynostratigraphic' preparation relatively freee of amorphous organic matter. Oxidised preparations were particularly necessary to accurately determine the palynomorph assemblages in the organic-rich sediments of the type Kimmeridge Clay. For the study on the Phillips 16/17-4A well and the Kimmeridge Clay Formation of the Piper Field,small numbers of previously prepared palynostratigraphic slides were loaned to me by Robertson Research.

Portions of the kerogen isolates were used to produce slides consisting of two strew mounts, one representing the unsieved kerogen isolate and the other that part retained by sieving with a 16µ nylon mesh. The sieved part of the unoxidised kerogen slides was used to determine the general character of the organic matter in the sample by making a minimum of 500 counts of the various particle species observed during several traverses over different parts of the slide. Following the counting the kerogen slides were examined under a Zeiss ultra-violet fluorescence microscope in order to provide a qualitative assessment of the character of the amorphous organic components. The count data (recorded on an automatic point counter device) was used to calculate the percentage particle abundances. For the most suitable samples the sieved parts of the palynostratigraphic slides were used to make a detailed study of the proportions of various morphological or otherwise significant groups of palynomorphs. Counts were recorded manually on a standard format data sheet and then converted into percentages of the total palynomorph count.

KEROGEN CATEGORIES

In its modern general sense the term "kerogen" is used to describe the whole of that "fraction of sedimentary organic matter which is insoluble in the usual organic solvents" (see Durand, 1980, p.25). As this is a palynological rather than geochemical investigation the above definition is for practical purposes equated with the palynological concept of kerogen as that fraction of the sedimentary organic matter which is insoluble in hydrochloric and hydrofluoric acids. The combination of a sediment's kerogen and palynomorph characteristics are here referred to as its "palynofacies" in accordance with the original definition of the term by Combaz (1964).

At present there is no single widely accepted terminology for dispersed sedimentary organic matter, but some of the more commonly used terms and their approximate inter-relationships are given in Table 3.1. These terms can be divided into two 'schools', the first derived from the long-standing discipline of coal petrography and based on reflected light microscopy, and the second development by palynologists using transmitted light microscopy (columns A to D in Table 3.1). Although the coal petrographic terminology may be regarded as the more objective system, the palynological method has many practical advantages and has grown in importance since the late 1960's, generating its own more general terminology in the process. In practice, the palynologist experienced in palynofacies studies can equate what he observes in transmitted light to the appropriate coal maceral with a fair degree of reliability.

My approach to palynofacies investigations has been a quantitative one utilising percentage data on the patterns of distribution of various kerogens and palynomorphs groups. Such a quantitative approach allows a much greater appreciation of spatial, facial and stratigraphic trends than is permitted by subjective general visual assessments (e.g. Habib, 1979). The main drawbacks of this approach are the tedious nature of the counting procedures and the sometimes misleading impression that can be conveyed by the apparently unequivocal numerical data (see later). For counts on kerogen slides the author uses a three-fold division with a total of ten categories which were selected on practical grounds for the rapid classification of any given kerogen particle. The three main divisions are palynomorphs, plant debris (phytoclasts) and amorphous organic matter (A.O.M.), of which the latter two normally comprise the bulk of the organic matter in organic-rich sedimentary rocks.

From experience I have found it useful to make some degree of differentiation of the palynomorphs during general kerogen counting, dividing them into three categories: sporomorphs, plankton, and undifferentiated forms (the latter being mainly simple sphaeromorphs and probably predominantly sporomorphs). Four groups of plant debris (phytoclasts) are recognised including fragments of cuticle, well lignified trachael material (from sclerenchyma and xylem tissues), poorly lignified, irregular fragments (probably from collenchyma and parenchyma cortex tissue) and carbonised, charcoal-like 'black wood'. The three remaining categories are amorphous organic matter, chitinous microforaminifera (or foraminiferal test linings) and 'palynodebris'. The term 'palynodebris' is here used for translucent, pale (usually yellowish) thin, membranous material which probably derives from the mechanical breakdown of palynomorphs and thin cuticular debris. It is felt that this usage of the term is more consistent with its derivation than that proposed by Manum (1976) and as subsequently adopted by Habib (1979).

It is necessary that the reader should fully appreciate the nature of the data calculated from the kerogen counts. The values given in the appendices and text are percentage particle abundances and as such cannot be simply translated into percentage proportions of different types of organic matter. There are several sources of error and variation which effect the data and its interpretation. Firstly the identification of particular particle types is not always easy and is often complicated by differing degrees of degradation and preservation. This introduces a degree of subjectivity into the kerogen classification which results in palynological kerogen typing being reliant on the self-consistency of the operator (e.g. see discussion in Batten, 1981). Secondly the data is not size-classed and therefore the different particles contain differing amounts of organic matter as well as organic matter of differing composition. However, comparisons between the sieved and unsieved parts of the kerogen slides suggest that sieving has little significant effect on the perceived character of the organic matter since the finer debris is volumetrically insignificant and mainly liberated from the margins of A.O.M. particles during the preparation treatment. During the counting a conscious attempt was made to record only the modal range of particle sizes present while allowing for the different dimensions inherent in contrasting particle types. The majority of particles counted lie in the range 30-100µ, with the larger particles (especially A.O.M.) having been broken up by short ultrasonic treatment during the preparation procedures. Since particle abundances are expressed as percentages it is necessary that the kerogen typing is calibrated with organic carbon values in order to assess the variations in net organic matter abundance observed between the samples. By its very nature percentage particle abundance data should only be compared with like data (preferably collected by the same operator) and used only for the detection and anlysis of trends.

Amorphous organic matter is frequently the dominant type of kerogen observed in organic-rich marine sedimentary rocks. It follows that it is crucial that this material be correctly identified and its nature assessed as accurately as possible. A.O.M. has often been referred to loosely as 'amorphous sapropel' but its composition can range from sapropelic to humic and therefore the general term A.O.M. is far more preferable. Collinite is the main form of humic A.O.M.. It is an amorphous cell-filling vitrinite maceral formed during the early diagenesis (humification and gelification) of plant debris and forms brown-coloured, irregularly shaped, homogeneous particles without internal structure. With experience it is possible to distinguish collinite from the more normal marine-derived and potentially sapropelic varieties of A.O.M. The latter shows a variable appearance and composition depending on its original composition and state of degradation and is invariably heterogeneous, forming an organic matrix to various mineral and organic inclusions including framboidal pyrite, phytoclasts and palynormorphs. When relatively undegraded it usually appears as irregular shaped particles of speckled yellow-brown gel which may exhibit a variably granular or 'clotted' character depending upon the abundance and nature of the included materials (e.g. see Timofeev & Bogulyubova, 1979; Combaz, 1980). During progressive degradation the amorphous particles becoming paler in colour (eventually grey), less cohesive and tend to break up during the preparation procedures, becoming disseminated across the slide and liberating their inclusions.

Although there are visual clues to the nature of A.O.M. which can be obtained using transmitted light, more objective optical determinations can only be made using ultra violet fluorescence. When examined under ultra violet light, exinitic and other lipid-containing organic matter exhibits variable degrees of green or yellow fluorescence while humic materials remain unaffected (e.g. see Teichmüller & Ottenjann, 1977; Powell et al. 1982 and Cook et al. 1981). This allows a very rapid simple way of estimating the sapropelic content of A.O.M. Although the general A.O.M. matrix may sometimes show dull fluorescence most of the sapropelic content normally occurs as variable amounts of included exinitic (liptodetrinite) or alginitic (bituminite) debris which is undetectable in plane light. As this material becomes degraded it becomes less and less fluorescent. When coupled with ultra violet fluorescence observations, I have found that percentage particle abundance data is usually sufficient to determine the nature of the total kerogen as a Type I, II or III sensu Tissot (1977). It should be noted that low H/C ratios (or hydrogen indices) are produced by increasing maturation, high contents of plant debris and 'syndepositional' degradation of marine materials, and that optical kerogen studies are, therefore, a vital adjunct to geochemical investigations.

The following diagnostic notes summarise the characteristics of the kerogen categories utilised in this thesis. Their approximate equivalents with pre-existing schemes may be determined by examining Table 3.1. The classification used was developed in the first few months of the project and subsequent experience has proven its usefulness. Additional categories were only required for the Kimmeridge oil shales in Dorset (see Chapter Five).

(a) Macrophyte debris

Wood (Wu): light to dark brown coloured, elongate angular particles commonly showing bordered pits (tracheal material) and dark, parallel-sided or fusiform thickenings giving a banded appearance. May appear splintery when physically degraded. These particles correspond principally with the tellinite coal maceral, but in the absence of reflectance data it is clear that they probably range from vitrinites to semi-fusainite.

Wood (Wd): light to dark brown coloured particles tending to be more equiangular than elongate and having irregular, often rounded but complex, composite outlines and usually lacking any internal structure.

The nature of the outlines generally indicates these particles are fragments (biochemically degraded?) but the more rounded material may represent floculated humic gels. Interpreted as vitrinite to semi-fusainite material derived from poorly lignified plant tissues.

Inertinite (I): Opaque elongate to equigranular particles which are generally significantly smaller than associated Wu or Wd particles and lack visible internal structure. Care was taken to exclude particles which were opaque because of their thickness and to prevent misidentification of pyrite or other opaque heavy minerals remaining in the kerogen preparation.

Cutinite (cu): Strongly cellular, pale yellow, translucent, platey particles of varying overall shape.

(b) Palynomorphs

Undifferentiated (UP) principally simple sphaeromorph palynomorphs (predominantly pollen). May include a small proportion of isolated foraminiferal chambers - particularly in more degraded or A.O.M.-rich samples where identification is not always definite

Plankton (P): dinocysts, acritarchs and prasinophycean algae.

Miospores (M): bisaccates, deltoid fern spores and *Classopollis* pollen.

(c) Amorphous organic matter (A): see descriptions above.

(d) Others

Microforaminifers: (F) self-explanatory (principally planispiral or trochospiral forms, more rarely serial or biserial)

Palynodebris (pd): translucent, pale (usually yellowish), thin, membraneous non-cellular material (exinite or cutinite coal maceral groups).

PALYNOMORPH CATEGORIES

Palynomorph counts were performed on the most suitable sieved kerogen slides and, where available, sieved, oxidised palynostratigraphic preparations. The selection of samples was primarily controlled by A.O.M. abundance, palynomorph density and the stratigraphic or facies significance of the sample point. The number of palynomorphs counted per slide was usually between 200 and 500 (mean >400, generally at least half the slide) and was partly dependent on palynomorph density and the quality of the preparation. Identification of palynomorphs was assisted by the preparation of a photographic 'atlas' culled from the appropriate texts on British Late Jurassic palynostratigraphy. The following categories of palynomorphs were used during the counts and recorded on a standard format data sheet for each slide (the capitalised letters are the same as those used in various data appendices). Some categories included on the data sheet were found to be unnecessary and are not included below.

A. Proximate dinocysts e.g. Gonyaulacysta, Acanthaulax

B. Cavate dinocysts e.g. Endoscrinium, Hexagonifera (monocavates)

C. Proximo-chorate dinocysts e.g. Cyclonephelium, Tenua

D. Chorate dinocysts e.g. Systematophora, Oligosphaeridium

E. Peridinean dinocysts e.g. Pareodinia, Metrelytron

F. Short-spined acanthomorph acritarchs e.g. Micrhystridium inconspicium

G. Long-spined acanthomorph acritarchs eg. Scolisphaeridium, M. fragile

H. Prasinophycean phycomas e.g. Tasmanites, Pterospermella, Cymatiosphaera, Crassosphaera

I. Undifferentiated dinocysts (identification impossible beyond this level; possibly A, B, C or E).

J. Bisaccate pollen e.g. Alisporites, Parvisaccites, Abietineapollinites

K *Perinopollenites*

L Simmple deltoid pteridophyte spores e.g. *Cyathidites*, *Dictyophyllidites*

M Zonate pteridophyte spores e.g. *Callialasporites*

N Ornamented or thick-walled deltoid pteriodophyte spores e.g. *Cicatricosisporites*, *Appendicisporites*, *Densoisporites*

O Undifferentiated palynomorphs (mainly sphaeromorph pollen but including any unidentified bodies as well)

	Particulate organic matter (POM)			Kerogen							
	Category	Source	Constituents	Coal maceral group	Coal maceral	A	B	C	D	Kerogen type: General	Kerogen type: E
Structured	Palynomorphs	Plankton	Blue-green algae and prasinophycean phycomas	Exinite or Liptinite	Alginite	Vegetable (MOV)	Algal	Phyrogen	Aqueous	Sapropelic (oil prone)	I
			Dinocysts and acritarchs		Sporinite		Herbaceous				II
		Sporomorphs	Spores and pollen						Spores and pollen		
	Phytoclasts	Higher plant debris (macrophyte tissues)	Cuticle		Cutinite				Terrestial		
			Ligno-cellulosic material	Vitrinite	Telinite	Tracheal (MOT)	Woody	Hylogen		Humic (gas prone)	III IV
			Carbonized material	Inertinite	Fusainite	Lignitic (MOL)	Coaly	Melanogen	Charcoal		
Unstructured	Amorphous organic matter (AOM)	Polygenetic and heterogeneous	Organic aggregates and flocs formed from dissolved organic matter, products of bio-chemical degradation of POM, and faecal pellets	Variable, but often with components of the following macerals:-	Collinite	Colloidal (MOC)	Amor-phous	Amor-phogen	Amor-phous	Variable: Humic	III
					{Liptodetrinite Bituminite}					Variable: Sapropelic	II/I

Tables 2.2/3.1 A classification of particulate organic matter found in marine sediments and the approximate equivalence of the most commonly used kerogen terms. Column A after Correia (1971) and McLachlan & Pieterse (1978); B after Staplin (1969) and Hunt (1979); C after Bujak et al. (1977); D after Masran & Pocock (1981); E after Tissot & Welte (1978). Coal maceral terminology (much abbreviated) after Stach et al. (1975). Fusainite should read fusinite.

CHAPTER FOUR

Sedimentological and palynological analysis of five cores from the Piper Field (Occidental Petroleum), northern North Sea

INTRODUCTION

This chapter contains the findings of a combined sedimentological-palynofacies study on Piper Formation cores from Wells 15/17-4, 5, 6, 7 and 8A (see Fig. 4.1). Some 877 feet of core were initially logged and described and 61 palynofacies and 60 petrographic slides were subsequently examined. This material provided the opportunity to examine the palynofacies characteristics of clastic sediments deposited in a relatively shallow (mixed layer) environment. Within the limitations of the study I have also attempted to appraise the published palaeo - environmental interpretation of the Piper reservoir as a "series of stacked and possibly imbricated barrier bar and other littoral and shallow shelf marine sandstone bodies (Williams et al. 1975, p.1593).

The lithostratigraphic units adopted in this chapter are the informal ones used in Williams et al. (1975). A considerably more detailed subdivision of the Piper Formation based upon analysis of the geophysical logs has been presented by Maher (unpublished, 1980, 1981a, 1981b). The older nomenclature was, however, considered more appropriate to the limited scope and core-based nature of this particular study. Chronostratigraphic data on the five wells was provided by Dr. L.A. Riley based on his review of the dinocyst floras (note the absence (?) of Callovian compared with the published accounts). All well depths cited are given in feet, and unless otherwise indicated, are uncorrected core depths. The electric log, lithological and chronostratigraphic correlation of the five wells and the distribution of the cored intervals are shown in Fig. 4.2.

Previous published accounts of the geology of the Piper field include Conner & Kelland (1974), Williams et al. (1975), Maher (1980, 1981a, 1981b) and Turner et al. (1984). Deegan and Scull (1977) formally designated the type section of the Piper Formation as the interval between 8444' and 8746' (below KCB) in the 15/17-4 well. Subseqent coring has, however, indicated that this well was not the most

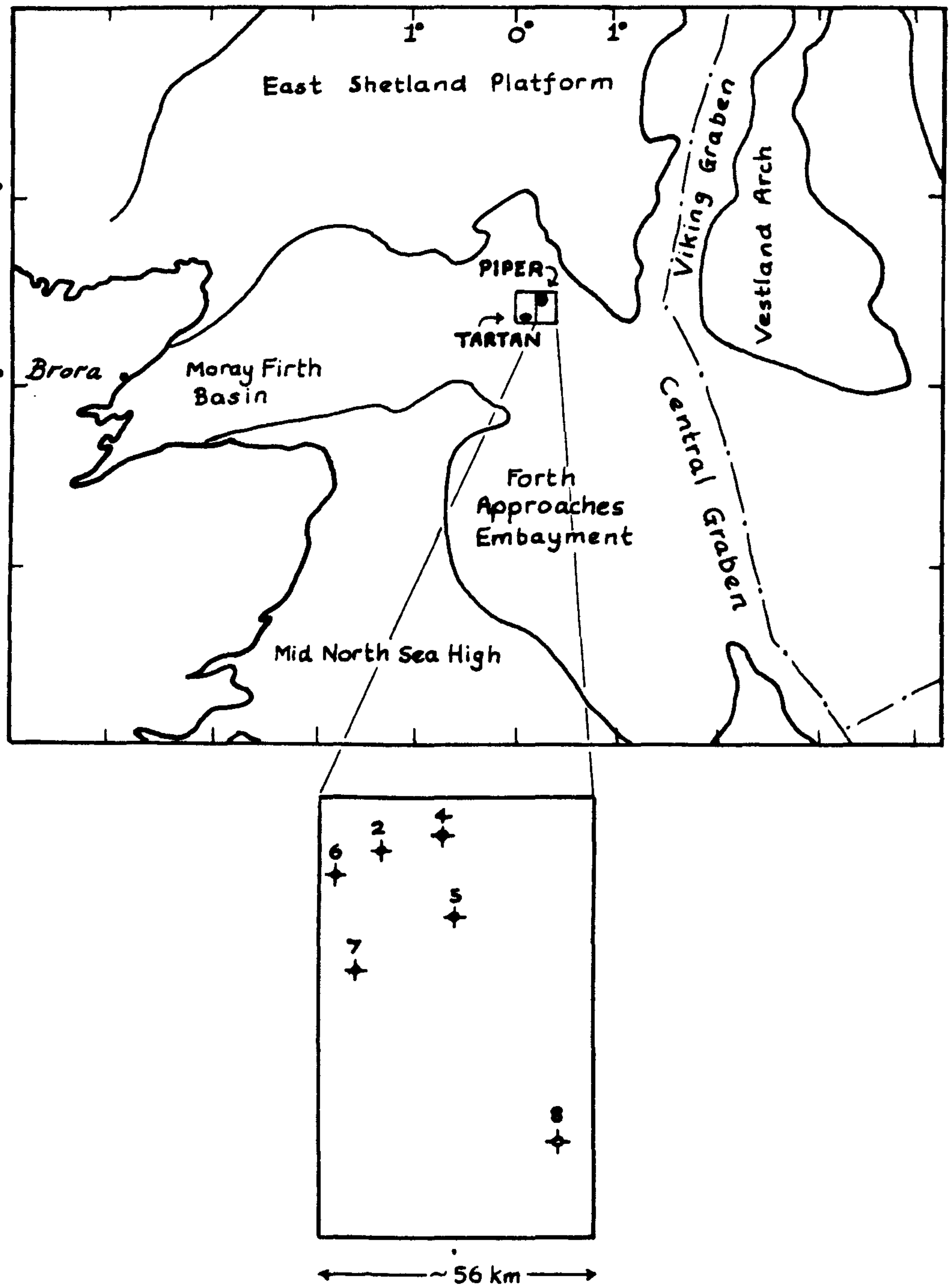


Fig. 4.1 Location of Block 15/17 and the Piper wells discussed in text

suited for selection of the stratotype (Conrad Maher, personal communication). Regional studies have indicated the greater extent of the Piper Formation and its analogues (e.g. Maher, 1980; Turner et al. 1984) and a remarkable regional continuity and uniformity of the general electric log signature (Conrad Maher, personal communication).

DESCRIPTION OF THE LITHOLOGICAL UNITS, WELL 15/17-4 TO 7

(i) The "non-marine shale"

This unit was only cored in the 15/17-4 well (see Table 4.1) where approximately ten feet of sediment was recovered. The top and bottom part of the cored interval consists of a medium to dark grey, unfossiliferous, non-calcareous mudstone containing pyrite and common to abundant plant debris. The centre part (8738.75 - 8743') consists of a pure, pale grey, waxy, non-fossiliferous, non-calcareous mudstone with a sub-conchoidal fracture. The lowermost 6" contains abundant plant debris and possible rootlet structures. XRD analysis of this pale grey mudstone unit (using an oriented sample technique) indicated a composition consisting of practically pure kaolinite. Three samples were submitted for organic carbon determinations, two from the normal mudstone (average 1.6%) and one from the pure, pale grey, mudstone interval (org.C 0.4%). Macrofossils and bioturbation were absent from this interval.

(ii) The "organic-rich shale"

This unit was also only cored in the 15/17-4 well where it is 35.75' thick. It is composed of silty to relatively pure, dark to medium grey, shaley, often pyritic, calcareous to non-calcareous

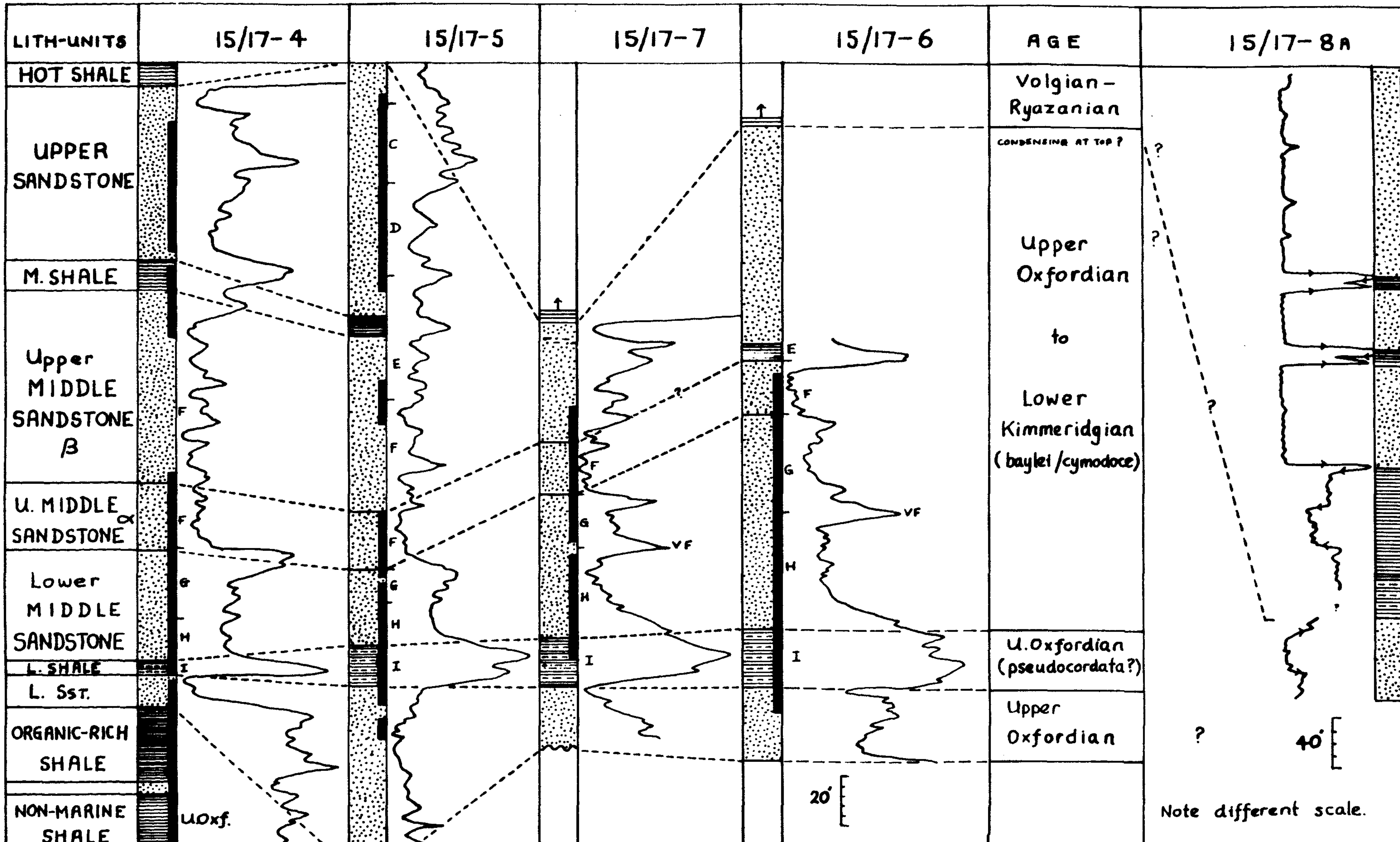


Fig. 4.2 Correlation of Piper wells by lithology, gamma ray log and age. Cored intervals indicated by vertical bars. Letters denote E-log units according to Maher (1980).

TABLE 4.1 Distribution of lithological units in cored intervals (uncorrected core depths in feet)

LITHOLOGICAL UNITS	15/17-4	15/17-5	15/17-6	15/17-7
Non-marine shale	8746 to 8735.5			
Organic-rich shale	8735.5 to 8715			
	8715 to 8699.5			
Lower sandstone	8699.75 to 8687	8588.5 to 8582	8720 to 8709.5	
		8575 to 8563		
Lower shale	8684 to 8680	8563 to 8552	8709.5 to 8680	9376 to 9359.5
Lower middle sandstone	8680 to 8632	8549 to 8511	8679.75 to 8625	9359.5 to 9321
			8623 to 8593	9315 to 9296
Upper middle sandstone (α)	8632 to 8600	8511 to 8942	8593 to 8572	9296 to 9282
		8491.5 to 8491		
Upper middle sandstone (β)	8584 to 8582	8451 to 8431		9282 to 9288
	8551 to 8528			
Middle shale	8528 to 8524			
	8516 to 8513.5			
Upper sandstone	8513.5 to 8490	8402 to 8384		
	8485 to 8460	8382 to 8346		
		8343 to 8315		

CORED INTERVALS	15/17-4	15/17-5	15/17-6	15/17-7
Core 1	8460 to 8485	8315 to 8343.1	8572 to 8623	9258 to 9276.5
2	8490 to 8516	8346 to 8352	8625 to 8679.5	9278 to 9315
3	8524 to 8551	8352 to 8383	8680 to 8720	9321 to 9376
4		8384 to 8402		
5	8582 to 8584	?		
6	8600 to 8630	8431 to 8451		
7	8630 to 8661	8461 to 8462		
8	8661 to 8684	8491 to 8491.5		
9	8687 to 8714	8492 to 8522		
10	8715 to 8746	8522 to 8549.75		
11		8552 to 8575		
12		8582 to 8588		

mudstone. Common irregular lenticules of grey to grey green, sometimes calcite cemented, silt to fine sand produce a mottled, bioturbated appearance with only relict bedding; where these are absent mottling is less conspicuous or possibly absent. Occasional microfaults and pyritic and calcareous concretions are present. Recognisable plant debris varies from relatively rare to quite abundant. Down to 8712.5' macrofaunal remains are common and occasionally abundant, including the bivalves Thracia, ?Trautscholdia and unidentified oyster debris. Apart from the general mottling and sporadic presence of pyritised horizontal tubes no particular burrow morphologies were identified. Five samples were submitted for organic carbon determinations and average 4.1% org.C. At the base of this shale unit is a four foot thick, poorly sorted, mottled, dirty, fine to medium grained sandstone containing common pyrite and carbonaceous debris.

Thin sections from this unit showed varying ratios of grains to matrix (the latter varying from 15-80%). Grains were predominantly coarse silt to very fine sand grade, angular to subrounded and composed of quartz; occasionally rounded, coarse quartz grains were also present. All the thin sections showed rare (1-3%) pale yellowish-green pellets of 'glauconite' (or possibly chamosite) and patchy calcite cementation was sometimes evident. One section showed occasional quartz grains with small scale overgrowths.

(iii) The "lower sandstone"

Cores through this sandstone were obtained from wells 15/17-4, 5, and 6 (see Table 4.1). The whole of this unit was cored in the 15/17-4 well where it appeared as a rapidly coarsening upward sequence overlying the 'organic-rich shale'. The sandstone passes upward from silt into a medium to coarse grained deposit, becoming increasingly poorly sorted and richer in pebble and granule grade material. The unit is generally mottled in this well and is topped by about two feet of argillaceous, medium-coarse, poorly sorted, bioturbated sandstone containing pebbles

($\leq$ 2 cm) and belemnite fragments. In the 15/17-5 well the cored interval forms the upper part of a coarsening upward sequence (determined from gamma log trend) and is a medium-coarse, poorly sorted sandstone with common pebble and granule grade material. Although generally mottled in appearance feint low angle planar bedding is sometimes visible. In the 15/17-6 well this unit is represented by a medium to fine grained sandstone with scattered rare granule material and occasional pebbly or granule-rich layers or laminae; low angle, planar bedding occurs throughout the cored section. Patchy calcite cementation occurs in the lower sandstone in the 15/17-4 well.

(iv) The "lower shale"

Cored sections of the lower shale were recovered from wells 15/17-4, 5, 6 and 7. Despite what the informal name of this unit would suggest the modal grain size is often silt rather than clay. In the 15/17-4 well the maximum thickness of the lower shale is seven feet, of which the upper four feet has been cored and consists of sandy mudstone overlain by argillaceous siltstone. The silty part contains lenticules of fine sand which may indicate relict ripple bedding. Bioclastic debris is common to abundant and includes Trautscholdia, Dentalium, very poorly preserved ammonite fragments and remains of a larger unidentified bivalve. The lower part of the core has been thoroughly broken up and other macrofossils may have been removed. The 'lower shale' is thicker in the 15/17-5 well ($\sim$13') but is split by a silty, fine grained sandstone whose position is clearly indicated by the gamma ray log (see Fig. 4.2 and log in Appendix). Silty, carbonaceous shale is the dominant lithology with varying amounts of medium-coarse grains as lenticules or mottles. No macrofossils were observed but macroscopic plant debris is common and some shows a 'coalified' appearance.

Out of the four wells the 'lower shale' is thickest in the 15/17-6 well where it consists of about thirty feet of grey, carbonaceous siltstone which is generally irregularly bedded to mottled, but occasionally shows relict ripple (?) bedding. No macrofossils were observed. Apart from the general mottling, some occasional Chondrites burrows were noted, along with sporadic low angle or horizontal, white-rimmed 'rind burrows' up to 1 cm in diameter (c.f. Chamberlain, 1978) and a few examples of Teichichnus. In the 15/17-7 well the 'lower shale' is a micaceous, variably carbonaceous, siltstone to silty, very-fine grained sandstone and is generally similar to the sequence in 15/17-6.

Eleven samples from the 'lower shale' were submitted for organic carbon analyses and yielded an average of 2.6% org.C. Little of note was observed in thin sections from this unit; rare glauconite (or possibly chamosite?) pellets were common accessory components (occasionally exhibiting a crude concentric, almost oolitic structure) and some bioclastic debris was also noted.

(v) The 'middle sandstone'

This unit is a coarsening upward sandstone body in all the wells examined but differences in modal grain size were noted. The most complete cored sections are those from wells 15/17-6 and 7 and although appreciable gaps in the core coverage occur in wells 4 and 5, examination of the geophysical logs suggests that the cored intervals are fairly representative for the unit. From studying the gamma ray and self potential (S.P.) logs it is apparent that the 'middle sandstone' can be divided into at least two parts, an upper and a lower division. In the 15/17-4 well the boundary between this upper and low division coincides with an anomalous gamma ray peak whose cause has been identified as a concentration of monazite grains in the heavy mineral suite (Trewin in Maher, unpublished). A similar gamma ray anomaly occurs in the same position in wells 15/17-5, 6 and 7 and probably also 15/17-3 (see Fig. 6 in Connor & Kelland, 1974). The S.P. log shows no variation in response across this boundary.

(a) The lower division

In 15/17-4 the lower division of the 'middle sandstone' consists of variably silty, fine grained sandstone with occasional lenses of coarser material. It is generally bioturbated and mottled in appearance but does contain some recognisable burrows (e.g. Ophiomorpha) and silty or carbonaceous layers sometimes indicate relict low angle (10-15^{o}) planar laminae or erosional surfaces. Coaly carbonaceous fragments and bioclastic debris (including belemnites) become more common towards the base where patchy carbonate cements are developed. The sequence in the 15/17-5 well is essentially the same. In wells 15/17-5 and 7 the top and basal parts of the lower division are fine grained sandstones separated by very fine grained intervals (14' and 10' thick respectively). The upper fine grained parts are structureless but exhibit good Ophiomorpha burrows (and a few biogenic escape structures in 15/17-6); the lower fine grained units are similar but become more carbonaceous downward, grading into the 'lower shale'. The intermediate very fine grained sandstones are variably mottled or cross-bedded. Three samples from the basal (more carbonaceous) part of the lower division in 15/17-7 gave an average organic carbon value of 0.8% org.C.

(b) The upper division

The contact between the upper and lower division is very sharp. Examination of the gamma ray log shows that the basal part of the upper division (the lowermost 15-30' denoted as α in Fig. 4.2) is characterised by a more even response and the remainder (β in Fig. 4.2) by an irregular 'saw-tooth' pattern; no variation is observed in the SP log. This effect is not observed in well 15/17-6 where the upper division is at its thinnest and where only the lowermost more regular log-unit (α) is apparently developed. The interpretation of this log pattern is problematical because of the poor core coverage in the upper division and the absence of an undisputable 'middle shale' interval in well 15/17-7. If the 'middle shale' in 15/17-7 is represented by the five and a half feet of very-fine-grained sandstone ($\sim$9820'), the top part of the core represents the 'upper sandstone'; if it does not the top of the core represents the top part of the upper division of the 'middle sandstone'. It may be supposed that the saw-tooth pattern at the top of the upper division (β) may be produced by the sporadic concentration of monazite grains within particular laminae or scour pockets (such as occur between 8547-8551' in 15/17-4).

In 15/17-4 the basal part of the upper division (α) is predominantly medium to fine grained and in the core interval between 8600 and 8640', contains common granule or pebble-rich erosional scour surfaces and coarse to granule grade laminae picking out low angle (10-20^{o}) bedding. The scour surfaces often define fining upward units, passing upward from pebbly bases to better sorted, structureless or planar laminated (more rarely rippled) sands with occasional tubular burrows. Bedding traces are generally picked out by carbonaceous, silty or coarse laminae. In 15/17-5 coarse layers in the medium to fine sand occasionally define planar cross lamination but the upper division in this well is dominantly mottled or structureless. The modal grain size in 15/17-6 is medium to coarse with abundant granules and pebbles; 10-20^{o} planar bedding is common and occasional trough cross-bedding is also present. The upper division in 15/17-7 is essentially the same, fine to medium grained with the ubiquitous scours, fining upward units like those in 15/17-4, and between 9284-9285-5' and 9286-9286.5' alternations of medium-coarse and very coarse-granule laminae strongly reminiscent of 'laminated storm sand layers' (c.f. Reineck & Singh, 1972 and see later).

Petrographic samples from the 'middle sandstone' yielded little of particular interest. Small amounts of rounded glauconite grains ($\leq$ 1%) were ubiquitous. Patchy sparite cements occurred in some thin sections (from both the upper and lower divisions) and very small amounts of authigenic kaolinite pore clay was also present. Occasional minor quartz overgrowths were observed and in some samples feldspar grains (0.5-4.0%) consistently exhibited overgrowths while quartz grains did not (e.g. samples SOP 35, 35A and 49C). The rounded pebble to granule grains in in the upper unit (often pale green to pinkish in hand specimen) were polycrystalline metaquartzite (including some gneiss or quartzitic mica schists) and constituted 2-20% of the sandstone. The 'middle sandstone' unit is dominantly well to moderately sorted, generally uncemented, often friable and has good porosity making it the best reservoir rock in the Piper Field (see Williams et al. 1975).

(vi) The "middle shale"

This unit was only cored in well 15/17-4. The top of the 'middle sandstone' in this well is a two foot thick, poorly sorted, burrow mottled, medium grained sandstone rich in oyster debris, which grades up into the 'middle shale', a fossiliferous, shaley, sandy, silty mudstone containing whole and fragmented oysters (Deltoideum?). Two samples gave an average organic carbon of 2.4%.

(vii) The "upper sandstone"

This sandstone was cored in wells 15/17-4 and 15/17-5. In 15/17-4 the top of the upper sandstone is very fine grained (8460-8476'), and generally mottled with a few clear burrows; the remainder is fine grained, more-or-less homogeneous or mottled with irregular relict bedding and contains four thin horizons rich in bioclastic debris (oyster>belemnites> gastropods). Patchy calcite cements occur in the lower part and carbonaceous material is occasionally fairly common. The burrows observed in this unit are predominantly ± horizontal (Ophiomorpha and/or ?Planolites) and often show burrowfills slightly coarser than the matrix. In 15/17-5 the 'upper sandstone' is medium to fine grained, dominantly mottled or structureless with occasional hints of low angle laminations. The lower half of the cored interval is less well sorted than the upper part and contains a one foot thick non-calcareous, shaley, silty, mudstone with fine grained sand lenticules (8389-8390'). Examination of thin sections from samples of the 'middle sandstone' showed little of particular note; rare authigenic kaolinite, occasional minor quartz overgrowths, small amounts of glauconite (and chamositic? 'ooliths' in muddier sediments) and some patchy sparite cementation was recorded.

(viii) The Kimmeridge Clay Formation

No cores were taken from this unit in the five wells examined in this study. The following comments are taken from an unpublished report by C.E. Maher of Occidental Petroleum (see also Maher, 1980):

> "Prior to the deposition of Kimmeridge Shale there was probably a long hiatus as indicated by a six foot zone of intensely bioturbated sand grading to silt and containing abundant shell material and glauconite in (well) P10. This long hiatus was accompanied by minor erosion of Piper sands and redeposition as "Unit IV". Unit IV is 0-31 m thick, argillaceous silty sand and occurs around the present crest of Piper". "Glauconitic sand and silt of the Kimmeridge Shale in P10 contains a fauna representative of fully marine conditions and grades upwards into a dark grey to black shale, laminated with no bioturbation and with silt in discrete laminae. Fish scales, fish debris, wood fragments, belemnites and lack of bioturbation indicate deeper, more stagnant marine conditions".

LITHOLOGICAL DESCRIPTION OF CORED INTERVAL, WELL 15/17-8A

The cored sandstone sequence in the 15/17-8A well is unlike that developed in the rest of the Piper Field and has therefore been described separately (see Fig. 4.2 and Table 4.1).

(i) Interval 12423' to 12490'

The cores from this interval consist of hard, grey, predominantly medium grained, moderate to poorly sorted sandstone. The sandstone, like most others in the Piper cores, is a quartz arenite (sensu Pettijohn, 1975). The interval is essentially uniform and structureless but a few low angle, planar, carbonaceous laminae occur in the lower part which is medium-fine grained and less well sorted. No bioturbation or bioclastic debris (with the exception of one possible small oyster fragment) was observed.

(ii) Interval 12490' to 12578'

This interval is composed of predominantly medium grained, moderately-poorly sorted sandstones arranged in a series of 'graded' units (seven of which are complete and have an average thickness of 11'). The grading is essentially represented by a downward decrease in sorting and is not usually reflected in the modal grain size of the matrix (although two units did show a shift from medium - coarse to medium sandstone and from medium to fine sandstones respectively; i.e. fining upward trends). The lower third of these graded units contain common but dispersed pebbles ($\leq$ 1.5 cm in diameter). The typical features of these units are compiled and represented in Fig. 4.3. Patchy calcite cementation occurs in the lower part of the interval.

(iii) Interval 12578' to 12894'

No cores were taken from this interval. The geological composite log records about 160' of grey, very fine to fine, poorly sorted, calcite cemented sandstones with two (approximately 10' thick) 'hot' shale intervals overlying about 125' of dark, brownish-black, silty, carbonaceous, calcareous 'hot' shale. Twenty feet of sandstone (in continuity with the unit described below) occurs at the base.

(iv) Interval 12894' to 12946'

This sequence consists of predominantly medium grained, poorly sorted sandstone, which has a rather mottled appearance. At least seven 'graded' units similar to those described above are present, differing only in that coarse to granule grade material in the medium grained matrix is often arranged in crude layers and that the poorly sorted bases are often carbonaceous and contain irregular silty agrillaceous lenses. A few rare tubular burrow sections were observed in the lower part. Pyrite is common.

As well as differing in their general organisation the sandstones in 15/17-8A are petrographically distinct from those in the other wells described here. They are generally less well sorted, more closely

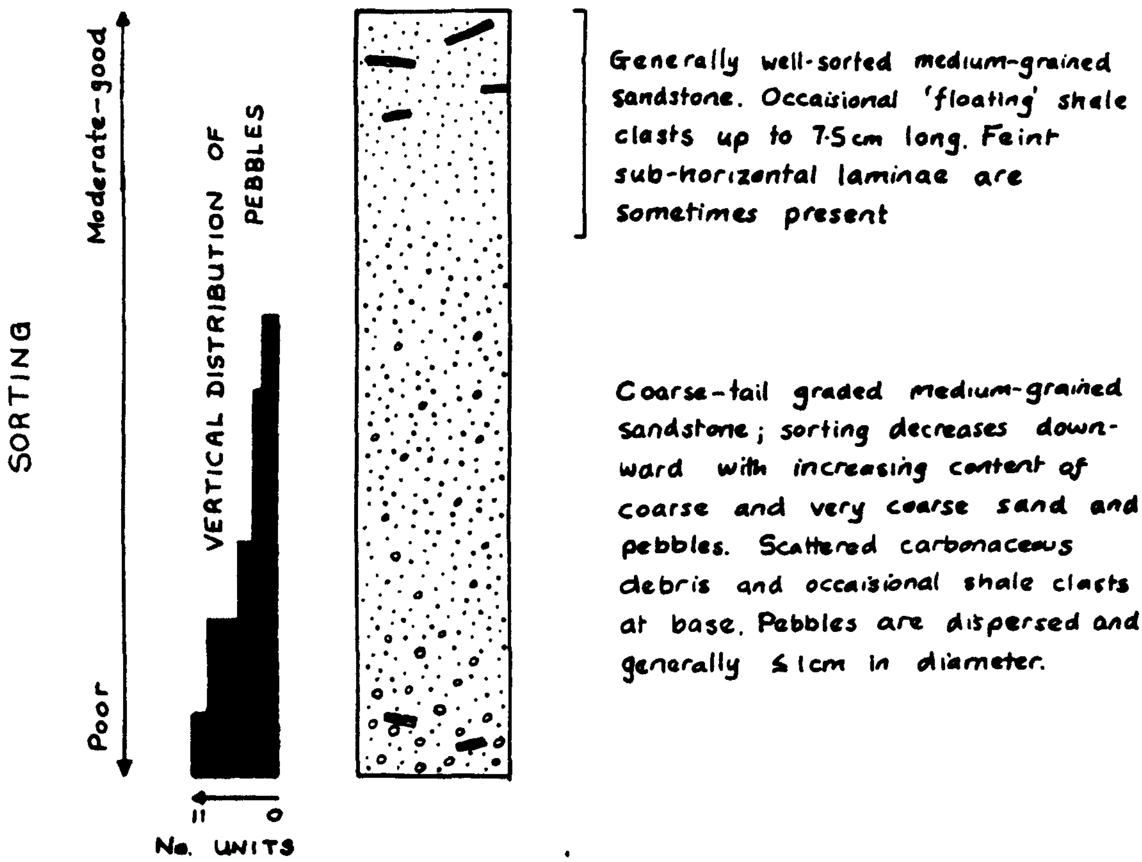


Fig. 4.3 Main features of graded units interpreted as grain flow-liquified flow deposits in cores from 15/17-8A.

packed and dirtier in appearance. Moderate and occasionally quite strong pressure solution is typical (as indicated by the abundance of concavo-convex and sutured grain contacts). Quartz overgrowths are generally common and there is little or no porosity. Clay lined, serrated styliolitic laminae were observed in the cores. Authigenic kaolinite pore clays are occasionally abundant, especially in the lower unit, ((iv) above) where pyrite is also common (≤ 10%). Very coarse to pebble grade material consists of rounded to subrounded polycrystalline metaquartzite grains and/or grains of sandstones showing moderate to strong pressure solution; mica-schist or gneiss pebbles were sometimes identifiable. The coarser intervals range in composition from quartz arenites to sublithic quartz arenites.

PALYNOLOGICAL OBSERVATIONS ON CORE SAMPLES FROM WELLS 15/17-4 TO 7

(a) Kerogen characteristics

The average kerogen composition of the 61 samples from the Piper cores is shown in Table 4.2. The kerogen is dominated by phytoclastic (i.e. woody) debris (averaging 57 %) and autochthonous organic matter (plankton, foraminiferal test linings and non-collinitic A.O.M.) constitutes less than 10% of the toal P.O.M. The only well in which all the lithological units of the Piper Formation were cored is 15/17-4; the most commonly cored units, and hence those best sampled, are the 'lower shale' (13 samples from 4 wells) and the 'middle sandstone' (25 samples from 3 wells).

Table 4.3 shows the average up sequence values for the proportions of wood, collinite, cuticle, palynomorphs and autochthonous organic matter in well 15/17-4. The transition from the 'non-marine shale' to the 'lower shale' (marine) is shown by a reduction in the total amounts of wood, collinite and cuticle and an increase in the relative proportions of palynomorphs and autochthonous organic matter. Table 4.3 and Table 4.4 (results from the 'lower shale' and lower 'middle

TABLE 4.2 Mean percentage particle abundances of Piper Formation kerogens

Wd	Wu	I	C	UP	P	M	F	A	Pd
45.2	11.9	8.2	0.9	4.0	3.1	1.2	0.3	11.8	14.8

TABLE 4.3 Stratigraphic trends in mean percentage particle abundances of kerogens in the lithological units in Well 15/17-4 (see also Appendix 4C)

Unit	Tw*	Collinitic A.O.M.	Cutinite	Autochth O.M.*	Tp*	No. Samples
Non-marine shale	56.5	27.2	5.3	0.8	1.9	4
Organic-rich shale	48.5	0.4	3.3	4.8	14.6	7
Lower shale	43.9	0	1.1	9.6	16.0	2
Lower middle sandstone	54.8	0	0.1	8.4	9.2	5
Middle shale	38.8	0	0	22.4	8.4	1
Upper sandstone	62.0	0	2.4	6.6	8.2	1

* Tw = Wd + Wu Autochth. O.M. = Non-collinitic AOM + F + P
Tp = UP + P + M

TABLE 4.4 Mean percentage particle abundances of kerogens in the Lower Shale and Lower Middle Sandstone Units of Wells 15/17-4 to 7 (see also Appendix 4C)

	Wd	Wu	I	C	UP	P	M	F	A	Pd	No. Samples
Lower shale	42.1	13.4	6.5	0.4	5.4	3.5	1.2	0.5	2.3	17.4	13
Lower Middle Sandstone	49.4	9.9	8.2	0.2	3.6	2.4	1.2	0.3	10.2*	14.3	21

* excluding samples with collinitic AOM = 3.1 (15 samples)

sandstone' averaged for the four wells) show that the marine sandstones generally contain somewhat higher amounts of wood but less palynomorphs and autochthonous organic matter than the shales (indicating hydrodynamic sorting).

When the average kerogen compositions of the 'lower shale' and lower 'middle sandstone' are considered with respect to the geographic location of the four wells an interesting pattern emerges (see Figs. 4.4 and 4.5). The proportions of phytoclastic debris in the kerogen are highest for both the 'lower shale' and 'middle sandstone' in well 15/17-6 where there is also an apparent minimum of autochthonous organic matter. The significance of these observations is discussed later when they can be considered along with the palynomorph data. Pyrolysis results confirm the overall terrigenous (Type III) character of the kerogen (Fig. 4.6).

(b) Palynomorph characteristics

The average palynomorph composition of 26 samples from the Piper Formation is recorded in Table 4.5. The plankton is dominated by dinocysts (proximates + proximo-chorates > chorates >> cavates) with only small amounts of acanthomorph acritarchs. The largest sporomorph group is the composite class consisting of simple sphaeromorphs and inaperturates, but bisaccates, 'light' deltoid spores and 'heavy' forms are also well represented. The averaged palynofacies results for the lower shale and the middle sandstone in the four wells are given in Table 4.6. It is apparent that the major differences between the two lithological units occur in the plankton with rather little contrast in the sporomorph catergories (see later).

The palynomorph data for the 'lower shale' in each of the four wells is plotted geographically in Fig. 4.7. The following features are of particular interest:

(i) the minimum value for plankton in 15/17-6

(ii) the higher values of bisaccates and light deltoid spores in

Fig. 4.4 Distribution of kerogen components in the lower shale

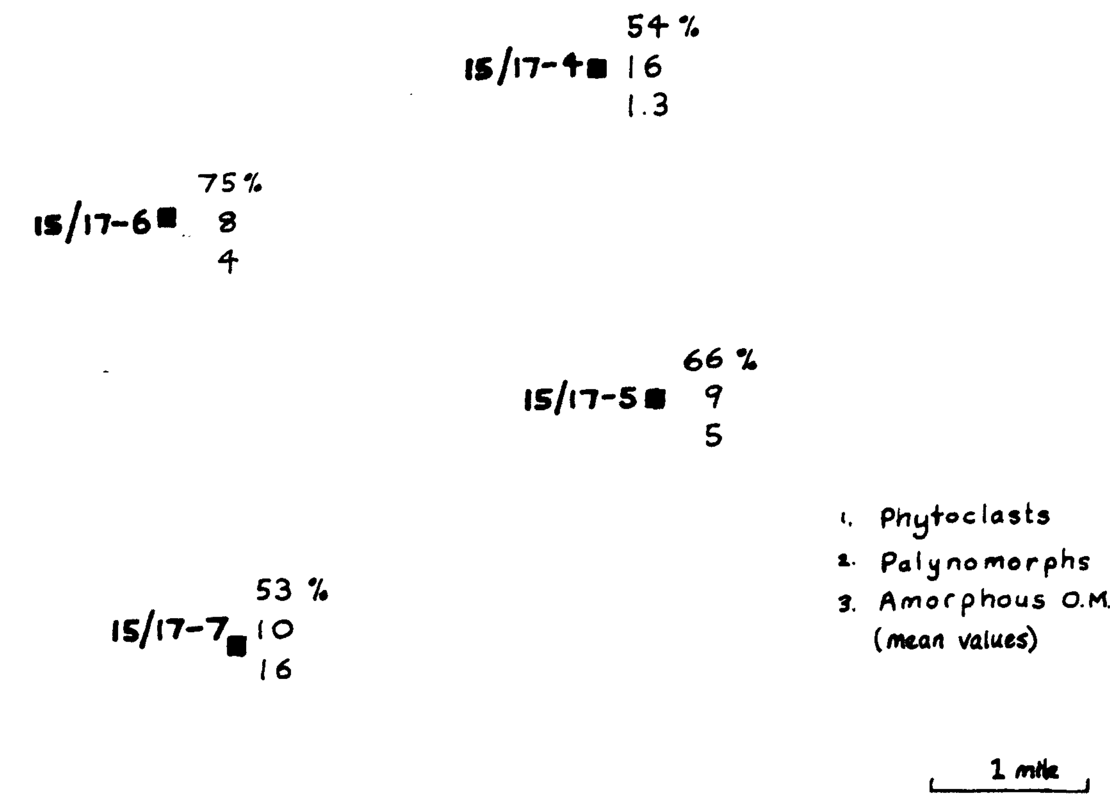
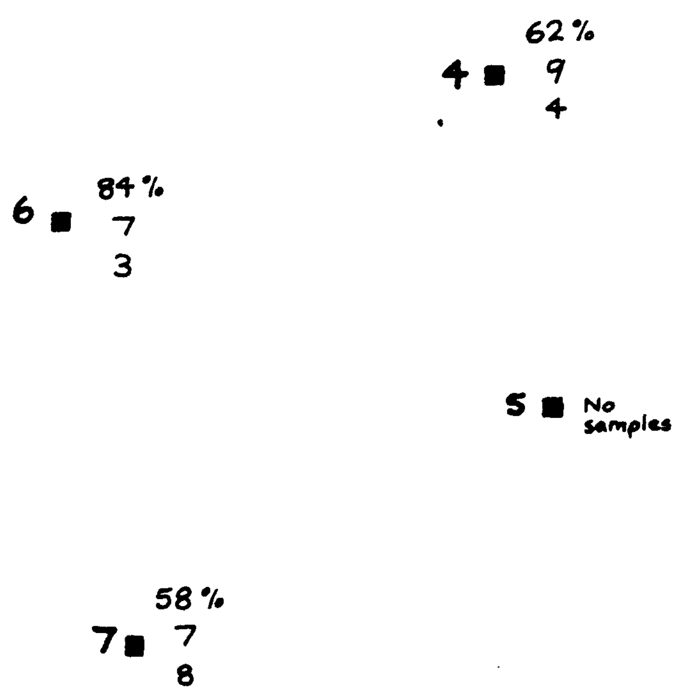


Fig. 4.5 Distribution of kerogen components in the lower middle sandstone



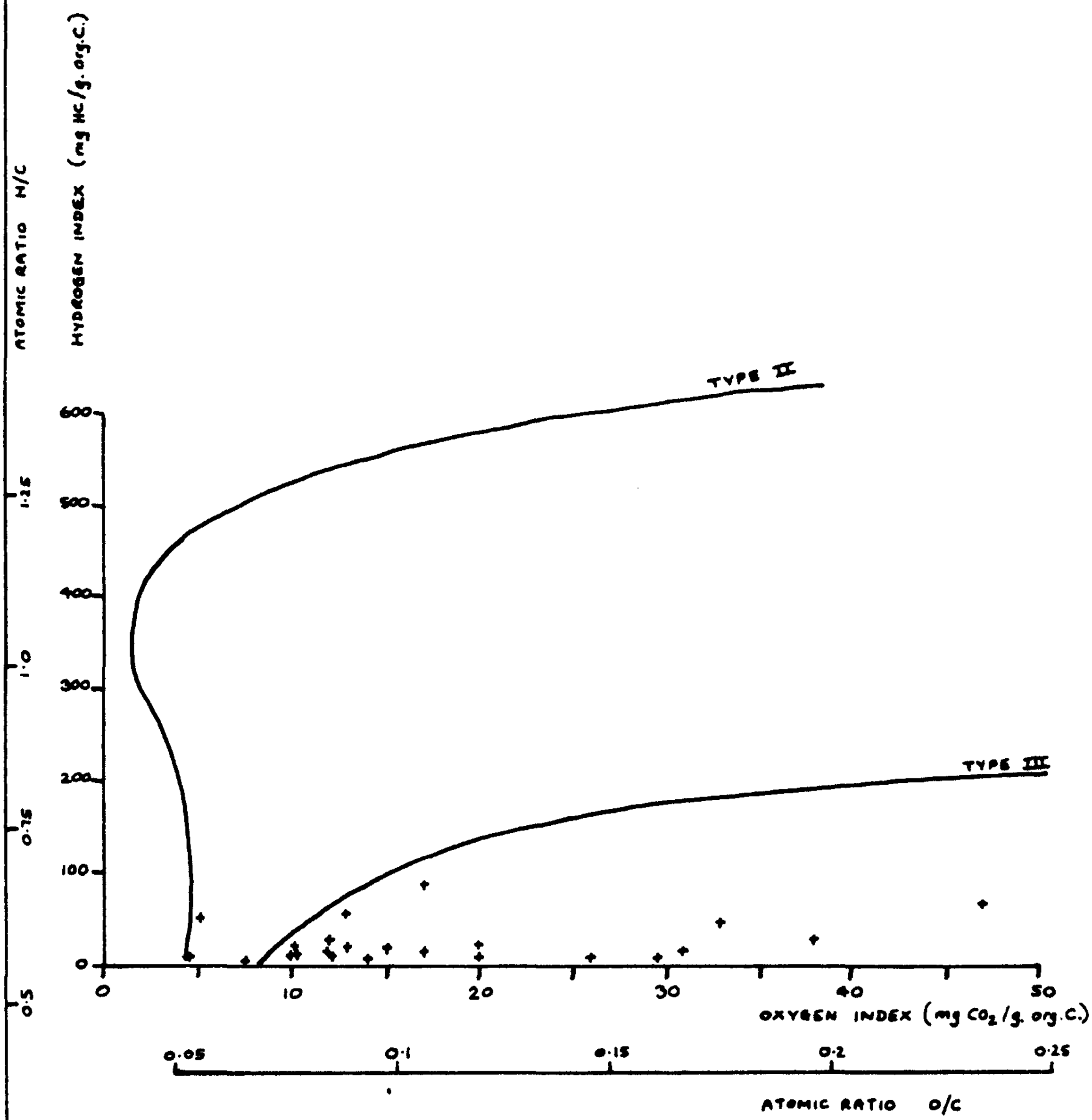


Fig. 4.6 Rock Eval. pyrolysis data (see also Appendix 4.F)

TABLE 4.5 Mean percentage abundance of major palynomorph groups in the Piper Formation

Proximate and proximochorate dinocysts	12.7
Chorate dinocysts	8.1
Total plankton (maximum)	32.7
Bisaccate pollen	6.9
Simple, unornamented deltoid spores	8.0
Ornamented and zonate spores	11.6

TABLE 4.6 Mean percentage abundance of major palynomorph groups in the Lower Shale and Lower Middle Sandstone (see also appendices 4D and 4E)

	Lower Shale	Lower Middle Sandstone
Proximate and proximochorate dinocysts	15.6	16.3
Chorate dinocysts	13.7	5.9
Total plankton (maximum)	52.9	43.5
Bisaccate pollen	5.5	8.1
Simple, unornamented deltoid spores	4.1	6.9
Ornamented and zonate spores	12.0	12.3

15/17-6

(iii) the higher values for heavy spores in 15/17-6 and 15/17-7

(iv) the greater abundance of chorate dinocysts in 15/17-4 and 15/17-5

When considered in conjunction with the kerogen data in Fig. 4.4 these results indicate that well 15/17-6 is most proximal with respect to the source of terrigenous (allochthonous) particulate organic matter and that wells 15/17-4 and 15/17-5 are the most distal. The modal grain size of the 'lower shale' division in 15/17-6 and 15/17-7 is silt whereas clay is predominant in the other wells; this also supports the inferred distal aspect of the eastern wells. The difference in the modal grain sizes in the 'lower shale' is reflected by the distribution of the heavier sporomorphs (iii above) which are often observed to be hydrodynamically equivalent to silt - very fine sand grade clastics (see Chapter 3).

The kerogen and palynomorph data (Figs. 4.5 and 4.8) for the lower part of the 'middle sandstone' suggest generally similar proximal-distal relationships to those in the 'lower shale' but is more ambiguous because of the lack of samples from 15/17-5. Wood is again most abundant in 15/17-6, and 15/17-4 shows significantly higher autochthonous organic matter components than the two western wells. Also worthy of note is that wells 15/17-6 and 15/17-7 show significant peaks in proximate + proximo-chorate dinocysts and also higher proportions of heavy sporomorphs (the latter suggesting that supply as well as hydrodynamic sorting is operative). The fact that the westerly wells are relatively more proximal is also supported by the observation that the upper 'middle sandstone' is coarser in these wells than it is in 15/17-4 and 15/17-5. The finding that bisaccate pollen were more abundant in the 'middle sandstone' than the 'lower shale' was somewhat surprising; in both cases, however, they were most abundant in 15/17-6 which was presumably closer to their source.

Fig. 4.7 Distribution of palynomorph components in the lower shale.

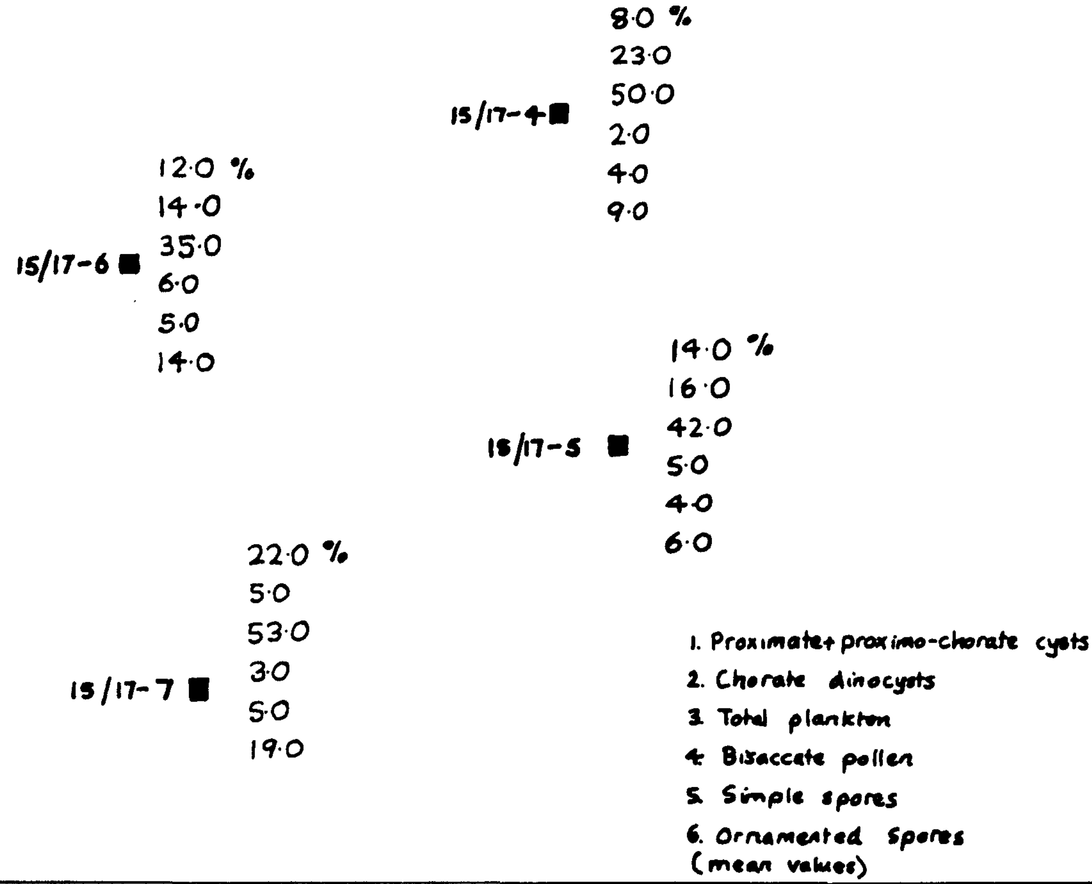
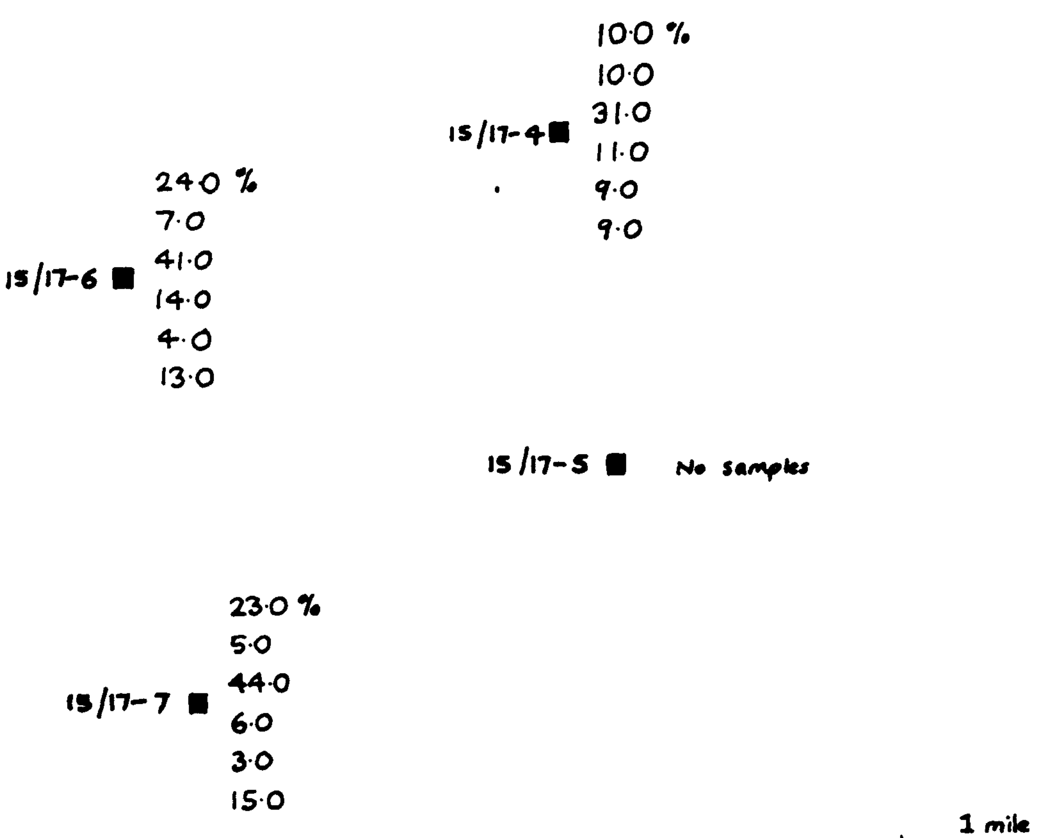


Fig. 4.8 Distribution of palynomorph components in the lower middle sandstone



(c) Comparison of palynofacies samples from the Piper and Kimmeridge Clay Formations (sensu Deegan & Scull, 1977)

Although no core samples of the Kimmeridge Clay Formation were obtained from the Piper Field, Robertson Research kindly supplied a few ready made slides that I could examine for the sake of comparison. The averaged palynofacies composition of 6 samples of the Kimmeridge Clay Formation (Volgian-Ryazanian) is shown in Table 4.7 along with the average results for the Piper Formation taken from Table 4.5. Significant differences are observed, in particular:

(i) the overall proportion of plankton is only slightly less in the Kimmeridge Clay Formation (K.C.F.) but is significantly different in its composition.

(ii) Chorate dinocysts are relatively more abundant in the K.C.F. (chorates > proximates + proximo-chorates >> cavates)

(iii) Prasinophycean algal phycomas are present in significant numbers in the K.C.F. (especially pterospermopsids). They are negligible in the Piper Formation.

(iv) Bisaccate pollen dominate the sporomorphs in the K.C.F. and heavy sporomorphs are only present as a minor component.

These changes are coupled with a very marked increase in A.O.M. in the Kimmeridge Clay Formation samples (no kerogen counts were made). The palynofacies character of the 'hot shale' interval clearly indicates a strong distal shift in facies (see ii and iv above and Chapter 3). It is difficult to interpret the significance of the appearance of the prasinophycean algae but this is typical for the Volgian North Sea Kimmeridge Clay Formation (L.A. Riley pers. comm) and characterises many 'black shale' sequences (see Singh, 1971; McLachlan & Pieterse, 1978; Davey, 1978; Hochuli & Kelts, 1979). It may indicate changing nutrient and plankton dynamics resulting from stratification and deoxygenation of the basin (indicated by dinocysts and A.O.M).

PALYNOLOGICAL OBSERVATIONS ON CORE SAMPLES FROM WELL 15/17-8A

Because of the general unsuitability of the core lithology for palynological sampling, only 8 slides were examined for kerogens and two for palynomorphs. This data is presented in Table 4.8. Compared with the other Piper samples it can be seen that the content of wood is generally higher (as is the autochthonous organic matter) and that the palynomorph content is considerably lower. The general palynofacies character of 15/17-8A is like that of the rest of the Piper Formation but the proportion of long spined acanthomorph acritarchs (and total plankton) is significantly higher and sporomorphs are less well represented. On the geological composite log the age of the cored interval is recorded as Volgian-Kimmeridgian. The absence of a good shale representative of the background sedimentation in this well makes any conclusions extremely tentative. The darker colours of the palynomorphs in 15/17-8A suggest higher thermal maturity and deeper burial than in the rest of the Piper field.

DISCUSSION AND SUMMARY OF PALYNOLOGICAL RESULTS

The up-sequence kerogen changes shown in Table 4.3 document the transition from proximal 'non-marine' to more distal marine conditions. This transition is recorded by the decrease in allochthonous organic matter and the relative increase in autochthonous materials. The greater abundance of cuticle material in the lower part of the sequence is characteristic of estuarine or prodelta deposits (see Chapter 3), and typically diminishes rapidly in an offshore direction. No palynomorph counts were carried out on the 'non-marine shale' because of the low relative palynomorph densities, but Robertson Research have recorded sufficient dinocysts to ascertain a Late Oxfordian age and so at least some of this unit must be marine or 'estuarine'. A counted sample from the base of the 'organic rich shale' proved to be devoid of plankton and higher samples from the same unit generally contain rather little

TABLE 4.7 Comparison between mean palynomorph assemblages of the Piper and Kimmeridge Clay Formations, Block 15/17

	Piper Fm.	Kimmeridge Clay Fm.
Proximate and proximochorate dinocysts	12.7 %	6.4 %
Chorate dinocysts	8.1	11.1
Prasinophycean phycomas	-	6.5
Total plankton (maximum)	32.7	30.6
Bisaccate pollen	6.9	23.3
Simple, unornamented deltoid spores	8.0	2.7
Ornamented and zonate spores	11.6	1.8
No. Samples	24	6

TABLE 4.8 Summary of mean kerogen and palynomorph assemblages in samples from Well 15/17-8A

Phytoclasts	67.4 %
Amorphous organic matter	20.2
Palynomorphs	1.6
Proximate and proximochorate dinocysts	11.5
Chorate dinocysts	5.3
Total plankton (maximum)	38.5
Bisaccate pollen	1.5
Simple, unornamented deltoid spores	3.8
Ornamented and zonate spores	6.9

plankton (av. total ~4.7%). Kerogens and palynomorphs, therefore, appear to define a gradient of increasing marine influence up toward the 'lower shale' which has a good marine plankton assemblage and marks the permanent establishment of marine conditions.

Examination of the kerogen and palynofacies data reveals significant lithological contrasts; relative to the sandstones the shales:-

i) contain proportionately somewhat lower amounts of phytoclastic debris

ii) have proportionately greater amounts of palynomorphs

iii) contain relatively more plankton

iv) have relatively more chorate dinocysts and relatively less proximate and proximo-chorate dinocysts

v) contain less 'heavy' sporomorphs (unless they are actually siltstones, q.v. 'lower shale').

These differences are probably predominantly the result of hydrodynamic sorting. Variations within the 'lower shale' and 'lower middle sandstone' indicates that well 15/17-6 is closest to the source of allochthonous organic matter and that wells 15/17-6 and 7 are relatively proximal when compared with 15/17-4 and 5. This may indicate a land area to the west or north-west and, although only four wells were studied, suggests the palaeogeographic polarity of the Piper Field. The nature of the dinocyst floras and the sporomorph distributions indicates that Piper sediments were deposited in a continuously mixed and/or seasonally stratified shelf regime with constantly aerobic bottom conditions. The A.O.M. (when not collinite) is of a degraded appearance and is dull under the ultra-violet microscope. Passing in to the Kimmeridge Clay Formation (the 'hot shale') there is a marked distal shift in the palynofacies and more persistent watermass stratification and dysaerobic to anoxic bottom conditions are indicated.

INTERPRETATION AND SYNTHESIS: PALAEOENVIRONMENT OF THE PIPER FORMATION

The Upper Oxfordian 'non-marine' and 'organic-rich' shale units of the Lower Piper Formation represent a transgressive sequence of increasing marine influence, possibly indicating an estuarine to shelf transition. Such an interpretation would be compatible with the presence of coals within the non-marine shales in other wells (Maher, unpublished), the up sequence gradients in cutinite and collinite (see above) and the rather low diversity of the macrofauna in the 'organic-rich shale'. Lack of core coverage precludes any further discussion of these units. Analysis of the 'lower sandstone' is hampered by the same problem; the general lithology and coarsening upward grain size trend may indicate a prograding submarine bar but there is insufficient data to assign any specific palaeoenvironment for this unit. The only part of the Piper Formation which has been studied and sampled in adequate detail to provide a palaeo-environmental diagnosis is the sequence from the base of the 'lower shale' to the top of the 'middle sandstone'. The remainder of this section will concentrate specifically on this interval and on the 'middle sandstone' in particular. From the 'lower shale' upward fully marine conditions are established, the 'lower shale' itself being deposited in a continuously mixed or seasonally stratified, aerobic shelf environment with both grain size and palynofacies indicating that wells 15/17-6 and 7 are relatively proximal to source and wells 15/17-4 and 5 relatively distal.

(a) ISOPACH DATA FOR THE 'MIDDLE SANDSTONE'

Since only five wells were actually examined in this study the isopach data used in this discussion is taken from the work of Maher (1980). Although the lithological divisions used here and those by Maher are somewhat different, it is possible to equate the two and hence derive the appropriate thickness data. The lower division of the 'middle sandstone' is represented by Maher's G and H sands, the basal part (α unit) of the upper division by Maher's F sand, and the remainder of the 'middle sandstone' by the F+E sands. This correlation is, however, only approximate and several discrepancies do occur, particularly in relation to the upper 'middle sandstone'. The hand-drawn isopachs are shown in Fig.s 4.9 and 4.10 (base map from Maher, unpublished). Although the data covers only a comparatively small area and some irregularity is evident, the isopachs of the lower 'middle sandstone' (Fig. 4.9) appear to define a linear feature running approximately NNE - SSW. The sparcity of data from the south west side of the structure makes it difficult to place the 15/17-7 well in context. The isopachs of the upper 'middle sandstone' in Fig. 4.10 show an easterly shift in the location of the thickest sections but an apparent N-S or NNE-SSW trend is still detectable. The pattern shown in Fig. 4.10 is in fact more complex than it at first appears. Earlier it was suggested that there is a natural two-fold division of the upper 'middle sandstone' into a basal part with a more-or-less uniform gamma ray response (α) and an upper part with a 'saw-tooth' gamma log pattern (β). Despite gaps in core coverage this division appears to be supported by lithological differences observed in the cores. The basal α division of the upper 'middle sandstone' is of more-or-less the same thickness (generally $\sim$25') in all four cores examined in this study and is the only part of the upper 'middle sandstone' present in well 15/17-6. It is suggested that this basal upper 'middle sandstone' unit forms a thin ($\sim$25') drape over the lower 'middle sandstone' isopachs and is overlain by the upper part of the upper 'middle sandstone' (β) only in the eastern part of the field - producing the isopach pattern visible in Fig. 4.10. The basal upper 'middle sandstone' unit (α) is equivalent to Maher's F sand interval and differs from the overlying 'E sand' which Maher describes as being like the H and G sands (i.e. like the lower 'middle sandstone'). It is therefore possible that the isopachs may be interpreted as the fusion of two similar sandstone bodies (see Fig. 4.11). Note that the top part of the upper 'middle sandstone' (β) in well 15/17-4 (8547-8551') is very like the basal upper middle sandstone (α) capping the lower 'middle sandstone'.

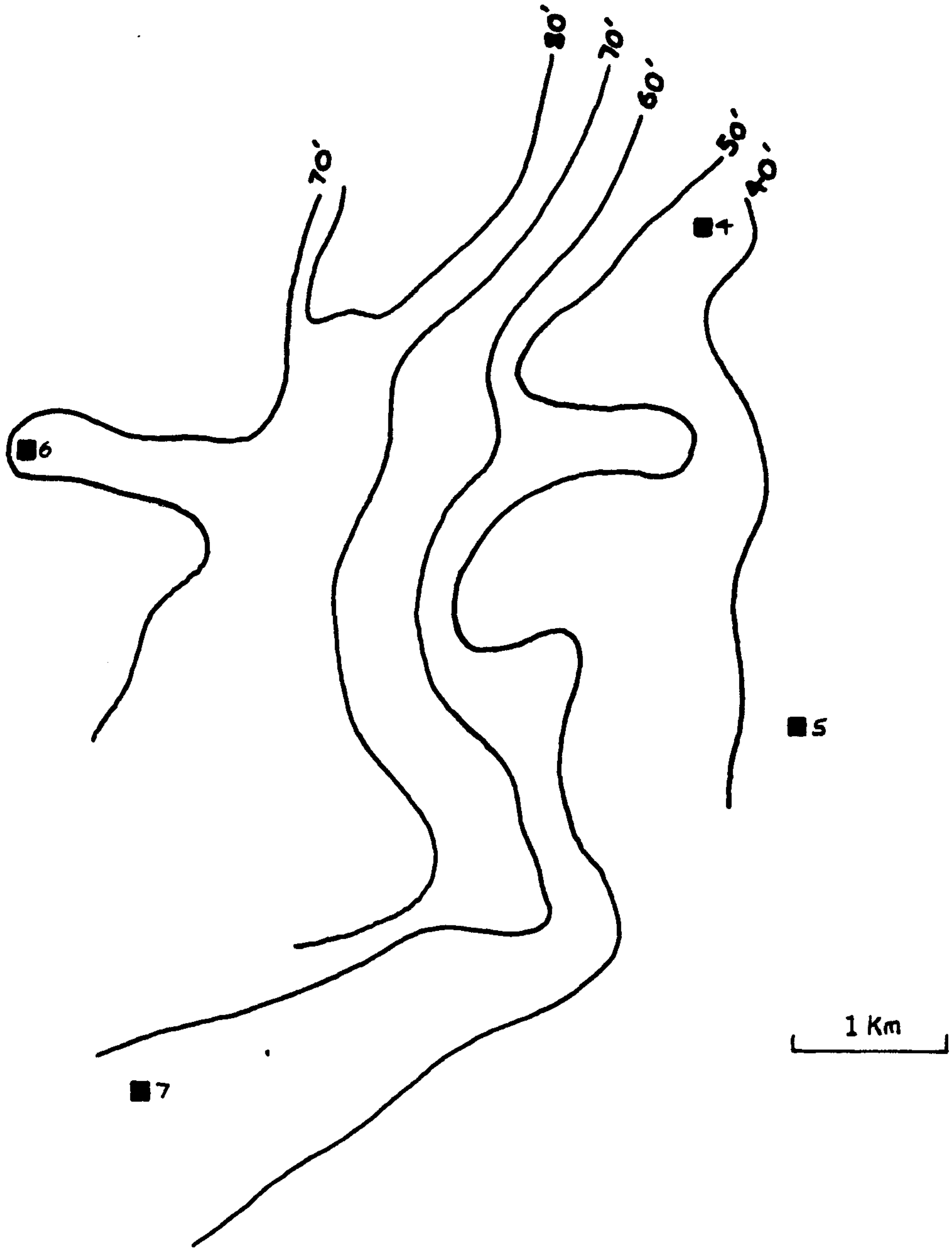


Fig. 4.9 Hand-contoured isopach data for the lower middle sandstone (values after Maher, unpublished)

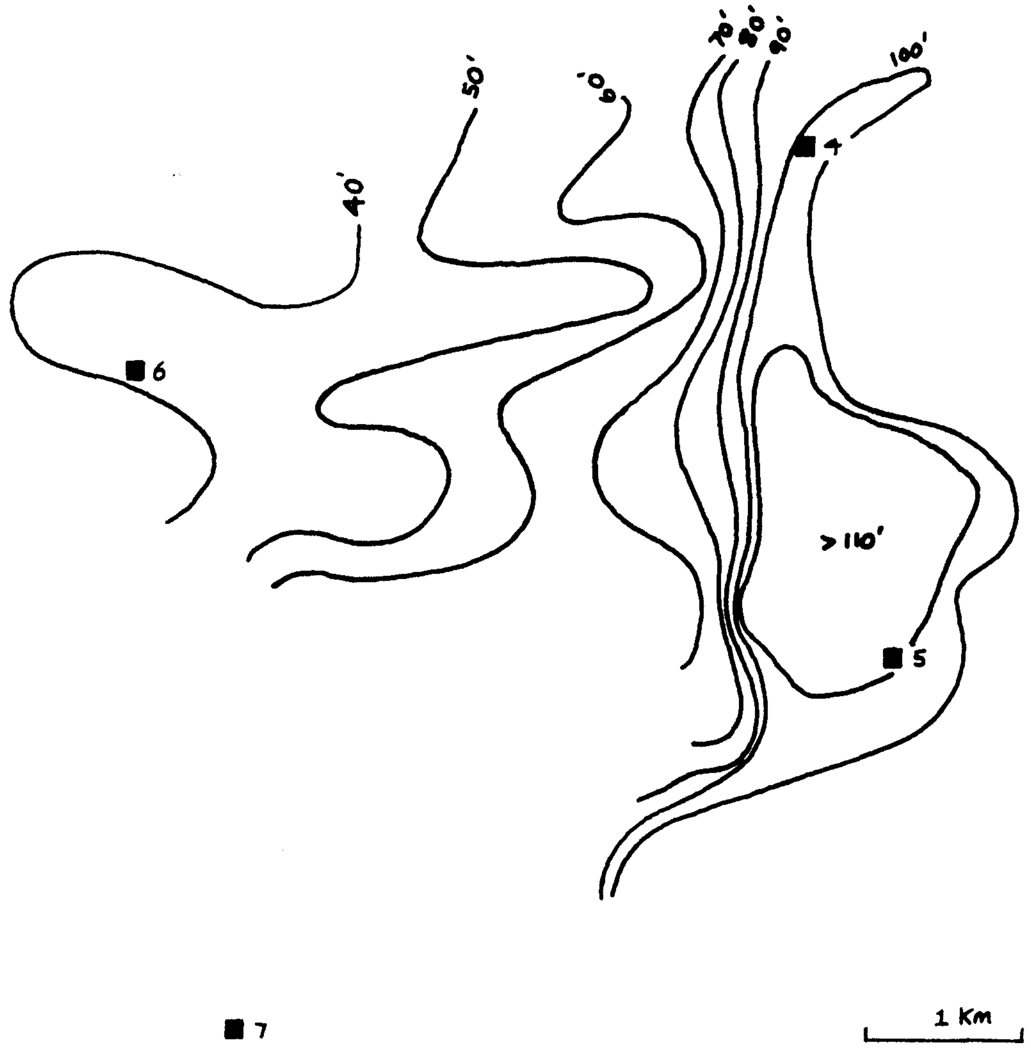


Fig. 4.10 Hand-contoured isopach data for the upper middle sandstone (α, β) (values after Maher, unpublished)

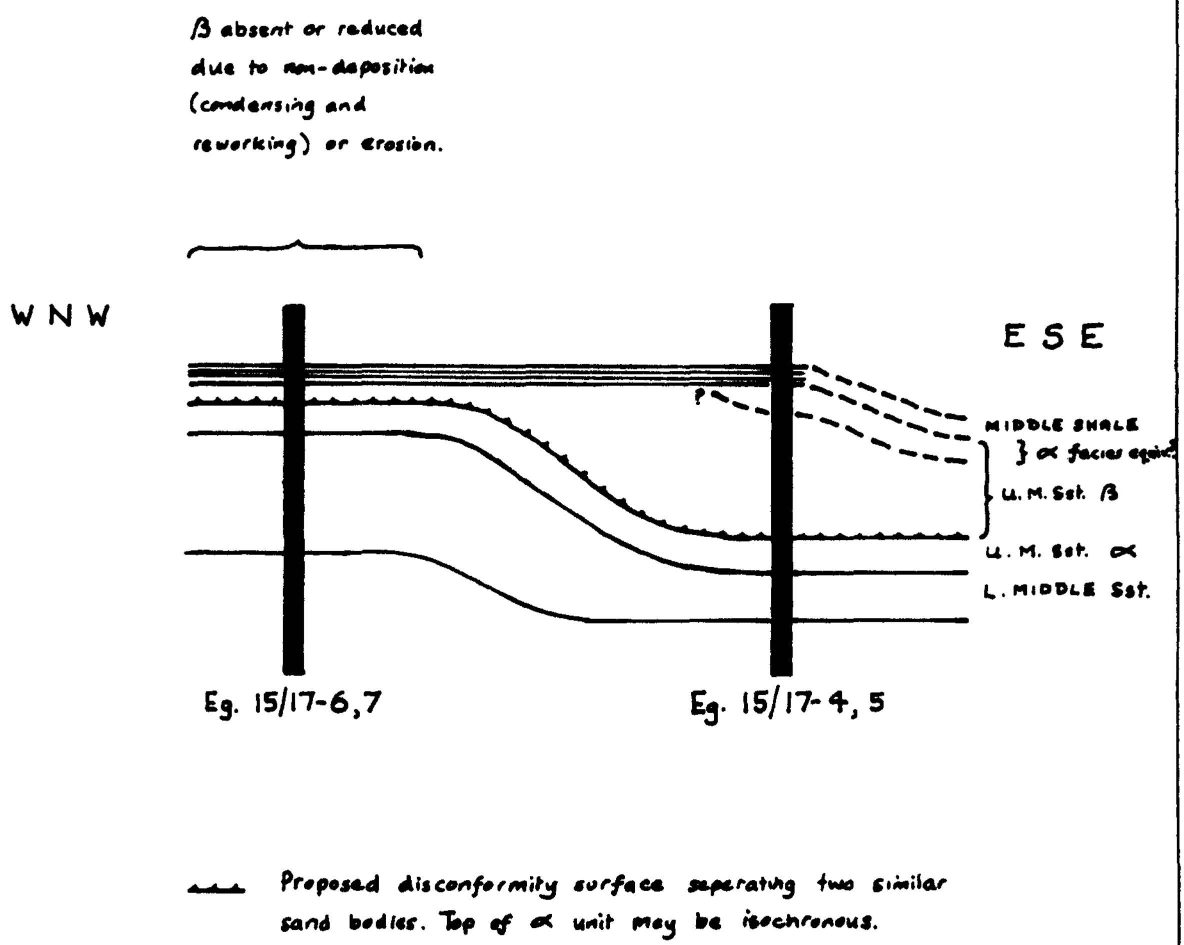


Fig. 4.11 Simplified schematic of middle sandstone interpretation. Isopach patterns controlled by synsedimentary tectonics (?) and fusion of sand bodies.

The linear trend indicated by the middle sandstone isopachs shows the correct orientation with respect to the proximal-distal relationships suggested by grain size and palynofacies. This suggests that the 'middle sandstone' is a linear sandstone body elongated 'parallel' to shoreline(?) and at 'right angles' to the palaeoslope. This is in broad agreement with the fact that the Shetland Platform is present to the north and northwest and that the main basinal area (Viking and Central Grabens) probably occurs to the east and southeast. Lithological correlations suggest that the middle sandstone may in fact consist of two imbricate sandstone bodies, a second body being 'welded' on to the east side of the one shown in Fig. 4.9.

(b) SEDIMENTOLOGICAL ANALYSIS OF THE 'LOWER SHALE' - 'MIDDLE SANDSTONE' SEQUENCE

Although modern work shows that a considerable complexity of sediment sequences can be produced within the barrier island depositional system (e.g. see excellent review by Reinson, 1979) it is apparent that this palaeoenvironment is not appropriate for the 'middle sandstone' which appears to lack aeolian deposits and is neither under- or overlain by non-marine 'lagoonal' facies. Despite these differences there are considerable parallels between parts of the barrier model and the Piper sediments. The best overall analogy for the 'lower shale'-'middle sandstone' sequence appears to be the shoreface zone of barrier bars, but the absence of emergent and or non-marine facies clearly points towards an offshore <u>submarine</u> bar environment. The various parts of the 'lower shale'-'middle sandstone' sequence are interpreted as follows (see Fig. 4.12):-

(i) The 'lower shale':

sandy, silty mudstones of the aerobic vital pantostrate (isostrate) biofacies. General bioturbate texture with rare recognisable <u>Teichichnus</u>, <u>Planolites</u> and <u>Chondrites</u> burrows and rare relict bedding (subhorizontal or ripple lamination). Interpreted as a lower shoreface-transition zone deposit (see Reinson, 1979; Reineck & Singh, 1973; Howard, 1971 and Tizzard & Lerbekmo, 1975).

(ii) Lower 'middle sandstone':

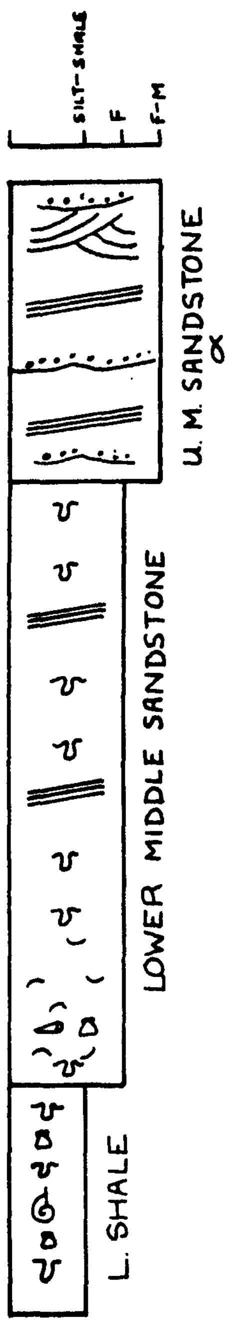
fine grained sandstones of the aerobic, vital lipostrate (heterostrate) biofacies. Generally thoroughly bioturbated and structureless except for occasional low angle planar laminae. Common Ophiomorpha burrows. Interpreted as a middle shoreface (lower bar) deposit (see Reinson, 1979; Tizzard & Lerbekmo, 1975; and La Fon 1981).

(iii) Basal upper 'middle sandstone' (α):

generally medium grained (fine to coarse) sandstones of the aerobic lethal lipostrate (heterostrate) biofacies. Primary sedimentary structures are dominated by low angle planar laminations with only rare trough cross-bedding and ripple bedding; bioturbation is insignificant. Scour surfaces, laminated storm sand layers (Reineck & Singh, 1972) and fining upward shoreface storm units (Kumar & Sanders, 1976) are probably present. Heavy mineral concentrations are probably present as indicated by the gamma ray anomaly at the contact with the lower middle sandstone. Low angle laminations are more likely to have been produced by storm processes than swash-backwash effects, but a combination of fairweather (normal wave action) and storm products is likely. Interpreted primarily as an upper shoreface (upper bar) deposit but may possible include equivalents of beach/foreshore sands (see Reinson, 1979; La Fon, 1981; and Tizzard & Lerbekmo, 1975).

(iv) Remainder of upper 'middle sandstone' (β):

where present, that part of the middle sandstone overlying the basal part of the upper division is herein interpreted as a repeated middle-upper shoreface sequence belonging to a second sand body present on the east side of the field.



INTERPRETATION	BIOFACIES	FEATURES
Upper bar. (Upper shoreface)	LETHAL HETEROSTRATE	Bioturbation rare. Scour surfaces with coarse grained lag deposits common. Laminated storm sand layers? Fining upward "shoreface" storm units?
Lower bar. (Middle Shoreface)	VITAL HETEROSTRATE	Mostly bioturbated but some primary bedding preserved. Conspicuous Ophiomorpha burrows. Bioclastic debris (bivalves and belemnites) fairly common toward base. Scattered carbonaceous debris common (org.C ≤ 1.0%).
Lower mixed-layer shelf mudstone. (Lower shoreface-transition zone)	VITAL ISOSTRATE	Heavily bioturbated (Teichichnus, Planolites, Chondrites burrows). Organic-rich (≤ 3.0% org.C.). Fauna includes Trautscholdia, Dentalium and ammonites.

Fig. 4.12 Facies interpretation of lower shale-upper middle sandstone (α) sequence. For explanation of biofacies see Chapter 2.

The sequence described above is very like that recorded by Tizzard & Lerbekmo (1975) for the Viking Formation in Alberta which they interpreted as an offshore bar deposit (see also La Fon, 1981). It differs significantly from the sequences interpreted as subtidal, tide dominated bars described by Berg (1975) and Nio & Siegenthaler (1978).

There is no direct evidence of significant tidal activity. The present interpretation of the 'middle sandstone' is largely without a modern analogue; this is representative of the primitive state of our current knowledge on modern shallow marine sands (see Johnson, 1978 and Walker, 1979).

(c) COMPARISON WITH THE BRORA BRICK CLAY - BRORA SANDSTONE SEQUENCE, BRORA, SUTHERLAND

These sediments were not examined during this study and all comments and descriptions are taken from Sykes (1975), except for the interpretation which has been slightly revised. The individual members of the sequence are described briefly below (see also Table 4.9 and Fig. 4.13).

(i) Brora Brick Clay Member (Brora Argillaceous Formation)

A 13-15m thick sequence of variously burrowed and laminated carbonaceous sandy siltstones that become sandier upwards. The member contains fining upward sequences (0.5-1.0m thick) consisting of carbonaceous silt grading upward into grey silty clay with *Entolium* and *Bositra*.

(ii) Fascally Siltstone Member (Brora Argillaceous Formation)

This member comprises a 31-44m thick coarsening upward, coarse siltstone to very fine sandstone. It contains no sedimentary structures and is generally homogeneously mottled but the ichnogenera *Thalassinoides*, *Chondrites* and *Planolites* are identifiable. It is poorly fossiliferous but does yield a mixed deep and shallow infauna including *Pleuromya*, *Pholadomya*, *Protocardia*, *Grammatodon* and *Corbulomima*. Some calcareous

Table 4.9 Lithostratigraphy and age of Brora sequence

<table>
<tr><th>FORMATION</th><th>MEMBER</th><th>AGE</th></tr>
<tr><td>Balintore</td><td>Ardassie Limestone</td><td>M. Oxfordian (plicatilis)</td></tr>
<tr><td rowspan="4">Brora Arenaceous</td><td>Brora Sandstone</td><td rowspan="2">L Oxfordian</td></tr>
<tr><td>Clynelish Quarry Sandstone</td></tr>
<tr><td>Fascally Sandstone</td><td rowspan="2">U. Callovian</td></tr>
<tr><td>Fascally Siltstone</td></tr>
<tr><td rowspan="3">Brora Argillaceous</td><td>Brora Brick Clay</td><td rowspan="3">M. Callovian to L. Callovian (calloviense)</td></tr>
<tr><td>Glauconitic Sandstone</td></tr>
<tr><td>Brora Shale</td></tr>
</table>

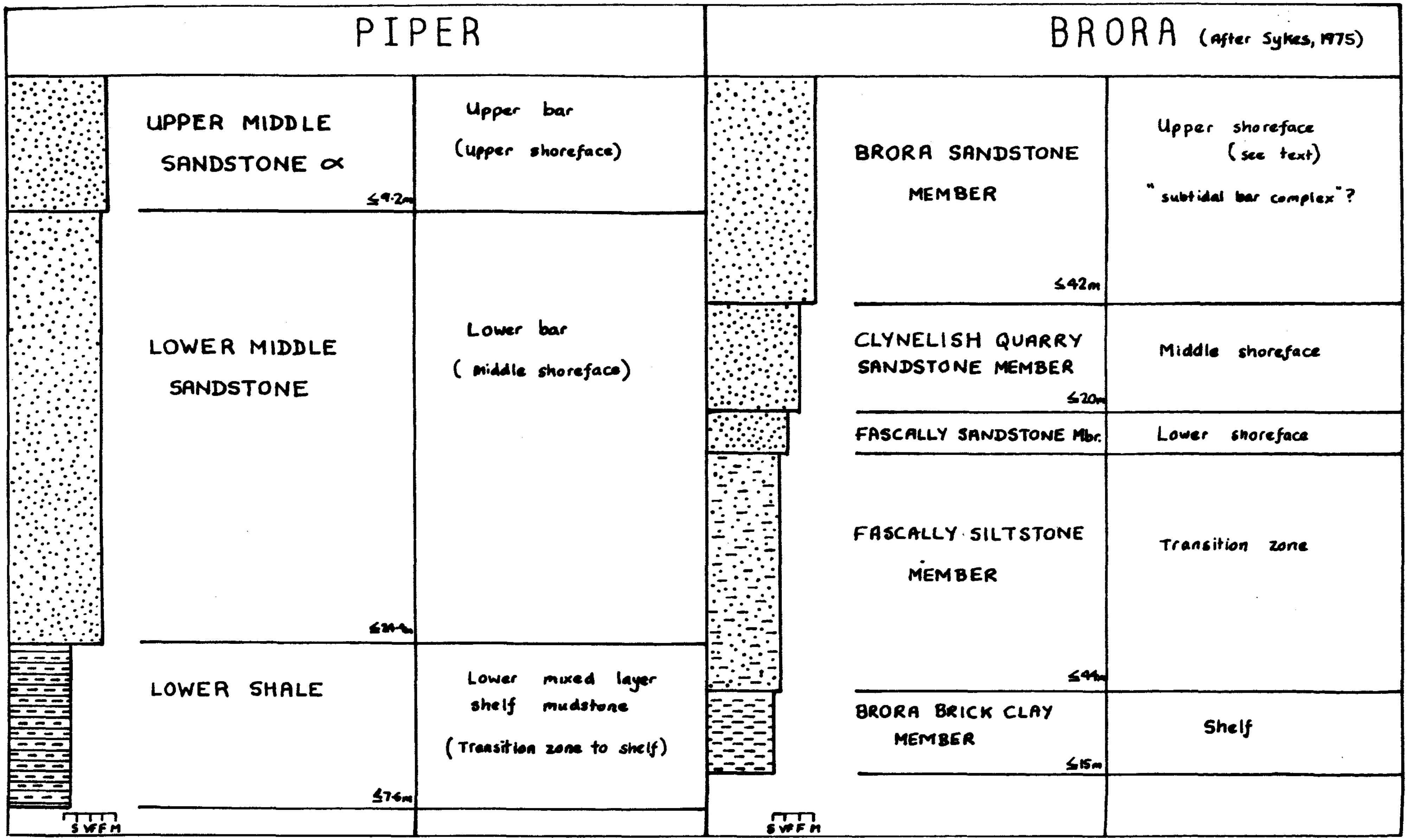


Fig. 4.13 Comparison of facies interpretations of the Late Jurassic coarsening-upward sequence from the Piper Formation and Brora.

concretions and sporadic glauconite is also present.

(iii) Fascally Sandstone Member (Brora Arenaceous Formation)

A 5-8m thick unit of well-sorted, very fine grained muddy sandstones occurring in 70cm thick beds with muddy partings. Occasional parallel and ripple-cross lamination occurs at the base of the member and bioturbation is in the form of mud filled burrows and Arenicolites and Teichichnus. Sparite cemented doggers are present. There is a mixed epifaunal-infaunal bivalve fauna containing Discomiltha and Chlamys.

(iv) Clynelish Quarry Sandstone Member (Brora Arenaceous Formation)

This member consists of about 20m of massive, friable, well-sorted fine grained sandstone with localised silica cementation. Laminae rich in carbonaceous, woody debris define ripples, trough cross-beds, slump and water escape structures and some megaripple bedforms ($\leq$ 50cm high) Occasional Ophiomorpha burrows are present.

(v) Brora Sandstone Member (Brora Argillaceous Formation)

This sandstone is at least 42m thick and its lower part consists of massive or cross bedded medium grained sandstone; silica cementation and burrows are absent. The upper part (top half) contains large scale inclined beds and trough cross bedding with scour surfaces overlain by quartz conglomerates or poorly sorted, pebbly coarse grained sandstones. Subordinate small tabular intrasets and parallel lamination is also present. Foreset beds are sometimes 'internally graded' passing upward from coarse to medium grained on a centimetre scale. The sandstone is decalcified but Meleagrinella moulds are quite common. The member is generally medium grained but the lenticular coarse scour pockets make it somewhat less well sorted than the Clynelish Quarry Sandstone.

From the above description and the aforegoing account of the Piper lithofacies it can be seen that there are strong parallels between the Brora sequence and the 'lower shale'-'middle sandstone' interval. A comparison between the two sections is shown in Fig. 4.13 along with the present interpretation and Sykes' previous interpretation of the Brora sediments. Several points of the comparison are discussed below.

(1) The similarity of the two sequences is probably greater than the data presented here suggests. This results from the fact that only one of these sequences was examined during this study and also from the natural biases inherent in outcrop-core comparisons.

(2) Although the Brora Brick Clay and Fascally Siltstone are genetically equated with the 'lower shale' in Fig. 4.13, it is in fact the Brora Shale Member which is most directly comparable to this unit. This reflects the slightly different positions of the two coarsening-upward cycles in terms of the overall sequences in which they are situated. The Brora Shale Member is a bioturbated sandy siltstone and organic-rich ("bituminous") shale unit that contains dense shell beds with Deltoideium, Trautscholdia and Thracia. The fauna is of low diversity and high density with Trautscholdia and Dentalium, the most important fossils, and cephalopods being comparatively rare; Sykes (1975) classified it as the "Trautscholdia-Palaeonucula bituminous shale subfacies". This faunal assemblage appears to be very like that found in the 'lower shale' (see earlier). Both the 'lower shale' and the Brora Shale occur near the base of transgressive sequences. It is interesting to note that both the siltstone unit at the base of the Brora Shale and the top part of the 'lower sandstone' (at the base of the 'lower shale') contain common belemnites.

(3) Sykes (1975) has interpreted both the Clynelish and Brora Sandstone Members as middle shoreface deposits. This has been revised as in Fig. 4.13 with the Brora Sandstone now interpreted as an upper shore-face deposit because of the absence of bioturbation and abundance of trough cross bedding (see Reinson, 1979). Planar bedding appears to be more abundant than trough cross bedding in the basal upper 'middle sandstone' in Piper but it is not possible to carry out a detailed comparison because of core-outcrop contrasts and the lack

of data on lateral variability.

(4) Neither the Brora sequence or Piper sequence are topped by aeolian sediments. The Brora Sandstone Member is overlain by the Ardassie Limestone Member (Balintore Formation), a muddy spicular sandstone, and the Piper sediments (allowing for the imbrication on the east side of the field) by marine shale (the 'middle shale'). Both units, therefore, appear to be shallow water submarine bar deposits. There is no data to substantiate the view that emergent facies were deposited and then eroded.

(d) INTERPRETATION OF THE SEDIMENTS IN WELL 15/17-8A

(i) Interval 12423' to 12490'

There is little in the way of diagnostic sedimentological features in this interval. For the greater part the sandstone is uniform and absence of structures and the lower sorting than is typical of the other Piper sediments suggests the general absence of currents or wave reworking. The only exception lies in the occasional low angle, planar, carbonaceous laminae in the lower part of the interval. Bioturbation is apparently absent. The presence of 'hot shales' in the underlying un-cored section, which are believed to represent dysaerobic-anoxic conditions (see Chapters 2 and 8) may indicate that low or absent bottom oxygenation precluded benthos and hence bioturbation. The absence of shales in this thick cored section could also indirectly imply that sedimentation rate may have been a limiting factor.

(ii) Interval 12490'-12578'

The characteristic feature of this interval is the presence of the cyclic units demonstrated in Fig. 4.3. Current structures and bioturbation are again absent. The distribution of the coarser materials (and in particular the pebble grade material) within these cycles indicates a mass emplacement mechanism was responsible for their sedimentation (the sandstones are not uniformly graded with the pebbles all at the base as in the appropriate parts of the other Piper wells). Mass emplacement mechanisms involve a whole continuum of hydrodynamic and mechanical processes, each of which can be defined and analysed in the laboratory but are not mutually exclusive in nature (Middleton & Hampton, 1976; Carter, 1975; Nardin et al. 1979). This has made it extremely difficult to develop ideal theoretical sequences of sedimentary structures for debris, grain, liquidised and fluidised flows comparable to the Bouma sequence for turbidity current deposits. However, a provisional attempt has been made by Middleton & Hampton (1976; a revised version of their 1973 paper). In terms of these hypothetical sequences (see their Fig. 9 on page 213) the 15/17-8A cycles appear to be something of a hybrid (see also Stauffer, 1967 for typical sequence types observed in possible grain flow units). The relevant features of the cycles observed in 15/17-8A are:-

1) The general uniformity of the medium grained matrix throughout the unit (more rarely with a coarser grained basal layer)
2) The grading of the coarser grade (very-coarse to pebble sized) materials which occur only in the lower third and increase in size and abundance downwards, but are dispersed within the matrix.
3) The occasional presence of floating shale clasts at the top of the units (and also sometimes in the basal pebbly parts).
4) Dish structures and other water escape structures were <u>not</u> observed - this may in part be due to the narrow diameter of the cores (2½") as such structures are often feint at the best of times. The narrow diameter also makes it difficult to comment on the nature of the tops and bases of these cyclic units. Core X-rays might yield much more information than can be obtained by visual observations.
5) No obvious reverse grading was observed in the cores but to a certain extent this could depend on where the base of the unit is taken; in outcrop weathering effects alleviate such problems. The apparent absence of reverse grading is considered to be real in this case.

With respect to liquified flow deposits, Lowe (1976, p.298) states that "Normal grading may develop if the sediment is poorly sorted and particularly if gravel sized clasts are present. In simple, non-turbulent resedimenting systems, this grading would probably appear as coarse tail grading". Liquified flows are those whose mobility results from partial support by escaping pore fluid in which the solid is settling through the liquid rather than the liquid flowing up through the solid as in fluidised flows (Lowe, 1979). The usage of the term and description of the process of fluidised flows in Middleton & Hampton (1976) is incorrect, refer to Lowe (1976, 1979) for discussion. From the characteristics of the cycles shown in Fig. 4.3 and the theoretical characteristics of the various mass emplacement processes described in the literature, the interval 12490'-12578' is here interpreted as a series of grain flow-liquified flow deposits with mobility resulting from dispersive pore pressures and/or partial support due to escaping pore fluids. Such flows probably resulted from failure on 'steep' slopes, possibly triggered by fault activity or sedimentary overloading. No wave or biogenic reworking is apparent at the top of the units. The well is probably relatively proximal with respect to the source of the flows.

(iii) <u>Interval 12894' to 12946'</u>

This interval is unique in that it has a general mottled appearance and contains relatively common irregular bedding. Planar laminae are sometimes indicated by grain size variations and rare tubular burrows are also present. For part of the interval the distribution of pebbles again suggests graded units as in Fig. 4.3. Although grain flow-liquified flow sediments might be present, the occurrence of mottling and structure <u>within</u> the units (e.g. common coarse-very coarse grained laminae within the medium grained matrix) suggests that at least some wave or current reworking,and definately biogenic reworking,took place. The shale at 1320' to 1335' shows a normal shale gamma-ray response and hence indirectly a pre-Volgian age for this, the lower core interval (which is below the hot shale in the uncored interval). The presence of bioturbation suggests aerobic bottom conditions were characteristic.

Overall the 15/17-8A sediments appear to have been deposited in deeper water than the rest of the Piper Formation, particularly in the case of the upper two core intervals where wave processes appear to be minimal and probable grain-liquified flows are present. These upper two core intervals are probably at least Volgian in age (as they overlie 'hot shales') and correlate with the transgressive deeper water Kimmeridge Clay Formation in the other wells. During this time the 'Upper sand' of the Piper Formation may have been reworked and redeposited into deeper water by mass emplacement processes. The lower core interval is probably a deeper water correlative of the Piper Formation on the Piper structure. Such an interpretation is in agreement with the position of the 15/17-8A well with respect to the determined palaeogeographic orientation of the Piper Formation in the type area (see Fig. 4.1).

CONCLUSIONS

Palynofacies analysis indicates that the lower argillaceous part of the Piper Formation marks a paralic (estuarine?) to marine, lower mixed layer facies transition. There is insufficient core coverage to comment on the nature of the 'lower sandstone', 'middle shale' and 'upper sandstone' units other than they are all shallow marine, aerobic, mixed layer deposits. The more complete core coverage allowed a detailed analysis of the coarsening upward 'lower shale'-'middle sandstone' interval which is interpreted as a lower-upper shoreface sequence of a submarine 'nearshore' bar; emergent facies are absent. Deposition occurred within a lower (seasonally stratified) to upper (continuously mixed) aerobic mixed layer environment. Strong parallels are noted between the 'lower shale'-'middle sandstone' lithofacies and those of the Brora Brick Clay Member to Brora Sandstone Member of Sutherland, Scotland described by Sykes (1975). Kerogen analysis of the Piper Formation revealed the predominance of allochthonous woody materials (i.e. a type III composition, see Chapter 3). Palynofacies and grain size trends indicate a relative proximal-distal trend from west or north-west to east or south-east during deposition of the 'lower shale' and 'middle sandstone'. Regionally, however, it is clear that the Piper facies were deposited within an extensive shelf area and that the shoreline during deposition of the 'lower shale' - 'middle sandstone' sequence was perhaps as much as tens of kilometers distant (see also Chapter 8). Examination of the isopach data suggests that the 'middle sandstone' consists of a linear, elongate sandstone bar or bars oriented approximately NNE-SSW. Lithofacies correlations and gamma ray logs together with isopach data suggest that the 'middle sandstone' consists of two separate sand bodies, a second body occurring on the east side of the field.

Palynofacies of the Kimmeridge Clay Formation (sensu Deegan & Scull, 1977) indicates that a strong distal shift in depositional environment concommitant with the development of oxygen defiency in the bottom waters occurred following a period of condensing at the top of the Piper Formation. The sandstones in the upper two cores of well 15/17-8A were deposited off-structure in a deep water (probably bottom water) environment and include grain flow-fluidised flow deposits which are probably intra-Kimmeridge Clay Formation. The relative distribution of in situ and redeposited shelf sands within the Piper 'province' is a matter which requires further investigation.

APPENDIX 4A SAMPLE DISTRIBUTION

Sample interval	Unit	Sample code			
		GOP	OP	POP	SOP
WELL 15/17-4					
8744 - 8744.5	N.M.Shl.	21		88	
8741.5				87B	
8740.5 - 8741		20A		87A	A56
8736.5 - 8737		20		87	
8730.5 - 8731	Org.rich Shl.	19		85	
8727 - 8727.5				84	83X
8724 - 8724.5		18		83	
8720 - 8720.5		17		82	54
8715 - 8715.5				81	53A
8710 - 8710.5		16		80	53
8702.5 - 8703		15		79	
8694.5 - 8695	L. sst			78	
8688		16A		80A	53X
8683 - 8683.5	L. Shl.	14		75	51A
8680 - 8681				74	51
8669.5 - 8670	L.M. sst			73	50B
8660 - 8660.5				72	
8640 - 8640.5				70	49C
8631 - 8631.5	U.M. sst (α)			69	
8624.5 - 8625				68	
8616.5 - 8617					48A
8550 - 8550.5	U.M. sst (β)				47
8540 - 8540.5				60	46B
8528.5 - 8529				58	46
8525 - 8525.5	M. Shl	13			45
8514 - 8514.5		12		56	44D
8495.5 - 8496	U sst				44A
8490 - 8490.5					44
8473.5 - 8474		11		50	43A
8464.5 - 8465					43
WELL 15/17-5					
8562 - 8562.5	L. Shl	3		13	18
8553 - 8553.5		2		12	17
8530 - 8530.5	L.M. sst				16
8500 - 8500.5	U.M. sst (α)				15
8440 - 8440.5	U.M. sst (β)				13
8400 - 8400.5	U. sst				12
8367 - 8367.5					10
8320 - 8320.5					8
WELL 15/17-6					
8718 - 8718.5	L. sst				42
8705.5 - 8706	L. Shl	10		47	41A
8700 - 8700.5		9		46	41
8690 - 8690.5		8		45	40A
8681 - 8681.5		7		44	40
8670 - 8670.5	L.M. sst			43	39A
8660 - 8660.5				42	39
8650 - 8650.5				41	38
8640 - 8640.5				40	37
8634.5 - 8635				39	36A
8630 - 8630.5				38	36
8606.5 - 8607					35A
8595.5 - 8596				35	35
8590 - 8590.5	U.M.sst (α)				34
8581 - 8581.5					33
8572.5 - 8573				32	32A

APPENDIX 4A (Cont)

Sample interval	Unit	GOP	OP	POP	SOP
		Sample code			
WELL 15/17-7					
9374 - 9374.5	L. Shl	40	40	1A	7
9369.5 - 9370		39	39		
9365.5 - 9366		38	38		6
9360 - 9360.5		37	37		
9358.5 - 9359	L.M. sst	36	36		
9355 - 9355.5		35	35		5
9350.5 - 9351		34	34		
9348.5 - 9349			33		
9330 - 9330.5					4
9314 - 9314.5			25	1B	
9311 - 9311.5			25		
9305.5 - 9306			24		
9303 - 9303.5			23		
9295.5	U.M. sst (α)		20		
9284.5 - 9285					2
9266.5 - 9267	U.M. sst (β)				1
WELL 15/17-8A					
12943.7 - 12946	Core 5			31B	
12940 - 12940.5				31A	
12930 - 12930.5				31	31
12922.5 - 12923				30	30
12902 - 12902.5					29
12899.8 - 12901.8				27	
12894.5 - 12895					28A
12570 - 12570.5	Core 4				
12556 - 12556.5				26	
12550 - 12550.5				25	27A
12534 - 12534.5					27
12514.5 - 12515	Core 3				26
12498 - 12498.5				22	
12488 - 12488.5					25
12482.5 - 12483					24A
12470 - 12470.5	Core 2				24
12460 - 12460.5					23
12443 - 12443.5	Core 1				22
12430 - 12430.5					21
	TOTALS	19	13	48	63

N.B. Samples were taken from pre-existing sample sets which consisted of six inch lengths of core, hence only the appropriate interval is given rather than the exact position.

APPENDIX 4B KEROGEN DATA

Sample	% Particle abundance									
	Wd	Wu	I	C	UP	P	M	F	A	Pd
WELL 15/17-4										
POP 88	26.4	30.2	1.6	14.6	0	0.8	0.6	0	24.0	1.6
87B	36.8	29.4	9.8	2.4	0.4	0.4	0	0	14.8	6.0
87A	34.6	22.4	2.4	1.2	0.6	0.6	0.6	0.2	36.4	1.2
87	22.0	25.4	9.8	2.8	1.4	1.2	1.0	0	33.4	3.0
85	25.2	25.8	13.8	9.2	2.6	6.6	6.6	0	0.6	9.6
84	13.6	17.8	19.6	3.2	9.4	9.6	2.2	0	0.4	24.2
83	45.0	15.6	19.4	4.4	2.3	3.0	1.2	0.2	0.2	8.8
82	40.6	11.8	20.4	1.0	3.6	4.2	2.2	0.2	0	16.0
81	51.4	12.0	11.2	0.8	5.0	2.4	1.0	0.2	0	16.0
80	30.6	15.2	11.0	1.4	18.6	3.4	0.8	0	0.8	18.2
79	27.0	8.2	15.0	2.8	12.8	1.6	3.0	0.2	1.0	28.4
78	21.6	6.0	20.0	0.2	4.0	0.6	2.6	0.4	18.0	26.6
80A	52.6	3.6	14.4	0	4.6	0.2	1.4	0.2	7.4	15.6
75	32.6	16.2	12.0	2.2	5.8	5.0	1.8	0.2	0.4	23.4
74	25.6	13.4	7.8	0	8.0	10.2	1.2	1.2	2.2	30.4
73	33.4	10.2	4.4	0	11.0	7.6	4.0	0.2	3.2	26.0
72	36.0	11.4	14.2	0.2	7.2	2.6	2.0	0.8	2.8	22.8
70	44.4	1.4	2.2	0	2.6	0.8	0.2	0	1.0	47.4
69	51.7	26.0	9.3	0	2.0	0.3	0	0	1.7	9.0
68	54.6	0	4.2	0.2	2.2	0	0.2	0	0.6	38.0
60	37.4	7.4	5.4	0	5.2	2.0	1.2	1.6	18.8	21.0
58	44.2	6.6	4.4	0	4.0	1.0	1.8	0.8	13.2	24.0
56	30.6	8.2	13.0	0	6.0	1.6	1.8	3.4	17.4	18.0
50	52.0	10.0	8.2	2.4	6.4	1.0	0.8	1.0	4.6	13.6
WELL 15/17-5										
POP 13	50.2	18.4	1.6	1.8	10.2	4.2	1.0	0.2	3.0	14.4
12	47.2	10.6	3.8	0.4	2.4	0.4	0.6	0.2	7.0	27.4
WELL 15/17-6										
POP 47	37.4	13.4	8.8	0.6	6.8	3.2	1.0	1.6	10.4	16.8
46	59.2	11.0	9.4	0.2	3.0	0.8	0.4	1.0	4.8	10.4
45	63.4	9.6	6.2	0	4.6	3.2	1.4	0	1.8	9.8
44	71.4	7.0	3.2	0	3.2	2.4	0.8	0	0.2	11.8
43	61.6	13.2	5.2	0.2	4.0	3.4	1.6	0.2	2.6	8.0
42	64.2	9.2	3.4	1.0	4.8	4.0	2.6	1.0	4.2	5.6
41	80.0	6.6	4.8	0.2	1.4	1.4	0.2	0.4	1.6	3.4
40	60.0	14.6	3.4	0.4	7.0	2.6	1.2	1.4	1.8	7.6
39	72.8	15.2	9.2	0.2	0.6	0.2	0	0	0.6	1.2
38	66.2	13.6	5.2	0	4.6	2.4	1.2	0.4	0.4	6.0
35	74.2	2.8	1.2	0	2.8	1.2	0	0	8.8	9.0
32	88.0	2.4	0.4	0	1.0	0	0	0	5.2	3.0
WELL 15/17-7										
POP 1A	28.2	14.6	3.8	0	6.6	6.2	0.8	0.6	22.0	17.2
OP 40	28.8	10.2	5.6	0	7.4	4.6	2.8	0.6	22.8	17.2

APPENDIX 4B (CONT)

Sample	Wd	Wu	I	C	UP	P	M	F	A	Pd
	% Particle abundance									
WELL 15/17-7										
OP 39	25.2	16.0	3.4	0.2	7.4	2.0	1.4	0.2	21.8	22.4
38	37.6	18.6	6.8	0	2.2	1.2	1.6	0.4	7.0	24.6
37	40.0	14.6	11.6	0	2.8	2.6	1.0	0.2	7.6	19.6
36	35.4	11.6	6.8	0	5.2	5.2	2.2	0.2	9.0	24.2
35	31.4	14.6	7.4	0.4	8.0	7.8	3.2	0.2	7.6	19.4
34	43.6	11.2	13.0	0	4.4	3.0	1.6	0	8.0	15.2
33	30.8	15.4	9.8	0	3.8	5.4	1.8	0.4	16.0	16.6
26	43.4	4.4	22.8	0.4	1.2	0.4	0.2	0	18.8	8.4
POP 1B	57.0	11.0	16.8	0	1.4	0.6	0.6	0	7.6	5.0
OP 25	59.0	6.4	11.8	0	2.0	0	0.2	0.2	13.4	7.0
24	17.8	1.1	8.3	1.1	3.3	0.6	0	0	52.8	15.0
23	19.6	2.0	7.2	0.4	3.6	0	0.4	0	38.8	28.0
20	53.0	0.6	2.6	0	0.8	0	0.2	0	38.4	4.4
WELL 15/17-8A										
POP 31B	62.2	21.6	1.0	0.2	1.6	1.0	0.2	0	7.0	5.0
31A	68.0	4.4	0.4	0.2	0.8	0	0.4	0	17.8	8.0
31	46.2	27.4	1.8	0	2.2	0.8	0.6	0	12.0	9.0
30	79.8	18.0	0.2	0	0	0	0	0	0.4	1.6
27	55.0	12.6	1.8	0.2	1.0	1.0	0.8	0	5.0	22.6
26	50.2	4.8	0	0	0.4	0.4	0.2	0	39.2	4.8
25	63.0	3.4	0	0.2	0.6	0.2	0.2	0.2	14.2	18.0
22	15.6	0.4	0.2	0	0.4	0	0	0	78.0	5.4

N.B. Underlined figures in column A indicate that the amorphous organic matter fraction is predominantly collinite

APPENDIC 4C: KEROGEN SUMMARY

Lithological unit	Mean percentage particle abundances: Phytoclasts		Inertinite		A.O.M.		Palynomorphs		No. Samples*
	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	
WELL 15/17-4									
Non-marine shale	62.4	8.3	5.9	4.5	-	-	1.9	1.2	4
Organic-rich shale	64.3	12.0	5.8	4.0	0.9	0.1	14.6	6.4	7 (2)
Lower shale	53.8	9.9	9.9	3.0	1.3	1.3	16.0	4.8	2
Lower middle sandstone	62.1	14.8	7.3	4.4	4.3	5.1	9.2	8.5	5
WELL 15/17-5									
Lower shale	65.9	6.1	2.7	1.6	5.0	2.8	9.4	8.5	2
WELL 15/17-6									
Lower shale (silt)	75.0	10.3	6.9	2.8	4.3	4.5	7.7	3.0	4
Lower middle sandstone	83.8	7.8	4.6	2.5	2.9	2.9	6.7	4.1	7
WELL 15/17-7									
Lower shale (silt)	53.0	10.7	6.2	3.3	16.2	8.2	10.1	4.3	5
Lower middle sandstone	57.8	19.9	11.5	5.3	8.2		7.3	5.9	9 (3)
WELL 15/17-8A									
All samples	67.4	24.4	0.7	0.8	20.2	26.3	1.6	1.3	8

* Figures in parentheses are the number of samples with non-collinitic A.O.M. used in calcuations

Phytoclasts = Wd + Wu + I $\qquad$ $\bar{x}$ mean

Palynomorphs = UP + P + M $\qquad$ δ standard deviation

APPENDIX 4D: PALYNOFACIES DATA (Key on page 176-177)

Samples	Palynomorph Groups (%)															Counts	Tp
	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O		
POP 85	-	-	-	-	-	-	-	-	-	15.6	1.1	45.5	0.3	19.3	18.2	358	15.8
POP 84	-	-	0.2	1.5	0.2	-	-	-	2.6	9.2	2.2	6.8	0.7	12.4	64.1	412	21.2
POP 80	1.7	0.5	0.5	0.5	-	-	-	-	2.7	14.2	2.0	4.5	2.2	7.0	64.2	402	22.8
POP 79	1.6	0.5	0.7	3.0	-	-	0.2	-	2.3	9.0	1.6	9.7	-	7.8	63.5	433	17.4
POP 78	2.6	0.4	2.2	4.4	-	-	-	1.6	-	17.8	1.3	3.6	-	8.4	52.4	225	7.2
POP 75	4.7	0.5	2.8	21.5	-	-	0.9	-	16.9	1.4	0.2	4.9	0.7	6.6	38.9	427	12.6
POP 74	5.7	-	2.8	24.0	0.6	-	1.1	-	17.6	1.9	0.6	2.3	1.1	8.7	33.5	470	19.4
POP 73	5.1	-	5.2	16.7	0.2	-	1.3	-	8.1	8.8	0.2	4.7	0.6	6.9	42.3	534	22.6
POP 72	7.3	0.9	3.1	3.5	-	1.2	3.5	0.2	5.2	12.2	0.2	12.4	0.5	9.6	40.1	426	11.8
POP 60	2.1	-	0.9	0.2	-	-	-	-	2.8	2.1	0.9	18.8	0.7	9.2	62.2	426	8.4
POP 56	3.4	-	1.3	4.0	0.6	-	0.2	0.2	4.4	3.2	0.8	19.0	1.3	6.7	55.1	476	8.4
POP 50	7.9	-	3.2	7.2	-	-	0.4	-	6.1	4.3	0.7	17.6	4.7	7.5	40.5	279	8.2
POP 13	7.6	2.4	7.3	15.2	-	-	-	-	10.7	7.8	0.7	1.4	0.2	5.5	41.2	422	15.4
POP 12	10.0	1.4	3.2	16.4	-	-	0.2	-	10.2	3.0	0.2	7.2	0.7	6.3	41.2	432	3.4
POP 47	0.9	0.5	0.5	16.9	-	0.2	0.7	-	4.1	2.1	0.7	9.6	1.4	15.8	46.8	438	11.0
POP 45	8.2	-	6.4	11.6	-	0.5	0.5	-	8.9	5.9	0.2	3.9	0.2	12.5	41.2	439	9.2
POP 44	18.2	-	3.0	13.0	-	-	0.8	-	8.1	10.0	0.3	0.5	-	12.0	34.2	369	6.4
POP 43	27.4	-	4.1	8.6	-	1.0	-	-	4.6	16.2	0.5	2.5	1.0	10.2	23.9	197	9.0
POP 40	21.9	0.9	2.7	3.6	-	0.6	0.3	-	9.3	19.2	0.6	4.2	0.9	15.3	20.7	334	10.8
POP 38	7.4	-	7.4	7.9	-	0.2	0.5	0.2	15.9	5.7	0.7	4.9	0.7	10.0	38.5	403	8.2
OP 40	19.0	3.2	-	4.4	0.5	-	0.7	-	15.9	3.9	-	4.9	1.0	21.4	25.1	410	14.8
OP 39	22.9	5.8	1.7	5.6	-	-	0.2	-	26.3	2.9	0.2	4.1	0.2	14.5	15.4	414	10.8
OP 36	24.9	2.5	1.1	3.7	-	-	-	-	22.4	1.7	-	4.5	-	16.4	22.9	354	12.6
OP 35	23.7	2.3	1.8	6.0	-	-	0.5	0.3	11.3	6.5	-	3.3	-	16.9	27.4	397	19.0
OP 34	10.2	0.5	5.7	5.5	-	-	0.5	-	8.9	8.6	-	2.2	0.2	10.5	47.1	401	9.0
POP 31B	7.9	0.8	0.8	4.0	-	-	3.2	-	24.6	1.6	-	3.2	0.8	4.8	48.4	126	2.9
POP 31	8.1	-	6.2	6.6	-	1.4	4.3	0.9	9.0	1.4	-	4.3	0.5	7.6	49.7	211	3.6

APPENDIX 4E PALYNOMORPH SUMMARY (KEY ON P.176)

Lithological unit	% Palynomorph Group												
	A+C		D		A-I		J		L		M+N		No.Samples
WELL 15/17-4	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	
Lower shale	8.0	0.5	22.8	1.8	49.6	3.2	1.7	0.4	3.6	1.8	8.6	1.8	2
Lower M. sst.	10.4	0.1	10.1	9.3	30.8	8.3	10.5	2.4	8.6	5.4	8.8	1.8	2
WELL 15/17-5													
Lower shale	14.1	1.2	15.8	0.9	41.8	2.0	5.4	3.4	4.3	4.1	6.4	0.9	2
WELL 15/17-6													
Lower shale	12.4	10.1	13.8	2.8	34.7	9.2	6.0	4.0	4.7	4.6	14.0	2.8	3
Lower M. sst.	23.6	8.4	6.7	2.7	41.3	3.8	13.7	7.1	3.9	1.2	12.7	3.0	3
WELL 15/17-7													
Lower shale	21.8	4.0	5.0	0.9	53.1	13.3	3.4	0.7	4.5	0.6	18.6	5.4	2
Lower M. sst.	22.5	5.7	5.1	1.2	43.9	11.8	5.6	3.5	3.3	1.2	14.7	3.4	3
WELL 15/17-8A	11.5	4.0	5.3	1.8	49.5	9.6	1.5	0.1	3.8	0.8	6.9	1.8	2
Kimmeridge Clay Fm	6.4	2.9	11.1	6.5	31.3	13.2	23.3	20.3	20.2	2.2	1.9	1.3	6

APPENDIX 4F: GEOCHEMICAL DATA (Analyses by Robertson Research)

		Organic Carbon (%)			Rock. Eval. Pyrolysis		
Sample	Lith Unit	Org.C	Unit $\bar{X}$	δ	Hydrogen Index	Oxygen Index	
GOP 21	N.M.Shl.	1.9	1.2	0.8	13	29	
20A		0.4			-	144	
20		1.4			19	888	
19	Org.rich Shl.	2.4	4.1	2.0	91	17	
18		4.7			17	10	*
17		7.3			7	4	*
16		3.1			21	15	
15		3.1			11	14	
16A	L. sst	3.7			14	12	*
14	L. shl.	4.7	2.6	1.1	49	5	+
3		3.8			15	17	+
2		3.5			12	20	*
10		2.7			28	38	+
9		3.3			45	33	+
8		2.3					
7		1.7			18	31	*
OP 40		1.1			19	13	*
39		2.1			25	12	*
38		1.4			12	12	*
37		1.8			16	10	*
36	L.M. sst	1.3	0.8	0.4	19	10	*
35		0.7			20	20	*
34		0.6			64	47	*
13	M. Shl.	2.4	1.9	0.8	4	7	*
12		2.5			55	13	*
11	U. sst	3.2			9	26	*

* denotes samples contaminated by migrant hydrocarbons (production index $\geq$ 0.3)

\+ denotes samples with production index $\leq$ 0.2; remainder undetermined.

APPENDIX 4G: PETROGRAPHIC POINT COUNT DATA*

Figures are percentages based on 1000 counts per slide; porosity % not inclusive

Sample	Unit	Qtz	Qpx	Fp	Falk	ACM	Other	Cmt	Porosity	Name
SOP 48A	U.M. sst (α)	92.0	4.2	0.1	3.0	0.5	0.2	-	13.0	Quartz arenite
32A	U.M. sst (α)	74.2	23.3	0	1.6	1.0	0	-	15.9	Sublithic arenite
2	U.M. sst (α)	95.1	1.2	0	0	2.5	0.3	-	0.9	Quartz arenite
13	U.M. sst (β)	80.7	2.7	0.3	2.8	13.0	0.4	0.2	8.4	Quartz arenite
44	U.M. sst	88.7	1.0	0.2	4.1	3.5	0.3	2.1	10.6	Quartz arenite
26	15/17-8A	90.1	7.3	0	0.5	1.9	0.1	-	4.8	Sublithic arenite

Qtz = quartz

Qpx = polycrystalline lithic quartz

Fp = plagioclase feldspar

Falk = alkali feldspar

ACM = authigenic clay minerals

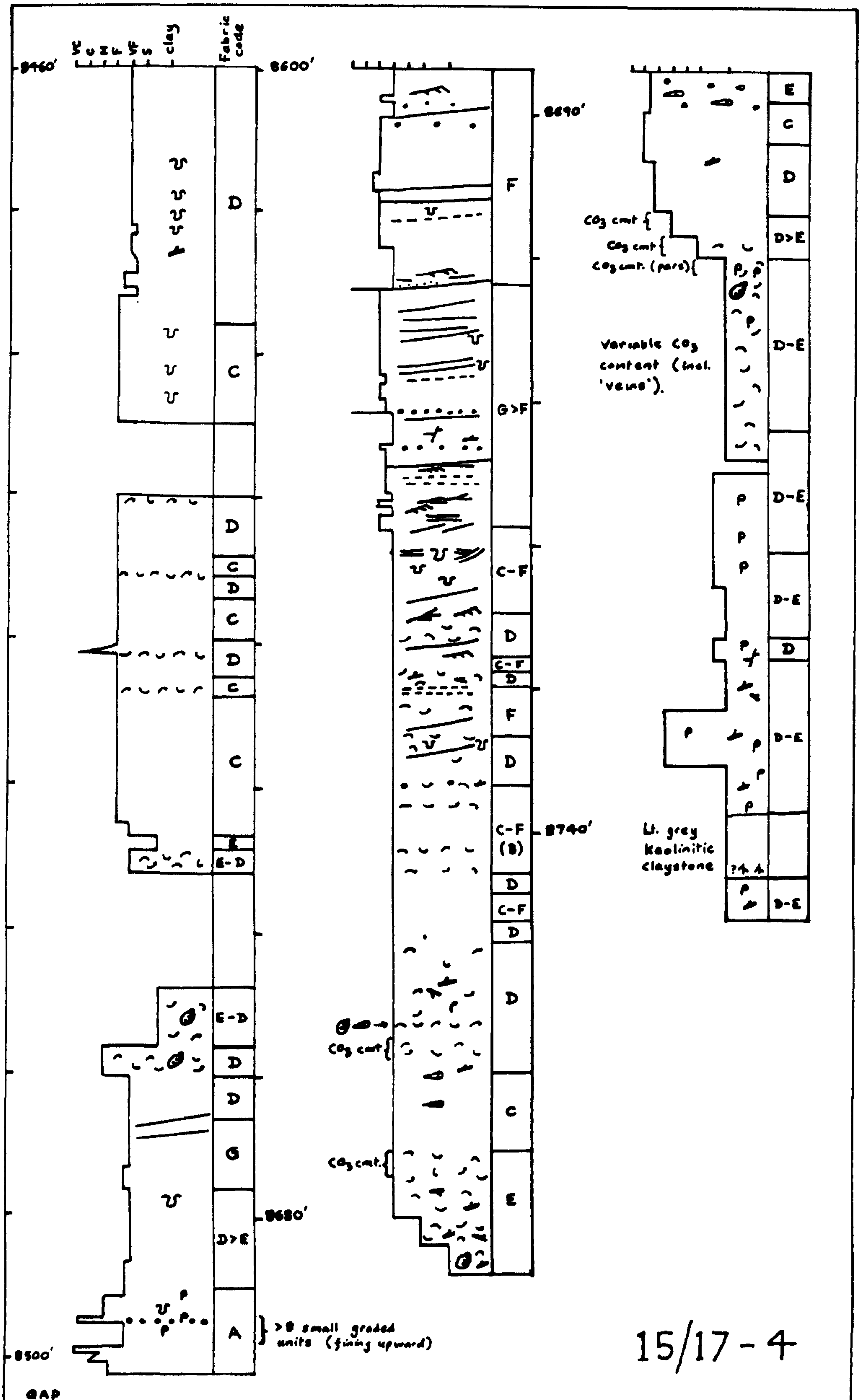
* N.B. only a few samples were point counted in order to check visual classification of sandstone type

PAGE
MISSING
IN
ORIGINAL

APPENDIX 4.H. SEDIMENTOLOGICAL LOGS

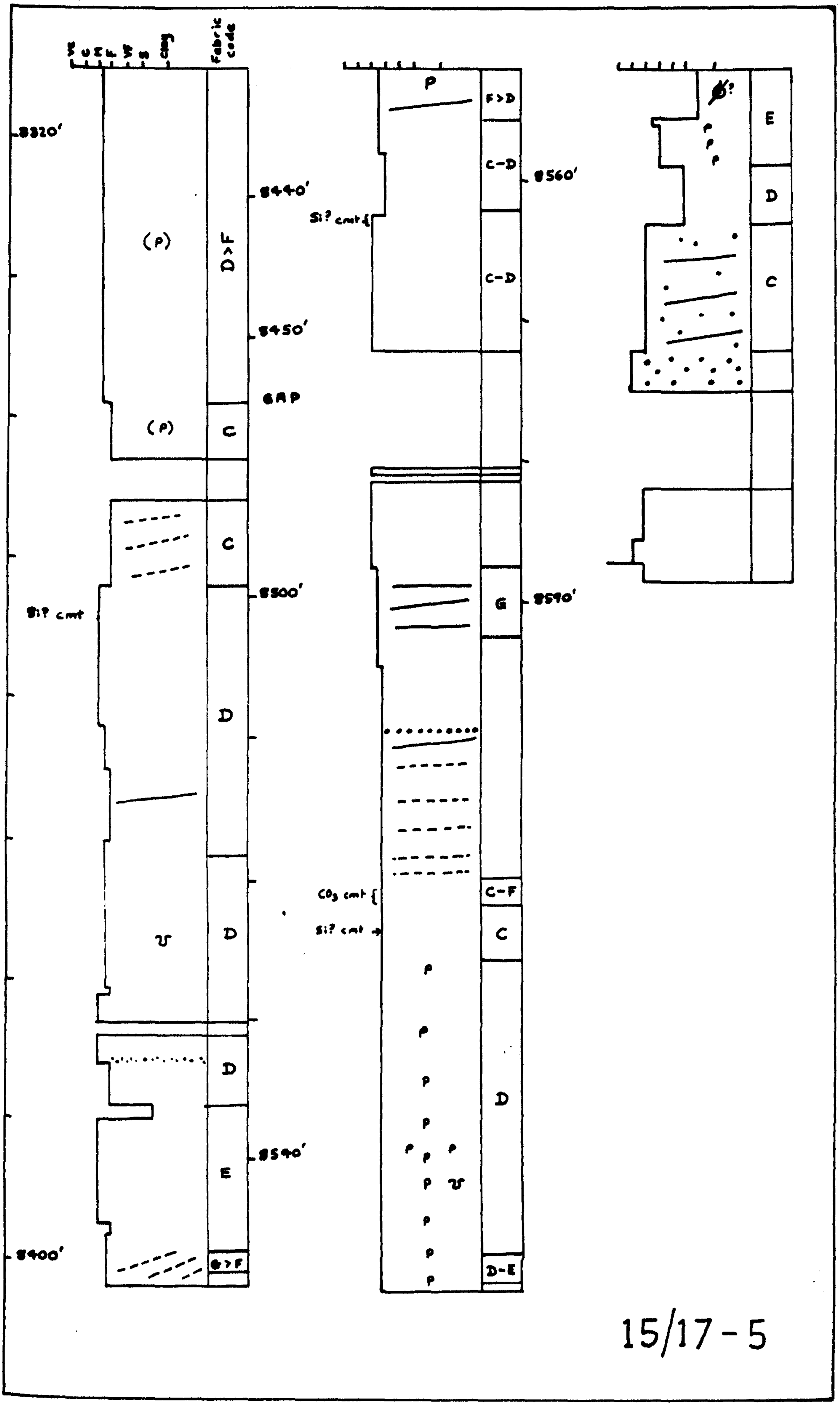
Key to codes denoting sedimentary fabric (as determined on core photographs). Where contradictory, observations indicated by symbols take precedence.

A	Structureless
B	Structureless except for isolated burrows
C	Slight mottling visible
D	Mottled
E	Bedding partially preserved (relict)
F	Indistinct feint primary bedding
G	Primary bedding (current bedding, lamination, etc.)

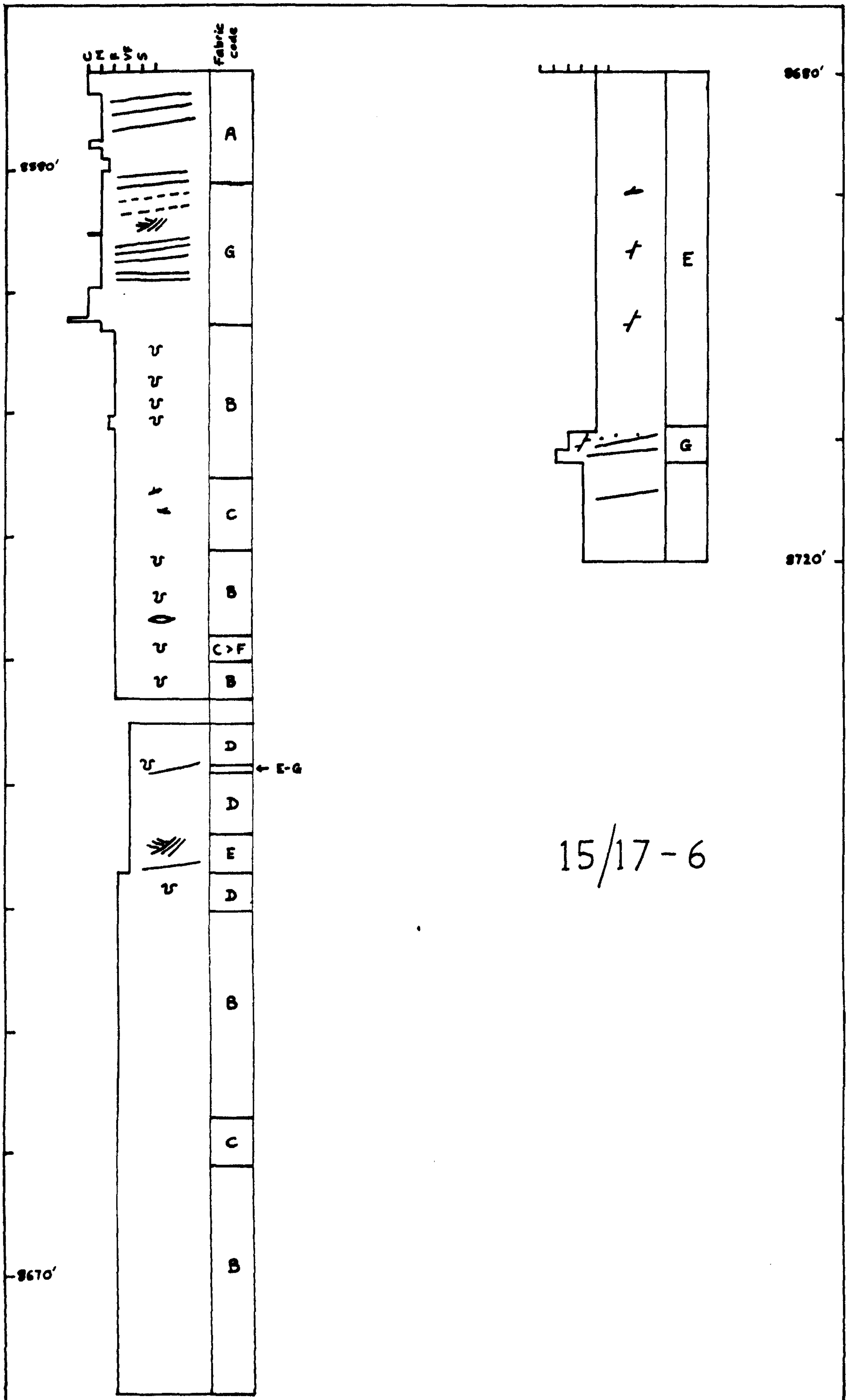


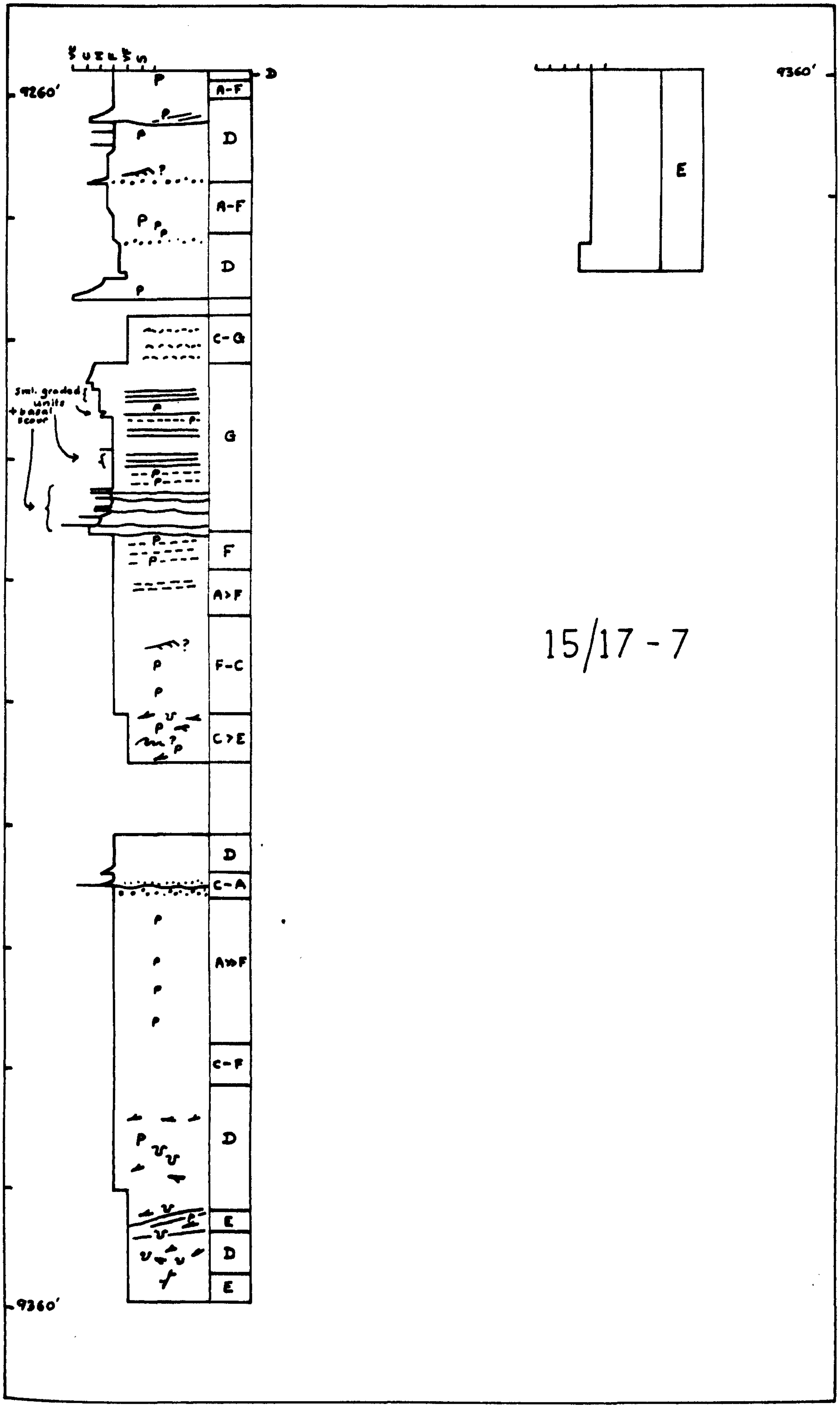
clay
Fabric code
8460'
8600'
8690'
8740'
8680'
8500'
GAP
>8 small graded units (fining upward)
CO_3 cmt
CO_3 cmt.
CO_3 cmt (pars)
Variable CO_3 content (incl. 'veins').
Lt. grey kaolinitic claystone
D
C
E
E-D
G
D>E
A
F
G>F
C-F
C-F (B)
D-E

15/17 - 4

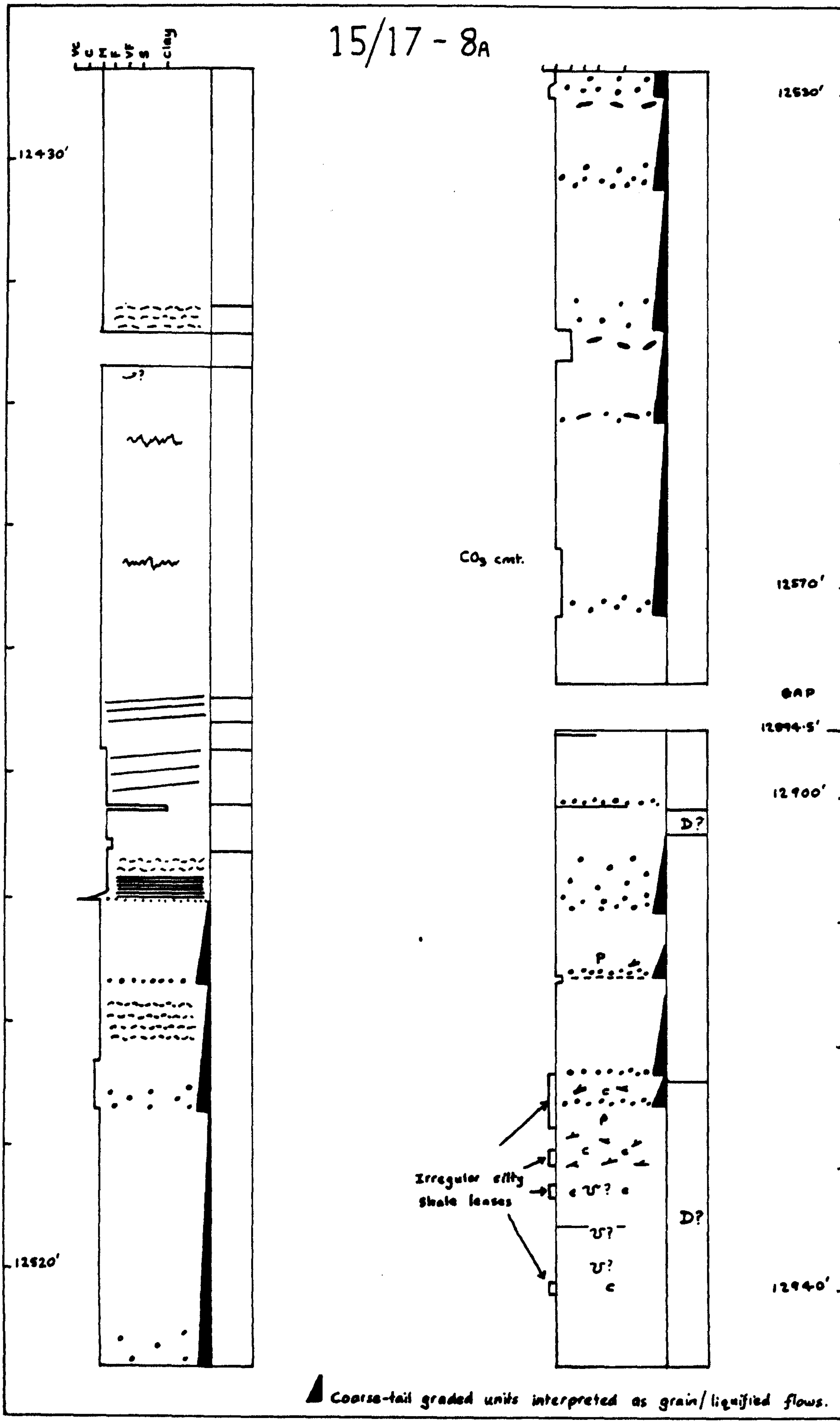

Fabric code
cmg
8320'
8440'
8450'
GAP
8500'
Si? cmt
8540'
8400'
(P)
D>F
C
D
E
G>F
P
F>D
C-D
8560'
Si? cmt
G
8590'
C-F
CO_3 cmt
Si? cmt
D-E
E
D
C

15/17-5


Fabric code
8590'
A
G
B
C
B
C > F
B
D
E-G
D
E
D
B
C
B
8670'
8680'
E
G
8720'
15/17-6


9260'
9360'
9360'
A-F
D
A-F
D
C-G
G
F
A>F
F-C
C>E
D
C-A
A>>F
C-F
D
E
D
E
E
sml. graded units + basal scour

15/17 - 7


15/17 - 8A
vc c m f vf si clay
12430'
12520'
12530'
CO3 cmt.
12570'
GAP
12894.5'
12900'
D?
P
C
Irregular silty
shale lenses
D?
12940'
Coarse-tail graded units interpreted as grain/liquified flows.

CHAPTER FIVE

Palaeoenvironmental synthesis of the type Kimmeridge Clay Formation, Dorset

INTRODUCTION

This study of the type Kimmeridge Clay Formation is based on a total of five weeks field work which was spent logging and sampling the sequence exposed between Kimmeridge Bay and Chapman's Pool in Dorset (see Fig. 5.1). About 30 petrographic and 60 palynofacies slides were examined and an extensive synthesis of the pre-existing relevent literature was also undertaken. Although this part of the thesis represents the least extensive investigation carried out during my work on the Upper Jurassic it is by no means the least significant. The importance of this particular study may be attributed to the following:-

(a) The sediments of the type Kimmeridge Clay represent the characteristic facies of the Kimmeridgian (sensu anglico) throughout large areas of England (Bristol Channel, Dorset, the south-east, English Channel, Lincolnshire, East Anglia, Yorkshire and parts of the southern North Sea ; see Chapter Eight).

(b) It provides interesting contrasts with the partly coeval sequences examined in Scotland and the northern North Sea.

(c) The sequence shows a consistent shallowing-upward trend in its upper part and is therefore useful for testing the possible onshore-offshore variations in several parameters.

(d) The cyclic sedimentation allows the detailed investigation of changing palaeoenvironments at a convenient scale.

(e) No overall integrated description and interpretation of the type Kimmeridge Clay has so far appeared in the literature, although the recent report by Cox and Gallois (1981) has made up for many of the deficiencies.

LITHOSTRATIGRAPHY

As far as I am aware the type Kimmeridge Clay has never been given a formal lithostratigraphic definition in Dorset. Although the name has received formation status in the North Sea (see Rhys, 1974 and Deegan & Scull, 1977) it has there been applied to the Kimmeridgian-Ryazanian

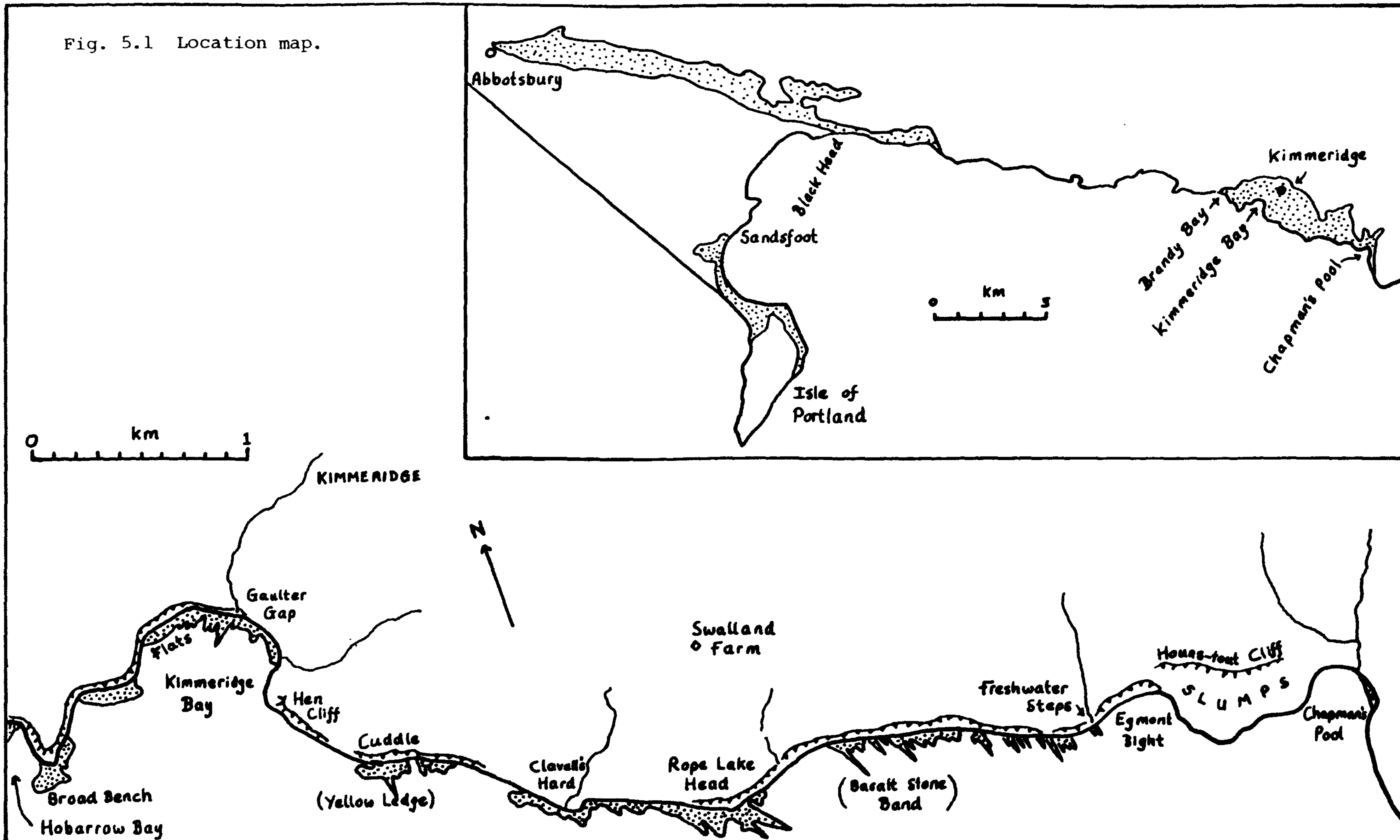


Fig. 5.1 Location map.

'hot-shale' interval in the graben sequences which, 'as shales go' is nothing like the type Kimmeridge Clay in Dorset from which the name has been extrapolated for convenience rather than accuracy. Since Arkell's classic memoir was published in 1947 the upper and lower boundaries of the Kimmeridge Clay have been somewhat modified following detailed investigations of the Portlandian and Oxfordian (Corallian) - Lower Kimmeridgian sediments. Townson's definition of the Portland Sand Formation is here accepted in unaltered form (see Table 5.1) with Arkell's Rhynconella Marls and succeeding units placed in the Black Nore Member and transferred to the Portland Group. The lithostratigraphic base of the Kimmeridge Clay is somewhat more problematical. Brookfield (1978a) has defined a "Passage Beds (Formation)" which includes all the sediments between the top of the cymodoce zone and the base of the Ringstead Waxy Clay (Brookfield, 1978a, p.6-7), and which forms the top part of an "Upper Calcareous Grit Group". The boundary with the Kimmeridge Clay ("Group") was taken somewhat arbitrarily (Brookfield, 1978a, p.27) at the horizon of large septarian concretions marking the cymodoce mutabils boundary (the "pumpkin-shaped" concretions of Cox and Gallois, 1981, p.33-34). Brookfield (1978a) resurrected the idea of a 'Passage Beds' unit from Blake (1875) and, in so doing, moved the base of the Kimmeridge Clay upward from its traditional position at the base of the Rhactorhynchia inconstans Bed (Arkell, 1947). Brookfield's arguments for making these changes include:

(a) "The natural sedimentary breaks within the beds studied .... do not correspond with the division between the Upper Calcareous Grit and Kimmeridge Clay" of Arkell. "The major breaks occur at the base of the Ringstead Waxy Clay and at the top of the Grey Clays of the upper cymodoce zone and these breaks were recognised by Blake (1875) who called the lower half of the beds between these two levels the Kimmeridge Passage Beds" (Brookfield, 1978a, p.5-7).

(b) "The Passage Beds contain similar fine-grained arenaceous facies,

TABLE 5.1 Revised Late Jurassic

Lithostratigraphy of Dorset

GROUP	FORMATION	MEMBER	INFORMAL UNITS	
PURBECK GROUP	Durlston Fm. Lulworth Fm.			
PORTLAND GROUP *	Portland Limestone Fm.	Winspit Mbr. Dancing Ledge Mbr. Dungy Head Mbr. Gad Cliff Mbr. Pondfield Mbr. Corton Hill Mbr.		
	Portland Sand Fm.	Black Nore Mbr.		Emmit Hill Beds Hounstout Marls Hounstout Clay Rhynconella Marls
	Kimmeridge Clay Fm. (this work)	Hounstout Mbr.	Arkell, 1947	Lingula Shales Rotunda clays and nodules Crushed ammonoid shales Shales and clays (un-named)
		Swalland Mbr.		Dicey clays Bituminous shales
		Kimmeridge Mbr.		Shales (un-named) Cattle Ledge shales Hen Cliff shales Maple Ledge Shales Gaulters Gap shales Washing Ledge shales Shales (un-named)
		Black Head Mbr.	Brookfield, 1978	Grey clays Sand clays Blue clays *Ostrea delta* bed *Exogyra nana* bed *Rhactorhyncia inconstans* bed Ringstead Coral Bed Ringstead Waxy Clay
OSMINGTON OOLITE group				

* After Townsen 1971, 1975.

have a uniform clay mineralogy throughout, distinct from beds above and below, and contain a macrofauna transitional between the Oxfordian and Kimmeridgian" (Brookfield, 1978a, p.7).

(c) "The junction between the Upper Calcareous Grit and Kimmeridge Clay was formerly placed at the top of the Ringstead Coral Bed, between the _pseudocordata_ and _baylei_ zones. However, this makes the junction lie within the Passage Beds sequence, and it has little lithological or faunal significance" (Brookfield, 1978a, p.7).

(d) The Kimmeridge Clay is "redefined to include the grey-black clays and shales above the Passage Beds on the Dorset Coast". "At Black Head, the junction between the Passage Beds and Kimmeridge Clay is marked by an impoverishment in macrofauna and microfauna. Only _Nanogyra virgula_, _Thracia depressa_ and ammonites are common in the Kimmeridge Clay. The clay mineral assemblages change from the illite-kaolinite assemblages of the Passage Beds into the illite-chlorite assemblages of the Kimmeridge Clay. The junction with the underlying Grey Clays is gradational but is conveniently taken above the horizon with large septarian concretions up to one metre in diameter" (Brookfield, 1978a, p.26-27).

(e) "A marked change in the invertebrate faunas ... occurs at the base of the _mutabilis_ zone, and is apparently ecologically controlled by the ubiquitous development of a black shale environment at this time in all British basins. However, nektonic forms show little differentiation and microplankton remains abundant. A continuous evolutionary series of the ammonites _Ringsteadia_, _Pictonia_, _Rasenia_, and _Aulacostephanus_ is traceable from the _pseudocordata_ zone to the _eudoxus_ zone" (Brookfield, 1978a, p.28).

Cox and Gallois (1981, p.1-2) have recently rejected Brookfield's 'Passage Beds (Formation)' on the following grounds: "Although Brookfield cited faunal, lithological and mineralogical evidence to support the case for the new formation, his main contention was that the Kimmeridge Clay (_mutabilis_ zone and above in his definition) is composed of uniform black shales with an impoverished macrofauna and microfauna, whereas the 'Passage Beds' are dominantly arenaceous and contain a typical Corallian fauna. These observations are not in accord with those of the present authors who regard the lithologies and faunas of the Kimmeridge Clay part of the 'Passage Beds' as typical of the Kimmeridge Clay".

Although there are valid criticisms that can be made of the Passage Beds, those above appear to be based almost entirely on misrepresentation. Brookfield does not claim that mutabilis and higher zones of the Kimmeridge Clay consist of "uniform black shales", but states only that a ubiquitous development of such environmental conditions occurred in the mutabilis zone, elsewhere describing the overlying Kimmeridge Clay as 'grey-black clays and shales' (see (d) above). Although Brookfield (1978a) does, perhaps unwisely, describe the Passage Beds as "dominantly arenaceous" in his abstract, it is clear that this was not an accurate generalisation of his observations since his sections show the sequence to be dominantly argillaceous and elsewhere (Brookfield, 1978b, p.260) he describes it as constituting a 'clay megafacies' and provides quantitative data on the sand fraction (see also (b) above). Nowhere does Brookfield (1978a) claim, as Cox and Gallois (1981) suppose, that the Passage Beds have a typically Corallian fauna, only that the macrofauna is transitional between the Oxfordian and Kimmeridgian. Since Brookfield's studies were confined to the cautisnigrae-mutabilis zones, I believe he was more concerned with emphasising the lithofacies and biofacies continuity across the Oxfordian-Kimmeridgian boundary than trying to deliberately overthrow the status quo of the Upper Jurassic lithostratigraphy. If anything, his "main contention" appears to have been that the lithostratigraphic boundary should be significant as well as convenient. His accounts are still among the most detailed for this part of the Jurassic in Dorset (see Brookfield, 1973a, 1973b, 1978a, 1978b).

The major question that must be asked of Brookfield's proposed

Passage Bed (Formation) is whether it is a sufficiently significant, distinct and mappable lithological unit. Certainly Cox and Gallois (1981, p.2) do not think so as they regard the whole baylei-eudoxus interval as being essentially of one lithofacies composition. By his own admission, Brookfield (1978a, p.27) regards the junction between mutabilis sediments and his underlying Grey Clay unit as 'transitional' and I am rather doubtful of whether a lithostratigraphic unit should have one of its boundaries taken at a horizon of concretions, however striking they may seem in the local area. Although I am naturally wary, not having worked on this lower part of the Kimmeridge Clay, it would seem a reasonable compromise to elevate the top of Brookfield's unit to the top of the uppermost of the two shelly siltstone bands occurring just above the base of the eudoxus zone (see Cox and Gallois, 1981; Blackhead Section p.33). Such a compromise would have the advantages of including all the silty mudstone-siltstone horizons of the Lower Kimmeridgian in one lithostratigraphic unit as well as separating out the baylei-mutabilis zones, which do not contain kerogenous shales, from the overlying kerogenous parts of the Kimmeridge Clay.

Although Brookfield (1978a) did not actually formally define lithostratigraphic units, he did indicate in parentheses the lithostratigraphic rank or status of his new units, hence "Passage Beds (Formation)". I believe the lithostratigraphic rank he gave to most of these units must be changed. The Passage Beds interval, even as enlarged above, is insufficiently distinct to separate it as a Formation from the Kimmeridge Clay, which in turn is sufficiently homogeneous to have only Formation rather than Group status. Internally, Brookfield's Passage Bed unit is also essentially homogeneous, consisting predominantly of variably silty and sandy mudstones, a degree of variation warranting member rather than formation rank. Although the term Passage Beds does have some historical significance, a geographically derived name would be more suitable for modern lithostratigraphic nomenclature; the name Blackhead Member is proposed as an alternative after the main section at this locality (see Brookfield 1978a, 1973a; Cox and Gallois, 1981). Having now manufactured a Blackhead Member ranging from the base of the Ringstead Waxy Clay (upper cautisnigrae zone) to the base of the eudoxus zone (an extended version of the Passage Beds unit) the next logical step is to make this the lowest member of the Kimmeridge Clay Formation. This manoevure lowers the base of the latter about 3m, rather than raising it about 9m by the inclusion of a separate Passage Beds Formation as proposed by Brookfield (1978a). Such a slight change solves many problems. Brookfield's emphasis on lithofacies continuity across the Oxfordian-Kimmeridgian stage boundary is retained and the Kimmeridge Clay Formation finally becomes a lithostratigraphic term in its own right without being confusingly coeval with the Kimmeridgian stage (sensu anglico). The tendency for British geologists to produce biostratigraphically defined lithostratigraphic units is a rather disturbing one which conceals the fact that they should be derived from different, independent sets of criteria. Details of the redefined lithostratigraphy of the Kimmeridge Clay Formation are given below, and are summarised in Table 5.1.

THE KIMMERIDGE CLAY FORMATION

(i) The Blackhead Member (after Black Head, Dorset)

Type localities: Black Head, Dorset (SY 722819 to SY 728819) and Osmington Mills, Dorset (SY 7336 8186 and SY 7342 8174), see Brookfield, 1973a, 1978a and Cox & Gallois (1981) and Fig. 5.1.

Age: upper cautisnigrae zone to lowest bed of eudoxus zone (Oxfordian-L. Kimmeridgian).

Base: contact between the Ringstead Waxy Clay and the Sandsfoot Grit (see Brookfield 1973a, 1978a and Table 5.1).

Top: prominent shelly siltstones situated approximately at the eudoxus-mutabilis zone boundary (see Cox and Gallois, 1981, p.33-34).

Thickness: approximately 41m at Black Head.

Lithological character: predominantly dark grey to pale grey, variably calcareous mudstones with variably silty and sandy bands (see Brookfield, 1973a, 1978a, 1978b and Cox & Gallois, 1981 for details). No kerogenous shales or kerogenous dolostones are present. Partly equivalent to the Abbotsbury Members (Abbotsbury Sandstone + Abbotsbury Ironstone sensu Brookfield 1978a).

(ii) The Kimmeridge Member (after Kimmeridge, Dorset)

Type localities: Hoborrow Bay, Dorset (SY 896790) eastwards around Kimmeridge Bay (SY 908791) to Clavell's Hard (SY 920777). Supplementary sections at Ringstead Bay and Brandy Bay (see Cox and Gallois, 1981, p.35-38). Most of the eudoxus zone is unexposed but recorded in the Broad Bench No. 1 borehole (see Cox & Gallois, 1981, p.5).

Age: eudoxus zone to wheatleyensis zone (pars).

Top: base of second coccolithic kerogenous laminite (fourth hard band) below the Blackstone oil shale (see log in appendix).

Thickness: 230m in the type area including about 93m of eudoxus sediments present only in boreholes.

Lithological character: essentially a series of alternating mudstone-shales and kerogenous shales (the clays and bituminous shales of Downie 1955). The carbonate content of the mudstone-shales is generally 5-15% (Dunn, 1972; Braide, 1976) but secondary, diagenetic kerogenous and non-kerogenous dolostones occur (see Bellamy, 1977, 1979 and later). The proportions of the lithologies in the exposed part of the member are ∿70% mudstone-shales, ∿20% kerogenous shales and ∿10% dolostone. The average organic carbon content for the exposed section is 6.4% (calculated from data in Davies, 1978). The individual lithologies present are partly discussed later but see also Aigner (1980). Litho-logical logs of this member can be found in Downie, 1955 and Cox and Gallois (1981).

(iii) The Swalland Member (after Swalland Farm, Kimmeridge, Dorset).

Type localities: Clavell's Hard (SY 920777) to Freshwater Steps

(SY 944772). Supplementary sections for parts of this member also occur in Brandy Bay and Ringstead Bay (see Cox and Gallois, 1981, p. 35-38).

Age: wheatleyensis zone (pars) to pectinatus zone (lower part of paravirgatus subzone).

Top: top of the 75 cm coccolithic kerogenous laminite band above the Freshwater Steps Stone Band (i.e. 2m above the latter); see log in appendix for details.

Thickness: 71.4m according to my own observations and log in the appendix, 79.1m according to that of Cox and Gallois (1981, p.42-43), and 76.8m according to the log of Downie (1955).

Lithological character: a more complex alternation of mudstone-shales, kerogenous shales, oil shales, coccolithic kerogenous laminites (kerogenous laminated marls), coccolithic marls, coccolith rhythmites and secondary non-kerogenous dolostones. The proportion of the major lithologies are ∿70% mudstone-shale, 12% kerogenous shale, 8% kerogenous coccolithic sediment, 4.5% non-kerogenous dolostone, 3% oil shales, 2.0% coccolithic marls and 0.5% coccolith rhythmites (figures calculated from the lithological log made during this study - see appendix). The average organic carbon content of the member is 6.3% (calculated from data in Davies, 1978); individual bands show much higher values - see later. The member is rendered distinctive by the abundance of kerogenous sediments, the presence of coccolith rich lithologies (including true limestones) and the absence of kerogenous dolostones.

(iv) The Hounstout Member (after Hounstout Cliff, Dorset)

Type localities: from Freshwater Steps (SY 944772) around Egmont Bight to Chapman's Pool (SY 955771) and including the lower slopes of Hounstout Cliff (SY 949771).

Age: pectinatus zone (pars) to fittoni zone (pars).

Top: the top of the member and the Kimmeridge Clay Formation is taken at the boundary btween the "Lingula Shales" and "Rhynconella Marls" (see Townson, 1975). The "Rhynconella Marls" at the base of the overlying Black Nore Member are "black, dolomitic, laminated, argillaceous siltstones" (Townson, 1975, p.622).

Thickness: the total thickness of the member is rather difficult to determine exactly owing to the slumped and presently poorly-exposed nature of much of Hounstout Cliff. The most accurate existing logs of this part of the sequence are those of Downie (1955; graphic) and Cope (1978; written and graphic). Based on these logs the estimated thickness of the member is 104m (Downie, 1955) to 120m (Cope, 1978). The lower part is also shown by Cox & Gallois (1981, p.43-44).

Lithological character: predominantly 'normal', usually calcareous, mudstones and shales becoming more silty and sandy upwards. The proportions of the various lithologies are: 92.5% mudstones, 6.4% kerogenous shales and 1.1% coccolithic kerogenous laminites. The kerogenous beds occur mainly as thin bands (15-30 cm) which form conspicuous seepages. The predominant organic carbon values lie in the range 1-2% (Davies, 1978). Concretions and nodules are quite common in this unit although rare in the Swalland Member.

Kimmeridge Clay Formation total thickness approximately 450-460m in the type area.

Comments on the proposed lithostratigraphic terminology

The objectives of the proposed lithostratigraphic scheme have been to formalise the type Kimmeridge Clay Formation, to emphasise the lithological variation that occurs in this thick sequence, and to highlight the contrasts between the type Formation and its name-sake in the North Sea Basins. The base of the Kimmeridge Clay has been lowered to include the Upper Oxfordian Ringstead Waxy Clay in order that the lower boundary of the Formation corresponds with a gross lithological change - from clays to the Sandsfoot Grit sandstones. This shift of the boundary also emphasises the gross lithological continuity across the Oxfordian-Kimmeridgian stage boundary. The top of the Kimmeridge Clay Formation has already been defined by default by Townson's work on the Portland

Group (Townson, 1971, 1975). The newly defined Formation has been divided into four members whose names have been taken from their main localities, the Black Head, Kimmeridge, Swalland and Hounstout Members. The key diagnostic features of these members are given below:-

(i) Blackhead Member: variably silt and sandy mudstones; faunally diverse, no kerogenous shales.

(ii) Kimmeridge Member: mudstones alternating with abundant kerogenous shale bands; faunally impoverished; contains prominent kerogenous dolostone bands (q.v. Bellamy, 1977, 1979 and Irwin, 1980).

(iii) Swalland Member: predominantly mudstones but with common kerogenous shales, true oil shales, coccolithic kerogenous laminites and coccolith limestones (marls and rhythmites); only non-kerogenous dolostones; faunally impoverished.

(iv) Hounstout Member: mudstones with only rare kerogenous shales and no stone bands (dolostones or coccolith limestones). Increasingly silty upward and with a less impoverished fauna and much lower organic carbon values than (ii) or (iii).

Only the Swalland Member was studied in any detail in this investigation. Some 132m of the section were logged which included 10m at the top of the Kimmeridge Member and 47m at the base of the Hounstout Member. For the greater part of the sequence Downie's (1955) log was found to be reliable and only in the Swalland Member was it considered necessary to add additional detail. Cox and Gallois' (1981) recent logs are correct for the most part but contain several lithological misidentifications and use different lithological terms to those in this work. Cope (1978) provides a useful log covering the difficult Hounstout Cliff section, which since it was logged by Arkell (1936-37) has significantly deteriorated due to slumping; the 'old road' shown on Arkell's diagrammatic section (Arkell, 1947, Fig. 1.6) has now been cut back to the sky line.

LITHOLOGIES OF THE TYPE KIMMERIDGE CLAY FORMATION

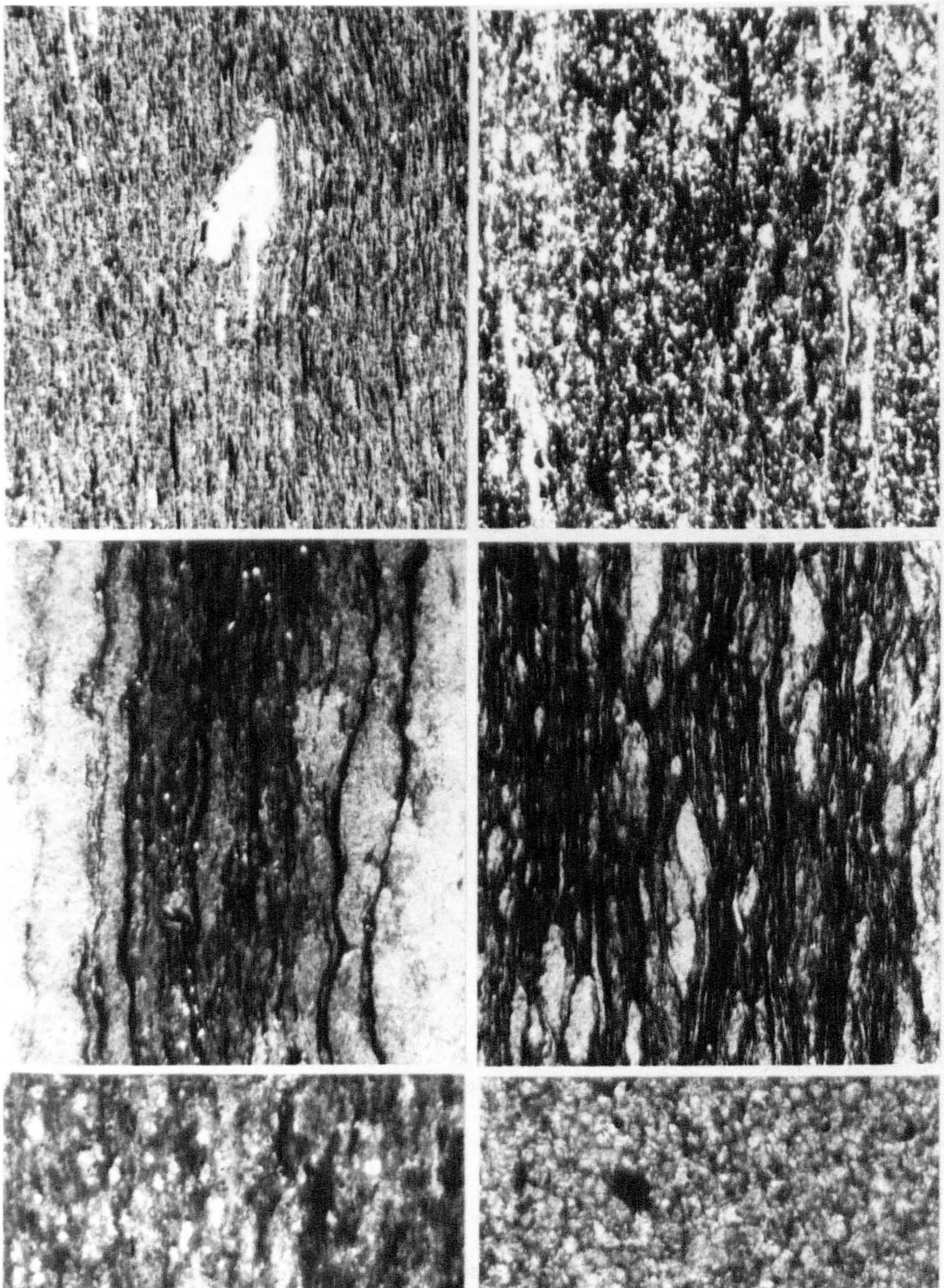
(A) Normal mudstone-shales ('Clays' of Downie, 1955)

The 'normal' mudstones of the Kimmeridge Clay vary in colour from pale grey to medium-dark grey. Their colour is largely determined by their carbonate and organic content relative to the clay matrix; carbonates give the clays a paler colour while high organic contents tend to make the rock a darker, browner colour. Carbonate and kerogen contents also have a strong influence on the fabric of the sediment; the more calcareous mudstones tend to be either 'blocky' with a sub-conchoidal fracture or 'dicey', and the organic-rich mudstones more shaley. The most typical 'dicey' clays occur between the Short Joint Coal and Basalt Stone (see log). The organic content of the mudstones generally lies between 1.0-8.0% Org.C in the Kimmeridge and Swalland Members and ≤4% Org.C in the Hounstout Member (Downie, 1955; Dunn, 1972; Davies, 1978). In the field the only indication of bedding is the parallel orientation of shell debris but in thin section it is also shown by the sub-parallel orientation of A.O.M. particles. No lamination exists as such (with the possible exception of the most kerogenous levels) but neither is bioturbation apparent except in some of the more calcareous mudstones.

The formation as a whole has a monotonous clay mineral assemblage consisting of illite>>kaolinite>mixed layer clays (Downie, 1955; Cosgrove, 1970; Braide, 1976; Davies, 1978; Bellamy, 1979). With the exception of the upper Hounstout Member sand grade material is rare. Preliminary analysis of sediments from the Swalland Member by J. Exton has shown sand contents generally <0.5%, dominated by pyrite and biogenic materials. Fine silt-sized material is fairly common (5-20%) and is dominated by quartz (Downie, 1955; pers. obs.). Bellamy (1979) has observed isolated siltstone lenses showing poorly developed ripple lamination at the base of the elegans zone. The silty and sandy parts of the Black Nore Member (see Neaverson, 1925, and Lloyd, 1959 for nature of the coarse fraction)

PLATE 5.1. Petrography of the principle Kimmeridge Clay lithologies. Figures in parentheses are the widths (long dimension) of the photographs in millimetres.

Top Left:	Oil shale ; predominantly orange coloured "flakes" of kerogen with a phosphatic skeletal fragment in the centre of the photograph. Sample BS30 (2.2)
Top Right:	Kerogenous shale; higher silt content and darker colour due to higher amounts of clay. Sample RVT 19 (2.2)
Centre Left:	Coccolithic limestone rhythmite showing the thicker, 'massive' and peletal varieties of laminae. Sample FWS 23 (2.2)
Centre Right:	Coccolithic kerogenous laminate exhibiting coccolithic faecal pellets in a laminated matrix rich in kerogen "flakes". Sample WS 17 (2.2)
Bottom Left:	Kerogenous dolostone. Sample YL 2 (0.85)
Bottom Right:	Non-kerogenous dolostone. Sample CL 4 (0.85)



also show occasional small scale primary current structures which are apparently absent or undiscernable in the Swalland Member. According to Bellamy (1979) coccoliths are the single most important source of primary calcite in the Kimmeridge Clay and Gallois and Medd (1979) report 5-12% coccoliths in the mudstones on a whole rock basis. Diagenetic carbonate nodules are most common in the Hounstout Member.

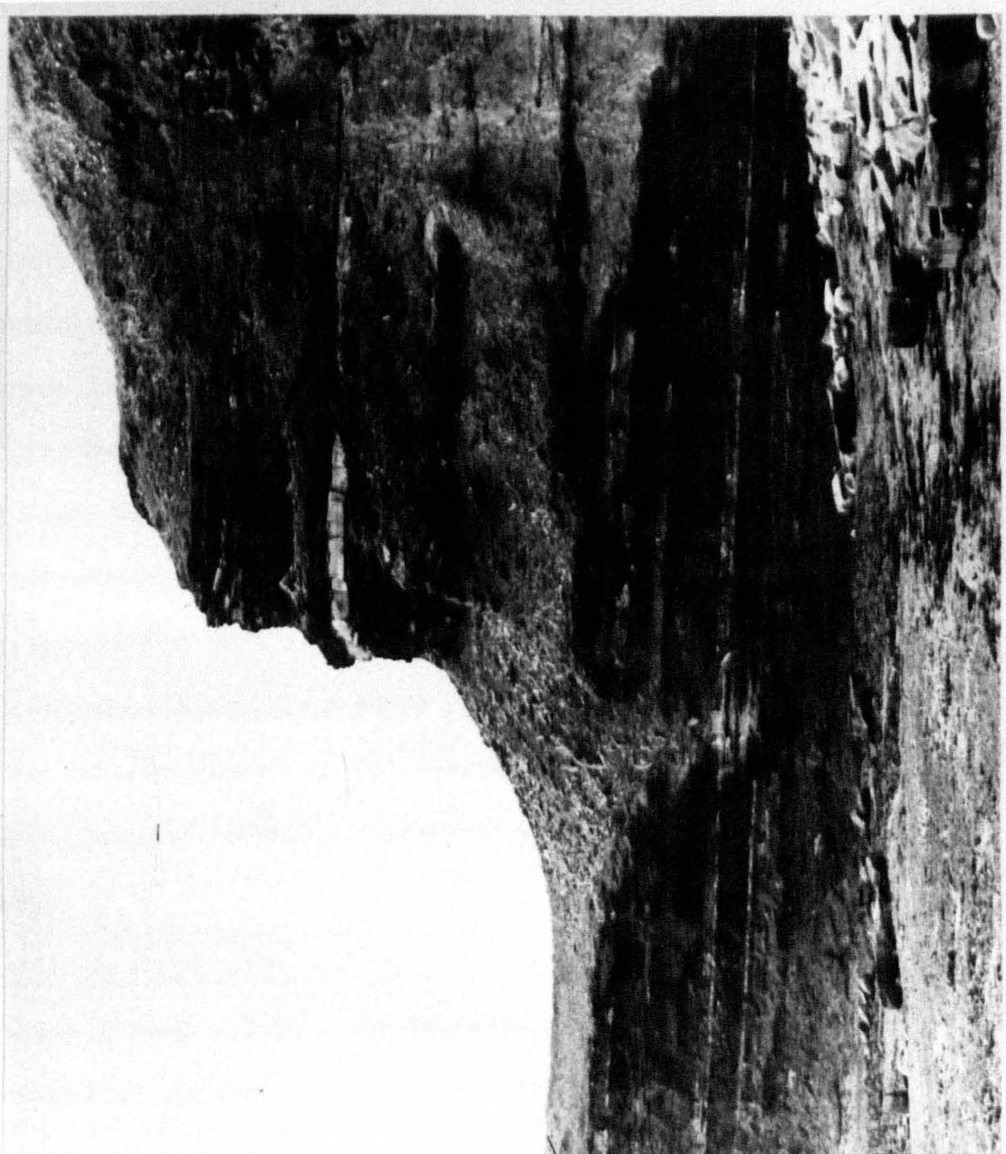
(B) Kerogenous Shales ('bituminous shales' of Downie, 1955)

The main lithological difference between the normal mudstone-shales and kerogenous shales is their higher organic content and its effects on the rock fabric. Organic contents range from 8.0-23.0% Org.C. and this is reflected in the abundance of orange-brown A.O.M. particles observed in thin section. Lamination and fissility are better developed but their palaeontological character is perhaps their most distinctive feature. It should be noted that Gallois (1976, 1978), Gallois & Cox (1974, 1976) and Cox & Gallois (1981) do not differentiate between kerogenous shales and oil shales as used here and in Downie (1955) and Tyson et al. (1979), but call all the kerogenous sediments oil shales. This discrepancy has led to conflicting accounts about the nature of the oil shales, in particular their faunal characteristics.

(C) Oil Shales

These distinctive sediments are mainly restricted to the Swalland Member, but there are a few occurrences of true oil shales in the Kimmeridge Member. The classic examples are the Blackstone (or 'Kimmeridge Coal' which is actually a composite unit when examined in detail), the Bubbicum (or 'Ruddicum' of Strahan, 1920) and the top and base of the Short Joint Coal (e.g. see Plate 5.2). Analyses of the Blackstone conducted on my behalf by Roberston Research have shown organic carbon values of 30-53%, and in the Short Joint Coal 26-38%. The exceptional abundance of kerogen in these rocks gives them a dull, dark brown colour and also tends to 'weld' them together making the bands massive and resistant. They are very tough and often splinter when

PLATE 5.2 Base of the Swalland Member at Clavell's Hard. For location see Fig. 5.1. The prominent light-coloured horizon is the Rope Lake Head Stone Band. The scree slope half-way up the section marks an old adit working of the Blackstone oil shale.



hammered producing fragments with knife-like edges. Their carbonate content is generally low but varies (< 10%) depending on the amount of cream-coloured coccolithic pellets (some of which are as large as 1cm in diameter). In thin section the oil shales are bright orange-brown in colour owing to the kerogen content which reaches up to 69% (derived by multiplying organic carbon by a factor of 1.3). They are microlaminated but show little indication of organic-mineral couplets; this is a result of the comparatively low carbonate and clay contents which are insufficient to interrupt what is virtually a continuous kerogen matrix. The statement made by Morris (1980) to the effect that "the oil shales which appear laminated in hand specimen are not finely laminated in thin section" is considered somewhat misleading. Micro-boudinage is a feature commonly seen in thin sections of the oil shales and larger plastic deformation structures are probably rather common but are difficult to make out due to the lack of clearly recognisable horizons within the bands. Compactional folding around the pyritic diagenetic carbonate nodules is well developed in the Blackstone. There can be little doubt that the oil shales probably represent at most 25% of their original thickness.

(D) Coccolith rhythmites (coccolith limestones of Downie, 1955)

Gallois & Medd (1979) have defined 'coccolith-rich bands' as sediments having greater than 50% calcite, of which 30% or more consists of recognisable coccoliths. While accepting this as a reasonable qualification of the descriptive prefix 'coccolithic' I have preferred to subdivide thse bands according to their gross character as observed in the field into the rhythmites, coccolithic marls and coccolithic kerogenous laminites. Only three really well developed coccolith rhythmites occur in the Kimmeridge Clay Formation; they are located within the White Stone Band, Middle Band and Freshwater Steps Stone Band, which all occur in the upper part of the Swalland Member. The classic example and the purest limestone of the three (98% $CaCO_3$) is the White

Stone Band; the Middle Band and Freshwater Steps Stone Band tend to be more argillaceous and kerogenous. All three are composite and it should be remembered that the term 'bands' refers more to the weathering character of these units than it does to their lithology. The calcareous core of the Short Joint Coal Band lower in the sequence was described by Downie (1955) as coccolithic and to contain calcite rhombs, calcispheres and faecal pellets. When examined under the S.E.M. it appeared to consist totally of euhedral calcite rhombs and the same observation was made by Bellamy (1979). If it was a coccolithic unit it may have been a rhythmite but all its internal structure was subsequently destroyed (see below). The 'microstratigraphy' of the White Stone Band and Freshwater Steps Stone Band is shown in Fig. 5.2 and Plate 5.3. The internal complexity and heterogeneity of these units should be a cautionary tale for anyone attempting to make sweeping generalisations about their characteristics.

The coccolith rhythmites are composed of three elements:-

(i) Finely microlaminated units consisting essentially of coccolith-kerogen alternations. In detail two sorts of laminae may be discerned, the first type consists of coccolithic pellets (0.2-1.0 mm long) inter-layered with kerogen 'flakes' (which have folded around the pellets during compaction) and the second type consists of relatively homogeneous, thicker laminae (0.25-2.0 mm) composed totally of coccoliths. The second type of laminae is often bound by thin planar kerogen laminae.

(ii) Dominantly microlaminated units which contain occasional burrows (described later) which often extend downward from a single datum. Such a unit is conspicuous within the White Stone Band and the whole of the Freshwater Steps Stone Band appears to be of this character. The two types of laminae described in (i) above are still recognisable and the bioturbation is too sparse to have caused much disruption.

(iii) Deformed units. Various scales and types of soft sediment deformation structures occur within the coccolith rhythmites. Microfaults

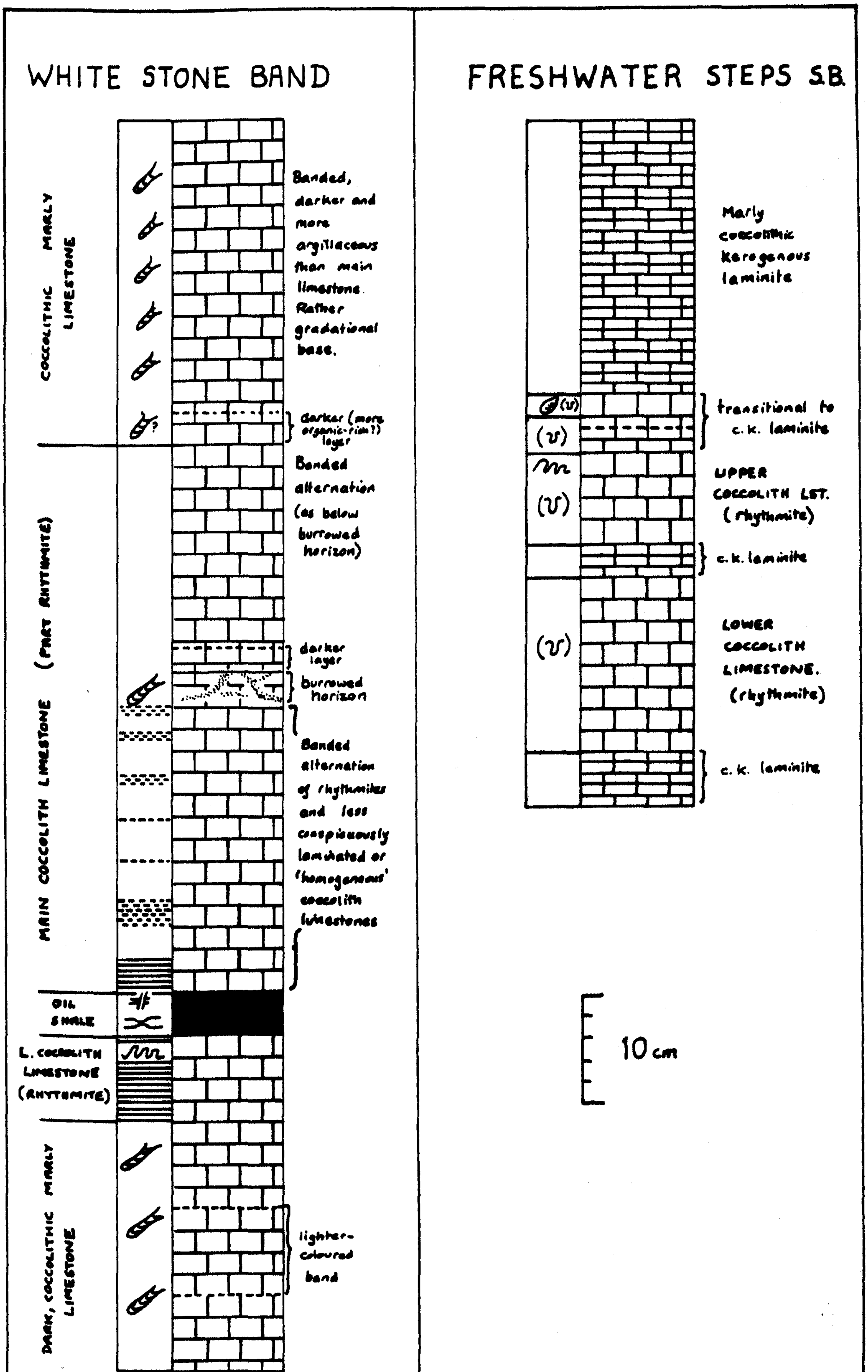
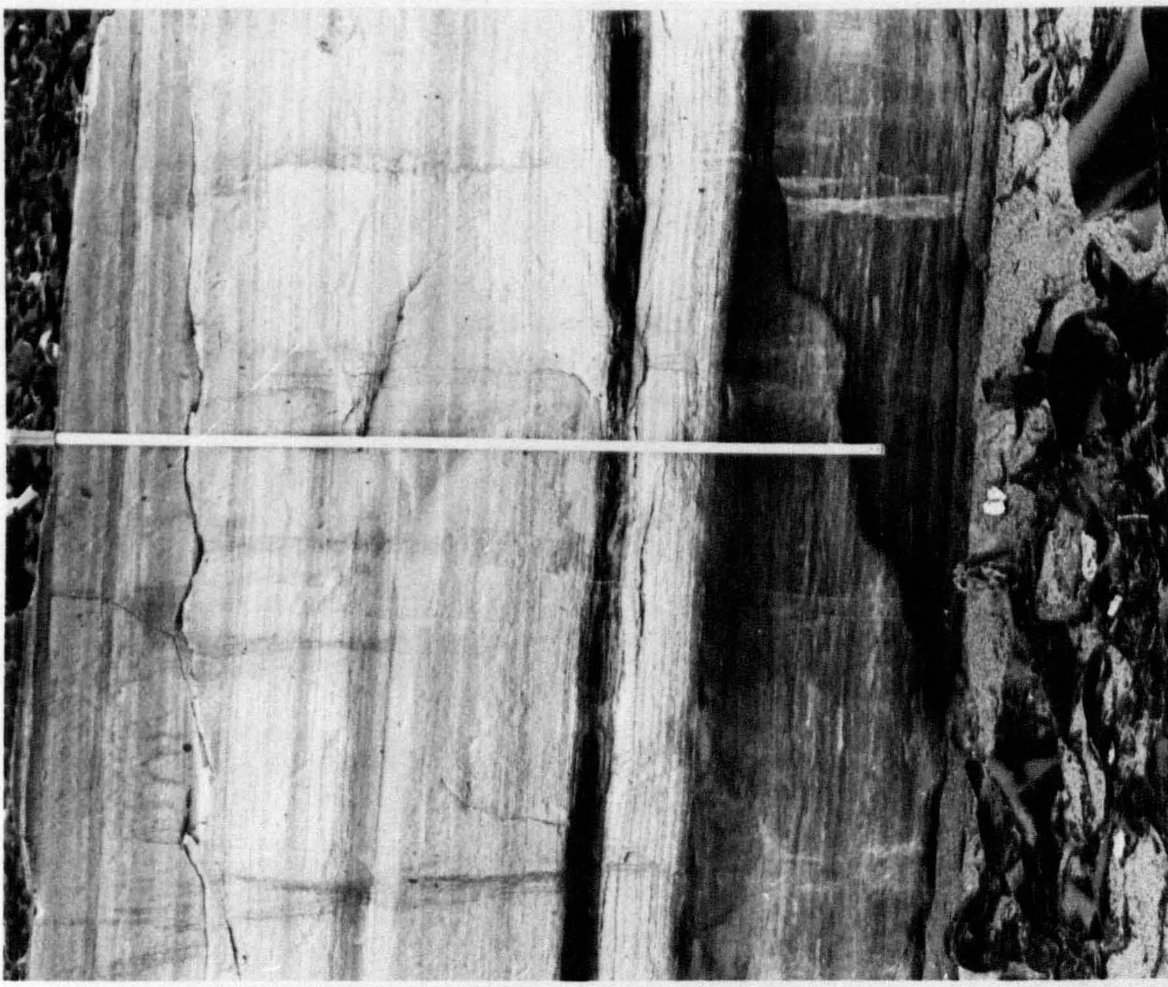


Fig. 5.2 Detailed sections of White Stone and Freshwater Steps Coccolith limestones

PLATE 5.3 Detail of the White Stone coccolithic limestone unit. (see also Fig. 5.2)

5.3A (TOP) General view of the White Stone Band. Note the sharp distinction between the light coloured ± rhythmitic coccolith limestone and the darker, marly, bioturbated coccolithic limestone at the base.

5.3B (BOTTOM) Close-up of the boudinage in the oil shale unit. Disruption of lamination in the adjacent coccolith limestone bands of uncertain origin (partly biogenic, partly fluid deformation?). Not the distinctly different appearance of the underlying marly limestone which is recognisably bioturbated.



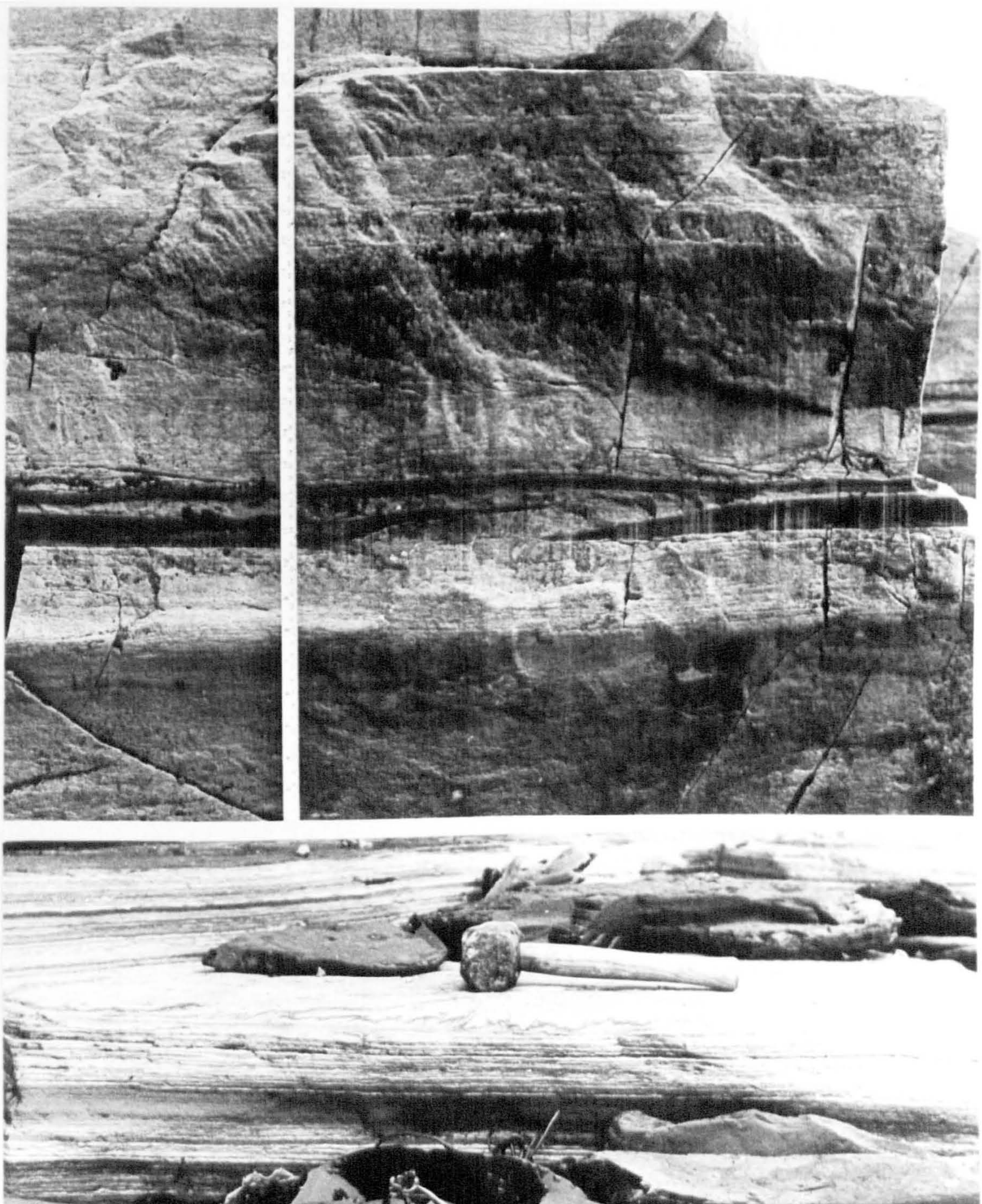
and boudinage (Plate 5.3b) are fairly common and injection features also occur (e.g. the coccolithic dykes through the oil shale within the White Stone Band; (Plate 5.4a). More rarely, vertically and laterally restricted deformation of individual (or 5-6) laminae occurs in the form of wrinkling or complex folding apparently related to localised lateral slippage and compression. This style of deformation has only been observed immediately below the oil shale in the White Stone Band and in part of the Freshwater Steps Band (Plate 5.4b). In places the rhythmites have been disrupted and homogenised with the destruction of all primary structure (though occasionally small isolated laminated 'rafts' occur). As no bioturbation is visible within these homogenised units, liquifaction seems to be the most likely process responsible for their formation (contrast with coccolithic marls described later). Most of the deformation features appear to have been produced during the compaction of the soft sediment.

Examination of coccolith rhythmite samples under the S.E.M. shows that coccospheres are common to abundant in the undisturbed laminae, while isolated coccolith plates dominate in the other units. The coccolith assemblages of the type Kimmeridge Clay generally show relatively high overall diversity (Gallois & Medd, 1979) but very poor species equitability and are overwhelmingly dominated by Ellipsogelosphaera britannica Stradner (apparently synonymous with Ellispagelosphaera frequens Noel; compare Gallois & Medd, 1979 and Busson & Noel, 1972). Framboidal pyrite is ubiquitous (pyritohedral crystals >> cubic forms) and occasional calcispheres were observed which were sometimes filled with pyrite or empty and exhibiting an internal organic membrane. In his more extensive study of the carbonate lithologies of the type Kimmeridge Clay, Bellamy (1979) found calcispheres to be relatively common. Further details may be found in his excellent thesis and are not duplicated here. Neither Bellamy (1979) or myself have observed any solution effects either in thin section or under the S.E.M.

PLATE 5.4 Deformation structures within the coccolith limestones

5.4A (TOP) Dyke of coccolithic material injected from below into the oil shale unit of the White Stone Band. These structures prove that liquifaction/fluidisation is partly responsible for disruption of the lamination within the rhythmites.

5.4B (BOTTOM) Localised wrinkling of lamination within rhythmites of the Freshwater Steps coccolith limestone. Such deformation also occurs locally within the White Stone.



(E) Coccolithic marls-limestones

These are relatively pure coccolithic sediments which appear to have been derived from the homogenisation of originally microlaminated sediments. Despite their general homogeneity, bioturbation is often conspicuous. The best examples are the units immediately below and above the White Stone Band rhythmites (see Fig. 5.2). They are pale in colour and have low organic matter contents. In their general appearance they resemble the Rope Lake Head non-kerogenous dolostone and are better cemented than the coccolith rhythmites and contain abundant calcite rhombs (Downie, 1955).

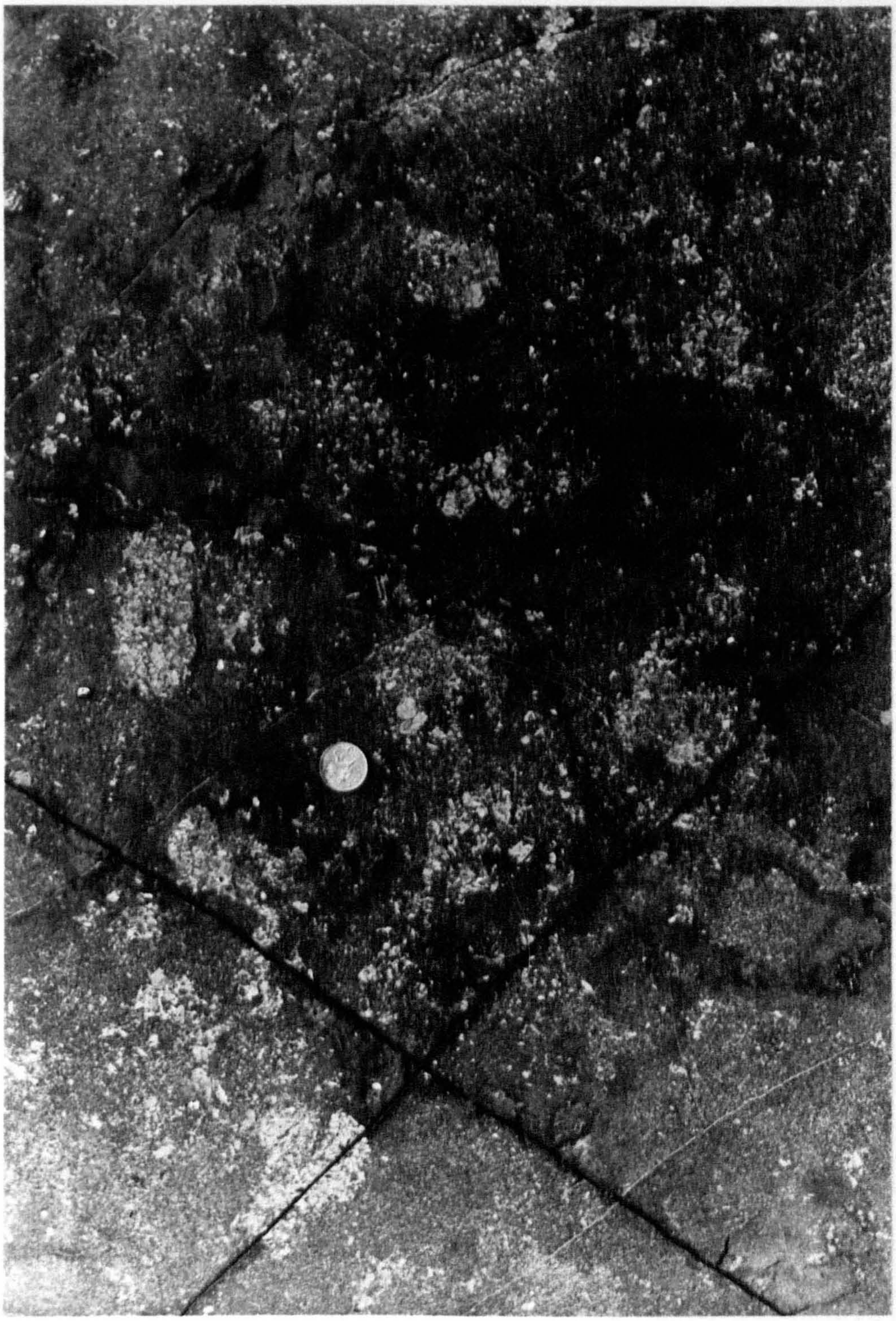
(F) Coccolithic kerogenous laminites (marls and limestones)

In the Kimmeridge and Swalland Members many of the more kerogenous bands contain appreciable amounts of coccolith material in the form of rounded, more-or-less ovate, pellets ranging in size from 0.1 to 10 mm. The pellets are generally cream or light brown coloured and when present in abundance give the sediment a speckled appearance (Plate 5.5). Coccolithic kerogenous laminites rarely contain distinct couplets but have a well developed, planar fabric. They are usually brownish-grey in colour and may contain as much as 80% carbonate; their organic content is generally comparable to normal kerogenous shales. Among the best examples are the two bands at the very base of the Swalland Member (2nd and 3rd below Bubbicum); the Middle Band also consists partly of coccolith kerogenous laminites. Coccospheres are commonly preserved within the pellets.

(G) Non-kerogenous dolostones (argillaceous dolostones of Bellamy, 1979)

These sediments form several conspicuous hard bands in the upper part of the Kimmeridge Member and in the Swalland Member. They are light to dark grey in colour but often exhibit a yellowish-grey weathering patina. In thin section the consist of a more-or-less equigranular mosaic of decimicrometre ferroan dolomite crystals; up to 10% calcite may be present and rare quartz silt also occurs (Dunn, 1972; Braide, 1976;

PLATE 5.5 Bedding surface showing pelletal nature of coccolithic material within one of the coccolithic kerogenous laminites. Some ammonite fragments are also visible. Largest pellets may be an artifact of how the bedding surface has weathered but large pellets do also occur on fresh surfaces.



Bellamy, 1979). Organic contents are low (generally ≤2% org.C.) and kerogen sparse in thin section, forming thin coatings around the dolomite crystals. The classic examples include the Cattle Ledge, Grey Ledge and Basalt Stone Bands and also the three unnamed bands between the White Stone Band and the Middle Band. In the field they may be rapidly distinguished from their kerogenous counterparts by their characteristic 'dicey', cuboidal weathering pattern which reaches its acme in the Basalt Stone (Plate 5.6). The Rope Lake Head Stone Band, mistakenly identified as a coccolith limestone by Gallois & Cox (1974) and Gallois (1976), is also a non-kerogenous dolostone but is more argillaceous than the others (Bellamy, 1979) and does not show the 'dicey' weathering pattern. From their general character and faunal content and by observation of some bands which appear to be incipient non-kerogenous dolostones (e.g. 4th hard band below the Bubbicum) these rocks appear to represent the diagenetic dolomitisation of calcareous mudstones.

(H) Kerogenous dolostones

Kerogenous dolostones are limited to the Kimmeridge Member and are petrographically similar to their non-kerogenous equivalents with the exception that they contain abundant, slightly disrupted, kerogen-rich laminae (organic carbon contents are ≤6%; Davies, 1978). In the field they are characterised by their conspicuous yellow weathering patina and the presence of incipient or well developed thrust polygon structures (see Bellamy, 1976, 1979). They are massive, composite units and do not show 'dicey' weathering. Their overall character, petrography, palynofacies (see later) and fauna indicates that they are predominantly dolomitised kerogenous shales (a conclusion shared by Gallois & Cox, 1974). However, they are clearly composite and although the dolomitisation makes them appear more homogeneous, more than one original lithology was often clearly present. The Yellow Ledge dolostone appears to have once consisted of both mudstone-shales and kerogenous shales and the Washing Ledge Band contains what is probably an oil shale or kerogenous shale in

PLATE 5.6 Characteristic dicey weathering pattern of the Basalt Stone non-kerogenous dolostone. The somewhat similar fracture pattern in the underlying mudstone may suggest partial dolomitisation.

its centre part. The classic examples are the Flats Stone Band, the Washing Ledge, the Maple Ledge Band and the Yellow Ledge Band. See also Irwin (1980).

(I) Short Joint Coal (calcite rhomb limestone)

As mentioned above, the carbonate unit at the centre of the Short Joint Coal appears to consist almost entirely of euhedral calcite rhombs. The unit has no internal structure and appears extensively disrupted giving rise to a series of parallel orientated, tapering dykes of carbonate material which project downward into the underlying oil shale at an angle of 20-30^{o} and are associated with pyrite nodules and low angle microfaulting (Plate 5.7).

PALAEOECOLOGICAL OBSERVATIONS

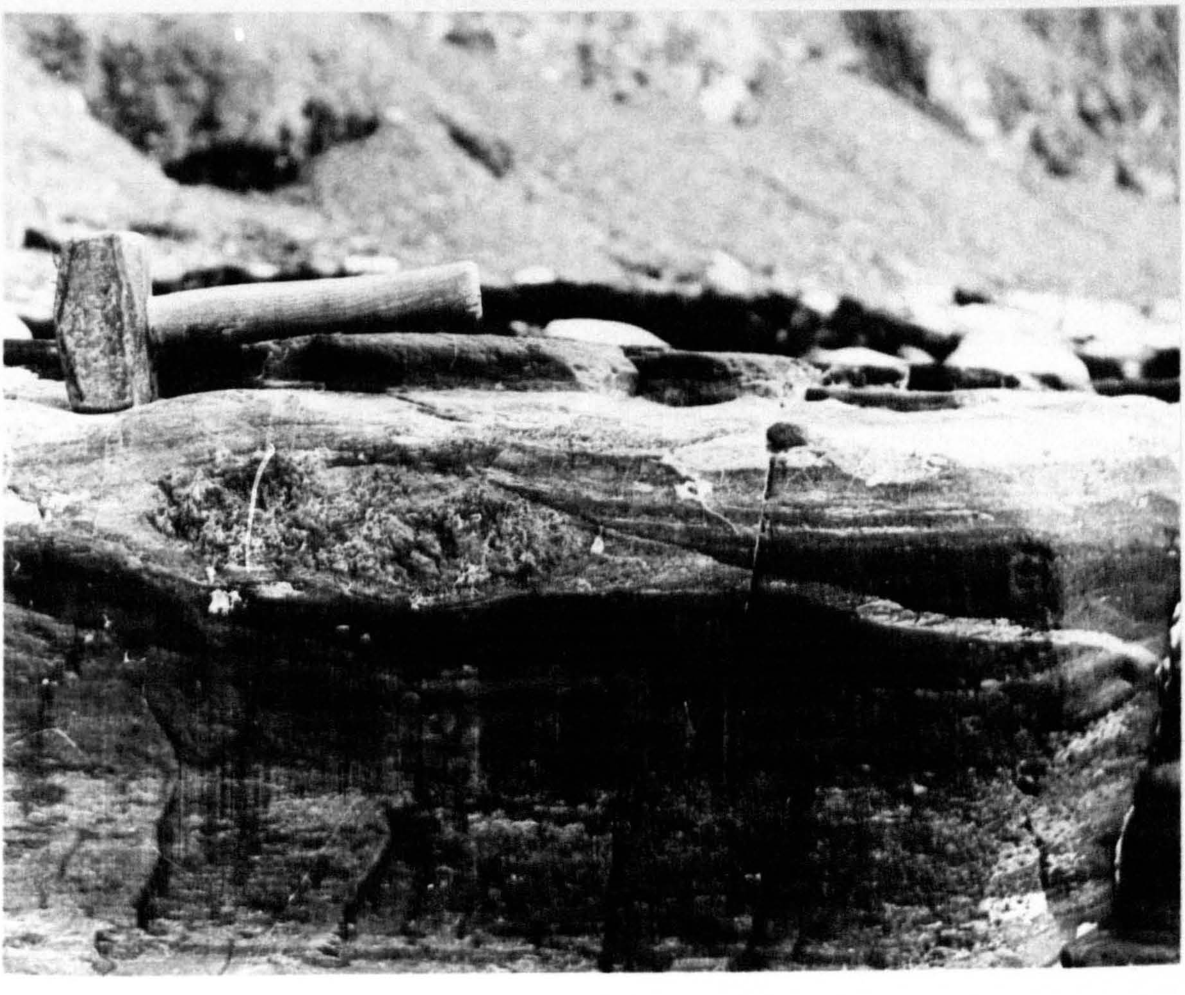
(1) MACROFAUNA

Unfortunately no detailed examination of the non-ammonite macrofauna of the type Kimmeridge Clay has ever been carried out; with the exception of Brookfield's study of the Upper Calcareous Grit Formation (Brookfield, 1973, 1978a) this situation pertains to the British Kimmeridgian in general. The most complete overall faunal list is probably still the one in Blake (1875) although the ammonite faunas are, of course, comparatively well known following the work of Arkell (1947, etc.), Ziegler (1962) and Cope (1967, 1978). Generalised comments on the fauna of the Kimmeridge Clay can be found in Blake (1975), Downie (1955) and Brookfield (1973, 1978a, 1978b); brief discussions also occur in Arkell (1947), Cope (1968), Townson (1971, 1975), Gallois & Cox (1974), Gallois (1976), Braide (1976), Tyson et al. (1979), Bellamy (1979), Morris (1980), Aigner (1980) and Cox & Gallois (1981). This section is based on the analysis of the data present in these references, coupled with my own observations, and is principally involved with the bivalve (pelecypod) fauna but includes some remarks on the distribution of ammonites .

PLATE 5.7 Structures observed within the Short Joint Coal oil shale and rhomb limestone unit.

5.7A (TOP) View showing low angle faulting and associated injections in the lower half of the Short Joint Coal. The hammer rests on the central rhomb limestone unit .

5.7B (BOTTOM) Close-up of one of the injection structures (sedimentary dyke) showing its association with large pyritic concretions. The exact origin of these features is not clear.



General comments

In the now classic paper in which the first measured section of the Kimmeridge Clay was presented, Blake (1875, p.197) observed that these sediments "contain a comparative paucity of species but an infinity of individuals". This feature also impressed Downie (1955) who records that the fauna is dominated by bivalves and ammonites and (that with the exception of the characteristic inarticulate forms Lingula and Discina) brachiopods only occur at the top and base of the sequence (i.e. in the Black Nore, Hounstout and Black Head Members). Curiously belemnites are rather rare and echinoderms in general are rather scarce (although some debris of the latter does occur in micropalaeontological residues, J. Exton pers. comm.). For the Upper Jurassic sequence as a whole Ager (1975, 1976) and Townson (1976) have discussed the gross faunal trends that may represent a progressive change in salinity (in particular between the Portland and Purbeck Groups), but the principal faunal controls within the Kimmeridge Clay appear to have been bottom oxygenation and substrate (see later). The cyclic sedimentation within the Swalland Member provides an ideal situation for examining the effects of these parameters on the fauna.

PALAEOECOLOGICAL OBSERVATIONS ON THE SWALLAND MEMBER

In comparison with the underlying Black Head Member, the Swalland Member shows a marked reduction in the diversity of its benthic fauna (Brookfield, 1978a). The most characteristic and abundant bivalve genera (and brachiopods) are listed in Table 5.2 along with their presumed mode of life, feeding method and taxonomic position. Downie (1955) has noted that there appears to be rather little difference in the actual genera present in the different lithologies but significant contrasts are seen in their relative abundance and pattern of occurrence. The faunal character of each of the major lithologies described previously is discussed below.

TABLE 5.2 Principal macrofaunal benthic elements of the type Kimmeridge Clay

Genus	Taxonomic Group	Feeding Method*	Mode of life
1 Astarte	Crassatellacea	S	Shallow infaunal
2 Camptonectes	Pectinacea	S	Epifaunal, byssate
3 Codakia ("Lucina")	Lucinacea	'S'	Shallow-'deep' infaunal
4 Corbula	Myoida	S	Shallow infaunal
5 Deltoideum	Ostreacea	S	Epifaunal, flat lying
6 Discinia	Orbiculoidea	S	Epifaunal
7 Entolium	Pectinacea	S	Epifaunal, flat lying or byssate
8 Grammatadon	Arcoida	S	Non-siphonate shallow infaunal
9 Lingula ovalis	Brachiopoda	LLS	Infaunal
10 Modiola	Mytiloida	LLS	Byssate, epifaunal semi-infaunal
11 Nanogyra	Ostreacea	LLS	Epifaunal, flat lying
12 Nicaniella (Trautscholdia)	Crassatellacea	S	Non-siphonate, shallow infaunal
13 Ostrea	Ostreacea	S	Epifaunal, flat lying
14 Oxytoma	Pectinacea	S	Epifaunal, pendant
15 Pleuromya	Pholadomyoida	LLS	Deep infaunal
16 Protocardia	Cardiacea	S	Shallow infaunal
17 Thracia	Pholadomyoida	MTS	Deep infaunal

* S suspension feeder; LL low-level; MT mucus tube constructor

Forms found in organic-rich shales or known to be resistant to oxygen stress: 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 12, 14, 15, 16, 17

Forms generally considered to be 'eurytopic': 2, 9, 10, 11, 12, 13, 14, 16

N.B. Ecological data from multiple sources including Brookfield (1973) Duff (1975) and Sykes (1975).

(A) Mudstone-shales (clays of Downie, 1955)

These sediments normally contain the most diverse benthic faunas that are to be found in the Swalland Member and in the Kimmeridge Clay Formation in general. Small, thin-shelled *Codakia*, *Protocardia*, *Modiola*, *Astarte*, *Corbula*, *Nanogyra*, *Deltoideum* and *Ostrea* are common to abundant and show a relatively uniform distribution, although some practically barren horizons do occur. No large autochthonous or allochthonous shell accumulations occur above the Black Head Member (cf. Brookfield, 1973, 1978a). All the bivalves, including normally relatively deep burrowers such as *Pleuromya*, are found parallel to bedding even though, according to Downie (1955), they are commonly articulated. Preservation is predominantly in the form of chalky, white calcite (although very rare lustrous aragonitic preservation does occur) and the fossils are generally flattened. Ammonites are common to abundant and are invariably crushed.

(B) Kerogenous shales

One of the most characteristic features of these sediments is the alternation of virtually barren laminae with bedding surfaces crowded with small shells ('shell plasters'). As there have been no detailed studies on the Kimmeridge Clay fauna it is not possible to comment on the total diversity of the benthos in the kerogenous shales in comparison with the other lithologies, but it is clear that faunal dominance is much higher than in the mudstone-shales. The shell plasters are almost monospecific, the most characteristic fossil being *Codakia* (*epilucina*) *minuscula* Blake (the species name invariably mis-spelt as *miniscula* by Gallois & Cox, 1974, 1976; Gallois, 1976), with *Astarte* and *Protocardia* the two other most common forms. These fossils tend to be smaller and if anything more fragile than they are in the mudstone-shales. The comment by Gallois (1976) that the oil shales contain a "richer and more abundant fauna" than any other lithology in the Kimmeridge Clay pertains to the kerogenous shales rather than the oil shales as defined here and is perhaps misleading since although the benthos undoubtedly shows higher densities, dominance is higher and diversity may be lower. Ager and Wallace (1966) have described the fauna from the Argiles de la Crèche in the Boulonnais which is in part coeval with the Swalland Member (wheatleyensis to hudlestoni zones according to Riley & Sarjeant, 1977) and is developed in a similar facies (mudstone-shales and kerogenous shales). These sediments, which were classified as a 'lethal-pantostrate' biofacies (see Chapter 2), are characterised by shell plasters of small, thin shelled *Astarte* sp. with subordinate *Protocardia* and *Pleuromya*. In Dorset the kerogenous shale fauna is preserved as chalky or non-chalky calcite and the benthos and ammonites (which are common) are again flattened or crushed. The phosphatic inarticulate brachipods *Lingula* and *Discina* are possibly more abundant in the kerogenous shales than the ordinary mudstone-shales.

(C) Oil shales

Downie (1955) has described the fauna of the oil shales as "frequently absent, only occasionally abundant". However, no detailed description of the oil shale fauna has appeared and apart from some preliminary results reported by Morris (1980) the following comments are based entirely on my own observations. During this study samples of the Blackstone and Short Joint Coal oil shales (minimum size 2250 cm^3) were collected, split and their contents examined to provide some qualitative information on their faunas and to supplement field observations.

(i) The Blackstone (see Fig. 5.3)

Comparison of the fauna in ± 2cm serial splits with the organic carbon values of these samples (determined by Roberston Research) reveals some interesting trends and emphasises the composite nature of this unit. Excluding the two thin oil shales at the very top (and the two intercalated kerogenous shales) the Blackstone consists of an upper oil shale ("the Blackstone", organic carbon 30-40%) a thin dividing kerogenous shale (organic carbon 14-24%) and a lower oil shale, the "Best Blackstone"

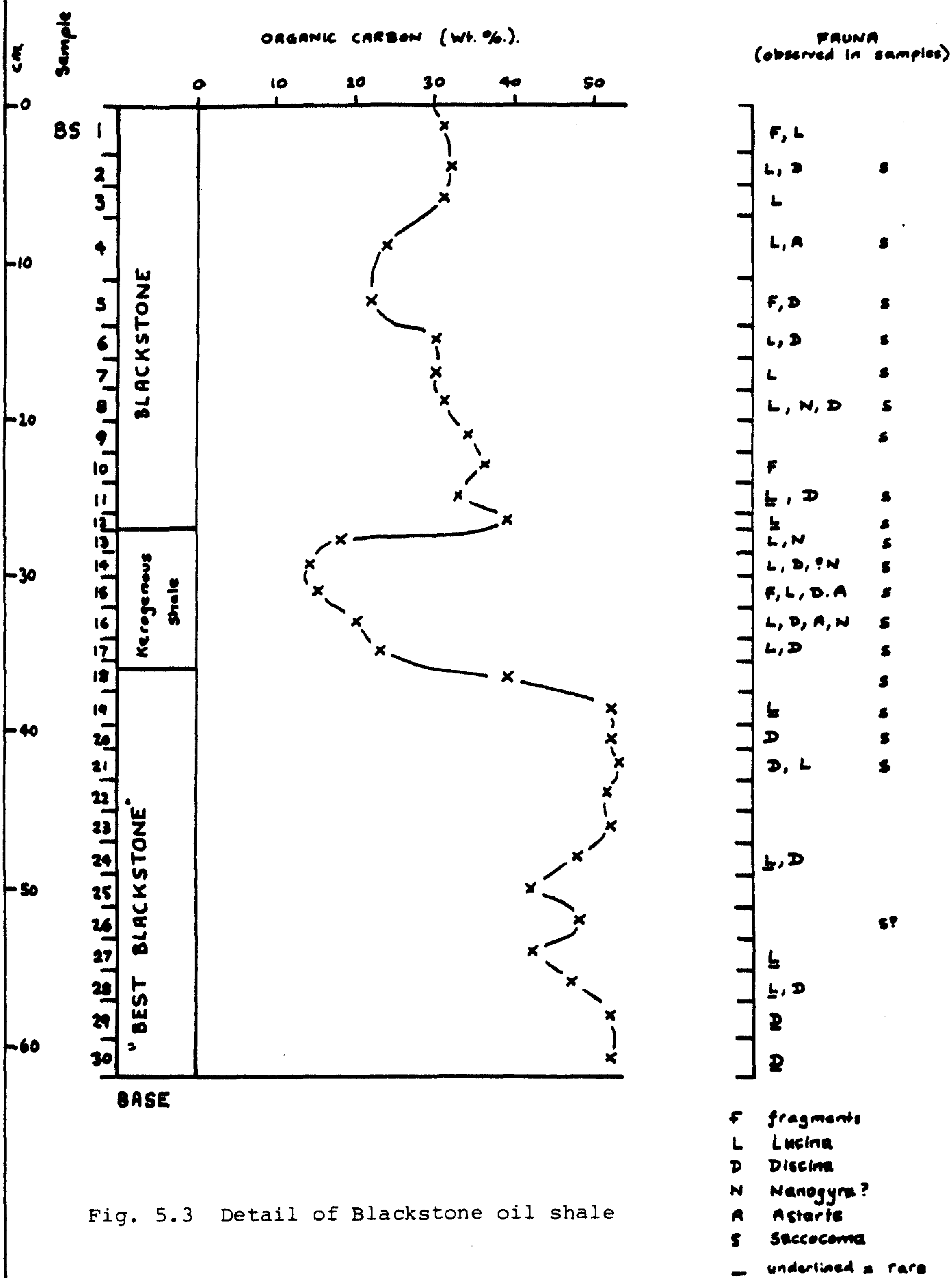


Fig. 5.3 Detail of Blackstone oil shale

of Strahan, 1920 (organic carbon 42-52%). The fauna of the kerogenous shale (samples BS 13-17) is fairly typical for this lithology; Codakia ('Lucina') is by far the most common bivalve, with only rare Astarte and very occasional Nanogyra. The upper oil shale is in many ways similar to the kerogenous shale with occasional 'Lucina' plasters, but in general is rather less fossiliferous and other bivalves (Astarte, Nanogyra) appear to be very rare. The phosphatic acrotretid inarticulate brachiopod Discina is farily common throughout the upper oil shale and underlying kerogenous shale. In contrast the 'Best Blackstone' has a very sparse fauna or is barren and tends to be characterised by its more speckled appearance and by the presence of sometimes fairly large coccolithic fecal pellets (<1cm). It contains only comparatively rare specimens of 'Lucina' and its meagre fauna is dominated by Discina. Throughout the Blackstone band the fossils are very fragile, very poorly calcified and often exhibit a pyritic sheen; they are also rather small (<1cm, often 5mm or less) but may tend to be larger (as well as more diverse) in the kerogenous shale.

(ii) The Short Joint Coal oil shales

The fauna in these oil shales was found to be comparable with the upper oil shale in the Blackstone described previously. 'Lucina' is again by far the most common fossil (although it does not form plasters) and only two possible specimens of Astarte and one of Nanogyra were found after splitting the samples. Discina was present but never very common. All the fossils were very poorly calcified, often completely or partially pyritised and were generally less than 5mm in size. The similarity with the fauna in the upper oil shale of the Blackstone Band may relate to the similar values of organic carbon, 30-40% in the former and 26-38% in the Short Joint Coal.

Whole specimens of ammonites do not appear to be particularly common within the oil shales although small fragments are sometimes rather abundant; during the examination of the Blackstone and Short

Joint Coal samples, only one (small) whole specimen was seen. This finding was in agreement with my field observations which also suggested the comparative rarity of ammonites within this lithology. As Cope (1967) has, however, pointed out, pyritised ammonites do occur in the uppermost part of the Blackstone (above the interval shown in Fig. 5.3). The only other pelagic fossil observed in the oil shales (at least in the Clavell's Head section) was the pyritised ambulacral (?) plates of the pelagic crinoid <u>Saccocoma</u>.

The massive, tough nature of the oil shales, which is such a useful characteristic when logging the sediments in the field, proves to be a significant hindrance when attempting to study their faunas. Fresh samples cannot be conveniently split along perfect bedding planes and this makes accurate quantitative investigations rather difficult. This may be one factor in the dissimilarity of the results obtained by Morris (1980) and those here. Morris (Table 4, p.164) shows an oil shale fauna with the following composition: <u>Exogyra</u> (i.e. <u>Nanogyra</u>) > <u>Liostrea</u> > <u>Lucina</u> (i.e. <u>Codakia</u>) > <u>Astarte</u>. No clue to which oil shale was analysed was given except that it was from the <u>wheatleyensis</u> zone and was thus probably either the Blackstone, the Bubbicum or the unnamed oil shale below it. In relation to my own observations the abundances of <u>Nanogyra</u> and <u>Liostrea</u> seem to be much too high (primarily at the expense of '<u>Lucina</u>'). The description of the Kimmeridge Clay given by Morris (1980) appears to have been taken primarily from Downie (1955) and papers by Gallois and it is possible that some degree of confusion between kerogenous shale and oil shale may have resulted. Morris's figures are most like the kerogenous shale within the Blackstone, but even here the importance of '<u>Lucina</u>' would have been underestimated. Morris' figures may have been biased by localised eipfaunal oysters on ammonites and differences between the two sets of observations may imply strong patchiness of the benthic faunas.

(D) Coccolith rhythmites

Downie (1955) and Bellamy (1979) do not report finding any body fossils (including ammonites) within these sediments and this conclusion is in general supported by this study. However, the Freshwater Steps Stone Band was found to contain rare flattened ammonites (only 2 specimens seen in the whole length of the outcrop) and rather rare oysters (Plate 5.8). One unidentified, apparently disarticulated, oyster was observed in the generally micro-laminated lower coccolithic band, and a patch of $\sim$10 relatively large ($\leq$7cm) pyritised oysters (Liostrea?) was found $\sim$5mm below the top of the generally micro-laminated upper coccolithic unit.

(E) Coccolithic marls-limestones

No body fossils were observed in these units (despite the presence of bioturbation - see later).

(F) Coccolithic kerogenous laminites

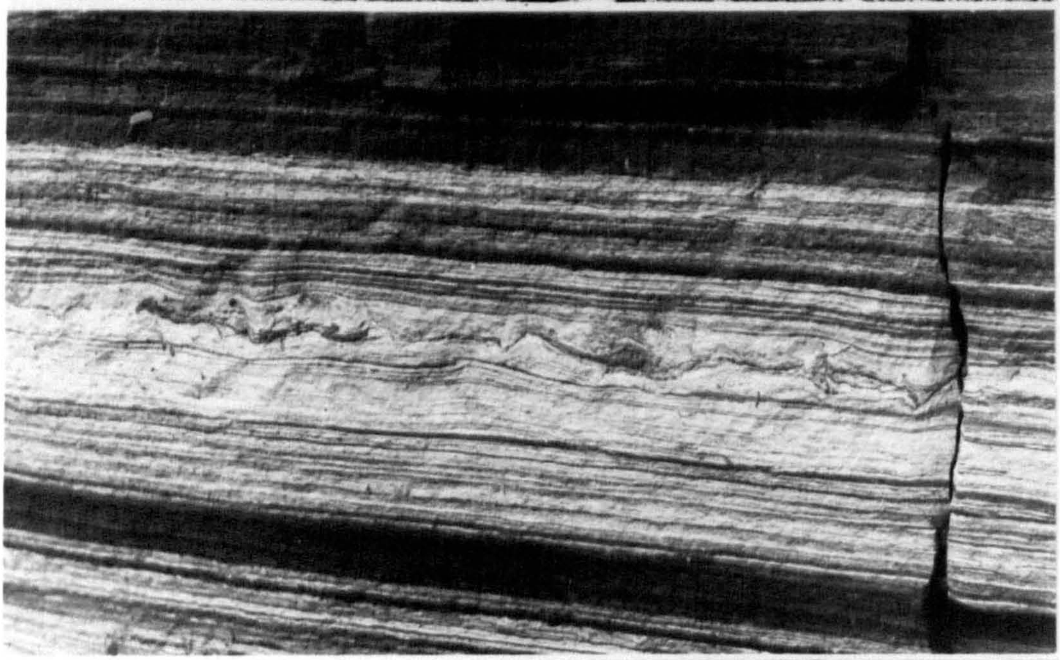
These sediments are predominantly devoid of benthos, although occasional tiny Discina may be present in some of the units. The coccolithic parts of the coccolithic kerogenous shales exposed in Clavell's Hard (3rd and 4th hard bands below the Blackstone) appear to be barren while their more normal kerogenous shale to oil shale parts were found to contain rare ammonites (mainly as debris but 2 small whole specimens recorded), 'Lucina', Discina and one specimen each of Deltoideum, Nanogyra, Modiola and Pinna (the latter being something of a surprise in this facies).

(G) Non-kerogenous dolostones

Only field observations were made on the fauna in this lithology. The 'dicey' weathering and fracturing of these sediments makes it rather difficult to find any recognisable fossils but a few unidentified (aragonitic?) oysters were seen in the Basalt Stone and two other bands between the White Stone and Freshwater Steps Stone Bands.

(H) Kerogenous dolostones

No specific observations were made on the faunas in these sediments



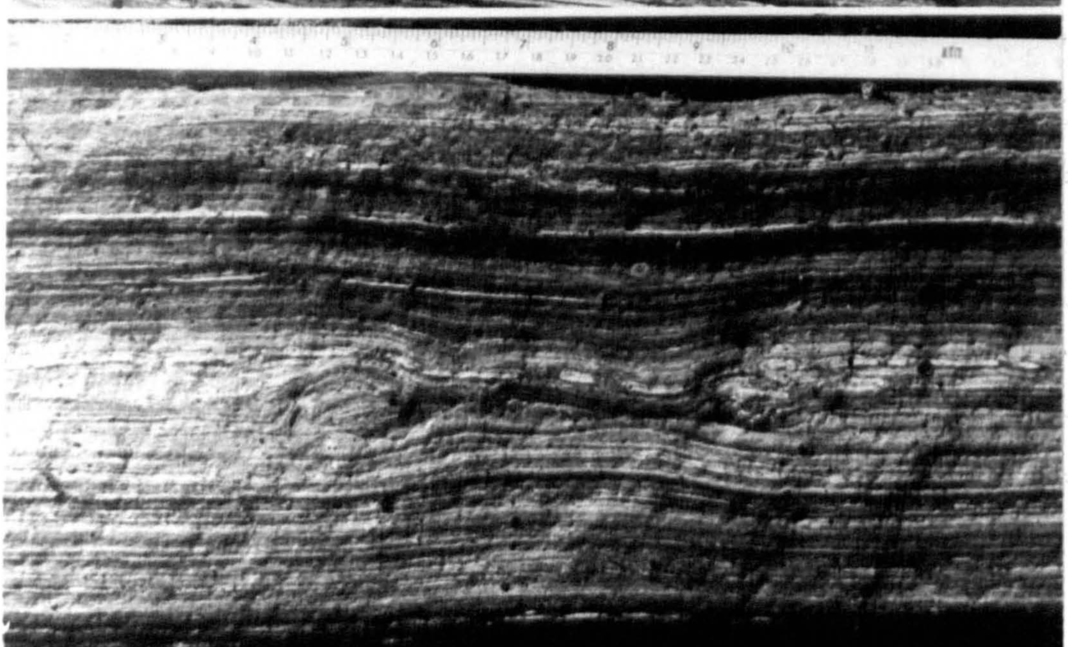


PLATE 5.8 Interesting features observed in the Freshwater Steps Stone Band

5.8A Patch of pyritised oysters from upper part of rhythmite.

5.8B Horizon of shell material (bivalve and/or ammonite debris?) within upper coccolith rhythmite. Note deformation of lamination during compaction.

5.8C Problematic structure within rhythmite. I have observed similar features on Leg 77 DSDP which resulted from compaction and dissolution of ammonites.

but kerogenous shale type assemblages were observed within the Yellow Ledge Band.

(I) Short Joint Coal rhomb limestone

No body fossils were found in splits of this lithology.

Comments on the macrofauna

The low diversity and generally high density of the faunas in the Kimmeridge, Swalland and Hounstout Members indicates that they were deposited in a stressed environment which was only successfully exploited by a few eurytopic benthic organisms. Both the high organic content of the sediments and the nature of their biofacies indicates that dissolved oxygen was a principal faunal control. This is also partially supported by the small size of the fossils which could be viewed as an adaptation to providing more favourable surface area: volume ratios (but note that while this is more favourable for low oxygen conditions, smaller body sizes may lower tolerance to sulphide; Groenendaal, 1980). Bulky density restraints were probably also an important contol on body size (see Thayer, 1975) and the poorly calcified and thin-shelled character of most of the fossils presumably reflects conditions unfavourable to calcification and partial dissolution during diagenesis.

The benthic fauna is dominated by infaunal suspension feeding bivalves (Table 5.2). Epifaunal suspension feeders are also quite common (*Nanogyra*, *Deltoideum*; *Ostrea* on ammonites) but are more restricted to the normal-mudstone shale substrates. Deposit feeding bivalves are present (e.g. '*Nucula*' sp.) but seem to be comparatively rare even in the organic-rich sediments, which appears to be somewhat unusual when compared with other Jurassic occurrences (Duff, 1975; Sykes, 1975; Morris, 1979). Bader (1954) has commented on the relationship between pelecypod faunas and sediment organic content (see also discussion in Purdy, 1964). Unfortunately there have been some rather misleading conclusions drawn from this work, for example Duff (1975) citing Bader (1954) concludes that "organic carbon contents of over 3% in recent mud deposits have been shown ... to cause a diminution in bivalve diversity, with infaunal deposit-feeding protobranchs becoming dominant" (p.452). As Purdy (1964) points out, Bader's data shows that above 3% organic carbon not only does the diversity of the assemblage decrease but the abundance of deposit feeders (which peaks around 2% org.C) also shows a marked decline. When this is taken into account it appears that up to around 3% org.C the organic matter encourages a dense deposit feeding fauna, which probably limits suspension feeders via the 'trophic group amensalism' effect of Rhoads & Young (1970), but above 3% org.C deposit feeders decline (due to the anoxic conditions within the sediment?) and the proportion of suspension feeders may subsequently increase again. As many of the normal mudstone-shales in the Kimmeridge and Swalland Members contain more than 3% org.C and those in the Hounstout Member 1-2% org.C, this may be one explanation for the 'rarity' of deposit feeders in these units. Detailed macrofaunal analyses will undoubtedly clarify the matter. The statement by Morris (1980, p.164) that 'infaunal deposit feeders are not abundant in the organic-rich shales but are found in greater numbers in the organic-poor clays" is based on the erroneous trophic classification of *Corbula* as a deposit feeder. *Corbula* is in fact a byssate, semi-infaunal suspension feeder (Stanley, 1970; Brookfield, 1973).

The most successful organism inhabiting the organic-rich sediments of the Kimmeridge Clay appears to have been '*Lucina*' *minuscula*. While this agrees with the general observation that lucinoids often dominate modern oxygen deficient environments (Valentine, 1973, p.168) it also presents some problems. Many present day lucinoids live in comparatively coarse sediments (Stanley, 1970; Kauffman, 1969) and Kranz (1974, p.255) notes that their vermiform foot would be "virtually useless" in soft muds. As all the fossils are usually found parallel to bedding (no doubt accentuated by the high compaction), and it is difficult to imagine any benthic organisms surviving in conditions hostile enough to preserve large amounts of marine organic matter, there is a strong implication that the fauna of the kerogenous lithologies is, to some extent at least, allochthonous. Rosenberg (1977) notes that the benthic infauna is constrained between the sediment-water interface and the R.P.D., and is forced upwards by declining bottom oxygenation, with even comparatively deep burrowers like Mya emerging from the sediment and lying on their sides with their long siphons extended into more oxygenated water (Jørgensen, 1980). The bedding-parallel occurrence of the infauna is therefore explicable in terms of an in situ mechanism, but since the activities of the infauna prior to their emergence would have been preserved as trace fossils or a general bioturbated texture, this explanation is only appropriate for the more normal mudstones and not the laminated kerogenous lithologies. For the latter it is possible that the shells are redeposited from other areas, where conditions forced the benthos to emerge from the sediment (a kind of pre-hydrodynamic, ecologic sorting) and currents subsequently transported the shells. Such transport, if it occurred, resulted in no appreciable scouring or erosion. The only alternative would be that lucinoids were capable of temporarily colonising the sediment under slightly improved conditions but were ecologically restrained from burrowing.

In relation to the rarity of ammonites in the oil shales and coccolith rhythmites it is worth recording the observations by Sykes on bituminous Callovian-Oxfordian sediments. Sykes (1975, Table 19) found that ammonites were 5-7 times more rare in 'bituminous shales' than other equivalent mudrocks and commented that "it is interesting to note that ammonites react in the same way as bivalves to such environmental changes, further strengthening the argument that they were mostly benthonic" (p.216). Ammonites were probably affected by the hostile bottom water conditions and the associated changes in the planktonic environment which would have been felt throughout the food-chain.

2. TRACE FOSSILS

Compared with other argillaceous Jurassic sequences trace fossils are not particularly common in the Kimmeridge Clay Formation. This effect is probably more apparent than real, resulting from the general homogeneity of the sediments which does not allow burrow fills to be readily distinguished, particularly after compaction. Lack of suitable outcrop is another significant problem that prevents accurate assessment of the extent of bioturbation in some parts of the sequence.

In the mudstone-shales bioturbation is rarely conspicuous but is indicated by the general homogeneity of the sediment. Chondrites is present (as recorded by Morris, 1980) but I have not observed any good examples of this ichnogenus in any of the lithologies. Rhizocorallium is occasionally recognisable in the mudstones (particularly at more calcareous levels). No identifiable burrows were observed in the kerogenous shales, oil shales and coccolithic kerogenous laminites during this study. The best trace fossils were found in the Rope Lake Head Stone Band (non-kerogenous dolostone), at certain horizons within the White Stone Band (Plate 5.9) and Freshwater Steps Stone Band ('coccolith limestones') and in the coccolithic marls above and below the White Stone Band.

The Rope Lake Head Stone Band contains common Rhizocorallium including some excellent examples ($\leq$ 40 cm long) showing clearly visible retrusive spreiten. The burrow fill is generally darker (more argillaceous?) than the dolostone matrix and may (at least in the upper part of the bed) have been continuous with the overlying kerogenous shale. Excellent examples of Rhizocorallium also occur in the medium-grey coccolithic marls associated with the White Stone Band, but here the burrow fill is white coccolithic material. A single(?) horizon of conspicuous Rhizocorallium burrows occurs within one of the rhythmite units in the White Stone Band. These rather small burrows extend downward for a vertical distance of about 3 cm from an horizon (marked by a

PLATE 5.9 Bioturbation within the White Stone Band

5.9A (TOP) Conspicuous horizon of *Rhizocorallium* burrows within the White Stone Band (see also Fig. 5.2)

5.9B (BOTTOM) Part of large coccolith-filled *Rhizocorallium* burrow from the dark marly limestone at the top of the White Stone Band.

prominent kerogenous laminae) 29 cm above the top of the central oil shale unit. Except in the actual burrow fill itself, disruption to the laminae is slight; only 13 burrows were measured over a lateral distance of 2.2m along the outcrop which gives some idea of the low density of these structures.

Morris(1980) records Teichichnus and Tigillites as well as Rhizocorallium in the White Stone Band. From his presentation at the Geological Society of London symposium on "black shales" in May 1979 it is apparent that no distinction between the marl and rhythmite units within the White Stone Band was attempted and so it is not possible to determine where these trace fossils were observed. I have not seen any undisputable examples of Teichichnus or Tigillites in the Kimmeridge Clay Formation and so if they do occur they are certainly not common. In addition to Rhizocorallium the White Stone Band and Freshwater Steps Band (which contains only rare Rhizocorallium) also exhibit another burrow form which is common at some horizons. This form is a horizontal tube (∿1cm in diameter) which is sometimes slightly sinuous but unbranched and shows a very limited (vertical) penetration of the sediment. Some occurrences (but clearly not all) may be one tube of a small retrusive Rhizocorallium. This may be the burrow that Morris (1980) attributed to Teichichnus but since the main characteristic of the latter is the vertical succession of retrusive or protrusive spreiten, and such features are not observed in this case, the name hardly seems appropriate. Townson (1971) reports a similar burrow from the Lower Portland Limestone Formation which he also (albeit tentatively) ascribed to the ichnogenus Teichichnus. In their study of the trace fossils in the Kimmeridgian of the Boulonnais, Ager and Wallace (1970) remark on the fact that Teichichnus was absent; had this trace fossil been present in the Dorset-Boulonnais basin one might have expected it to be best developed in the Boulonnais facies (e.g. see Simpson, 1975, Table 4). However, Simpson (1975) does note that in distal shelf deposits Teichichnus burrows are scarce, narrow and shallow. Teichichnus does not occur in genetically similar coccolithic sediments in the Cretaceous of the U.S.A. (Hattin, 1971). It is possible that this type of burrow in Dorset may be a horizontal, unbranched form of Planolites affinities.

Townson (1971) found that Rhizocorallium was abundant in the Gad Cliff Member of the Portland Sand Formation and gives the following description (p.18) - "although the burrows (i.e. Rhizocorallium) are common the sediment between them still retains depositional laminations" and "the lack of complete homogenisation of slowly deposited sediment suggests that burrowing was a rare occurrence in spite of the abundance of the fossil". This description could equally well be applied to many occurrences in the Kimmeridge Clay. Body fossils other than ammonites are rare in the basinal development of the Gad Cliff Member (Townson, 1971, 1975) and the sediments are believed to have been deposited below a 'critical level' in the sea where oxygen deficient, stagnant conditions prevailed. The occurrence and extent of bioturbation by the Rhizocorallium organism (crustacean?) seems to have been strongly influenced by bottom oxygenation throughout the Upper Jurassic sequence. This may also apply to the Boulonnais where Rhizocorallium was only recorded at calcareous levels within the Schistes de Chatillon (Ager and Wallace, 1966). Townson, (1971) also notes that at some levels vertically retrusive Rhizocorallium simulates "small scale cross stratification" (p.19); I have observed this effect in the Kimmeridge Clay (e.g. in the coccolithic marls associated with the White Stone Band) where the burrows give the sediment a 'flecked' appearance (the latter possibly misleading Irwin, 1979).

The absence and/or rarity of trace fossils in the kerogenous lithologies appears to be due to the ecological preclusion of the burrowing habit and the un-cohesive nature of the substrate (see previous section). In the case of the coccolith rhythmites the bottom conditions appear to have been either predominantly anaerobic or unsuitable for dense infaunal populations. Even where oysters were observed in the Freshwater Steps Stone Band the micro-laminations were not obliterated by infaunal activity. The presence of Rhizocorallium within the White Stone Band, however, clearly indicates that the nature of the substrate was not the predominant control; bottom oxygenation (and its temporal variability) was probably the most significant factor. Some of the less-well or non-laminated intervals within the rhythmite units of the White Stone Band may have resulted from:

(a) intervals of more constant sedimentation

(b) inorganic homogenisation of the sediment (liquifaction, etc.)

(c) brief interludes of better oxygenation when metazoan benthos was present.

Calvert (1964) has noted that soft bodied metazoan organisms can be abundant under anaerobic (to dysaerobic) conditions but they are rather inactive and cause little disruption of the sediment. In the case of the coccolith rhythmites the temporary establishment of anaerobic to dysaerobic conditions may have been just sufficient to cause slight homogenisation but too low (or too brief) to allow colonisation by fossilisable benthos (see Chapter Two). Even if bioturbation could be proved at certain levels it would be impossible to say whether a 'homogeneous' unit represented continuous 'oxygenated' conditions or a brief episode which resulted in the homogenisation of previously accumulated micro-laminated sediments. From the gross character of the White Stone Band one suspects the latter would be more likely. Such intervals were possibly more common in the Freshwater Steps Band but again their effects on the gross sediment structure were negligible (Plate 5.10).

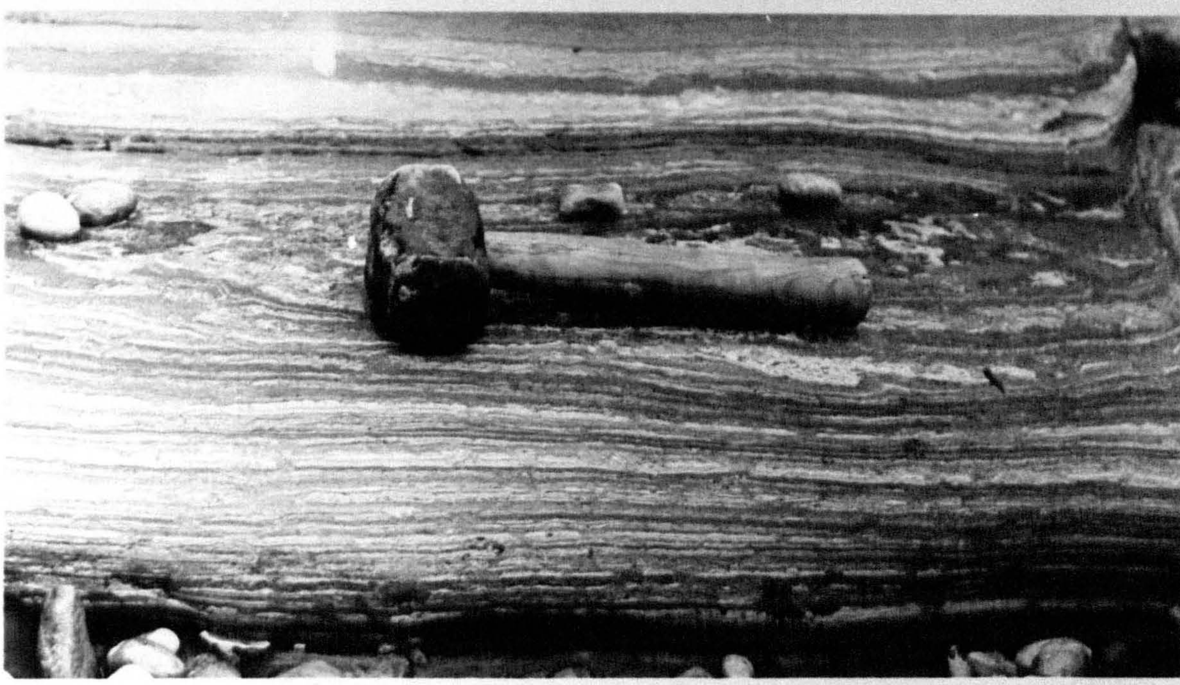
(3) MICROFAUNA

With the exceptions of the papers by Lloyd (1959) and Kilenyi (1969) very little has been published on the general character of the

PLATE 5.10 Bioturbation within the Freshwater Steps Stone Band

5.10A (TOP) Bioturbated bedding plain; burrows of *Planolites* affinity and of limited vertical extent.

5.10B (BOTTOM) Freshwater Steps coccolith rhythmite unit showing the widespread but slight disruption of the lamination by vertical restricted, small burrows.



Kimmeridge Clay microfaunas. What is particularly lacking is the analysis of microfaunal variation between lithologies and up-sequence trends within the same lithology. Many of the kerogenous lithologies (including some of the more organic-rich mudstones) are virtually unprocessable using ordinary techniques and much of the microfauna may be damaged or lost by the application of harsher methods or longer processing times (J. Exton pers. comm.).

The foraminiferal assemblages are dominated by a low diversity assemblage of eurytopic agglutinating forms such as Trochammina, Haplophragmoides, Ammobaculites, Reophax and Textularia (Lloyd, 1959; J. Exton pers. comm.) which are ubiquitous in the Upper Jurassic. Although diversity is low, densities are often high and the agglutinating foraminifera are small (often predominantly VF sand grade), thin-walled, virtually transparent and difficult to identify (J. Exton pers. comm.). Calcareous foraminifera are dominated by nodosariids which although not particularly abundant are often well preserved (possibly indicating an R.P.D. close to the sediment surface - see Chapter 2). According to Kilenyi (1969) the ostracod fauna is of low diversity and low density; he reports that the autissoidorensis to wheatleyensis zones are "almost completely barren" (p.155) but this largely due to the difficulties of processing these organic-rich sediments. Good faunas were found in the Black Head Member, a faunal break then occurred at the base of the elegans zone and reasonable faunas did not then re-appear until the Hounstout Member was reached. Bottom oxygenation may again be invoked as the principal faunal control.

ASPECTS OF THE TRACE ELEMENT GEOCHEMISTRY, SWALLAND MEMBER

The principal source of trace element data on the type Kimmeridge Clay is Dunn (1972) with additional information in Dunn (1974) and Cosgrove (1970). Only that part of this data which is relevant to the nature of the Kimmeridge palaeoenvironment will be discussed here, namely the patterns of distribution of chromium, vanadium, molybdenum and uranium. Dunn (1972) notes that according to the criteria given by Krejci-Graf (1964, 1972; see also Chapter Two) the sediments of the Kimmeridge Clay are predominantly of sapropel affinities. Calculation of vanadium:chromium ratios from Dunn's data shows values for the oil shales and coccolith kerogenous shales of between 1.5 and 5.2, whereas values above unity are supposed to indicate sapropels. Chromium values show quite clear minima at the levels of the most kerogenous bands. Calvert (1976) has reviewed the trace element character of organic-rich sediments and found that molybdenum is preferentially concentrated in sapropels due to adsorption on organic matter and coprecipitation with FeS. In the Kimmeridge Clay molybdenum shows a strong correlation with the most organic-rich lithologies: the Blackstone contains up to 92 ppm (Cosgrove, 1970) whereas normal mudstones contain 0-10 ppm. These trace element distributions suggest that anoxic conditions may have extended above the sediment-water interface for at least part of the time during which the oil shales were being deposited. The high levels of iodine in the Blackstone reported by Cosgrove (1970) are also compatible with such an interpretation (see Calvert, 1976).

Dunn (1972) has given trace element data for close-spaced samples through the White Stone Band (see his Fig. 48, p.280). The rhythmite parts of the band show a sapropel type V:Cr distribution but there is a sharp increase in Cr values in the overlying coccolithic marl indicative of improved bottom oxygenation. Molybdenum values are high in the rhythmites (particularly when the purity of the White Stone Band is taken into account), again quickly diminishing in the coccolithic marl. Molybdenum shows a sharp peak of 120 ppm in the 'central' oil shale. It is interesting but not surprising to note that the trace element data are in good agreement with the palaeo-environmental conditions as determined by the distribution of trace and body fossils.

Cosgrove (1970) reported that the level of uranium in the Kimmeridge Clay was below the limit of dectability for the method of analysis that he used on samples (i.e. less than 10 ppm). During the present study a preliminary analysis of uranium and thorium in the Kimmeridge Clay Formation (Kimmeridge and Swalland Member) was undertaken using a Geometrics Exploranium portable gamma ray spectrometer. The highest value recorded for uranium was 7.5 ppm (an average of several readings summing more than 1000 counts for the uranium window) which tends to support Cosgrove's findings. The uranium and thorium values determined by this method (and the calculated Th:U ratios) are shown in Table 5.3. Theoretical considerations would predict (a) an inter-relationship between Th + U contents and total gamma flux (since these two elements contribute about 90% of the latter; see Davies, 1978; Bjorlykke et al. 1975) and (b) a correlation between uranium and organic carbon contents (see Calvert, 1976; Swanson, 1960). In the Kimmeridge Clay in Lincolnshire the highest gamma responses occur in oil shale and kerogenous shale bands and are lowest in calcareous sediments (see Gallois, 1973) and one would expect this trend to be repeated in Dorset.

From an examination of Table 5.3 several points are immediately obvious:-

i There is little relationship between uranium contents and organic matter.

ii The highest uranium contents occur in the more kerogenous coccolithic sediments.

iii Non-kerogenous dolostones show higher Th + U contents than their kerogenous counterparts.

iv The three clay bands sampled show higher Th + U values than either the dolostones or oil shales.

The low uranium contents are something of a surprise in the oil shales since one would have thought their high organic carbon contents (and high phosphorous levels - see Dunn, 1972) would have favoured the accumulation of uranium. The highly sapropelic (as opposed to humic) contents of these sediments may be partly responsible for this anomaly

TABLE 5.3 Gamma-ray Spectrometer Data

Sample lithology or position	K_2O%	ppm U	ppm Th	Th/U	% Contribution* to total γ U	Th	K
Flats Stone Band (K.D)	1.3	1.5	4.0	1.9	49	36	16
Yellow Ledge (K.D)	1.4	2.1	4.4	2.1	55	32	14
Cattle Ledge (N.K.D)	2.3	2.0	6.9	3.5	42	40	18
Shaley clay	3.7	6.5	12.6	1.9	57	41	12
Clay	2.7	3.3	9.0	2.7	49	36	15
Coccolithic kerogenous laminite	4.6	6.6	15.0	2.3	53	33	14
Blackstone (O.S)	2.4	3.8	7.0	1.8	58	29	13
Rope Lake Head (N.K.D)	2.0	1.9	5.7	3.0	45	37	18
Short Joint Coal (O.S)	2.2	3.2	6.9	2.2	54	32	14
Kerogenous shale	4.1	4.1	12.2	3.0	46	37	17
Basalt stone (N.K.D)	2.0	1.2	4.9	4.1	37	41	23
White Stone Band (C.L)	1.0	3.8	1.8	0.5	81	11	8
Middle Stone Band (C.L)	2.7	7.1	7.0	1.0	71	19	10
Freshwater Steps Band (C.L)	1.6	7.5	3.0	0.4	84	9	7
Clay	3.5	2.7	10.4	3.9	39	42	19

K.D Kerogenous dolostone

N.K.D Non-kerogenous dolostone

O.S Oil shale

C.L Coccolith limestone

* From formula in Bjorlykke _et al._ 1975

and supply effects may be indicated. Davies (1978; Log 6) gives a diagrammatic plot of total residual gamma activity for samples collected at about ten foot intervals through the Kimmeridge Clay Formation which also shows a rather anomalous pattern. Prominent peaks in gamma activity only occurred at the levels of the Yellow Ledge and Flats Stone Band kerogenous dolostones (disagreeing with (iii) above) and at a horizon ∿3m above the Short Joint Coal; the Blackstone correlated with a conspicuous low. A rather less well developed peak corresponded to the Middle Band coccolith limestone. Although these findings can only be regarded as preliminary they cast some doubt on the accuracy of lithology-gamma ray log correlations in the Kimmeridge Clay. The thorium:uranium ratios, however, do indicate highly reducing depositional conditions where these would be expected (shown by Th:U <2; Bjorlykke et al. 1975) except in the coccolithic kerogenous laminite, but this divergence is explained by the heterogeneous nature and thinness of the band that was analysed.

PALYNOFACIES CHARACTER OF THE KIMMERIDGE CLAY FORMATION

(i) Overall trends in the Kimmeridgian-Portlandian sequence

The only truly palynofacies orientated accounts of this sequence are those of Riley (1974) and Mebradu (1976, 1978) but additional information can be found in Ioannides et al. (1976) and Norris (1969). General comments on the nature and origins of organic matter in the Kimmeridge Clay Formation may be found in Downie (1955), Cosgrove (1970), Dunn (1972) and Bellamy (1979). No attempt was made during this study to verify the nature of the palynofacies of the overall sequence as recorded in these references, but this data was re-evaluated in accordance with the principles formulated in Chapter 3. My own observations were confined to an examination of the cyclic sedimentation in the Swalland Member.

Riley (1974) notes that for most of the Kimmeridge Clay the abundance of A.O.M., the relatively high dinocyst diversity (including a high proportion of chorate cysts) and lack of acritarchs, implies a low energy, open marine environment with oxygen deficient bottom conditions. Passing upward through the Hounstout Member, Riley (ibid) records increasing abundances of miospores and a decrease in dinocyst diversity (with reduced chorate cysts and an increase in peridinacean forms), coupled with a progressive increase in the proportion of acritarchs. In the Black Nore Member he found a low diversity miospore assemblage dominated by Classopollis and inaperturate forms, rather few dinocysts, but abundant ($\leq$ 80%) long-spined, acanthomorph acritarchs. Phytoclast debris and reworked spores were also more common in this unit. The rest of the Portland Sand Formation contained only a few poorly preserved palynomorphs, with a generally similar miospore assemblage, few dinocysts, quite abundant short-spined, acanthomorph acritarchs, abundant phytoclasts and locally abundant microforaminifera.

Riley (1974) attributed the low diversity miospore assemblage of the Black Nore Member and Portland Sand Formation to winnowing in a shallow water, high energy environment. Norris (1969) has previously recognised the low diversity of the sporomorphs in this part of the section, referring to this assemblage as 'suite A'. These miospores are dominated by 'light' morphotypes - Cerebropollenites, thin walled deltoid spores, bisaccates, Classopollis, Peripollenites and inaperturate forms, with only spasmodic occurrences of heavier pteridophyte spores. Classopollis and bisaccates are the most numerically dominant forms, in general indicating a 'distal' character. Norris (1969, p.603) comments: "Both the Upper Kimmeridgian and Portlandian comprise marine sediments deposited in an offshore environment. Considerable sedimentary sorting of the miospores by wind and water is likely and may have resulted in only a fraction of the total available spore-pollen population from the adjacent land areas reaching the depositional site. This relative impoverishment may have been accentuated by the extremely limited flora that occupied the coastal sites at the time". Winnowing in the normal sense would have left a 'heavy' sporomorph assemblage behind and removed the light fraction; the evidence is not compatible with this viewpoint. The data suggests a relatively distal environment for the Black Nore Member with no immediate fluvial sources of spores, such that even when the energy of the environment increased higher up in the Portland Sand Formation only a relatively 'light' miospore assemblage was present (although the more turbulent conditions are reflected by the plankton). This example serves to emphasise the importance of proximity to source rather than proximity to shoreline.

Riley (1974) also carried out a detailed study of the palynofacies variation through the White Stone Band (see his Fig. 17). He found that the coccolith limestone contained a palynomorph assemblage dominated by miospores (78-99%, mostly bisaccates, simple sphaeromorphs and Classopollis) while chorate dinocysts were the most abundant representatives of the plankton, with only <5% acritarchs. The central oil shale was found to be 'virtually devoid of microplankton' and consisted mostly of A.O.M., phytoclastic debris and miospores. Passing from the marl bands on either side into the limestone the numbers of microforaminifera decreased. The significance of these distributions will be discussed in detail later, but a distal depositional environment of poor bottom oxygenation is indicated.

Mebradu's work on the Dorset Section, though of greater stratigraphical extent is of more limited value (see Mebradu, 1976, 1978). This is mainly due to the small number of counts employed for each sample (often less than 100!) but is aggravated by the poor labelling and calibration of his diagrams. However, his general findings include:-

(a) The Kimmeridgian has a less dense and less diverse miospore assemblage than the underlying 'corallian'.

(b) Spores are, on average, least abundant (but most variable) between the base of the scitulus zone and half way up the hudlestoni zone.

(c) Microplankton are, on average, most abundant (but again most variable) between the base of the scitulus and top of the pectinatus zones (biased by incorrect records of tasmanitids?).

(d) The proportion of miospores increased in the Upper Kimmeridgian (allegory of (b) and (c)).

(e) The Kimmeridge Clay was characterised by a distal, open sea environment with reducing bottom conditions.

These observations generally confirm those of Riley (1974). The variability in samples from the scitulus to pectinatus zones (points (b) and (c) above) may reflect the cyclic sedimentation observed in this interval (see later).

From trace element data (mainly the distribution and abundance of I and U) Cosgrove (1970) and Dunn (1972) considered the principle source of organic matter in the Kimmeridge Clay Formation to be phytoplanktonic organisms. Most of Riley's (1974) discussion on the nature of the Kimmeridge Clay palaeo-environment is taken almost verbatum from an unpublished manuscript by Downie (essentially a re-write of his 1955 thesis), including the view that the A.O.M. (and therefore the bulk of the organic matter in these sediments) represented flocculated globules of dopplerite (humic) gel. Although the C, H, O data in his own thesis suggested the contrary, Downie apparently again voiced the same opinion in Ioannides et al. (1976). Needless to say, neither Downie or Riley still hold this view, which apparently came about because the dominant structured component of the kerogen associated with the A.O.M. is of terrigenous origins (i.e. phytoclasts and sporomorphs). Gallois (1976), more impressed by the association of coccolithic and kerogenous sediments and the evidence for modern A.O.M. being of marine derivation, concluded that phytoplankton (principally dinoflagellates) were the main source of organic matter. Davies (1978) has since used a variety of organic-geochemical techniques to conclusively demonstrate that the kerogen is predominantly of a marine origin (verified by pyrolysis results obtained by Robertson Research that indicate a sapropelic 'Type II' composition ;

see also Williams & Douglas, 1981). In good agreement with the palynomorph distributions, Davies (1978) notes an increase in allochthonous organic materials in the Hounstout and Black Nore Members.

(ii) Lithofacies-palynofacies correlations in the Swalland Member

Gross kerogen characteristics

When examining palynological residues from these sediments it was found to be necessary to modify the categories normally used in the quantitative description of the overall kerogen that were outlined in Chapter Three. Several types of amorphous particles were found to be present (Plate 5.11):

(a) Normal 'fluffy' amorphous particles containing various micro-inclusions.

(b) Platey, brown amorphous particles with angular outlines but no internal structure (save inclusions)

(c) Platey or flake-like, thin,yellowish,lamellar particles without internal structure. Almost membranous in appearance; usually lacking inclusions. Possibly a thinner version of (b).

(d) Brown particles similar to (b) with a fibrous structure best seen at the edge of the particles, but running all the way through them. Such particles tend to have elongate shapes and are pseudo-amorphous.

This diversity of 'amorphous' particles is most pronounced in the kerogenous shales and oil shales but these categories (along with the normal ones described in Chapter Three) were used in the typing of all the Dorset samples for the sake of uniformity. The (b) and (c) types ('platey' and 'lamellar' A.O.M.) may simply represent compaction artifacts produced from the normal A.O.M. by the compression of the kerogenous sediment. The 'fibrous' organic material (type (d) above) is less easily accounted for. Considering the internal structure of this material (even more apparent under U.V. light) I have come to the opinion that it may represent rather degraded,non-lignified phytoclastic debris. If this origin is accepted it suggests that the Blackstone

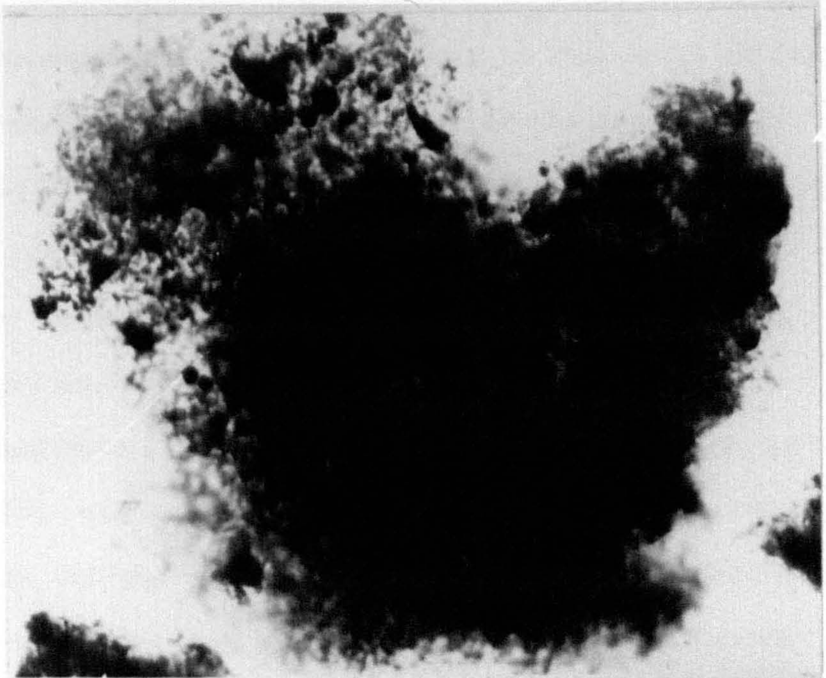
PLATE 5.11. Varieties of amorphous and pseudo-amorphous organic matter observed in Dorset oil shales (Blackstone, sample BS15).

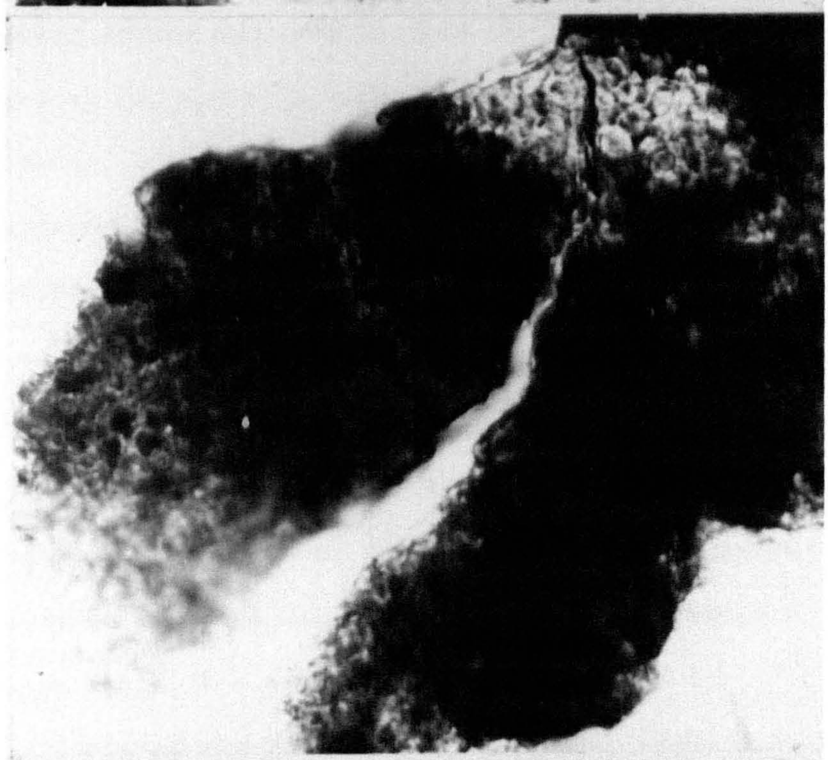
Top: Normal 'fluffy' A. O. M.

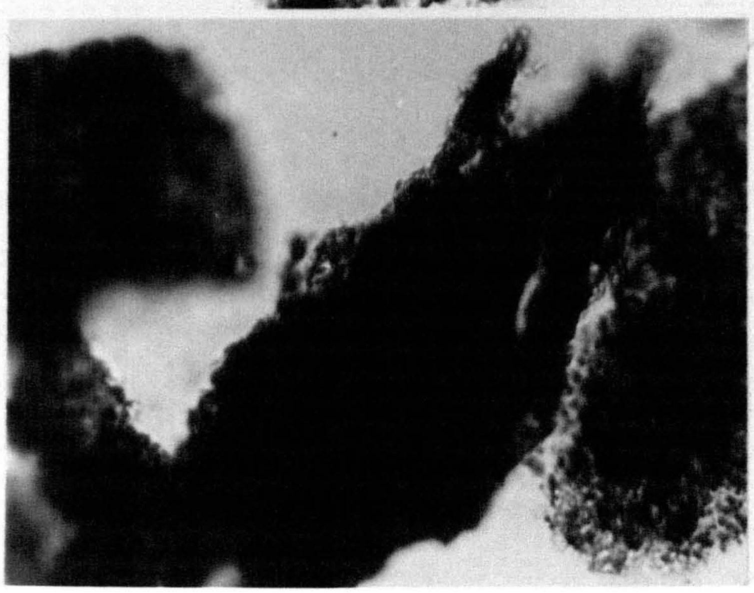
Centre: Platey pseudo-amorphous organic matter

Bottom: Fibrous organic matter interpreted as debris of poorly lignified marine macrophytes.

All photographs to same scale; width of photographs approximately 0.15 mm.







contains up to 30% or more of macrophyte material in some parts of its thickness, which seems an apparent contradiction of the organic geochemistry (Davies, 1978) and is certainly not reflected in the uranium content. Given the comments of Masran and Pocock (1980) reported in Chapter Three, it is suggested that this fibrous material represents poorly lignified, possibly bacterially reworked, marine macrophyte detritus. Such an explanation would help to account for the high iodine values (Cosgrove, 1970) and the unusual carbon composition reported by Murchison and Douglas (1979). The fibrous organic particles observed in palynological preparations may correspond to the relatively long red-orange strips of organic matter observed in thin sections from the kerogenous lithologies in the Swalland Member (sometimes longer than the size of the thin section). Marine macrophytes may have been the substrate for the epifaunal brachiopod Discina which is sometimes the only conspicuous macrofossil in the more kerogenous beds; it is also interesting to note that modern Codakia dominates low diversity mollusc assemblages in recent Thalassia seagrass communities! (Jackson, 1977; Wanless, 1981). It is difficult to estimate to what extent this type of organic matter may be occurring in situ; it is quite possibly derived from shallower, more marginal areas of the basin. Similar fibrous organic matter has been identified amongst the normal A.O.M. in comparable coeval sediments of the Volga Basin (Strakhov, 1969, p.344), where it has been interpreted as seaweed or cyanobacterial debris. Aigner (1980) has provided evidence for bivalve shell 'plasters' in the Kimmeridge Member being formed by redeposition of shells during 'hydro dynamic events'; it is therefore possible that both Lucinids and fibrous organic matter may have been derived from shallower water seagrass communities.

The results of the palynological kerogen typing of the Kimmeridge samples are given in Table 5.4. The oil shales can be clearly distinguished from normal mudstone shales by the relative porportions of

TABLE 5.4 Kerogen data

		% Particle abundances										
Sample	Lith.	FM	Wu	I	UP	P	M	PA	LA	A	Pd	F
B.S.1	Oil shl	21.0	0.6	2.0	0.4	0	0	42.0	30.2	3.0	0.8	0
2	"	19.2	0.4	2.4	0	0.8	0.2	47.4	23.8	4.0	1.8	0.2
3	"	17.6	1.4	3.2	0.2	0.2	0.2	42.9	20.4	10.7	2.2	1.2
5	"	19.4	0.2	0.8	0.4	0	0	52.0	16.8	6.9	2.4	0.8
7	"	24.4	0.4	2.4	1.6	0.6	0	38.2	20.4	8.0	3.8	0.6
9	"	32.4	0.4	1.8	0.6	0.2	0	44.0	17.4	2.0	1.2	0.8
11	"	32.4	0.2	1.4	0	0	0	46.8	15.2	3.2	0.8	0.2
13	"	23.0	1.8	3.0	2.0	0.4	0.4	23.4	4.2	37.8	4.0	1.0
15	"	33.0	1.8	2.4	1.4	0.2	0.8	15.6	5.8	36.0	3.0	0.8
17	"	29.4	0.8	3.0	2.8	0.6	0.8	18.8	9.6	32.4	1.8	0.2
19	"	34.3	0	0.5	0.5	0	0	58.5	5.2	0.5	0	0.2
23	"	32.9	0	0.5	0.2	0	0	53.9	7.5	4.0	0.7	0.2
25	"	18.4	0.2	0.6	0.2	0.2	0	64.4	10.2	5.4	0.4	0
27	"	17.0	0.8	0.2	0.2	0	0	65.2	12.6	3.4	0.6	0
29	"	17.6	0	0.2	0	0	0	78.2	3.8	0.2	0	0
30	"	12.8	0	0.6	0	0	0	82.6	2.8	1.0	0.2	0
SJC 1	"	20.4	1.8	0.2	0.6	0	0	38.0	2.0	32.6	4.2	0.2
4	"	13.6	1.0	0.6	0.8	0.4	0	59.7	0.6	19.4	3.0	1.6
8	"	17.8	0.6	0.6	0.2	0	0	61.0	17.0	1.2	1.4	0.2
14	"	8.8	0.4	0.4	0	0	0	70.0	14.8	0.4	4.2	1.0
RVT 1	Mdst	6.2	1.8	2.6	4.8	1.2	1.0	12.4	4.0	64.0	3.6	0.6
3	"	11.0	0.8	2.2	4.4	1.0	0.2	31.6	9.8	37.2	2.6	0.2
5	"	10.0	1.4	1.2	3.2	0	0.2	41.4	21.0	16.0	4.6	1.0
10	"	1.0	6.6	2.8	4.2	1.0	0.4	8.4	17.2	52.0	4.8	2.6
12	"	7.0	4.4	3.4	7.4	0.8	0.2	6.4	1.8	64.6	3.8	1.0
14	"	10.4	8.3	4.1	7.0	3.3	1.7	3.1	1.9	51.1	8.1	1.0
16	"	12.7	9.8	3.3	6.6	2.3	1.4	3.9	0.4	54.7	4.1	0.8
18	"	11.0	4.3	1.1	12.3	6.5	2.6	2.1	6.5	38.9	13.1	1.5
KE 1	Mdst	2.8	4.2	2.4	4.7	1.3	0.4	0.3	0.1	74.8	8.6	0.4
2	"	2.0	7.4	8.2	8.4	1.8	1.4	0	0	38.6	33.6	0.4
5	K. Shl	5.4	4.0	2.2	1.4	0	0	16.6	5.2	61.8	3.0	0.4
15	Mdst	0	12.8	4.5	24.8	3.1	5.0	0.6	0	35.9	10.3	3.1
22	"	0.4	8.8	0.6	8.6	1.4	0.4	0.8	0.4	72.2	2.2	0.6
30	"	1.8	9.8	1.0	6.8	0	0.2	7.4	9.4	61.0	2.6	0
YL 2	K.D	2.8	2.4	1.8	0.6	0	0	35.4	5.8	49.2	1.6	0.4
FWS22	C.L	4.4	2.2	0.8	4.8	0.8	2.2	0.4	0	79.8	4.8	0.6
23	"	4.2	0.8	0.6	10.8	2.0	2.0	0	0	64.6	17.0	0

Mdst = mudstone
K.Shl = kerogenous shale
K.D = kerogenous dolostone
C.L = coccolith Lst.
FM = fibrous material (pseudo-amorphous)
PA = platey amorphous
LA = 'lamellar' amorphous
A = ordinary amorphous

normal and platey-lamellar A.O.M. and by palynomorph abundances. The low palynomorph abundances of the oil shales are not just a masking effect but hold true even for oxidised palynopreparations with A.O.M. largely removed. Ioannides et al. (1976) report that sporomorphs dominate the palyno-assemblage of the Blackstone, being more abundant than in the mudstones above and below; this is certainly true in the relative sense since plankton is very rare in the oil shales. The kerogenous shales have a composition intermediate between those of the oil shales and normal mudstones. The Yellow Ledge kerogenous dolostone has a kerogen composition like that of the kerogenous shales while the non-kerogenous dolostones are characterised by 'pellicular organic matter' (sensu Combaz, 1980) and were not suitable for counting.

Some significant differences were observed in the responses to ultra violet light between the kerogens of the different lithologies (see also Fig. 5.3). These differences are summarised below:

(a) Oil shale kerogens showed strong fluorescence under U.V. with a more-or-less homogeneous response.

(b) Kerogenous shale kerogens showed moderate fluorescence under U.V. and were characterised by a heterogeneous response that picked out common-abundant strongly fluorescent inclusions within the A.O.M. particles (probably alginitic debris).

(c) Normal mudstone-shale kerogens showed only weak, dull fluorescence with a more-or-less homogeneous to slightly heterogeneous response.

(d) The dolostones and coccolithic sediments also showed dull fluorescence; modern subaerial oxidation may be an important factor in the case of the porous 'coccolith limestones'.

These differences are most likely produced by the differing lipid contents of these lithologies which relate back to the original nature of the organic matter and the bottom oxygenation in the depositional environment (see Chapter Two). The latter factor seems the most important control, determining how much of the original lipid content in the autochthonous organic matter survived to be preserved in the sediment and hence available to fluoresce under the U.V. light. The high sulphur content of the (A.O.M.) kerogen in the Blackstone oil shale reported by Downie (1955) would, according to Kendrick (1979) indicate anoxic bottom conditions which would be most suitable for lipid preservation. The kerogen data is shown in Fig. 5.4 and Table 5.4.

Palynofacies characteristics

The palynofacies character of 11 of the Dorset samples are shown in Table 5.5 (see also Plate 5.12). No oil shale samples are present because of the rarity of palynomorphs in this lithology, and the 'pellicular' nature of the non-kerogenous dolostone residues made them unsuitable for detailed counts. The most significant facts to be drawn from Table 5.5 are:-

(a) Proximate dinocysts are only abundant in the normal mudstone shales.

(b) Chorate dinocysts are the single most abundant representatives of the microplankton in the coccolithic sediments (predominantly the Systematophora morphotype).

(c) the dinocysts are practically all gonyaulacean forms with only rare ($\leq$0.6%) peridinacean types (generally Pareodinia or Netrelytron sp.).

(d) Acritarchs are only poorly represented but long-spined acanthomorph acritarchs are the most common.

(e) Prasinophycean algae are uncommon (tasmanitids>pterospermopsids >cymatiosphaerids). The proportions of tasmanitids in the assemblages reported by Mebradu (1976, 1978) are considered by Dr. L.A. Riley (pers. comm.) and myself to be too high and may include the small dense sphaeromorph sporomorphs that are common in some samples.

(f) Microplankton is generally most abundant in the mudstones but miospores dominate all the assemblages (i.e. $>70\%$).

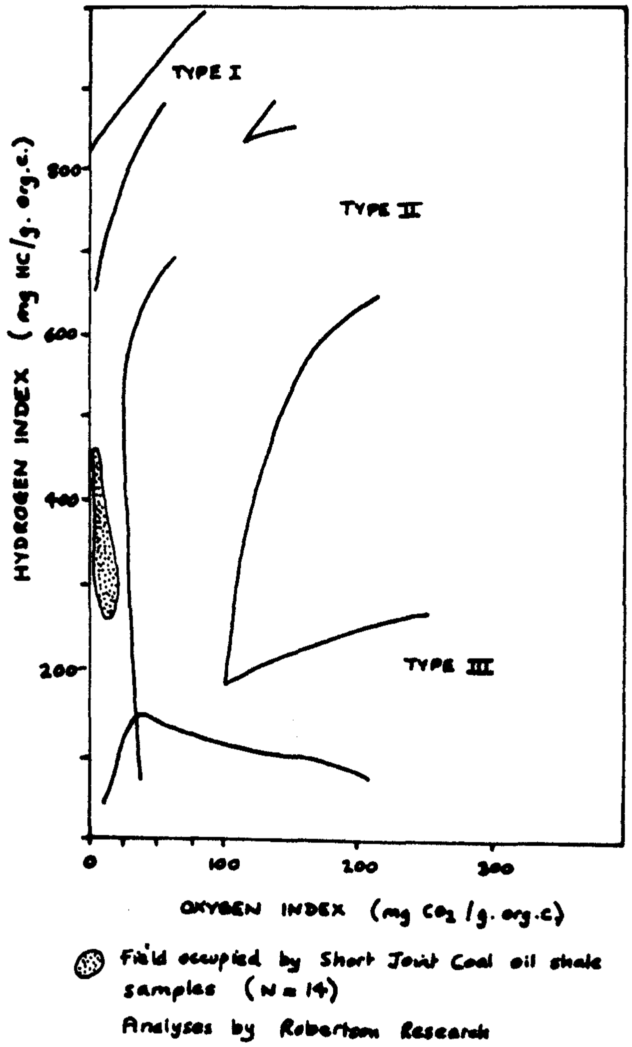


Fig. 5.4 Pyrolysis results for Short Joint Coal oil shale samples

TABLE 5.5 Palynomorph data (Key on page 176-177)

Sample		Lith	Palynomorph groups (%) A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	Classopollis	Microf	No. Counts
KE	1	Mdst	8.0	0.4	1.8	1.2	0.2	0	0.4	1.8	7.2	7.2	0.8	3.1	0	2.0	62.2	3.7	15	489
	2	"	11.5	1.1	4.9	1.8	0.2	0	0.9	2.5	10.8	6.7	0.9	3.8	0.7	4.7	53.5	0.4	4	445
	15	"	4.8	1.3	3.1	2.3	0.2	0	0	0.6	2.7	12.7	2.1	2.9	0	1.0	60.7	5.6	27	519
	22	"	7.7	0.6	0.9	4.9	0.6	0	0	2.1	2.1	10.1	0	0	0	0.3	65.0	5.2	4	326
RVT	14	"	7.2	0.4	2.2	3.6	0	0	0.2	0	1.8	3.1	0.2	3.8	0.2	15.4	61.7	0	36	447
	18	"	19.5	2.1	2.1	12.5	0	0	0	0	3.5	3.7	2.1	2.1	0.9	2.8	48.7	0	48	431
KE	5	K.Shl	0	0	0	0	0	0	0	0.7	2.0	14.9	0	0.7	1.4	0	79.7	0.7	71	148
WS	14	CL (B)	0	0.2	0	2.6	0	0	0.5	0	3.6	4.5	0.2	0.5	0.2	0.5	75.5	11.8	13	424
	16	C.L (R)	2.3	1.8	0.8	3.1	0	0.3	0	2.6	5.7	7.0	0.3	3.9	0.5	3.1	54.9	13.3	5	384
FWS	22	K.C.L	0.4	0.2	0.2	2.0	0.2	0	0.7	0.7	1.1	19.3	1.1	3.1	0.2	2.6	68.0	1.3	7	455
	23	C.L. (B)	0.6	1.0	0.2	17.5	0.2	0.2	0.8	0	3.6	3.6	0.4	0.8	0.2	0	68.1	3.0	2	526

Key to Lithological codes as Table 5.4 plus B (Bioturbated) R (Rhythmite) K.C.L (Kerogeneous coccolith limestones)

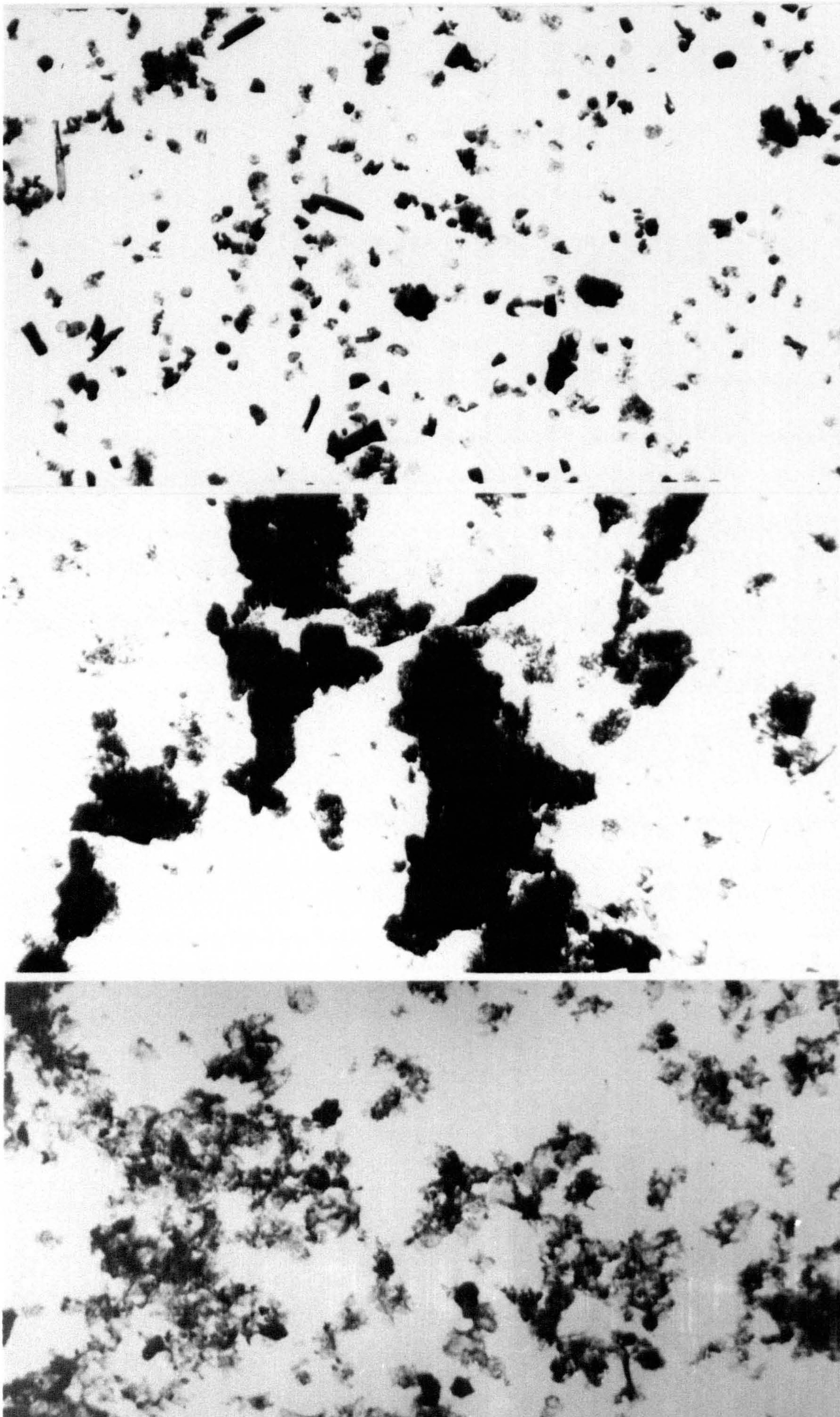


PLATE 5.12. Palynofacies characteristics of some Kimmeridge Clay lithologies. Widths of photographs in parentheses.

Top: A distal shelf mudstone assemblage characteristic of predominantly aerobic conditions (high palynomorphs, low phytoclasts, low A.O.M). Sample KE 15 (1.8 mm)

Centre: Typical appearance of an unoxidised preparation of dysaerobic-anoxic sediment facies (high A.O.M., low palynomorphs, low phytoclasts). Palynomorphs partly masked. Sample FWS 23, a coccolithic kerogenous laminite (2.9 mm).

Bottom: Pelicular A.O.M. (sensu Combaz) typical of dolostones. Kerogen bears imprints of the dolomite crystals, sometimes giving a pseudo-cellular appearance. Sample CL 4 (0.8 mm).

(g) Bisaccates are the most abundant miospores, the next most abundant group being light trilete spores or Classopollis. The latter is most abundant in samples from the pectinatus zone upward and Riley (1974) notes it becomes the dominant form in the Upper Hounstout and Black Nore Members. Unclassified sphaeromorphs dominate all the samples, however (see Table 5.5).

(h) In the coccolithic sediments Classopollis commonly occurs in groups of 3-8; only separate pollen occurred in the mudstone samples.

(i) Heavy sporomorphs are rare but most common in the normal mudstone shales.

(j) Microforaminifera:palynomorph ratios are generally highest in the mudstones but the highest value was recorded in the kerogenous shale sample (KE5) which had both an abundance of microforaminifera and a dearth of palynomorphs.

(iii) Discussion

It is not possible to give an authoritative account of the palynofacies of the Kimmeridge Clay Formation as a whole. Although there are two previous studies on this section (i.e. Riley, 1974; Mebradu, 1976, 1978) there is not an adequate coverage of sufficiently detailed, combined, quantitative kerogen and palynomorph analyses. The few results presented here cannot provide a realistic picture of these broader trends. Riley's palynomorph data is adequate for the top part of the sequence but Mebradu's data is less useful, being confined to total miospore and total plankton estimates without morphotype break-downs. Accepting these difficulties a general description of the sequence is given below.

Brookfield (1973a, 1978b) describes the baylei-mutabilis interval as consisting of fluvially influenced, low-energy shelf deposits. Although there is no evidence for the existence of any deltas, such a fluvial influence would be compatible with the high miospore percentages observed by Mebradu (e.g. see Mebradu, 1978, Fig. 1) in this part of the succession. The miospore input apparently declines (partly due to higher sediment dilution?) in the central part of the sequence (scitulus to lower hudlestoni zones). In this part of the Formation the miospore assemblage has a markedly distal aspect with bisaccates the most abundant forms (other than the simple sphaeromorphs and inaperturate component). It would be interesting to know whether the heavier morphotypes were progessively filtered out through the Lower Kimmeridgian, in particular the deltoid pteridophyte spores which were poorly represented in my samples. The erratic nature of the miospore distribution in the central interval observed by Mebradu may relate to variations in accumulation rate as well as fluctuations in supply, as is suggested by the cyclic sedimentation in this part of the sequence.

In the Swalland Member the palynofacies of the normal mudstone shales indicates a distal, open marine environment that was either seasonally (or continuously) mixed to the bottom. The major kerogen component is (degraded) A.O.M. and the allochthonous component generally varies between 10 and 30%. The kerogenous shales and oil shales contain mainly undegraded or relatively undegraded A.O.M. and yield very few palynomorphs but appear to contain proportionately more miospores than the mudstones (see Ioannides et al. 1976). The palynofacies of the coccolithic sediments suggests a distal open marine environment that was 'continously' stratified with stagnant, probaly poorly oxygenated bottom conditions. The paucity of dinocysts in the coccolithic and kerogenous sediments suggests hydrographically more stable conditions than during the deposition of the normal mudstone-shales and does not necessarily imply any changes in dinoflagellate productivity.

The increasing importance of Classopollis (cheirdepidacean conifer pollen) from the pectinatus zone upwards, and its apparent replacement of the pteridophyte spore component,probably reflects changes in the land floras produced by the Late Jurassic transition to a warm, dry climate. The growth in importance of the Classopollis flora in the Late Jurassic is a global phenomenum.

In the Hounstout Member the palynofacies indicates a gradual proximal shift with decreasing A.O.M. and increasing amounts of allochthonous organic matter. Although this shift is reflected in the plankton (as recorded by Riley, 1974) little change occurs in the miospore assemblage (bisaccates, inaperturates, Classopollis and other light sporomorphs) except in terms of relative abundances, suggesting an impoverished parent flora and lack of fluvial inputs. Turbulent, high stress conditions in the Portland Sand Formation are reflected by the replacement of dinocysts by acritarchs. Nearly all of the Hounstout and Black Nore Members appears to have been deposited in a seasonally or continuously mixed environment with well oxygenated, in part turbulent, bottom conditions unsuitable for any appreciable preservation of A.O.M.

CYCLIC SEDIMENTATION IN THE KIMMERIDGE CLAY FORMATION: INTERPRETATION AND SYNTHESIS

(i) Modern analogues: Quaternary sapropel sequences, Eastern Mediterranean and Black Sea

The Late Pleistocene-Holocene sediments of the Black Sea (the type area of the 'euxinic' environment) were briefly described in Chapter 2. The sequence of normal sediments followed by sapropels,and subsequently coccolithic sapropels,also occurs in the eastern Mediterranean basin (including the Adriatic and Aegean), although the two occurrences are separated in time (e.g. see Ryan, 1971; Kidd et al. 1978 and Stanley & Blanpied, 1980). From the modern Black Sea environment we can determine the nature of the depositional conditions which produced the sapropel sequence and this appreciation has been applied to the interpretation of the Mediterranean sequences. In terms of the present study the Mediterranean sequences are most similar to the type Kimmeridgian because several cycles are present, but the interpretation is again based on the Black Sea where we can see one of these cycles still in progress. The importance of these modern analogues is that they establish and explain the genetic sequence of normal sediments followed by sapropels and then coccolith oozes.

The Mediterranean sapropel sequences have been studied in detail (see Van Straaten, 1971; Hesse et al. 1971; Nesteroff, 1973; Maldonado & Stanley, 1975, 1976; Kidd et al. 1978; Sigl et al. 1978; Dominik & Stoffers, 1978; Stanley, 1978; Stanley & Maldonado, 1975; Mangini & Dominik, 1979). The typical features of the sapropel cycles as recorded in these references are shown diagrammatically in Fig. 5.5 and some of the more salient points are itemised below:

(a) Despite the nomenclatural vagaries of D.S.D.P. and other parties the eastern Mediterranean "sapropels" are mostly sapropels in the sense of sediments deposited under anoxic bottom conditions (see Chapter 2), but some may be gyttjas (see Kidd et al. 1978).

(b) In the eastern Mediterranean the sapropels are underlain (and often overlain) by 'precursor' sediments (the grey marls of Nesteroff, 1973 and organic oozes and protosapropels of Maldonado & Stanley, 1976, etc.) which are not seen in the Black Sea. These sediments apparently indicate dysaerobic to anaerobic bottom conditions prior to complete stagnation - which was evidently more sudden in the euxine basin.

(c) The sapropels are confined to depths 800-1,000m below present sea level (Stanley, 1978).

(d) The sapropel sequences accumulated during rising and falling sea levels and during both warming and cooling trends in palaeoclimate (Thunnell et al. 1977; Stanley & Maldonado, 1979).

(e) The accumulation rate of the sapropels is up to two times lower than the Quaternary average (Mangini & Dominik, 1979). See also the discussion in Bottema & Van Straaten (1966).

(f) Nesteroff (1973) considers the low carbonate contents of the sapropels and their precursors to indicate low primary productivity and/or dissolution. Sigl et al. (1978) note that microfossil preservation is in fact greater in the sapropels than the background sediments and that the organic content of the sapropels is not related to their micro-

"FACIES"	LITHOLOGY	CHARACTERISTICS
AEROBIC	CALCAREOUS OOZE	Unusually carbonate-rich ooze frequently tops sapropel sequence.
AEROBIC	OXIDISED LAYER	Light orange-yellow-beige sediment. Usually thin (~5-7cm) often absent. No pyrite, gypsum or pteropods are present.
ANAEROBIC TO DYSAEROBIC	ORGANIC OOZE TO PROTOSAPROPEL	Usually thin if present. Not all sequences are symmetrical.
ANOXIC	SAPROPEL	Sharp contacts. Usually 5-10cm thick (≤20cm). Black, organic-rich (≤20% Org.C, usually < 10%), microlaminated. General alternation of thick clay and thin coccolithic laminae. Discrete lamina of planktonic forams may also occur. Low carbonate content usual, but good carbonate preservation. Contains conspicuous amounts of pteropods, gypsum crystals and siliceous microplankton. Enriched in Molybdenum and other trace elements. Kerogen is dominantly marine and amorphous. Benthos absent.
DYSAEROBIC TO ANAEROBIC	PROTOSAPROPEL	Often laminated; no burrows. Contains benthic and abundant planktonic forams. Low diversity benthic forams dominated by Globobulimina and Chilostomella sp. Usually thinner than organic ooze.
DYSAEROBIC	ORGANIC OOZE	Olive-gray. Characteristic pyritised worm burrows; intensely bioturbated. Gypsum crystals present. Very few forams. Low carbonate preservation. Benthic forams dominated by same forms as above. Contains ~1.0% org.C.
AEROBIC	OLIVE-GRAY MUD (MARL)	Low carbonate preservation where bioturbated. Benthic foram population diverse but decreasing toward sapropel. <0.5% org.C.

Fig. 5.5 Typical features of eastern Mediterranean Quaternary Sapropel sequences. Organic ooze to oxidised layer generally 30-50 cm. Sources cited in text.

fossil content. Van Straaten (1971) does not believe the occurrences of sapropels is related to increases in palaeo-productivity, as once stagnation was attained this would have resulted in higher org.c. values than are commonly observed.

(g) During sapropel deposition reduced bottom circulation led to the accentuation of local sources of clay minerals (Dominik & Stoffers, 1978). Van Straaten (1971) notes that shallow water displaced faunal remains are absent from the sapropels and believes that low-density turbidity currents were absent at these times and that only normal turbidites reached the basin floor.

(h) The depositional model for the sapropel sequences invokes widespread salinity stratification of the 'Mediterranean' due to changes in the precipitation-evaporation balance, glacial meltwaters (both related to climate) and watermass exchange with the Black Sea basin (freshwater prior to 9,000 y.b.p.). Details may be found in Bradley (1938); Olausson (1961), Vergnaud-Grazzini et al. (1977), Kidd et al. (1978) and Stanley & Blanpied (1981). This stratification resulted in the stagnation and deoxygenation of the bottom water throughout the basin.

(i) In detail a fair degree of variability is observed within the sapropel sequences. Pelagic and re-sedimented varieties are recognised (see Nesteroff, 1973; Kidd et al. 1978), and other characteristics apparently varied with the stability of the depositional environment in different parts of the basin.

(j) Sapropels are relatively enriched in radiolaria in comparison with normal sediments (Caulet, 1974).

Most of the features described above agree well with the theoretical character of stratified (anaerobic-anoxic) basins discussed in Chapter Two. The fact that salinity plays a crucial role in the depositional models of the Black Sea and Mediterranean sapropels is irrelevent to this discussion; watermass stratification (whatever its cause) is the common theme. Although climatic and oceanographic factors are known to be involved in the causation of sapropel deposition in these areas, no clear overall pattern has yet emerged. Clearly, meromixis,whatever its cause,is the main factor.

The coccolithic units of the cycles appear to be somewhat poorly represented in the Mediterranean when compared with Unit One in the Black Sea (see Chapter Two). For the Black Sea I considered the change sapropel- coccolithic sapropel to indicate the arrival of the chemocline-nutricline at a level where seasonal (?) convective processes could mix nutrients into the euphotic zone and encourage brief plankton blooms. It is apparent that in the Eastern Mediterranean the anoxic interface never reached as shallow depths as in the Black Sea (note (c) above) which necessarily implies lower rates of nutrient release because of the lower levels of hydrodynamic energy available for mixing. The anoxic nutrient pool could not have been directly inter-active with the primary production in the euphotic zone. This in turn implicates the processes controlling the depth of the main pycnocline or indicates that the meromictic periods were too short to allow the chemocline-nutricline to achieve its full ascent.

Ryan and Cita (1977, p.198) conclude of oceanic sapropels that "anomalously low surface ocean fertility is suggested by small concentrations of planktonic skeletal components and of abnormal ratios of siliceous to calcareous microfauna".

(ii) Ancient anologues: Upper Cretaceous, North American interior

The Turonian-Campanian interval of North America contains several developments of kerogenous and coccolithic sediments in the Colorado Group including the Greenhorn Limestone Formation (and laterally equivalent "second white speckled shale"), the Niobrara Chalk Formation (and laterally equivalent "first white speckled shale") and the Pierre Shale (see Hattin, 1971, 1981; Simpson, 1975; Byers, 1979). Many features of these sediments are akin to those in the Kimmeridge Clay Formation.

Byers (1979) has described the Upper Niobrara-Lower Pierre kerogenous coccolithic sediments in the following terms. "The only structure is a planar lineation produced by the alignment of fine carbonaceous flakes parallel to each other and the bedding. Coccolith aggregates also form discontinuous linear horizontal bands. However, true laminae (i.e. couplets) are not present, only lenslike bands of darker and lighter brown matrix". He also notes that "there is no sign of any burrowing structures" and that the absence of couplets "indicates a greater uniformity of deposition, a lack of regular (perhaps annual) sedimentation events". These sediments were classified as a lethal pantostrate (isostrate) biofacies (see Chapter Two). This description is very similar to the coccolithic kerogenous shales in the Kimmeridge Clay Formation.

Hattin (1971) has described similar sediments from the Greenhorn Formation as laminated, impure shaley chalks containing flattened fossils and up to 6.3% organic carbon. He records that "the only ubiquitous benthonic fossils are epifaunal inoceramids that apparently lay exposed on the sea floor with the plane of commissure parallel with the sediment water interface. The thin and seemingly fragile shells of these abundant bivalves reflect a quiety, deep-water habitat.... In all the beds the diversity of benthonic macro-invertebrates is low consisting only rarely of more than three or four species" (p.420). He concludes that "the more-or-less laminated (and virtually unburrowed) character of shaley chalk, the scarcity ... of sedimentary structures and textures indicative of wave or strong current action and extraordinarily low diversity of the shell benthos suggests that the bottom waters were poorly circulated and probably low in dissolved oxygen content. Under such conditions organic matter accumulating on the sea bottom would create interstitial reducing conditions ... inimical to the development of a burrowing infauna" (p.421). He also notes that "with the exception of certain beds ammonites are very scarce in shaley chalks" (p.428). This description again has much in common with the Kimmeridgian sediments, perhaps in particular the Freshwater Steps Stone Band with its 'absence' of ammonites and rare oysters.

Simpson (1975) in his jargonised account of the Colorado Group classifies both the first and second 'white speckled shales' as consisting predominantly of the lethal pantostrate (isostrate) biofacies and invokes the Black Sea analogy for these kerogenous and coccolithic sediments. Hattin (1975) has interpreted the coccolithic speckles in the Colorado sediments as faecal pellets of Zooplankton (av. diameter 0.12mm). This interpretation probably applies equally well to the Kimmeridge Clay. Roth et al. (1975) note that modern copepod faecal pellets often contain whole, undamaged coccospheres as has been noted in the Kimmeridge examples. It is the irregular distribution of these coccolithic pellets that simultaneously creates a laminated appearance but fails to define clear laminae and couplets - possibly indicating the absence of clear seasonal peaks in primary productivity (as implied by Byers, 1979) or a dearth of zooplankton. Even the apparently finely laminated nature of the Unit One coccolithic sediments in the Black Sea is on closer inspection quite irregular because of this effect (see Fig. 5.A in Müller & Stoffers, 1974).

(iii) The Kimmeridge Clay cycles

(a) Discussion of Gallois' 'bloom' hypothesis (q.v. Gallois, 1976)

Gallois (1976) has summarised his model for the origin of the organic-rich sediments in the Kimmeridge Clay in the following way. "By analogy with the origin of the coccolith-rich beds, with which they are so intimately associated, the Kimmeridge Clay oil shales appear to have resulted from algal (probably dinoflagellate) blooms. The combined faunal and sedimentary evidence suggests that the organic content of the shales is largely derived from blooms, and that the blooms themselves, by de-oxygenating the water, provided the temporary anaerobic bottom conditions required to ensure the preservation of the organic content" (Gallois, 1976, p.475). Gallois rejected the 'stagnant bottom conditions hypothesis' on several grounds: "first there is evidence of a plentiful benthonic fauna in the oil shales, both in the form of burrow-fill structures and bivalve and gastropod shells. Second the oil shales are thin seams (generally <0.1m thick) which occur over a wide area (>200,000 km^2), apparently independent of distance from land and the overall shape of the sea floor. Thirdly the oil shales occur in juxtaposition (commonly inter-burrowed) with a variety of lithologies which must represent a variety of differing bottom conditions ... The stagnant bottom conditions hypothesis could explain very local occurrences of oil shale within a single basin of deposition containing variable bottom conditions, or could explain widespread occurrences within a basin containing uniform conditions. It does not, however, explain the widespread occurrences in the Kimmeridge Clay which embrace several depositional basins containing variable conditions (Gallois, 1974, p.474). Gallois (loc. cit. p.473) considered that the "Kimmeridge Clay oil shales ... were formed by algal blooms in a environment between open ocean and enclosed marine basin".

Although dinoflagellate blooms may well have been an important source of organic matter for the sediments in the Kimmeridge Clay

Formation they cannot be regarded as the key factor in the causation of kerogenous shale deposition. Organic matter supply is only one of several important controls, and its importance is usually exceeded by preservational factors (see Chapter Two). In general, persistent bottom de-oxygenation is required to preserve significant amounts of marine organic matter in the sediment; it is unlikely that the 'temporary' episodes described by Gallois would achieve this effect (q.v. Falkowski et al. 1980). Gallois' model does not make it clear whether the decay of the dinoflagellate bloom itself,or the decay of benthos poisoned by toxic dinoflagellates, or a combination of the two, is responsible for the envisaged temporary de-oxygenation. It should be noted that comparatively few dinoflagellates actually lead to benthic mass mortalities through the production of toxins (P.S.P. etc., see Chapter Two), and that most large dinoflagellate blooms occur over rather 'small' areas (maximum affected area 14,000 square miles?, Steidinger, 1973) and are notoriously patchy.

Gallois (loc. cit. p.476) states that "oil shale fauna, although limited in variety, is fully marine and clearly does not represent an anaerobic bottom environment comparable to that of any modern closed or restricted basin". This statement is remarkable since the fauna of the kerogenous intervals has a composition, character and distribution typical of an oxygen stressed environment. Accepting the different usages of the term 'oil shale', Gallois' comments on the fauna are still highly misleading. The eurytopic aspect of the faunas indicates that the environment was stressful before any 'mass mortality' occurred and the situation was not one of a normal fauna suddenly succumbing to a toxic dinoflagellate bloom.

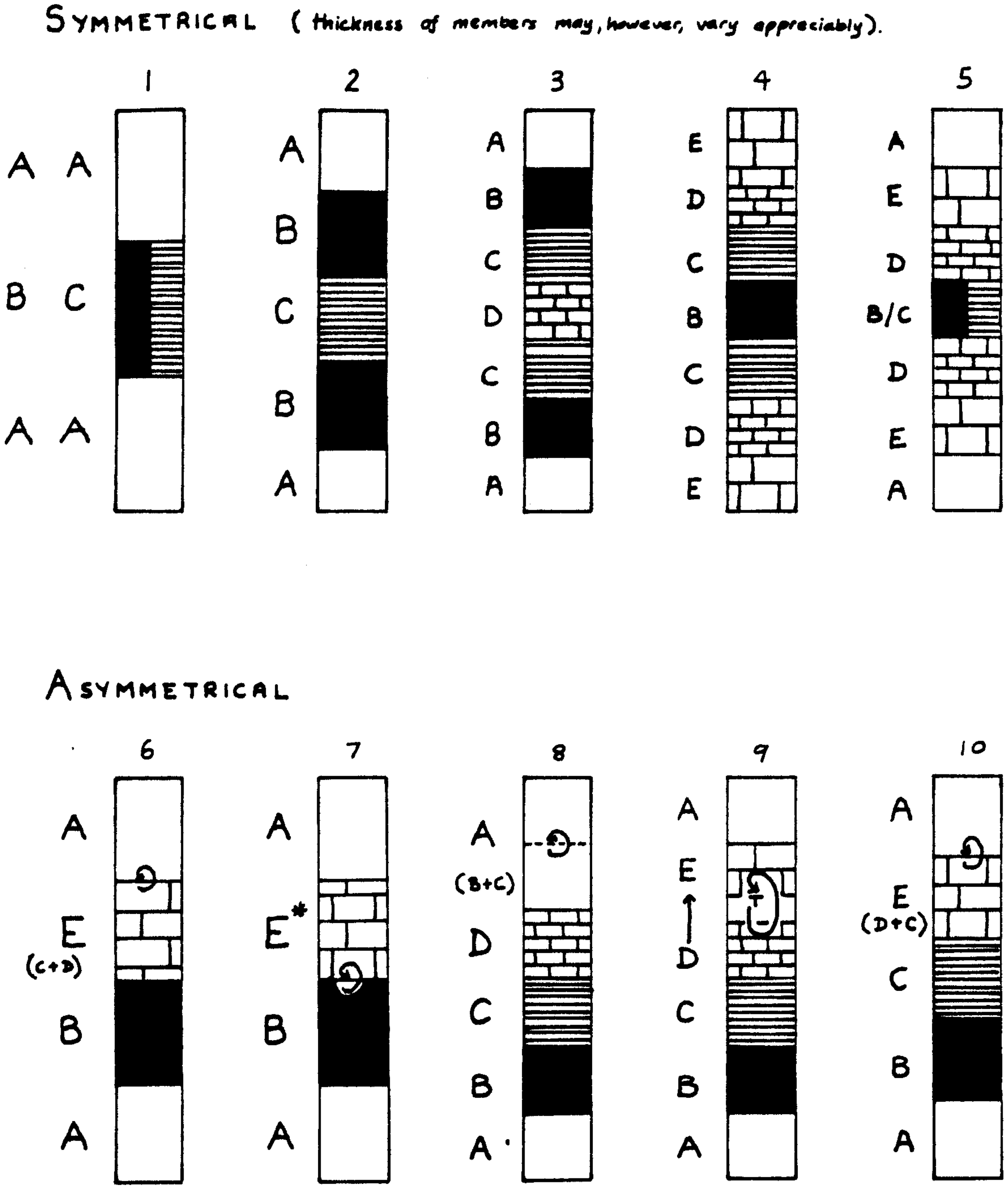
The uniformity and areal extent of the Kimmeridge Clay kerogenous facies implies a 'pelagic' mechanism. Although apparently appreciating this point, Gallois over-emphasised production rather than preservation as the source of uniformity. I believe that the general uniformity of the Kimmeridge Clay indicates a widespread homogeneous bottom water environment developed under a regional pycnocline (as in the modern analogues). The fact that the Kimmeridge Clay maintains its gross uniformity over large areas despite variations in thickness and hence accumulation rate implies a mechanism which influences both the preservation of organic matter and the general pattern of sedimentation. If this were not the case differential dilution and greater lateral heterogeneity would be anticipated. Watermass stratification provides such a mechanism while synchronous algal blooms would not (see Chapter Two). The lateral continuity of the coccolith limestone bands does, however, suggest widespread 'synchronous'plankton blooms. What are the nutrient sources for such a bloom? Gallois (1976) only mentions upwelling and land-derived nutrients; both of these are directional and/ or areally confined and would not be expected to produce lateral uniformity. On the contrary, seasonal lowering of the pycnocline in a stratified watermass could produce a regionally uniform supply of nutrients; such a mechanism has been proven on a small scale in lakes (see Stauffer & Lee, 1974), and is a major feature of the oceans.

(b) Present interpretation

The modern-Quaternary analogues described earlier clearly establish the validity of argillaceous, kerogenous-coccolithic sediment sequences as a product of changing watermass conditions in stratified basins. The perfect sequence of clay-sapropel-coccolithic sapropel is, however, rarely reproduced in ancient sediments (at least on a comparable scale) but similar cycles are known (see Fig. 5.6). These variants on the theme are produced by several factors:-

(1) Palaeogeographic position within the basin of deposition and the differing geological characteristics of different basins (particularly tectonic stability, subsidence rates etc.).

(2) Varying rates of basin stratification and overturn, determining the symmetry of the cycles.



A bioturbated clay B 'black shale' C 'coccolithic black shale'

D coccolith rhythmite E bioturbated coccolith-rich limestone (often cemented)

horizon of turn-over event in asymmetric sequences.

( ) denotes units assimilated into aerobic facies by bioturbation.

E* calcareous deposit resulting from turn-over productivity spike.

N.B. E may replace A as background sediment as in 4.
Contacts may be sharp or gradational.

Fig. 5.6 Examples of some varieties of distal shelf and pelagic black shale - coccolith limestone rhythms. Varieties 1,2,4,5,6 and 10 recognised in the Swalland Member.

(3) Differing water depths between basins.

(4) The stability of the watermass stratification and the scale of the event producing the subsequent overturn.

(5) The nature of the normal indigenous background sedimentation.

Weissert et al. (1979) have described cycles consisting of a black shale overlain by coccolithic rhythmites that become progressively more bioturbated upwards. It is possible that this might indicate a progressive dispersal of the bottom water by a 'slow' lowering of the pycnocline. Nutrients released from the top of the bottom water zone would promote coccolith blooms whose products were sedimented into the still anaerobic, basal bottom water environment. As the circulation improved coccolithic rhythmites would then give way to bioturbated coccolithic marls. In oceanic basins the sea floor will always lie in the bottom water zone and overturn will indicate a weakening of the stratification (though not necessarily over the whole basin) and the renewal of thermohaline circulation, whereas in shallower basins lowering of the pycnocline to the bottom may be indicated. In the Black Sea and eastern Mediterranean cycles the sapropel unit appears to record the interval between the O_2:H_2S interface rising above the sea floor and its subsequent arrival at a shallow position where it could influence plankton dynamics. Since these basins are of oceanic depths this progressive ascent of the O_2:H_2S interface takes some time (see Deuser, 1974) and sapropel layers of significant thickness can accumulate. In shallow basins this time interval would not be available or would be very brief and so true sapropels (undiluted by coccoliths or diatoms) would be expected to be rare or very thin. Significant thicknesses of sapropel type sediment in shallow basins are more likely to represent repeated transient anoxic episodes within generally anaerobic or dysaerobic settings (in which the O_2:H_2S interface may only rise a few centimetres above the sediment) and are genetically distinct from the single (?) long-term Black Sea type cycles.

As might be deduced from the preceeding discussion, no ideal Black Sea type cycles are present in the Swalland Member of the Kimmeridge Clay Formation. With the exception of the thin unit in the White Stone Band (and ignoring the possibility that the Short Joint Coal rhomb limestone was originally coccolithic) the oil shales, which are the best contenders for sapropel equivalents, are separated from the coccolith rhythmites. The "Best Blackstone" and possibly also the White Stone Band oil shale, are the only probable sapropel type sediments, although anaerobic (rather than anoxic) conditions often occurred during the deposition of the Swalland Member.

The general character of the Swalland Member (lithological, palaeo-ecological and geochemical) suggests a cyclic alternation between lower mixed layer (seasonally stratified) and bottom water (anaerobic, 'permanently' stratified) environments. What factors resulted in the apparent periodic establishment of 'permanent' stratification, i.e. meromixis? There is no firm evidence for the episodic existence of a low-salinity surface layer during the deposition of the type Kimmeridge Clay Formation, although the nekton and plankton were stressed during the 'meromictic' episodes represented by the oil shales, coccolithic kerogenous shales and coccolith rhythmites. Temperature appears to have been the main agent of watermass stratification for most areas in the Late Jurassic, the main variables being climate, water depth and palaeo-geography (see Chapter Two).

What evidence is there to suggest that this thermal stratification periodically became so stable as to result in meromixis and anaerobic to to anoxic bottom waters? For the kerogenous lithologies (kerogenous shales, oil shales and coccolithic kerogenous laminites) the abundance and character of the preserved organic matter indicates that conditions at the sediment water face were largely in the range anoxic to dysaerobic; even relatively brief aerobic episodes would have resulted in the large scale destruction of this organic matter (see Chapter Two). The laminated, unbioturbated nature of these sediments also implies anoxic to anaerobic conditions (with the R.P.D. very near, if not above, the sediment water interface), while the occasional presence of impoverished faunas indicates that dysaerobic conditions were sometimes established. The presence or absence of laminations and macrofauna is, however, a somewhat ambiguous criterion of meromixis in that it may fail to record transient aerobic episodes that were too brief to allow benthic recolonisation (see Chapter Two). Palynofacies data, in particular the presence and abundance of dinocysts, is another criterion for estimating the stability of the watermass stratification (see Chapter Three). The paucity of dinocysts in the oil shales, coccolithic kerogenous laminites and coccolith rhythmites (see Table 5.5) suggests stable (meromictic) stratification without seasonal bottom mixing. Where stratification was probably seasonal (i.e. in the clays) dinocyst densities are high. It is interesting to note that the Freshwater Steps Stone Band which has common but limited bioturbation and rare oysters, suggesting periodic aerobic conditions, has far more dinocysts than the coccolith rhythmite unit in the White Stone Band. Dinocysts may therefore be quite a sensitive indicator of hydrographic stability.

Although the watermass stratification may have sometimes become meromictic (viz. not seasonally disrupted) this does not imply that there was no mixing below the pycnocline at these times. However, the evidence does suggest that whatever level of mixing was present, it was insufficient to totally replace the oxygen utilised by organic decay. Variations in mixing rate and B.O.D. will have resulted in conditions varying from anoxic to dysaerobic at the sediment-water interface and have determined the relative vertical extents of anoxic, anaerobic, dysaerobic and aerobic conditions within the bottom water. The O_2:H_2S interface will have fluctuated in position from just below the sediment surface to a few centimetres above it and may, though not necessarily, have even reached the pycnocline. Meromixis in a shallow setting like the N.W. European late Jurassic epeiric sea will have been a much more dynamic and generally variable state than that in the deep Black Sea and eastern Mediterranean basins. One might also imagine that thermal (endogenic) stratification may have tended to be somewhat less stable than salinity (ectogenic) stratification, despite the more equable Jurassic climate, and that this would have contributed to the more dynamic, delicately balanced conditions inferred for the Swalland Member.

Given that the style of meromixis in the Black Sea and Kimmeridgian cycles was somewhat different, how does the genetic relationship between oil shale and coccolith limestone compare with that for sapropel and coccolithic sapropel? In both cases the coccolithic units appear to indicate better nutrient supply to the euphotic part of the mixed layer. In the shallow Kimmeridge Clay sea primary productivity was probably nutrient-limited during the meromictic phases, with coccolith blooms only developing during seasonal lowering of the pycnocline (resulting in entrainment of nutrient-rich waters). If such seasonal fluctuations in the pycnocline level were common place why are the oil shales not coccolithic too? It may be that the coccolith blooms only developed when, prior to mixing, the nutricline occurred just below the pycnocline, or only during larger mixing events (when the pycnocline was lowered further than usual), or only when a sufficiently large nutrient pool had accumulated in the bottom water. The relative positions of the nutricline and pycnocline may have played a crucial role like that in the Black Sea cycles, although the nutricline may not have correlated with the O_2:H_2S interface as in the latter case. During the normal mudstone sedimentation coccoliths probably alternated with dinoflagellates as the predominant plankton in accordance with the seasonal stratification cycle. The palaeo-environmental interpretation of each of the major lithologies in the Swalland Member is summarised in Fig. 5.7.

LITHOLOGY	ENVIRONMENT (WATERMASS)	CONDITIONS AT SEDIMENT-WATER INTERFACE	FAUNA	COMMENTS
COCCOLITH RHYTHMITE	Monimolimnion. Predominantly stable meromixis.	Anoxic-anaerobic » dysaerobic > aerobic	Ammonites absent or very rare. Usually no preserved benthic body fossils (see comments). Low 'diversity' coccolith assemblage.	Coccolith blooms stimulated by partial mixing of upper part of bottomwater (?). Occasional mixing to bottom permits brief benthic recolonisation (eg. Freshwater Steps S.B.). Productivity conceivably very high but episodic/periodic. Extremely low clastic dilution.
COCCOLITHIC KEROGENOUS LAMINITE	Monimolimnion Stable meromixis	Anoxic-anaerobic » dysaerobic	No ammonites observed. Benthos absent except derived (?) Discina.	Like kerogenous shale but more carbonate-rich due to conspicuous amounts of coccolithic faecal pellets; intermediate to rhythmite. Low to very low clastic dilution. Appear to 'replace' kerogenous shales up-sequence (possible water depth effect?)
OIL SHALE	Monimolimnion Stable meromixis.	Anoxic (O_2:H_2S interface probably close to bottom).	Ammonites not observed. Rare, small benthos probably derived; dominated by 'Lucina' and Discina.	Productivity was probably limited by nutrient availability. Conversely eutrophication may have occurred if nutrient-rich bottom water extended up into euphotic zone. Contains some allochthonous o.m. (derived marine macrophyte debris?). Very low clastic dilution. Probably a true sapropel.
KEROGENOUS SHALE	Lower mixed layer to monimolimnion. (short term meromictic episodes).	Alternating (?) dysaerobic and anaerobic/anoxic	Ammonites common to rare frequently fragmented. Small, low diversity benthos (as above) commonly as 'plasters'. Derived and in situ?	Shelly horizons due to redeposition and/or recolonisation events interrupting predominantly anaerobic-anoxic conditions. Transitional between a sapropel and gyttja.
MUDSTONE-SHALE	Lower mixed-layer. Seasonal stratification	Aerobic ≥ dysaerobic » anaerobic-anoxic.	Ammonites common (variable) Low diversity benthos (see Table 5.2.)	Distal shelf sediment often very rich in organic matter. Productivity was probably moderate to good. Probably includes kerogenous shales assimilated during bioturbational mixing. Probably higher sedimentation rate than other lithologies. Gyttja.

Fig. 5.7 Palaeoenvironmental summary for principal lithologies of Swalland Member

ADDENDUM : DISCUSSION OF AIGNER (1980)

Aigner has recently published a description of some of the mudstone and 'bituminous shale' (kerogenous shale) units from the Kimmeridge Member (Lower Kimmeridgian). This extra information is a welcome addition to our knowledge of this part of the section and in general confirms my own observations on the equivalent lithologies in the Swalland Member. Several points are worthy of particular comment:

(a) Aigner reports that articulated bivalves are rare in the sequence that he studied. This contradicts the view of Downie (1955) reproduced in Tyson et al. (1979). I have not personally made any rigorous observations on this matter but would be more inclined to believe Aigner's more recent work with the proviso that a significant amount of variability might be anticipated between the Swalland and Kimmeridge Members.

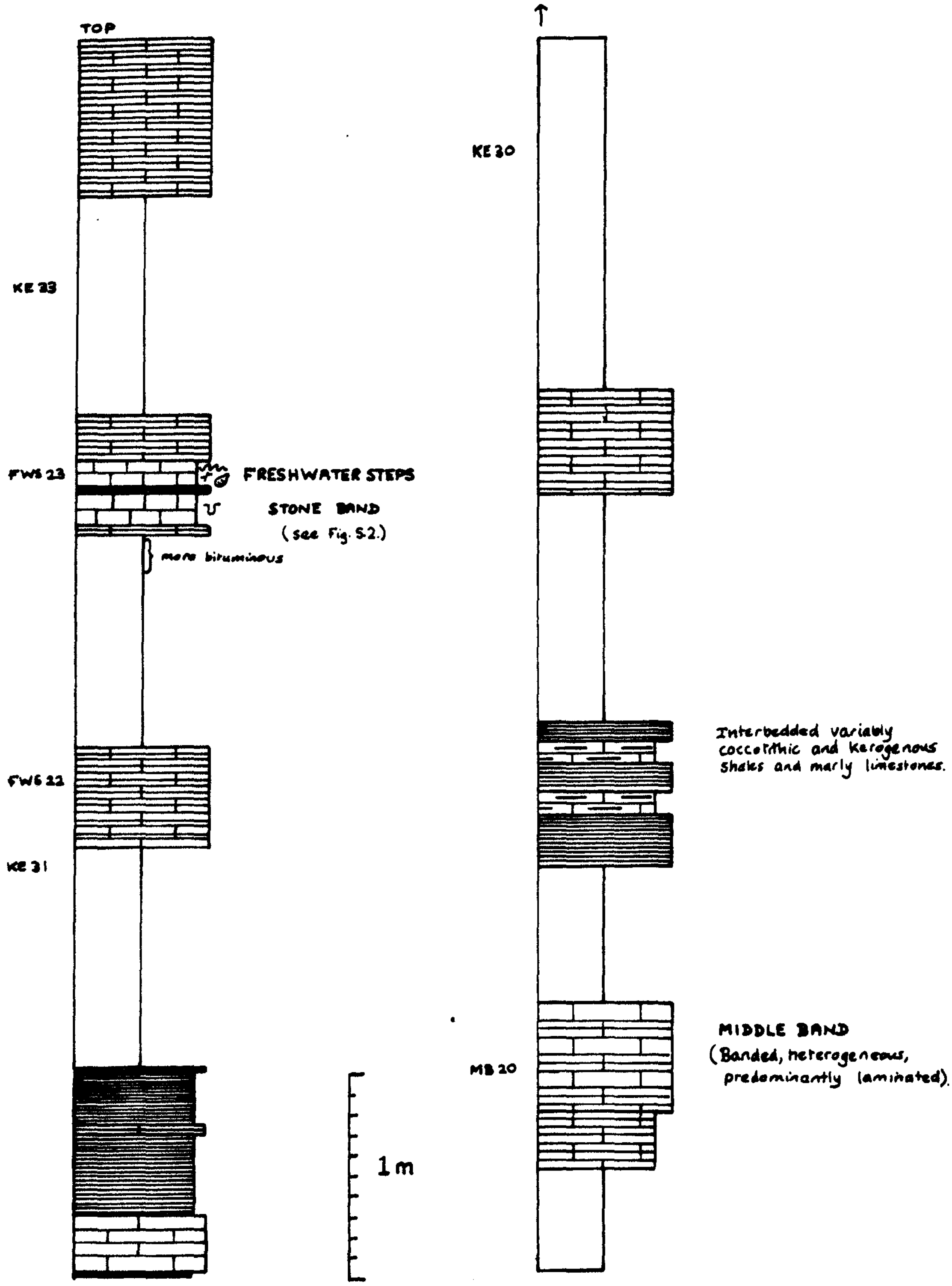
(b) Aigner found that, with the exception of Ostrea and Exogyra (present as an encrusting epifauna on the upper surfaces of ammonites), the kerogenous shales lacked an autochthonous infauna. However, infaunal forms were present as shell pavements or 'plasters' associated with levels of minor, low-relief scour yielding evidence of preferred (current) orientation. Aigner considers these horizons to represent 'hydrodynamic events' that led to temporarily improved bottom oxygenation and introduced shells and coccolith material. However, he does admit that "it cannot be excluded that occasionally even infauna was able to colonise the sea floor during short intervals" (p.334). From my own observations on the kerogenous shales in the Swalland Member I have interpreted these sediments to indicate fluctuating dysaerobic-aerobic conditions where the R.P.D. was normally very close to the sediment surface and periodically rose slightly above it (seasonal deoxygenation and short-term meromixis). It seems quite acceptable that the change in circulation regime between the various oxygenation states might be associated with bottom currents capable of producing slight scours and winnowing. The very shallow position of the R.P.D. in the sediment was considered the prime cause for the absence of infaunal organisms during much of the time when kerogenous shales were being deposited (see earlier); the presence of epifauna on raised surfaces is not incompatible with this view.

(c) Aigner's comment that "Tyson et al. (1979) considered that anaerobic conditions developed in a stratified water column and that during bituminous shale sedimentation conditions were similar to those of the present day Black Sea" (p.325) is a misleading misquote. Only the oil shales and coccolith rhythmites were associated with a depositional environment that showed a fundamental similarity with that of the Black Sea (c.f. Tyson et al., 1979).

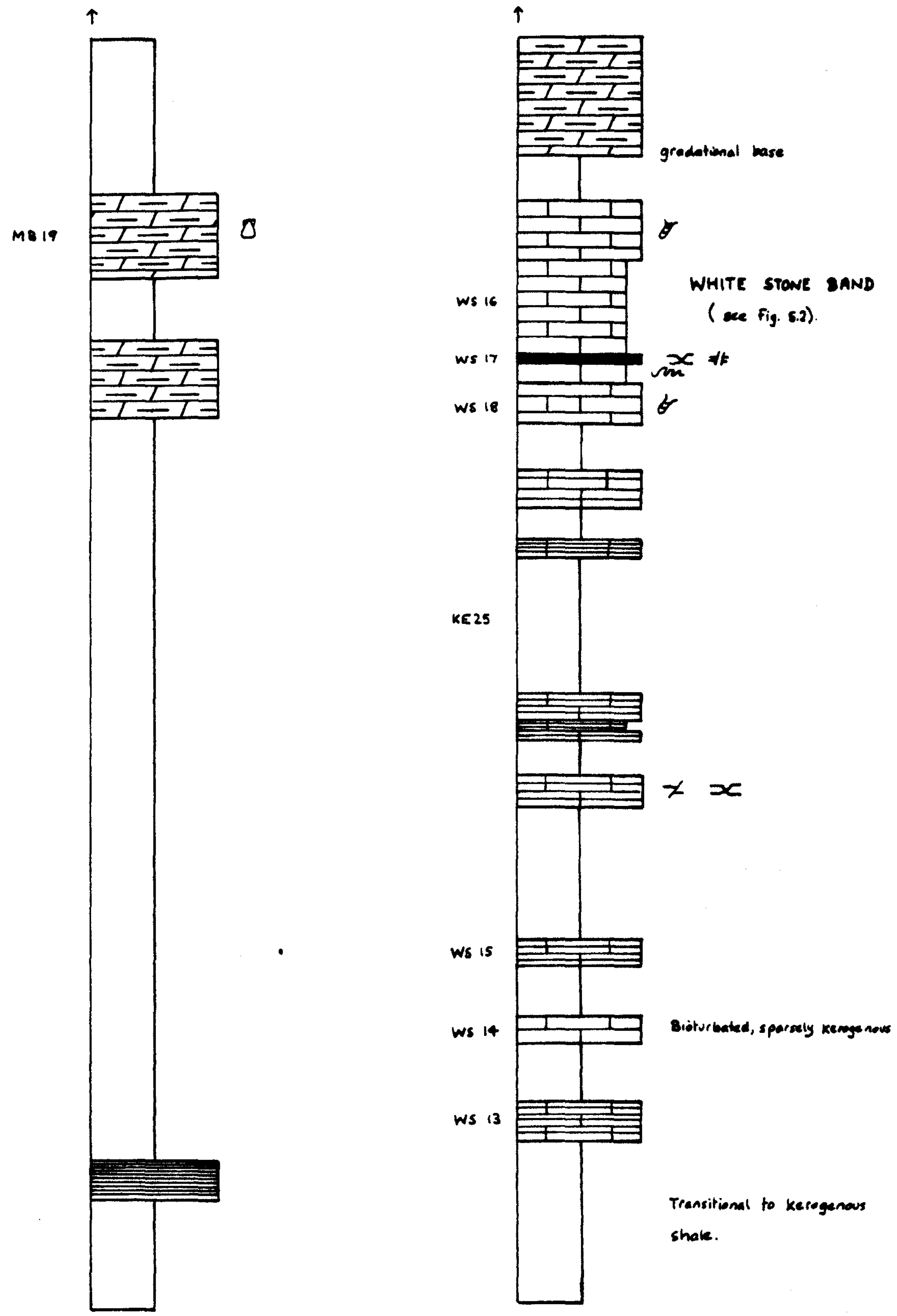
(d) Another comment to the effect that the "lateral passage between mudstones and bituminous shales puts a regional constraint on any interpretation of the cyclicity and probably questions a largely anoxic water column as proposed by Tyson et al. (1979)" (p.335) is based on the same misconception. In any case, the cited palaeontological similarities between the mudstones and kerogenous shales do not conclusively prove local lateral transitions occurred between their two respective palaeoenvironments, and transitions over larger areas may be attributed to tectonic influences on basin palaeobathymetry (e.g. over the Portland Swell, q.v. Townson, 1971). In addition, Tyson et al. (1979) stressed the fact that although the term $O_2:H_2S$ interface was used in the discussion, this did not necessarily imply persistant total bottom water anoxia.

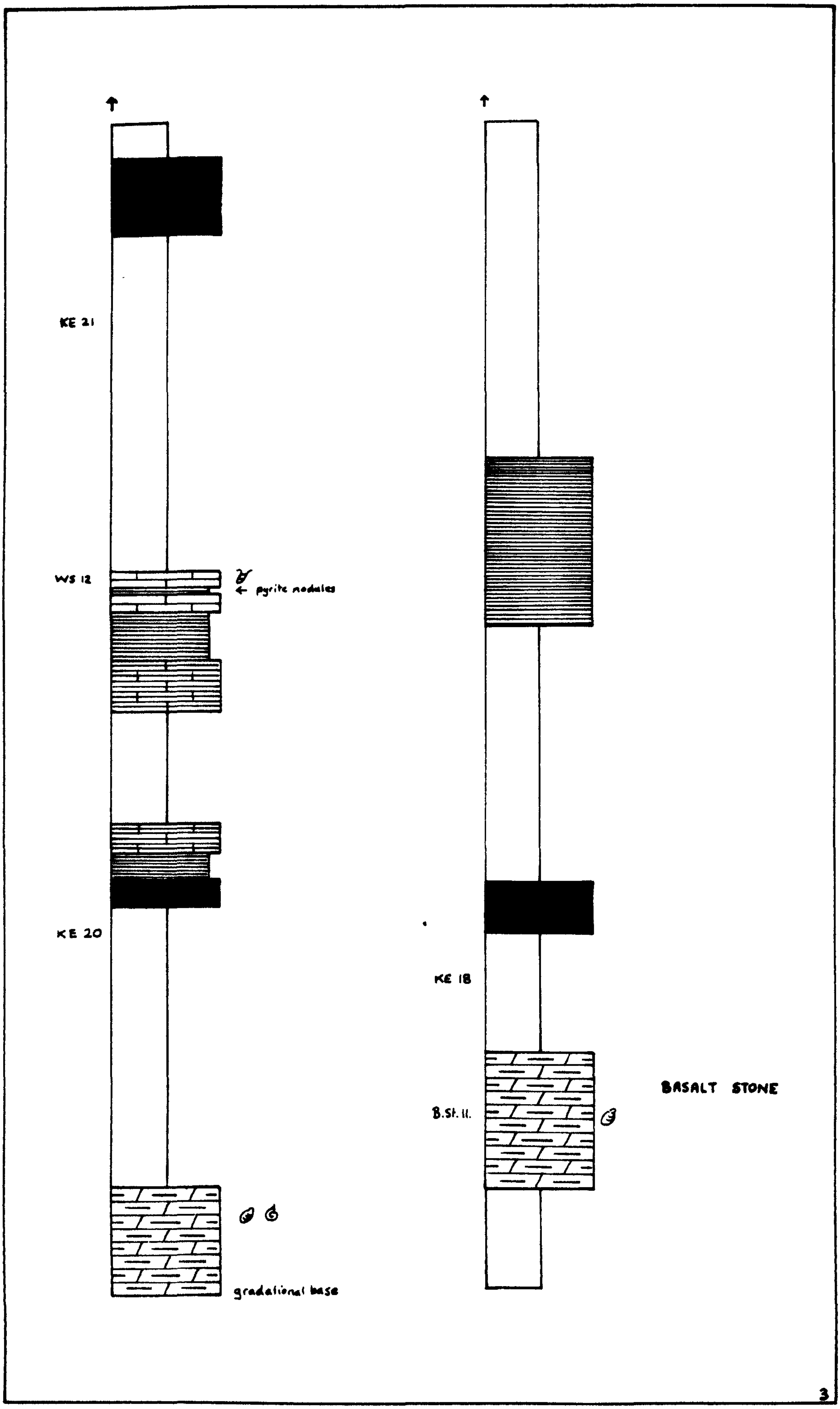
(e) Some of Aigner's arguments are based on the acceptance of the erroneous suggestion of Ioannides et al. (1976) that the bulk of the organic matter in the Kimmeridge Clay is of terrestrial derivation (see earlier comments).

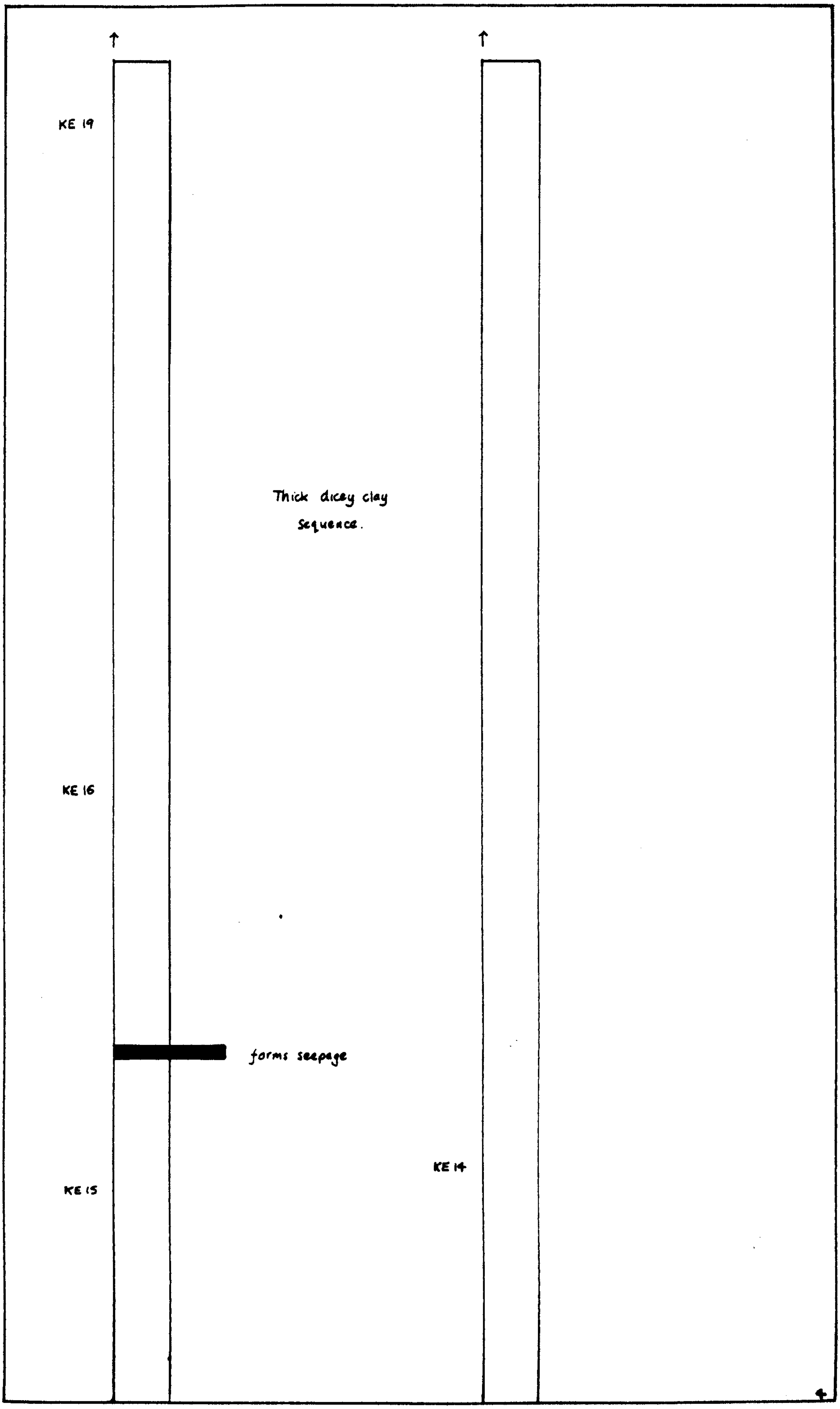
APPENDIX 5.1. LITHOLOGICAL LOG OF THE SWALLAND MEMBER

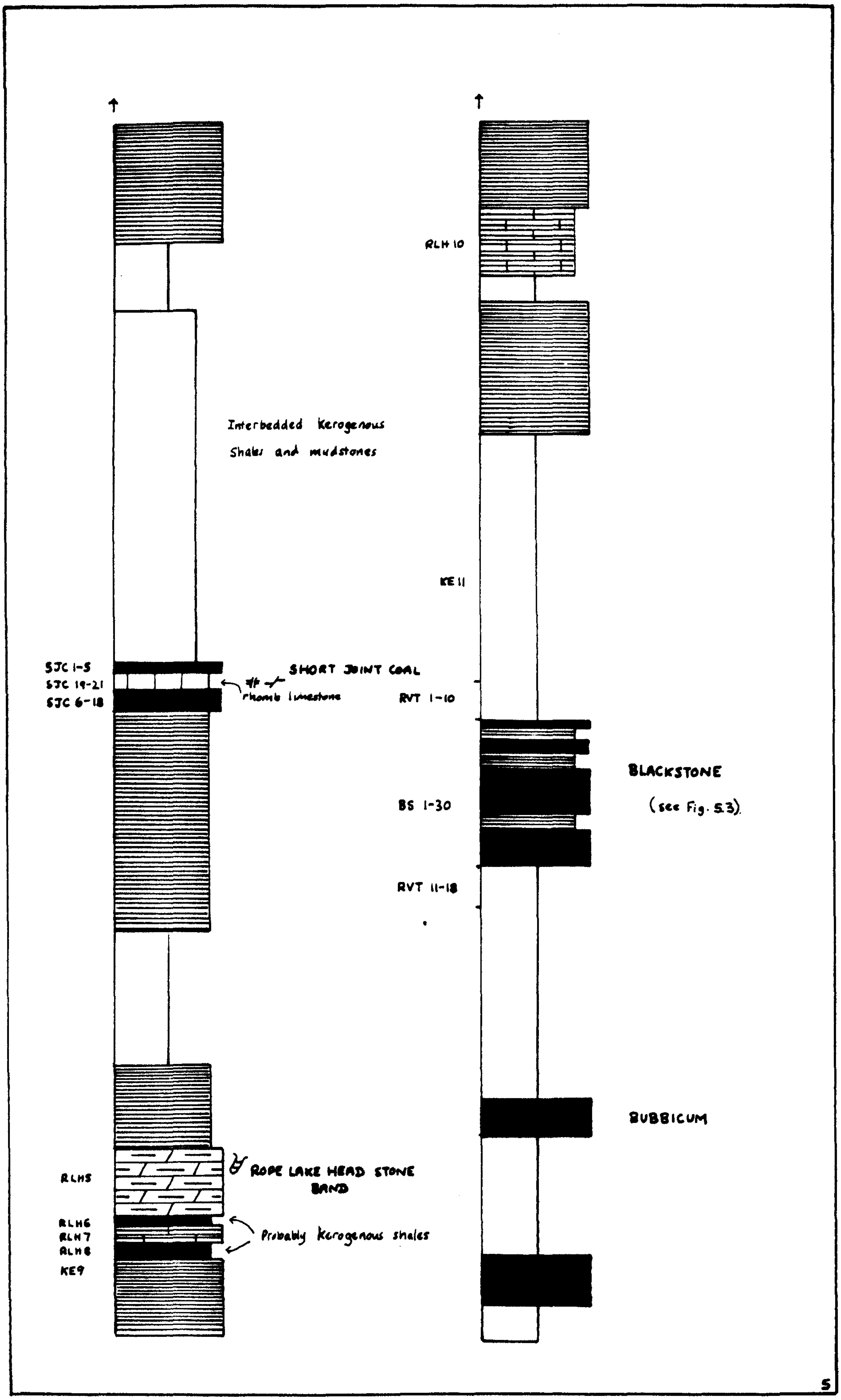


N.B. LITHOLOGICAL SYMBOLS AS IN FIG. 5.7. EXCEPT [symbol] WHICH IS NON-KEROGENOUS DOLOSTONE

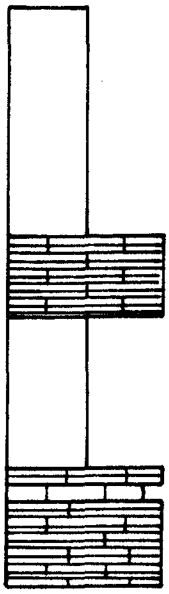

MB 19
gradational base
WHITE STONE BAND
(see Fig. 5.2).
WS 16
WS 17
WS 18
KE25
WS 15
WS 14
Bioturbated, sparsely kerogenous
WS 13
Transitional to kerogenous shale.


KE 21
WS 12
pyrite nodules
KE 20
gradational base
KE 18
B.St. 11.
BASALT STONE


KE 19
Thick dicey clay
sequence.
KE 16
forms seepage
KE 15
KE 14



RLH 10
Interbedded Kerogenous
Shales and mudstones
KE 11
SJC 1-5
SJC 19-21
SJC 6-18
SHORT JOINT COAL
rhomb limestone
RVT 1-10
BLACKSTONE
(see Fig. 5.3)
BS 1-30
RVT 11-18
BUBBICUM
RLH5
ROPE LAKE HEAD STONE
BAND
RLH6
RLH7
RLH8
Probably Kerogenous shales
KE9
5



BASE OF THE SWALLAND MEMBER

CHAPTER SIX

Sedimentological and palynological analysis of the Kimmeridgian sediments exposed between Kintradwell and West Garty, Sutherland, N.E. Scotland

INTRODUCTION

This chapter is based on the results of eight weeks field work which was spent logging, studying and sampling the Kimmeridgian sediments exposed between the northern end of Brora links and the beach below West Garty, Sutherland (see Fig. 6.1). Some 360m of section were logged (although considerably more was examined in outcrop) and 170 petrographic and 231 palynological slides were subsequently described. Robertson Research also processed 74 of my samples for organic carbon determinations and 29 of these for Rock. Eval. pyrolysis. These observations form a considerable data base for comparison with the other sections studied during this research project.

The Kintradwell-West Garty section forms part of the Brora outlier, a down thrown fault block of Mesozoic sediments which includes the only Upper Jurassic onshore outcrop in the U.K. north of Yorkshire which is in the general vicinity of the northern North Sea. The Kimmeridgian-Volgian section is a thick sequence of submarine fault scarp-fan associated sediments consisting of turbiditic and channelised sandstones and debris flow ('boulder bed') deposits interbedded with dark, grey to black, organic-rich shales. The section was studied because of its obvious contrasts and, in part, similarities with the other sequences examined in this thesis, and to extend the databank on the palynofacies characteristics of Upper Jurassic shale environments. The sediments described in detail in this chapter are all Lower Kimmeridgian (sensu anglico) and are therefore in part time equivalent with the Black Head and Kimmeridge Members of the type Kimmeridge Clay Formation of Dorset (see Chapter Five). The upper, Volgian part of the section between Portgower and Ord Point was only examined in reconnaissance and was neither logged nor sampled.

Several factors conspire to make a geological account of the Kintradwell-West Garty section a formidable task. Like any other field area, the nature of the outcrop and exposure poses the first major

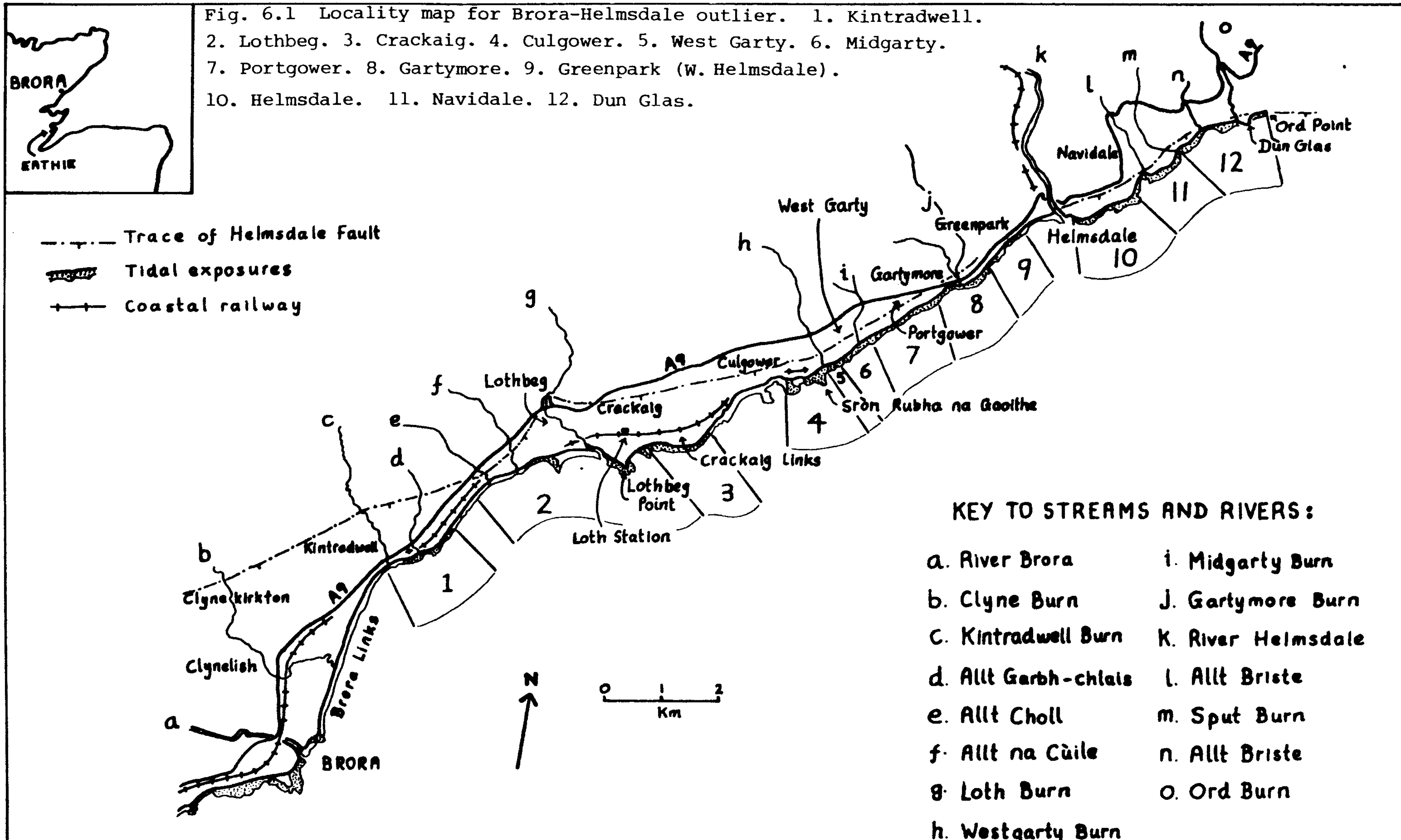


Fig. 6.1 Locality map for Brora-Helmsdale outlier. 1. Kintradwell. 2. Lothbeg. 3. Crackaig. 4. Culgower. 5. West Garty. 6. Midgarty. 7. Portgower. 8. Gartymore. 9. Greenpark (W. Helmsdale). 10. Helmsdale. 11. Navidale. 12. Dun Glas.

limitation on any subsequent understanding of the geology. Unfortunately, the bulk of the Mesozoic outlier is covered by raised beach and drift deposits and the outcrop is essentially limited to discontinuous tidal exposures and a few small, scattered inland outcrops. This is hardly an ideal situation for attempting to reconstruct three-dimensional lithofacies relationships.

In addition, the rather poor biostratigraphic resolution in this area, the gaps in exposure,and the original complexity of the facies organisation combine to introduce significant degrees of uncertainty into any geological interpretations. The policy followed here has been to log the most representative sections from each of the main coastal exposures and to describe (and log where appropriate) every inland exposure encountered. Unfortunately, the facies complexity of these sediments also defies any truly meaningful lithostratigraphic treatment. Despite the long-standing usage of such names as the 'Loth River Shales' and 'Allt na Cùile sandstone', only lithofacies units are recognised in this account.

PREVIOUS WORKS

Only a very brief review of the most relevant previous works is given here. A more complete, historically oriented account can be found in Linsley (1972). The first noteable paper on the Brora-Helmsdale sequence is that by Macgregor (1916), the author of the now famous 'fallen stack' and shoreline hypothesis for the origin of the boulder beds. This was followed by the more descriptive treatment in the geological survey memoir by Lee (1925) and the brief Geologist's Association excursion report by Macgregor et al. (1930). In 1932 the classic paper by Bailey and Weir was published in which the submarine fault scarp origin of the boulder beds was recognised for the first time. Their excellent account, which is still the most informative, single published work on the section, was not followed by a more recent study until that of Crowell (1960) who discussed the provenance and mode of emplacement of the boulder beds. Apart from brief comments on the 'Loth River Shale' section in a paper by Arkell and Callomon (1963) no further information on this highly interesting sequence appeared prior to the thesis by Linsley (1972).

The main contributions of Linsley's thesis include a provisional ammoníte zonation of the sequence and the production of a series of detailed, plane-tabled, topographic maps of the foreshore exposures. Although general geological descriptions of the various outcrops were offered, only synthetic sections were presented (without any actual logged vertical sections) and with only very sparse petrographic details, and often only very brief and general sedimentological descriptions. Brookfield's (1973) thesis on the Oxfordian-Lower Kimmeridgian of Dorset contains a brief appendix on the Brora sequence (p.526-531) which was based on his B.Sc. thesis, and he followed this in 1976 with a short paper on the age of the 'Allt na Cùile Sandstone' and 'Loth River Shale'. Sykes' (1975) thesis on the Scottish Callovian-Oxfordian also contains a few comments on the Lower Kimmeridgian ('Loth River Shale' and 'Allt na Cùile Sandstone') including a brief comparison with the Hareelv Formation in Eastern Greenland. Neves and Selley (1975) have provided a recent summary of the whole Jurassic succession in the Brora outlier which includes some updated interpretations of the fault scarp-fan sediment sequences in the Kimmeridgian-Volgian (based primarily on Linsley, 1972). The only other significant works are those by Lam and Porter (1977) who describe some of the major palynological features of the whole Jurassic sequence in the Borora outlier, and Riley (1980) who has used dinocysts to demonstrate that the Upper Jurassic sediments range upwards into the basal Portlandian (middle part of the Middle Volgian).

LITHOLOGICAL SECTIONS

The measured sections described in the following account are an attempt to record the most representative facies developments in each of the main areas of outcrop. Since all the coarse-grained lithologies (sandstones and 'boulder beds') are laterally discontinuous and laterally variable units, the measured sections cannot be considered typical of the sequence, and as such are not a basis for lithological correlation. The positions at which the measured sections were taken were in part selected with reference to Linsley's maps of the foreshore exposures, and where appropriate the relevant parts of these maps are reproduced here. It should be noted that changes in beach cover in some parts of the outcrop make the usage of these maps rather difficult, particularly since they do not include any landmarks above the high water mark. In the following description of the lithological succession the various measured sections and described outcrops have been grouped into five main localities:-

A Kintradwell

B Lothbeg (includes Allt Choll, Allt na Cùile, Lothbeg, Loth River, Lothbeg Point and Loth Station)

C Crackaig Links

D Culgower (including Sròn Rubha na Gaoithe)

E West Garty

The positions of these localities and other important features and place names are shown on Fig. 6.1. In addition to these main areas two other small outcrops are briefly described, one at Clynekirkton (Clyne Burn) north of Brora, and the other near Sput Burn, east of Navidale; both of these are discussed in the description of the Lothbeg Area.

KINTRADWELL AREA

<u>Section A1 Kintradwell Burn (NC921070)</u> <u>Cymodoce</u> Zone?

As far as I am aware this section represents the most westerly outcrop ever recorded from the Kintradwell area. It occurs at the north east end of Brora Links, just south west of the mouth of

Kintradwell Burn and is located to the south west of Linsley's last map (XVII) and is therefore of uncertain age but is assumed to belong to the cymodoce zone along with the other beach exposures. Approximately 6.8m of section were measured (the maximum permitted by tides and beach cover) of which two thirds consisted of shale (see Fig. 6.2). A strike and dip of $248^{\circ}/16^{\circ}$N was recorded.

The two amalgamated, calcite cemented sandstone beds near the top of the section are worthy of note. The upper bed is fine grained and shows convolute lamination in its upper half and poorly defined ripples in its lower part. The underlying bed is graded from medium grained, moderately to poorly sorted and sparsely bioclastic at its base,to fine grained, moderately to well sorted and bioclast free in its upper part. The very top of the latter bed is partly convolute laminated and contains up to 10% wood and rare shale clasts ($\leq$ 20 cm in length). In both of these sandstones the calcite cement has clearly corroded the quartz grains and constitutes 40-50% of the rock, resulting in excellent 'poikolitic' (= poikoiloptic, poikiloblastic) lustre-mottled textures (see later). The base of the section shows a typical example of the rapid lateral changes in lithology and cementation which are occasionally conspicuous in the Kintradwell-West Garty succession.

Sandstones appear to be more common just to the south-west of section A1 (about 50% as opposed to 25% as in Fig. 6.2). These sandstones exhibit relatively common clasts of quartizitic sandstone of pebble to boulder grade, are matrix-supported and exhibit common bedding surfaces rich in bioclastic material (including pelecypods, brachiopods, ammonites and belemnites). The belemnites sometimes exhibit preferred orientations in an approximately N-S direction (q.v. Crowell, 1960). The general lithologies of the section are virtually identical with those indicated on Linsley's map XVII.

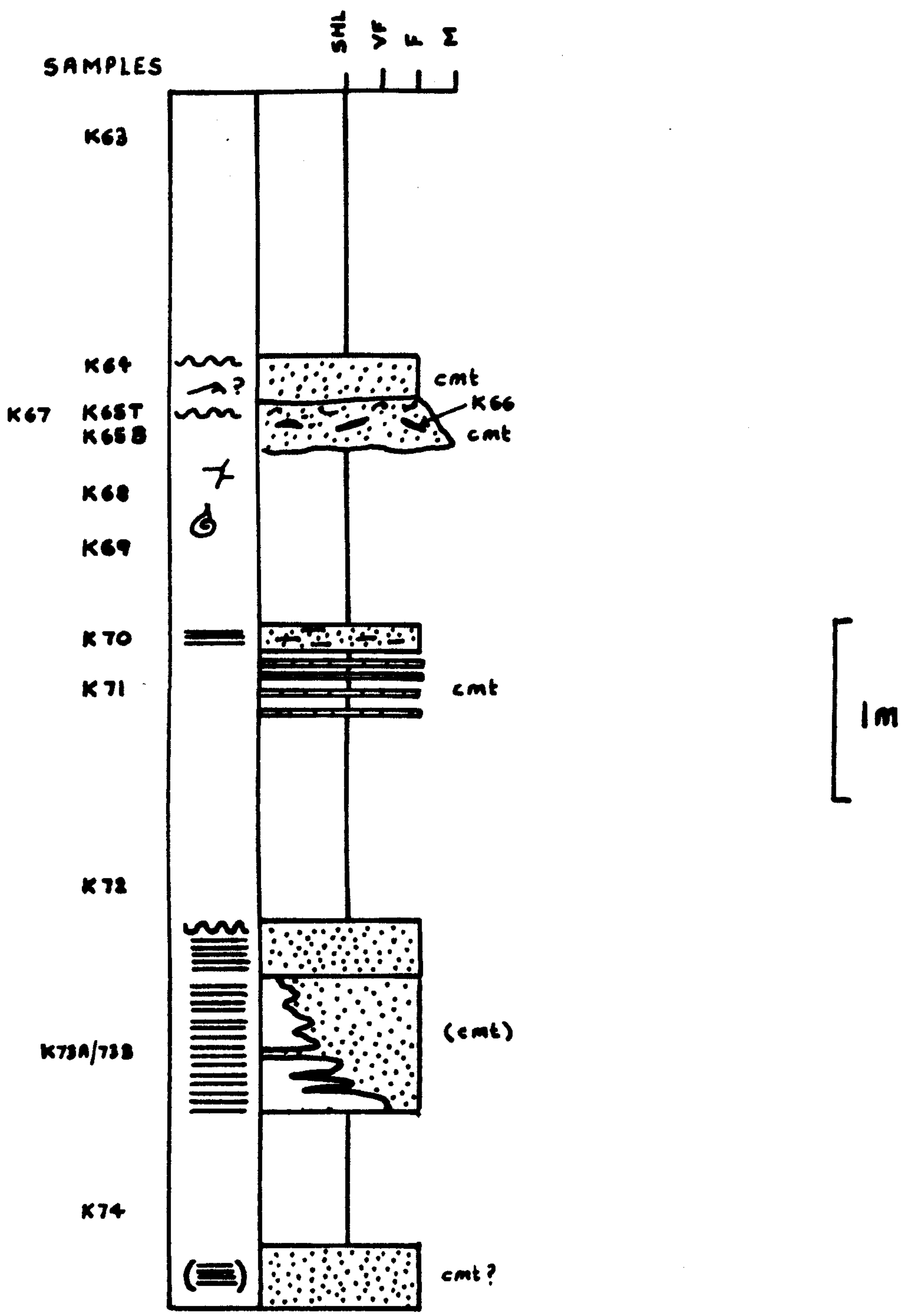


Fig. 6.2 Section A1 Kintradwell. For explanation to symbols see Fig. 1.

Section A2 Kintradwell beach west (NC 926074) *cymodoce* zone

This section is located south of the point where Allt Garbh-chlais cross the high water line on Kintradwell beach (see Fig. 6.3 from Linsley's map XVI). About 37m of sediment were logged, the greater part consisting of shale but with about 16% sandstone, the latter mainly at the base of the section near low water mark (see Fig. 6.4). The sequence is rather atypical of the west Kintradwell outcrop and represents a conscious attempt to select a line of section showing the nature of the background sedimentation at Kintradwell (in part, for the purposes of the palynofacies study). Irregular, coarse-grained, lenticular units and deformed, slumped horizons were avoided, and as such are under-represented in Fig. 6.4. Only specific features of this section are discussed below.

The lower sandstone-bearing 8.6m of section A2 continues laterally westwards as a series of discontinuous, but in part, overlapping lenticular sandstones and subordinate intercalated boulder bed lenses. The boulder beds are matrix supported and consist of a sand matrix containing subangular to subrounded, cobble to boulder grade clasts of bedded, grey, quartz sandstone. Some of the sandstones and/or boulder beds contain abundant rounded quartz pebbles ($\leq$ 3cm) and conspicuous coarse bioclastic debris (including belemnite rostra up to 6cm in diamter!) but such material is rare if the outcrop is considered as a whole. The pebbly facies also occurs in the overfolded slump block at the top of the section (see Figs. 6.4 and 6.4 and Plate 6.1). This slump consists of very poorly sorted, calcite cemented (lustre-mottled) sandstone which contains abundant, well rounded, polycrystalline metaquartzite pebbles ($\leq$ 2.5cm) and somewhat less abundant subrounded sandstone clasts ($\leq$ 16cm) in a fairly shelly sandstone matrix. The block is internally deformed and complexly overfolded toward the SSE.

The shales composing most of section A2 are grey to black, organic-rich and carbonaceous (see palynofacies section). They contain appreciable amounts of coarse silt and very fine sand ($\leq$ 50%) which is

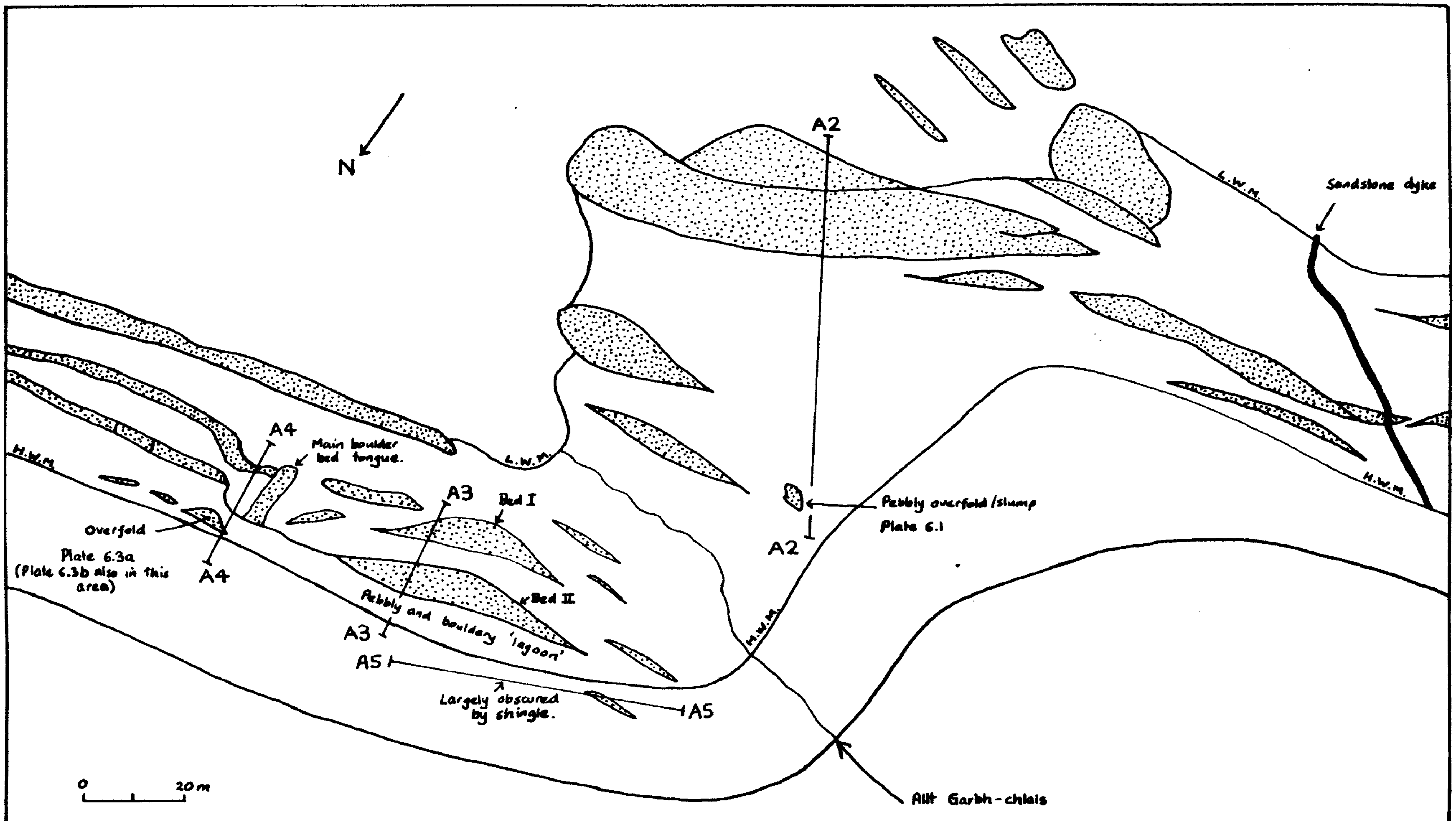


Fig. 6.3 Foreshore exposure at east Kintradwell (after Linsley, 1972).

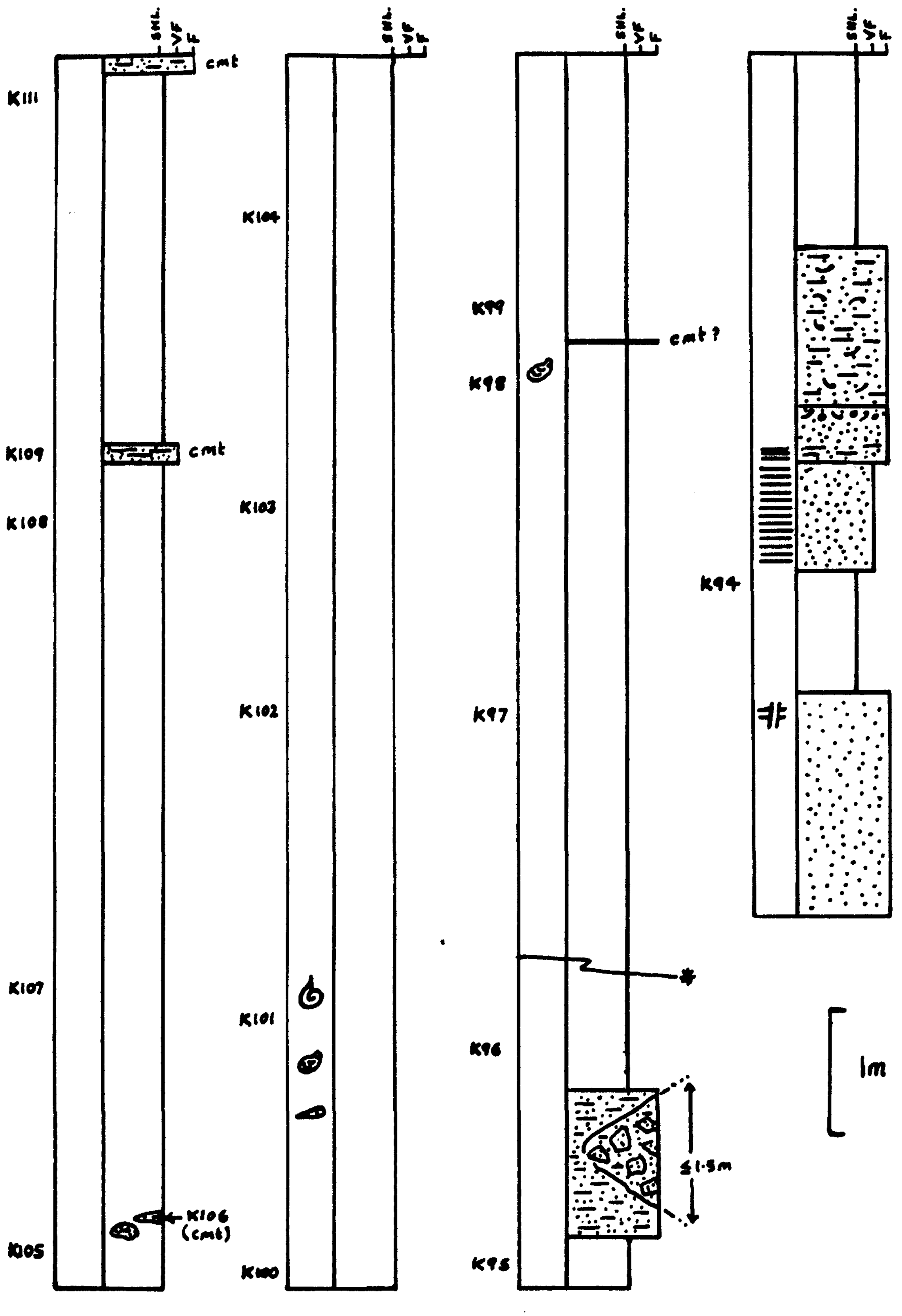


Fig. 6.4 Section A2 Kintradwell.
* denotes horizon of dislocation.

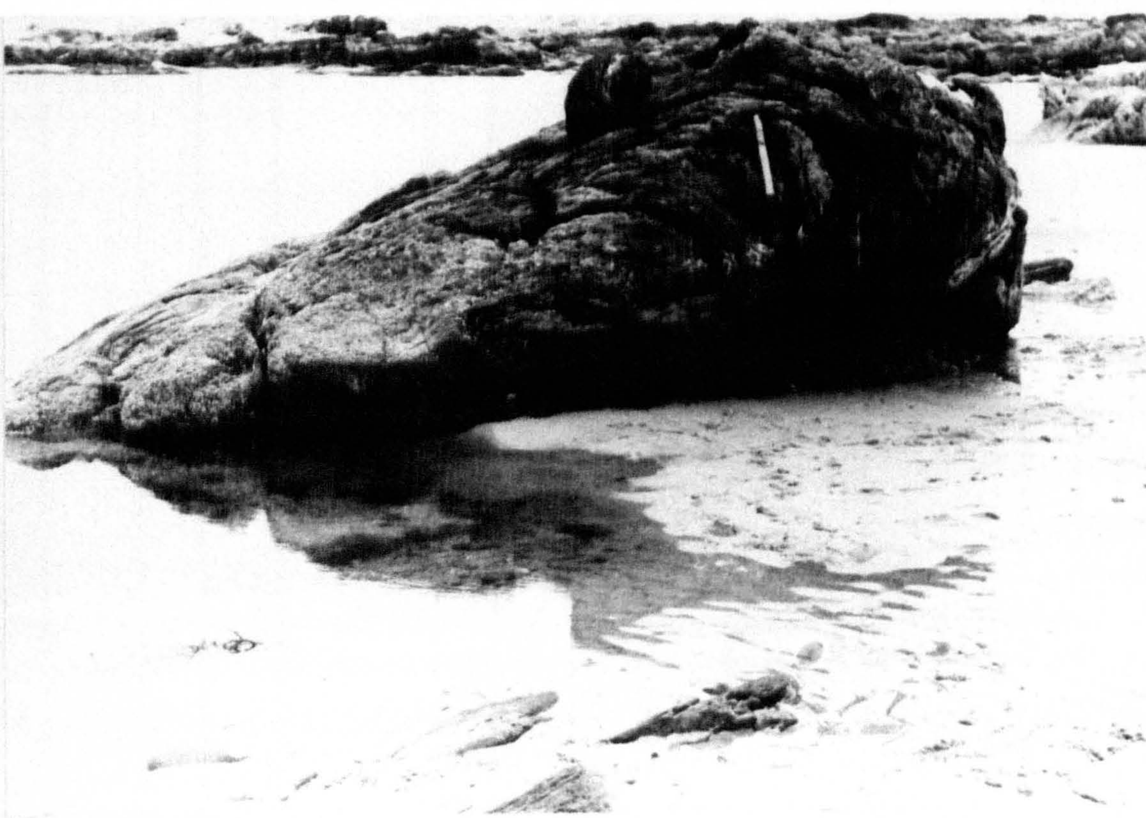


PLATE 6.1 Pebbly overfolded slump unit at the top of section A2 Kintradwell (see also Fig. 6.3) mainly concentrated in discrete laminae and occasional sandstone stringers (< 1cm). The shales also exhibit patchy calcite cementation and contain a scattered fauna of pelecypods, ammonites and belemnites. The ammonites and many of the pelecypods are often little more than decalcified impressions, but other, more robust, fossils are often well preserved, including occasional large oysters $\leq$ 9cm (some with a pearly lustre and possibly aragonitic). Bioturbation was only very rarely observed and was usually more-or-less bedding parallel. A few lenticular calcite-cemented beds are present in the section (see Fig. 6.4); they appear to be calcite cemented shales or argillaceous sandstones, their original nature obscured by the replacive cement. A thin section of one of these bands (K109 on Fig. 6.4) shows incipient cone-in-cone structures like those described in Carboniferous marine cements by Füchtbauer (1971); the calcite crystals grew adjacent to, and disrupted laminae of carbonaceous material.

Sections A3-A5 Kintradwell beach east (NC 927075) *Cymodoce* zone

These sections are centred around the prominent boulder bed tongue located on Kintradwell beach about 100m east of the point where Allt Garbh-chlais crosses the high water mark (see Fig. 6.3). The detailed relationships between the three sections are shown diagrammatically in Fig. 6.5 (see also Plates 6.2 to 6.7). This area was studied in detail for three reasons. Firstly, apart from a small gap, sections A3 and A4 are essentially continuous with section A2; secondly the sections show several lithological variants that are not well developed to the west, and thirdly they illustrate several significant aspects of lateral variation and 'syn-sedimentary tectonics'. The sections are shown in Fig.s 6.6 to 6.8.

Section A3 is located on the west side of the boulder bed tongue where there is comparatively little disruption and deformation of the bedding. Some 13.47m were logged of which about 20% is represented as discrete

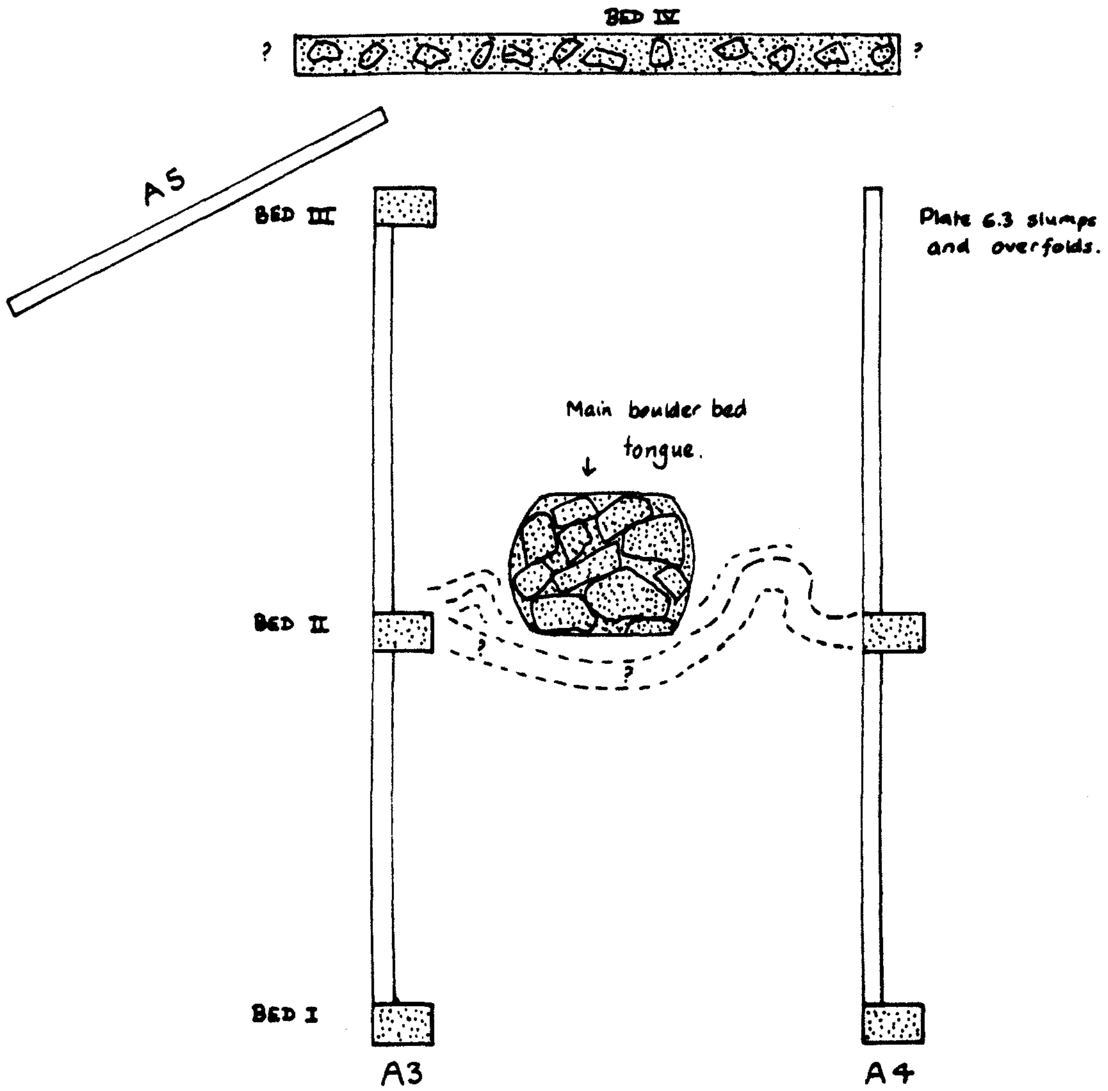


Fig. 6.5 Schematic showing the relationship of sections A3 to A5.

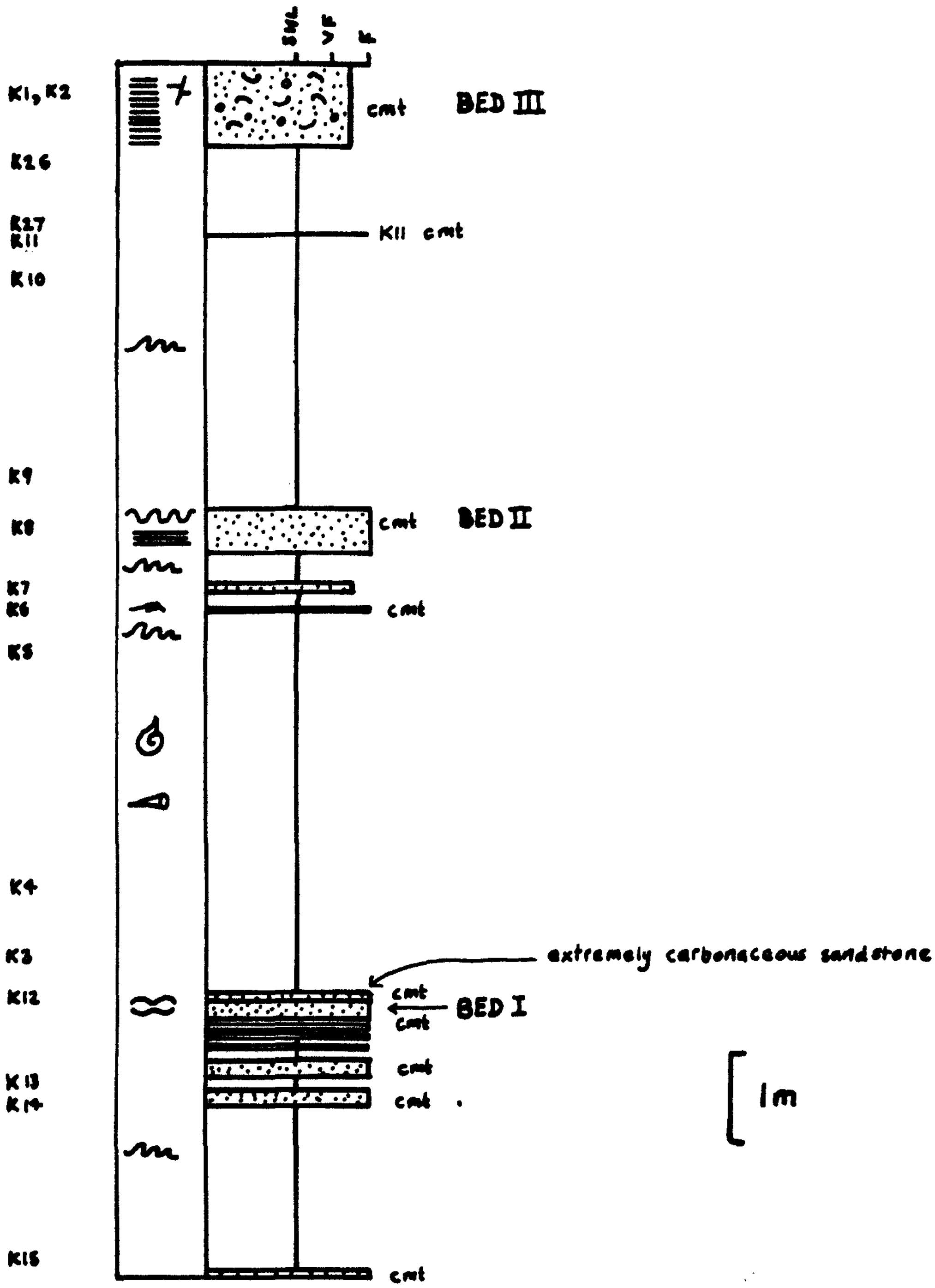


Fig. 6.6 Section A3 Kintradwell.

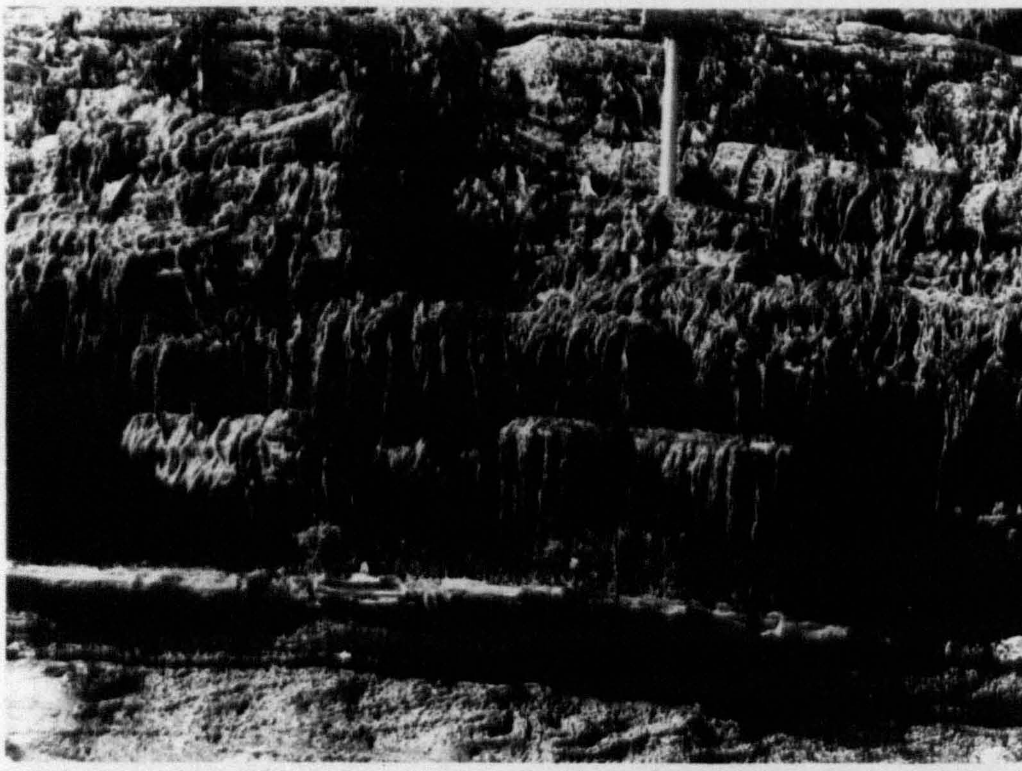


PLATE 6.2 Penecontemporaneous deformation features within section A3, Kintradwell (see also Figs. 6.5, 6.6)

6.2A Incipient pull-aparts in Bed I (near hammer). Note also the prominent horizon of overfolding and dislocation (partly obscured by mud) at the base of the photograph.

6.2B Overfold in Bed II; note also the prominent horizon of dislocation at the base associated with deformation and total disruption of interbedded sand laminae.

6.3C Deformed bedding (resulting from dewatering?) in Bed II.

sandstone beds. The key horizons in the section are the three beds labelled I, II and III on Fig. 6.6; bed I continues un-interrupted below the boulder bed tongue and forms the base of section A4. Beds I and II are easily identified in outcrop as they form prominent reefs and show conspicuous penecontemporaneous deformation structures including convolute bedding and pull-aparts (incipient ball and pillow) as shown in Plate 6.2a (bed I) and Plates 6.2b and 6.2c (bed II). Also apparent in these photographs are examples of small-scale slump folding and at least one dislocation surface (Plate 6.2b). The fine to very-fine grained calcareous sandstones comprising these beds are often highly carbonaceous (3-25% woody material); they are all calcite-cemented (containing 30-60% total calcite) and usually show good lustre-mottled textures.

Section A4 is located on the east side of the boulder bed tongue in an area which is generally much more deformed, making accurate logging more difficult. The lateral equivalent of bed II is recognisable but generally less massive and more shaley in appearance and was thrown up into a broad arch (now breached) by the emplacement of, and compaction around, the boulder bed tongue. The shale interval above the bed II equivalent contains a few thin boulder bed lenses, and in its upper part a prominent over-folded, shelly, calcareous, fine grained sandstone (see Plate 6.3a), and one of the classic, oft-photographed, Kintradwell boulder beds (Plate 6.3b), the latter more-or-less level with bed III in section A3. The total thickness logged in this section was 12.69m (Fig. 6.7) - slightly greater than the same interval in section A3; the difference is probably due to small errors in measuring the thickness of the shale units and also differential compaction effects resulting from variable cementation of the sand 'stringers' in the shales.

Section A5 is located on the west side and stratigraphically above the boulder bed tongue (see Fig. 6.5). It represents an allochthonous slump unit which was emplaced between the deposition of bed III in section A3 and bed IV which overlies section A3 and A4. Most of section A5 is

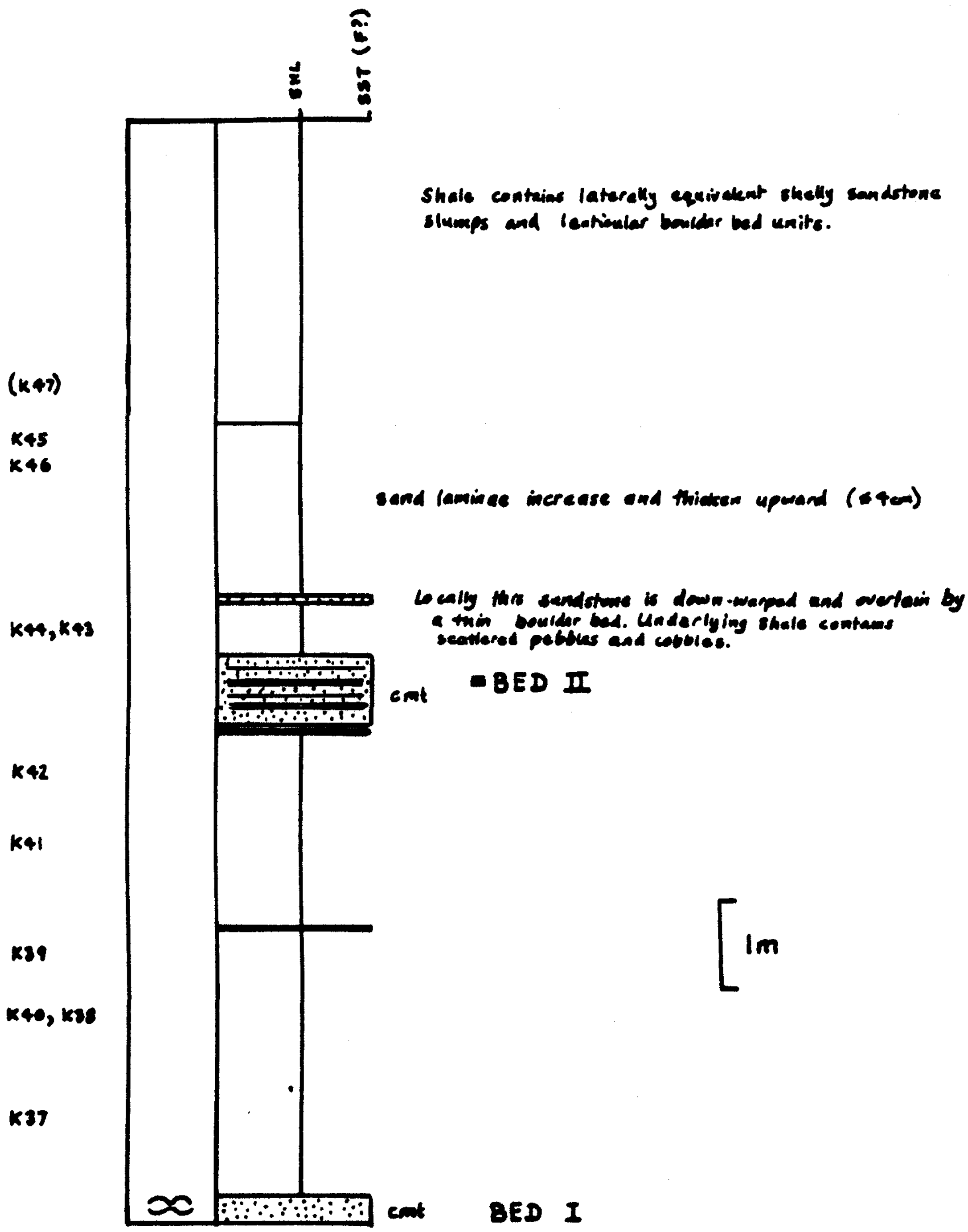
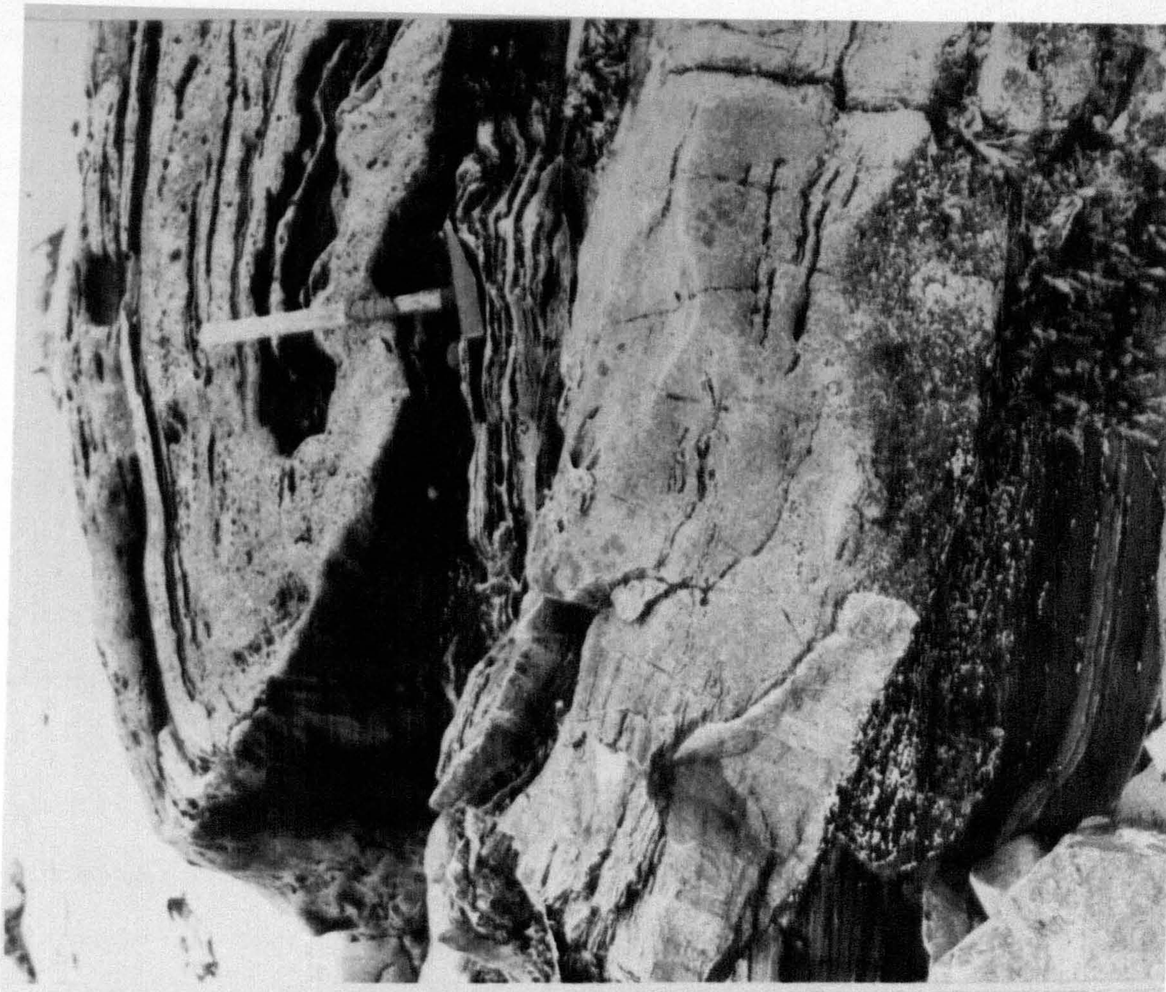


Fig. 6.7 Section A4 Kintradwell.

PLATE 6.3 Prominent overfold (A) and boulder bed (B) in section A4
Kintradwell



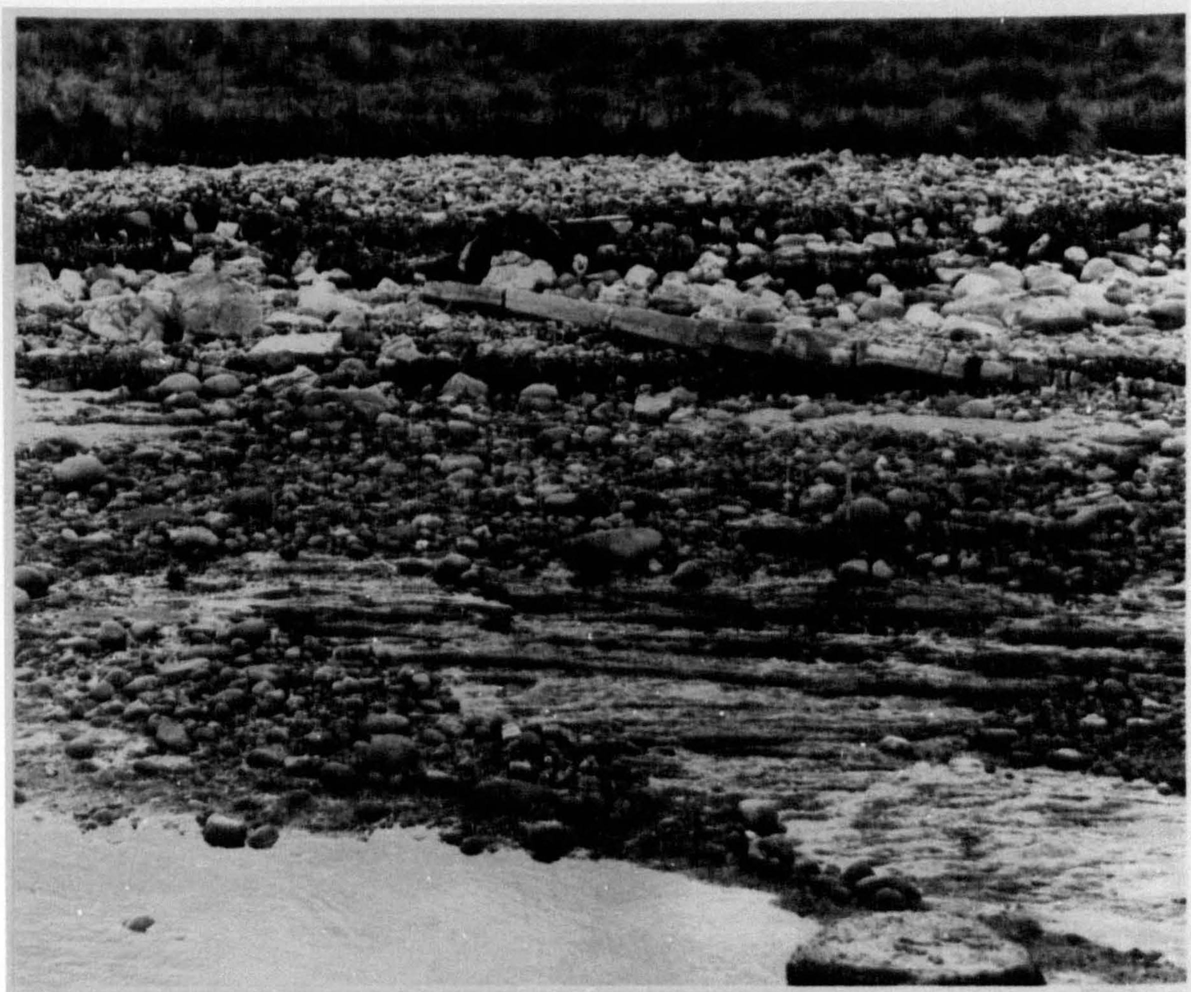
obscured by the shingle beach with only the sandstone beds projecting through the loose cover, the strike of these beds emphasising the angular discordance between this unit and section A3 (see Plate 6.4a). The top of section A5 shows a prominent overfold (see Plate 6.4b) which suggests this allochthonous unit was derived from the east-north-east. Lack of adequate exposure makes it impossible to assess the areal extent of this slump sheet.

The lithofacies of the sediments in this section is somewhat different from that seen in the more autochthonous sections (A3 and A4). Of the 9.43m measured, 20% is represented by bands of white, light grey or cream-coloured calcite cemented sandstones. These beds can be clearly seen to thin in both directions along strike and are, therefore, lenticular units with a maximum estimated lateral extent of about 10m. Many of these sandstones are, at least in part, rich in bioclastic debris (3-25%) and are all lustre-mottled and therefore have overall calcite contents (estimated from thin section) as high as 40-65%. They are all fine-grained. Some of the beds contain interesting sedimentary structures such as the beautiful water-escape structures shown in Plate 6.5a (from the thick bed at the base of the section) and convolute bedding, channeling and microfaulting. Plate 6.5b shows the internal complexity of one of these sandstones, which at its thickest point contains a strongly deformed basal layer rich in bioclastic debris and angular sandstone clasts (generally $\leq$ 8cm, but one clast, indicated on the photograph, measured 17cm in long dimension). The upper part of the bed is more regularly bedded, shelly and convoluted in its upper 5cm. Intra-clasts were also observed in the bed shown in Plate 6.5b whose lower 11cm is rich in belemnite and bivalve fragments and rounded to subrounded sandstone clasts ($\leq$ 2cm). This bed was found to be unique in that out of all those examined in this field area it contained a single in situ 'granite' clast. The top of the bed contains common to abundant, randomly distributed, bioclastic debris and clasts $\leq$ 16cm; since the

PLATE 6.4 Interesting features at beach end of sections A3-A5, Kintradwell

6.4A View showing disconformity between cemented sandstones and shales of section A5 and shales of section A3

6.4B Overfolded sandstone and boulder bed at top of section A5



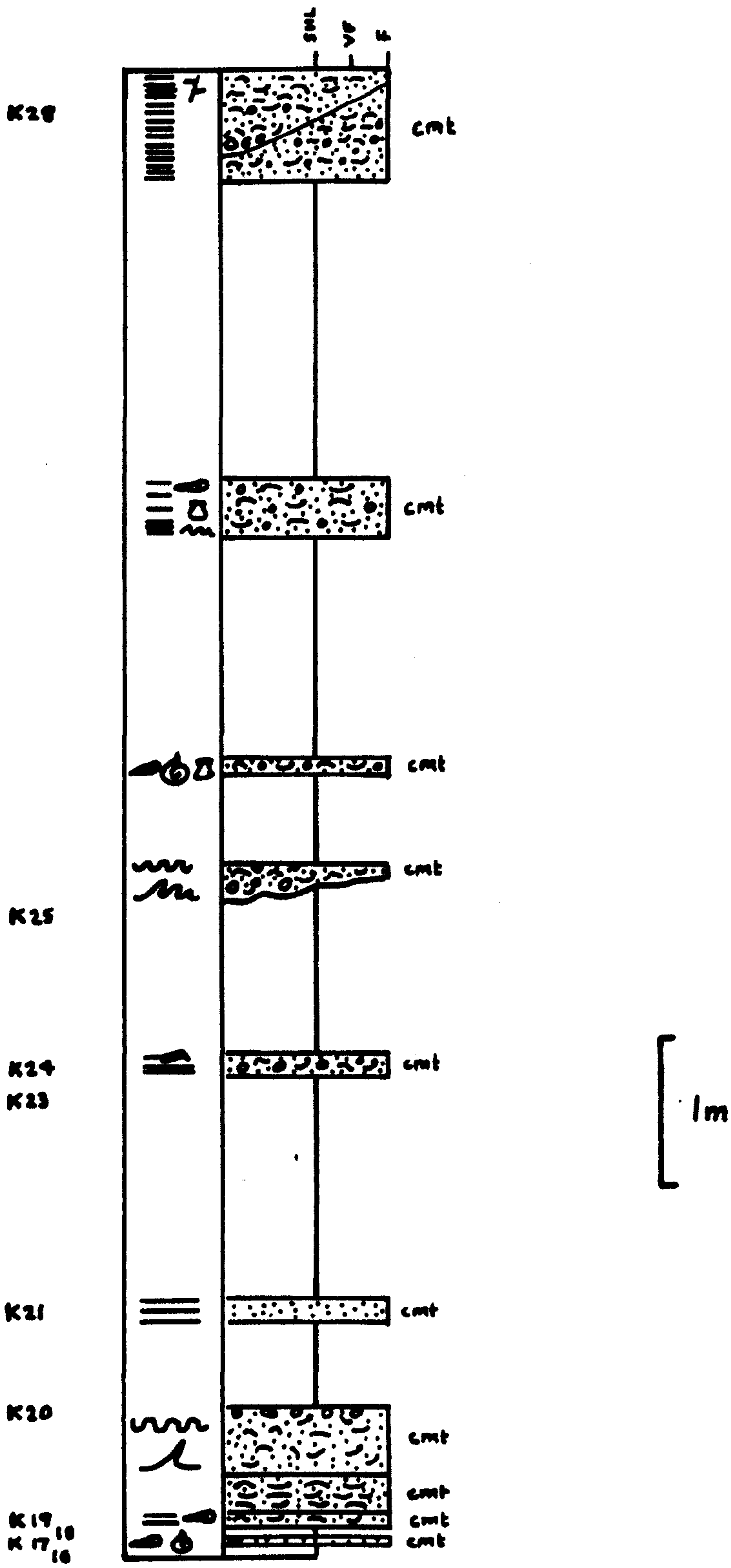
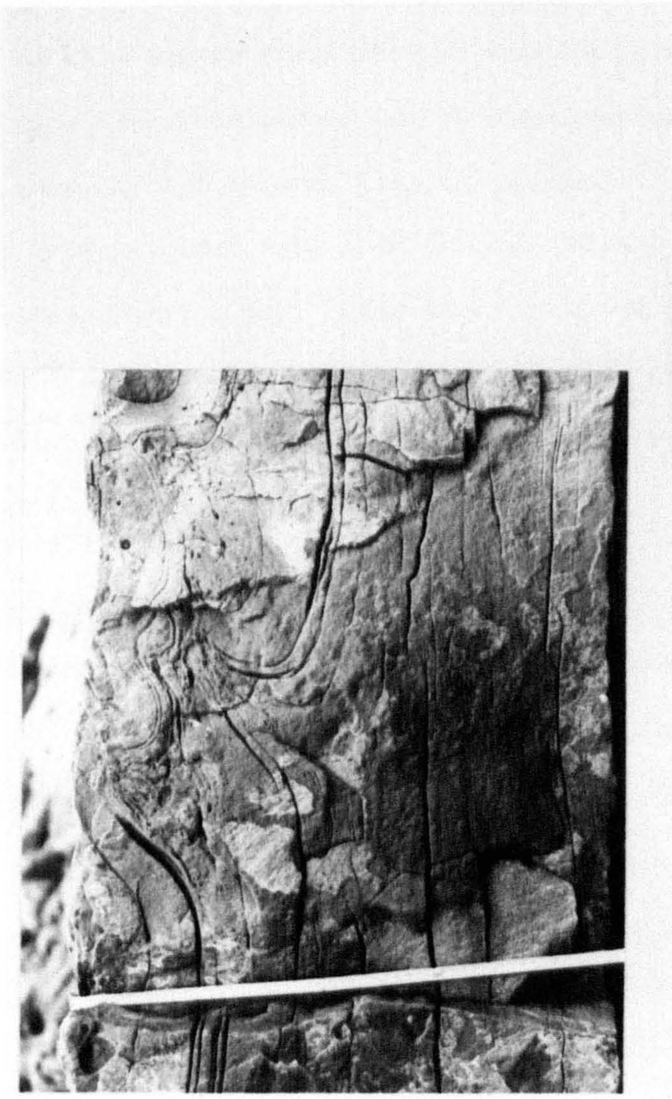


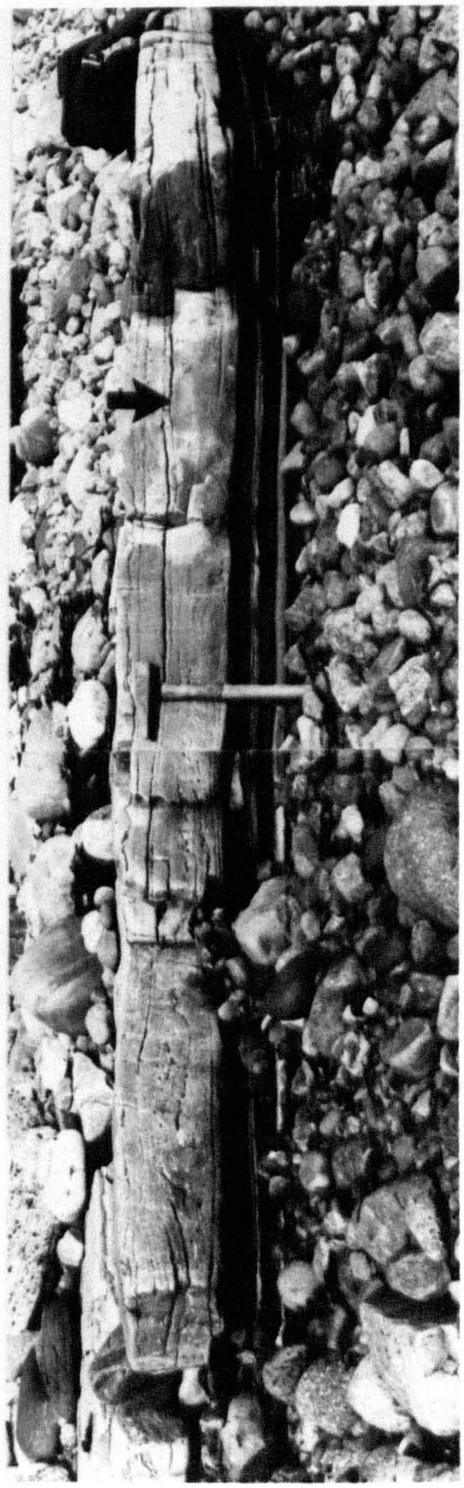
Fig. 6.8 Section A5 Kintradwell.

PLATE 6.5 Detail of Section A5 Kintradwell

6.5A Prominent flame structures in sandstone bed K20 (see Fig. 6.8)

6.5B Lenticular cemented sandstone (above K25 shale) showing basal slumping and convolution (at top and base). Note also the large clast indicated by arrow.





water escape structures occur immediately below, this may represent reverse grading due to dispersive pore pressures in a temporarily fluidised bed. Very coarse to granule grade clasts of quartz cemented, pure or micaceous sandstones were commonly observed in thin sections. The uppermost bed in section A5 contains several instances of channeling. Most of this bed consists more-or-less of parallel-bedded sandstone with bioclast-rich layers, but at least one channel (60cm x 3m) is cut down into it; the channel fill has a basal 5cm coarse bioclastic lag which grades upward into fine grained, finely bedded, microfaulted sandstone (Plate 6.6a). Near where it forms the closure of the plunging overfold shown in Plate 6.4b,this bed contains an intercalated (channelised?) boulder bed unit which is 24cm in thickness and contains sandstone clasts up to 22cm in diameter.

Boulder bed tongue and bed IV

The main boulder bed tongue shown in Plates 6.6b and 6.7 has, as pointed out by Linsley (1972, p.43), "a density uncommon in the Kintradwell succession". It is an elongate mass (1-2m x 3m x 15m) formed by a clast-supported monomict 'breccia', composed of angular to subangular clasts of fine grained, cemented quartz sandstones with a sandstone matrix. The clasts range in size from less than 5cm to 1.3m (long axis) and all exhibit parallel bedding or planar cross-bedding. As well as being unusually clast-rich for this locality, the tongue is also remarkable in that its emplacement appears only to have deformed, rather than eroded, the underlying sediments.

Bed IV is a more modest boulder bed forming the top of sections A3-A5. It is a clast-rich, matrix-supported, monomict boulder bed containing clasts of angular to subangular, planar cross-bedded, cemented, grey quartz sandstones (maximum dimensions $\leq$ 1m, average $\sim$ 30cm). The matrix consists of very fine to fine grained, calcite-cemented sandstone, which is pure and contains some bioclastic debris at the base, but becomes dirtier and more carbonaceous at the top. The sandstone clasts,

PLATE 6.6 Boulder bed units from sections A3-A5 Kintradwell

6.6A Channelised boulder bed in overfold at top of section A5

6.6B Main boulder bed tongue at east Kintradwell (see also Fig. 6.3). Note arching and folding of underlying shaley sandstone.

PLATE 6.7 General view of east Kintradwell beach. Section A5 largely obscured by beach shingle on left; main boulder tongue visible just below horizon; prominent reef formed by edge of Bed II at right.

some of which weather red, show a slight tendency for a preferrred orientation (long axes 180°-190°); calcite veins cut both matrix and clasts in the same direction. The largest clasts appear to occur at the top of the bed, but this may be only apparent and result from the poor exposure. The clasts in both these units may be Jurassic and/or Devonian (q.v. Macgregor et al. 1930, p.82), the latter of the middle Old Red Sandstone John O'Groats Sandstone facies.

Section A6 Allt Garbh-Chlais (NC 925075) ? *cymodoce* zone

A small section (5.19m) of shales and sandstones is accessibly exposed in the stream below the small road bridge southwest of Kintradwell Farm. A little over 9% of the section is represented by thin sandstone beds. The shales contain abundant thin laminae of uncemented fine sand. Some good plant fossils and oysters were noted and a strike and dip of 273°/20°N was measured. The section is shown in Fig. 6.9.

Section A7 Kintradwell Bluff (NC 926077) ? *cymodoce* zone

This section is located about 150m to the north east of section A6 in the low cliff running just to the south east of the road. About five and a half metres were measured (Fig. 6.10), much of which was poorly and only patchily exposed due to grass cover. The most conspicuous bed is that at the base, a calcite cemented, fine grained sandstone with belemnite and oyster debris and pebbly lenses containing sandstone clasts. The presence of a few thin shales ($\leq$ 3cm) intercalated in the sandstone suggests the bed is composite. The remainder of the section is shale but there are common bands of uncemented fine sand varying in thickness up to 4cm (mainly $\sim$ 1cm). Some of these bands show slight cementation by iron oxides but are still friable.

DISCUSSION OF THE KINTRADWELL REGION

One of the few comments that Bailey and Weir (1932) made on the Kintradwell sequence was that the beds were "involved in comtemporaneous

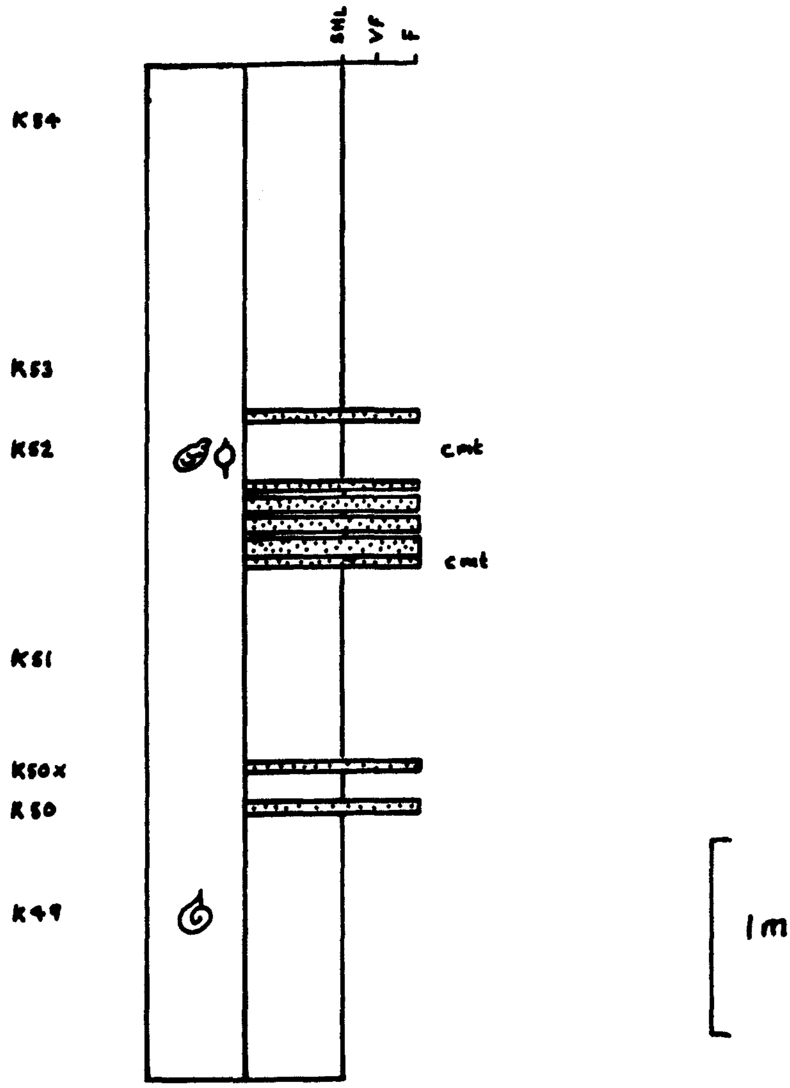


Fig. 6.9 Section A6 Allt Garbh-Chlais (Kintradwell).

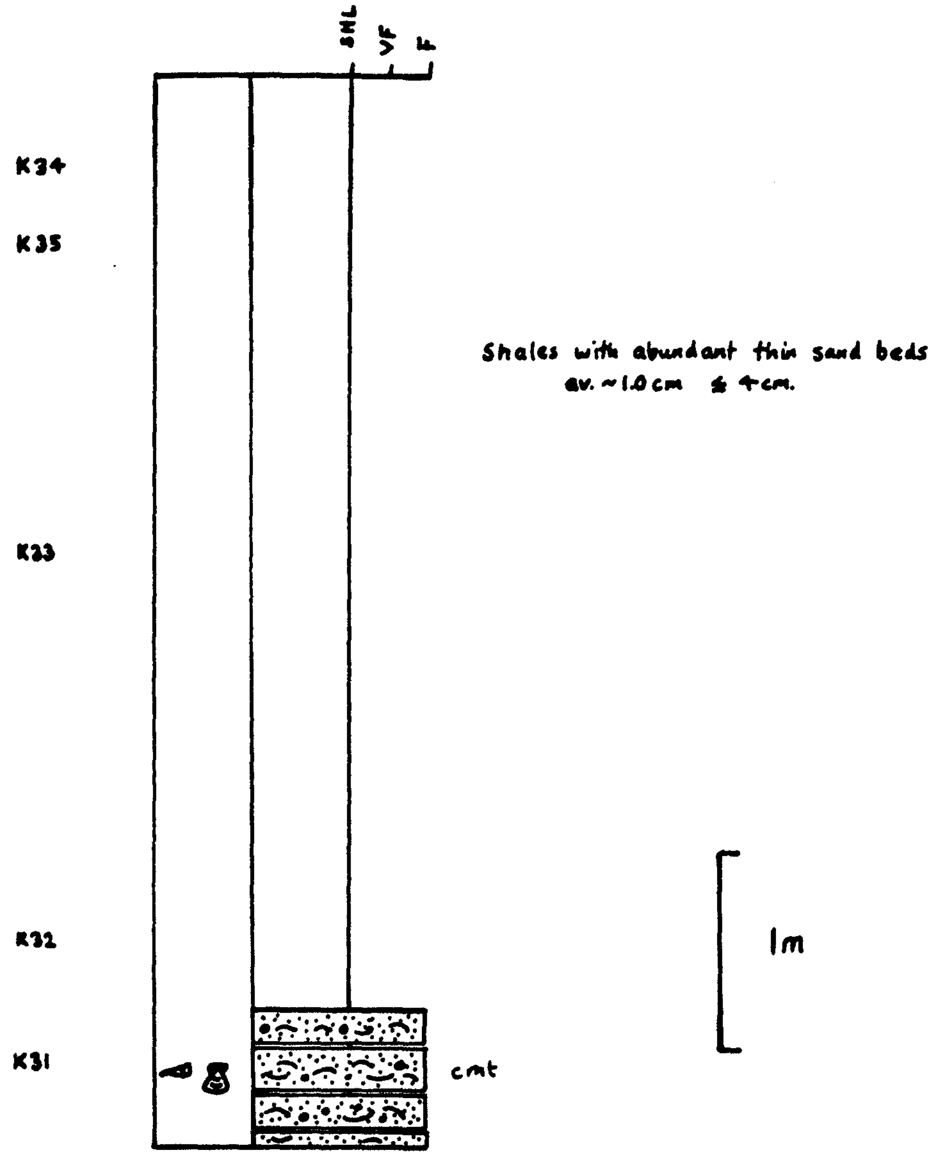


Fig. 6.10 Section A7 Kintradwell.

contortions" (p.443). In retrospect I also have come to consider this one of the most significant features of this locality. The several noteable slumps and overfolds (e.g. the boulder bed tongue on east Kintradwell beach, the overfold at the top of section A4, the overfolded pebbly block at the top of section A2) which have been noted by Bailey and Weir (1932) and Linsley (1972), actually represent only a fraction of the evidence on which this conclusion was based. In my initial observations on the west Kintradwell outcrop I was perplexed by the lateral variability and complexity of the lithological sequence. However, once I had come to establish the irrecontrovertable presence of slump sheets in the sequence (section A5, east Kintradwell), I soon appreciated that the complexity was not just a case of simple lateral facies variation and that slumps and dislocations of variable scales were quite common throughout the outcrop. Many of these phenomena are not conspicuous since they appear to have involved slabs of sediment which slid down the palaeo-slope on surfaces more-or-less parallel to bedding without causing any great disruption (although sometimes slight rotation was apparently involved). Given that the facies relationships are, in the first instance, already very complex it is impossible to ascertain how much lateral displacement has been involved. Probably most of the phenomena are quite local.

Linsley (1972, p.43) describes the 'wedging out' and truncation of (sand?) laminae in the shales as being the result of erosive bottom currents. Although I observed many such features in the Kintradwell outcrop I could not definitely attribute any of them to erosive surfaces in the normal sense. Most appeared to represent low angle dislocation surfaces produced by lateral slippage and sliding; a good example is visible in Plate 6.2b from the shale below bed II in section A3. Laterally these dislocations often appear to pass into thin bands ($\leq$ 20cm) of wrinkled, overfolded or complexly deformed laminae which either represent the deformed toes of such slides, or drag folding in the sediment underlying the slide. These small scale features tend to be most common in the vicinity of sandstone and boulder beds, possibly suggesting that rapid deposition and loading produced high pore pressures in the shales, reducing friction and encouraging the formation of dislocation surfaces.

Another interesting aspect of the slumping and sliding in the Kintradwell area is their bearing on palaeoslope directions. There are two conspicuous directions; one more-or-less north-south at right angles to the strike (shown by slides, clast orientation, etc.) and the other NE-SW to ENE-WSW which is often demonstrated by the larger over-folds (in sandstone beds). These observations (although lacking detailed quantitative studies) imply that the maximum palaeoslope lay in the quadrant south to south-west, a finding compatible with the data presented by Crowell (1961) for the Upper Jurassic sequence as a whole. The magnitude of the palaeoslope was probably only a few degrees - easily sufficient for slide formation, particularly if the area was exposed to periodic seismic tremors as has been widely supposed by earlier workers.

Sedimentation in the Kintradwell area was dominated by shale deposition with only about 20% (less by volume) of the succession being made up of sandstones and boulder bed units. Most of the sandstones are parallel-sided, featureless units, their only often-visible structure being the presence of carbonaceous laminae in their upper part. Very few exhibit anything approaching a bouma sequence, some of the noteable exceptions being in sections A1 and A5 where A-E or B-E sequences are sometimes recognisable. The latter beds also variously show erosive bases, amalgamation, matrix grading (rare), coarse-tail grading of clasts and bioclastic material (common) rare cross-lamination, convolute lamination, pull aparts and loading structures. All these features are characteristic of rapidly deposited sediments of the turbidite association. Many of the more featureless sandstones are also probably turbidites but their true nature has at least in part, been obscured by the widespread replacive calcite cements. Sedimentary structures can often be seen only when they involve the distribution and or orientation of bioclastic or carbonaceous debris.

The cement represents such a large proportion of the rock volume that it tends to limit the accentuation of structures during weathering. The apparent paucity of sandgrains coarser than fine grained in the depositional environment has also made grading rarely visible except where it is illustrated by the distribution of clasts, carbonaceous or bioclastic debris. The depositional processes of the boulder beds are discussed at the end of the chapter.

LOTHBEG AREA

Section B1 Allt na Cùile (NC 941093)

This section is situated in the cliffs on the north side of the railway line approximately 70m north east of the mouth of the Allt na Cùile gorge (see Fig.s 6.11, 6.12 and Plate 6.8a). The section consists of about 10m of grey, yellowish-grey and yellow, variably ferruginous, fine to very fine, moderately to poorly sorted,'veined' quartz sandstones. The sandstone varies in character from weak and friable to relatively hard, the degree of cementation varying mainly with the degree of iron 'staining', but also resulting from slight pressure solution (and in some beds minor authigenic quartz overgrowths). The upper part of the sandstone is well bedded, the lower part more massive, but some crude layering appears to be present in most units. The sandstones are loosely packed and porosity determined in thin section varies from 25-40% in the cleaner, more well-sorted beds, to 5-20% in the dirtier, less well-sorted ones.

The lower part of the section consists of an alternation of carbonaceous and pure sandstones which contain common ($\leq$ 25%) dispersed, but often crudely layered, rounded quartz granules and pebbles. The

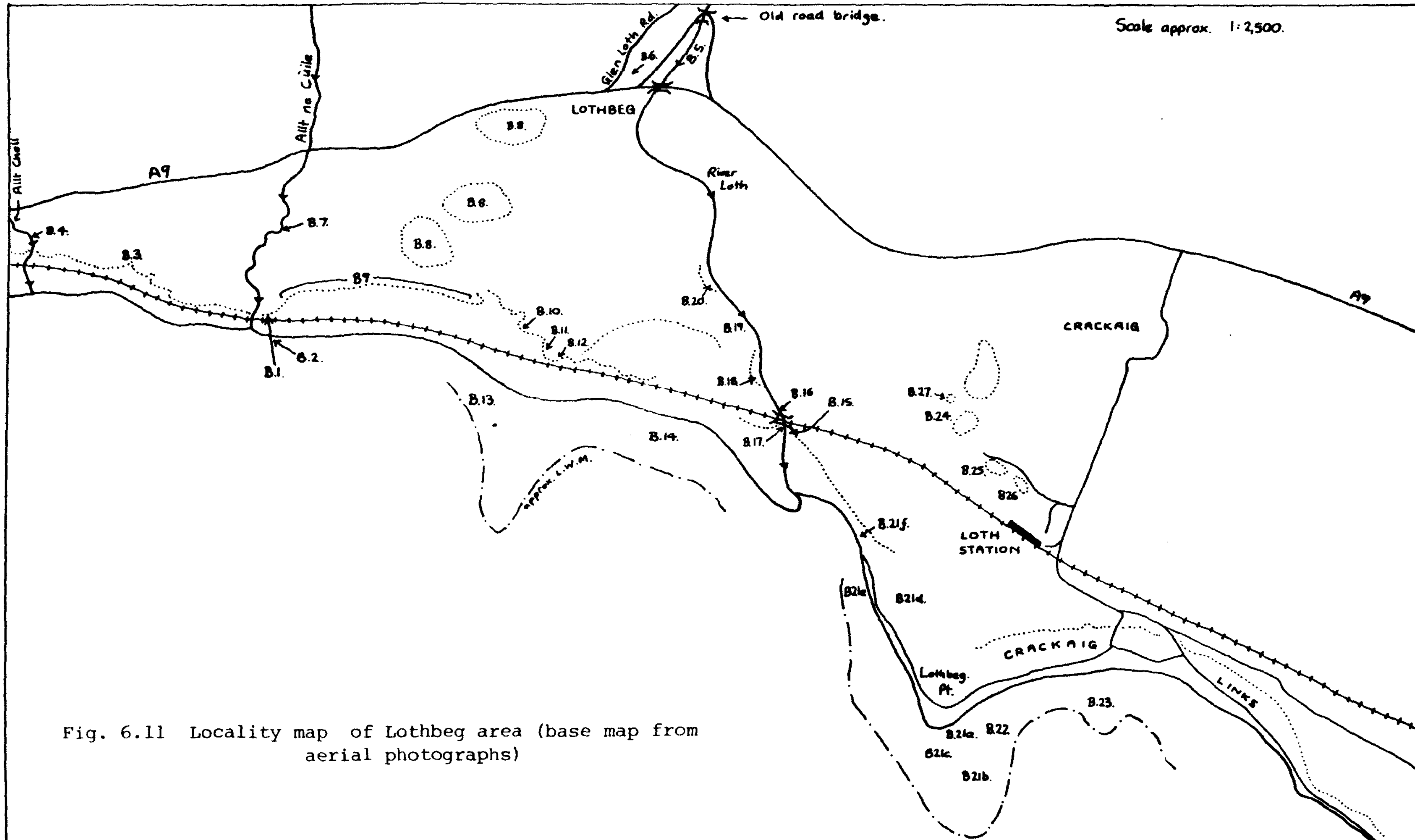


Fig. 6.11 Locality map of Lothbeg area (base map from aerial photographs)

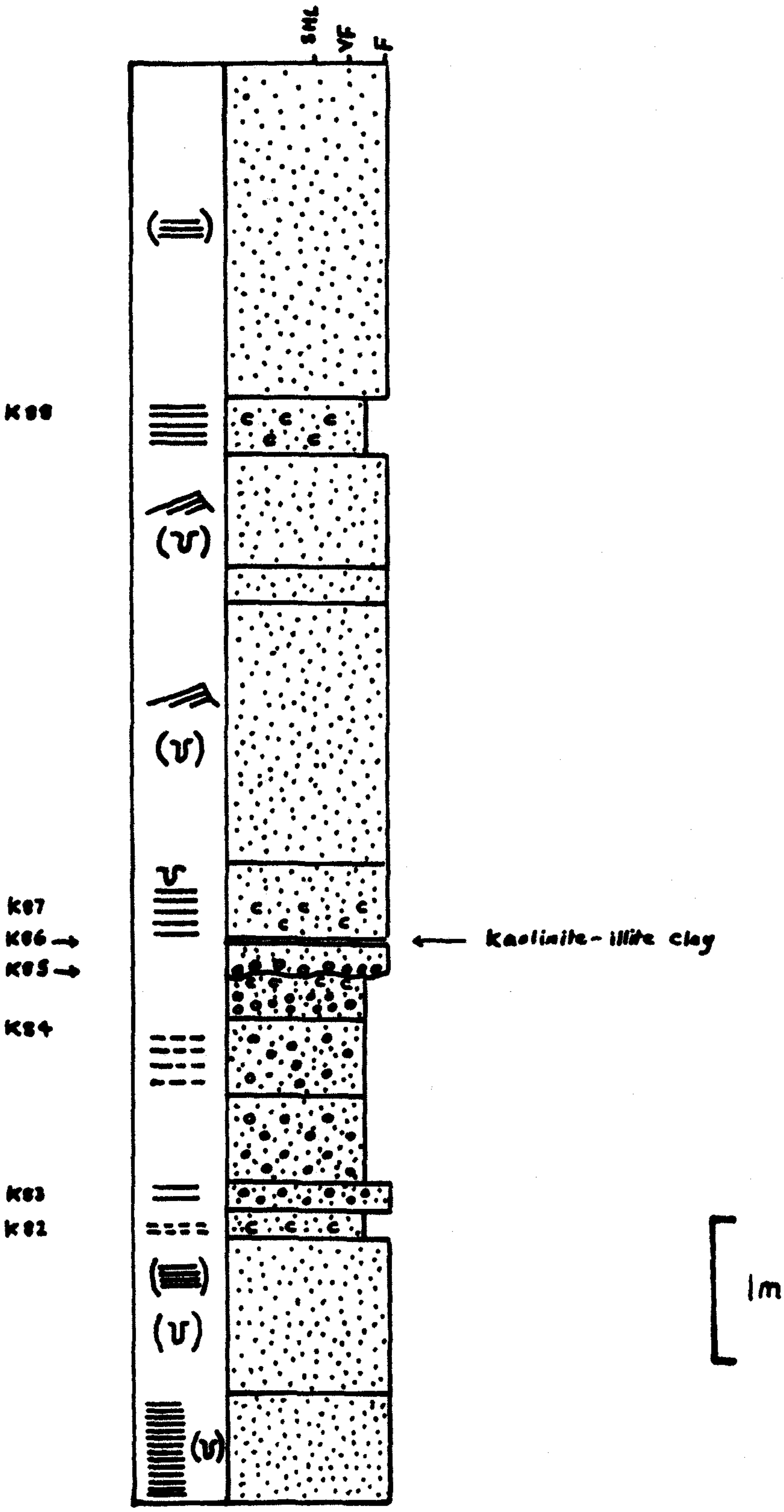
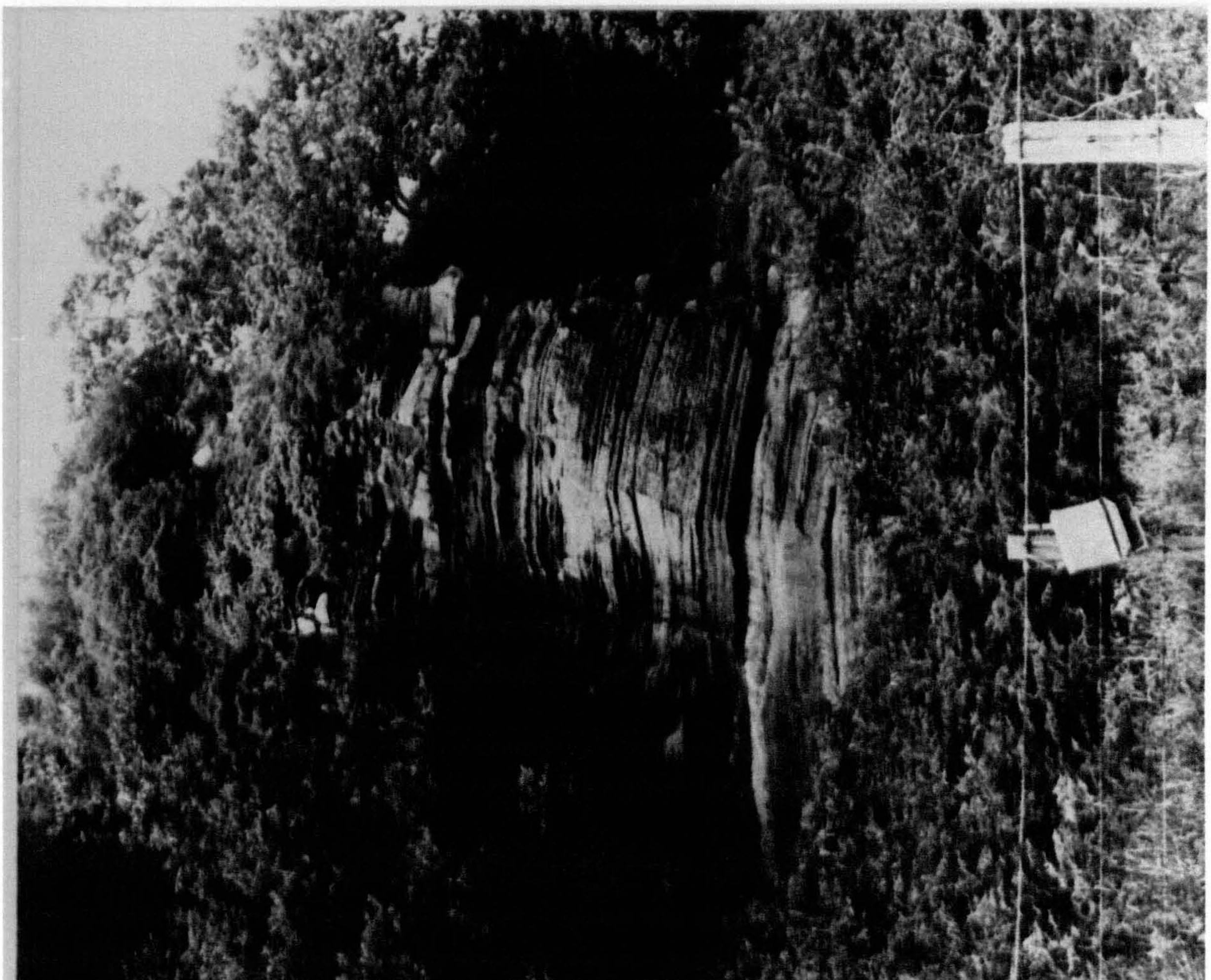


Fig. 6.12 Section B1 Allt na Cùile Sandstone body.

PLATE 6.8 Cliff exposures of the type Allt na Cùile Sandstone, locality B1

6.8A General view of locality B1 taken from shore side of railway line. Note upper well bedded sandstone and lower more massive unit. Plate 6.9 taken under overhang (shadowed area) to left.

6.8B Detail of current bedding exposed on right side of cliff in 6.8A.



upper part of the section is much better bedded and consists of pebble-free, better sorted sandstones arranged in 1-3cm thick, planar or tangential beds in sets bound by planar or slightly concave-upward surfaces (see Plate 6.8b). The individual beds sometimes weather into a rather irregular 'knobbly' appearance and occasionally (where the weathering has been more discriminating) it can be seen that many of the beds are highly bioturbated and contain numerous trace fossils (including Chondrites, Thalassinoides and Ophiomorpha-like forms). This is particularly well illustrated in Plates 6.9a and 6.9b taken under the overhang (shadowed area) visible on the left hand side of Plate 6.8a.

The two uppermost beds in the lower part of the section are also worthy of particular mention (see Fig. 6.11). The lower of the two beds contains a basal 7cm division rich in quartz pebbles ($\leq$ 3cm) and scattered better cemented, angular, intraformational (?) sandstone clasts ($\leq$ 15cm) lying on an erosional contact. The top bed is interesting because its lower 5cm consists mainly of a relatively pure, waxy, light grey clay. Subsequent XRD analysis of this clay showed it to be composed mainly of kaolinite with illite. Within the vicinity in which section B1 was measured, the lower part of the sequence showed rapid lateral changes in lithology; within about 10m ($\pm$ along strike) thinly bedded sandstones pass into massive ones; hard, cemented, apparently massive sandstones, pass into friable, laminated, carbonaceous sands and vice versa.

Lee (1925, p.104) gives a measured section of 27-28m from the gorge of the Allt na Cùile, just 70m south west of section B1. This has been considered the type section of the so-called Allt na Cùile sandstone, although as Linsley rightly points out, the facies variation is such that it can hardly be considered representative. From the contemporary sedimentologist's view, Lee's section unfortunately leaves much to be desired and so has not been reproduced here. However, Lee does note the presence of poorly preserved fossil casts, emphasising the decalcified

PLATE 6.9 Bioturbation within bedded sandstone of the type Allt na Cùile. Burrow forms include *Ophiomorpha*, *Planolites* and *Chondrites*.


nature of the sandstone, which are absent or were overlooked in my section. At the time of my field work, section B1 was the only one in this sandstone which was effectively accessible; it is perhaps more detailed but no more representative than was Lee's. The only real common observation between the two sections is that they both indicate that the pebbly sandstones are, with the exception of the higher conglomeratic units, limited to the lower part of the outcrop (first ± 5m above the railway), even if only for that part of the sandstone exposed between these two points. In addition, Linsley (1972, p.54) reports that there is a "twenty feet high exposure of silty carbonaceous shales" at the 'start' of the Allt na Cùile gorge (NC 941093) which is not recorded in Lee's section and was not visited during this study. Plate 6.10a shows the cliff face exposed immediately to the east of the mouth of the Allt na Cùile gorge; note the lateral variation in the expression of the bedding and the apparently lenticular, rubbly (conglomeratic?) horizon half way up the face on the right hand side.

Section B2 Allt na Cuile beach (NC 941092)

The 6.58m of section shown in Fig. 6.13 was recorded in beach exposures about 125m south east of section B1. The sandstones exposed on the beach were soft and friable, partly obscured by algae and sand,and could not be described in detail comparable to section B1. It was estimated that the gap between the top of section B2 and the base of section B1 was in the order of 2-3m. The section shows the continuation of the pebbly sandstones characteristic of the lower part of section B1. Although exposure did not allow detailed observations, the pebbles often appear to be arranged in 'graded' units of about 20cm thickness, the pebbles dispersed in the matrix, but increasing in abundance toward the base.

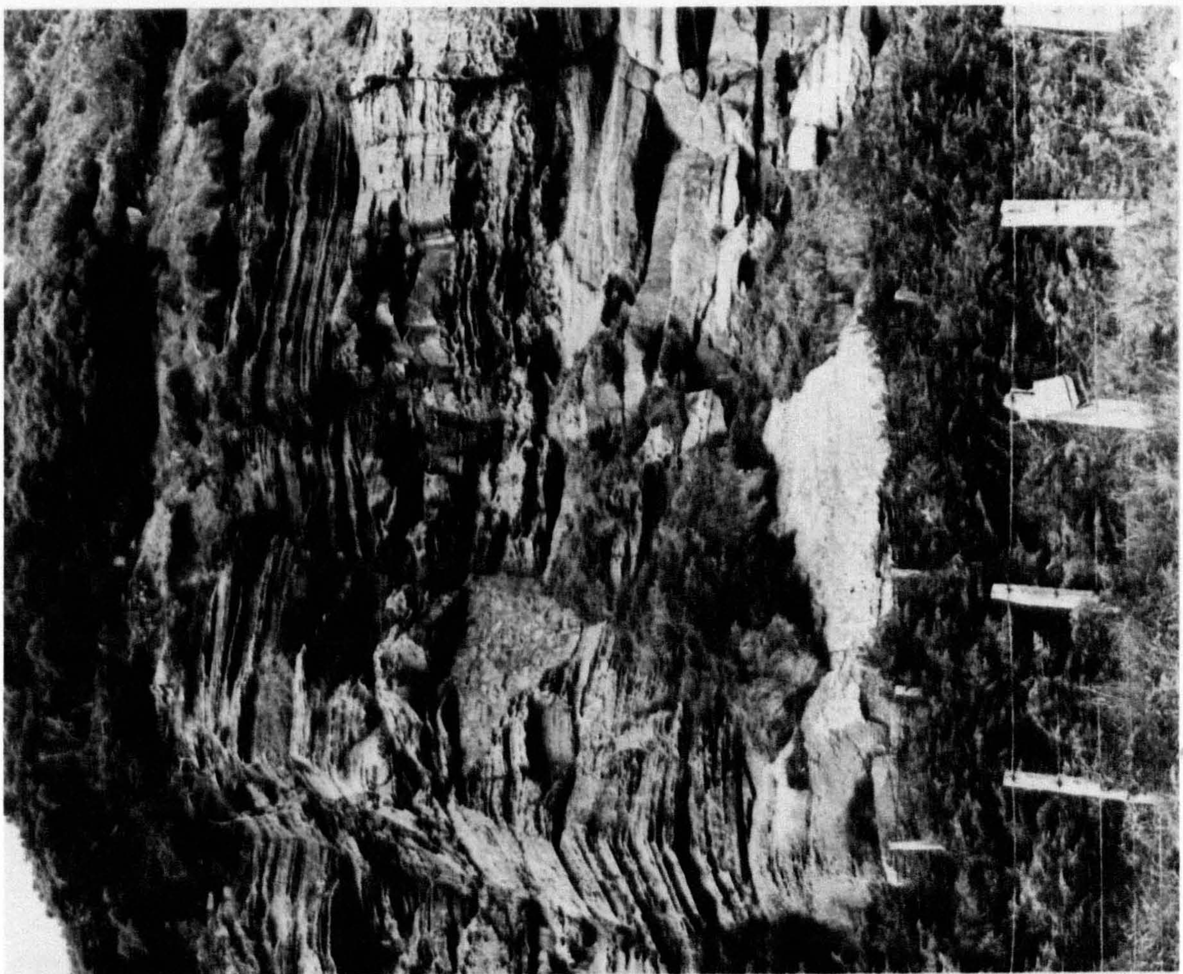
Locality B3 un-named stream south west of Allt na Cuile (NC 937091)

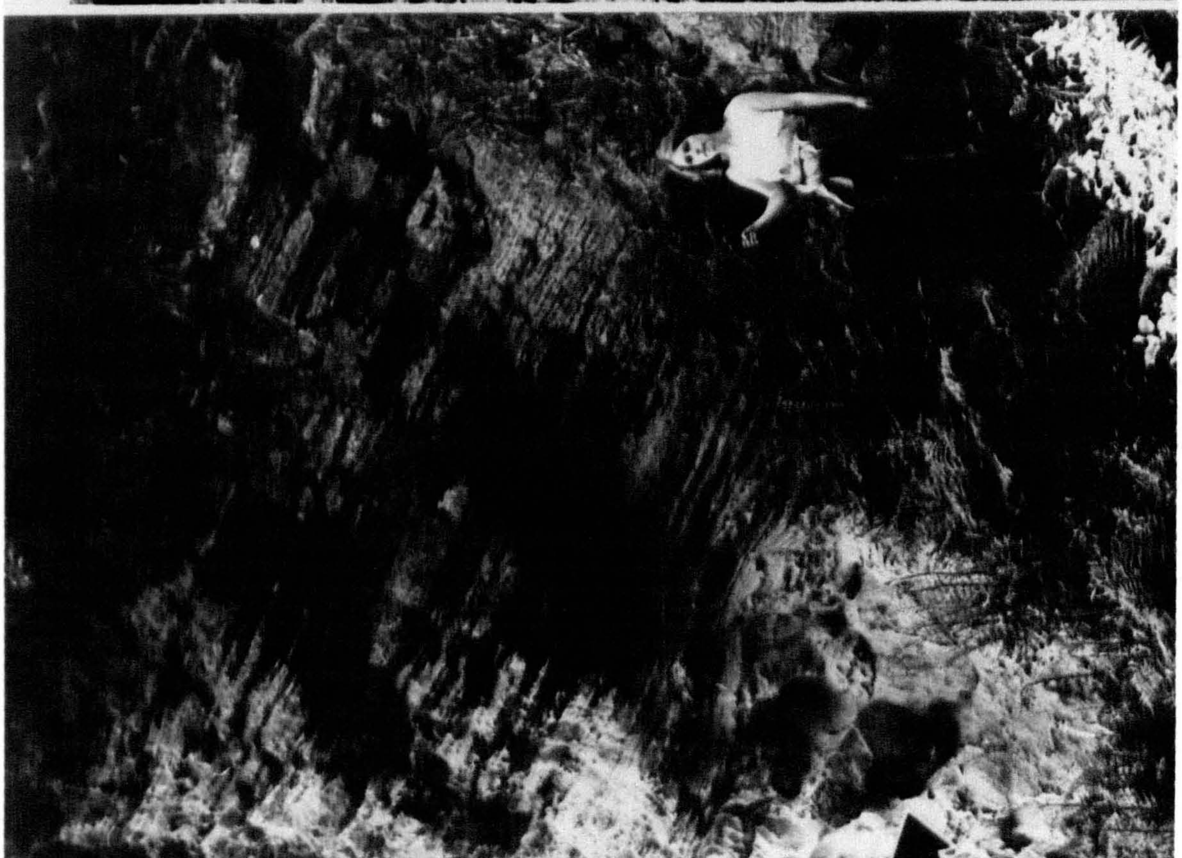
This locality occurs in the small, most-westerly valley of the two

PLATE 6.10 Additional views of the type Allt na Cùile sandstone

6.10A Cliff exposure west of locality B1; note rubbly horizon centre right

6.10B Bedded Allt na Cùille sandstone overlying massive breccia horizon locality B2. Angular dark clast is probably Moinian. Sally Barritt for scale.





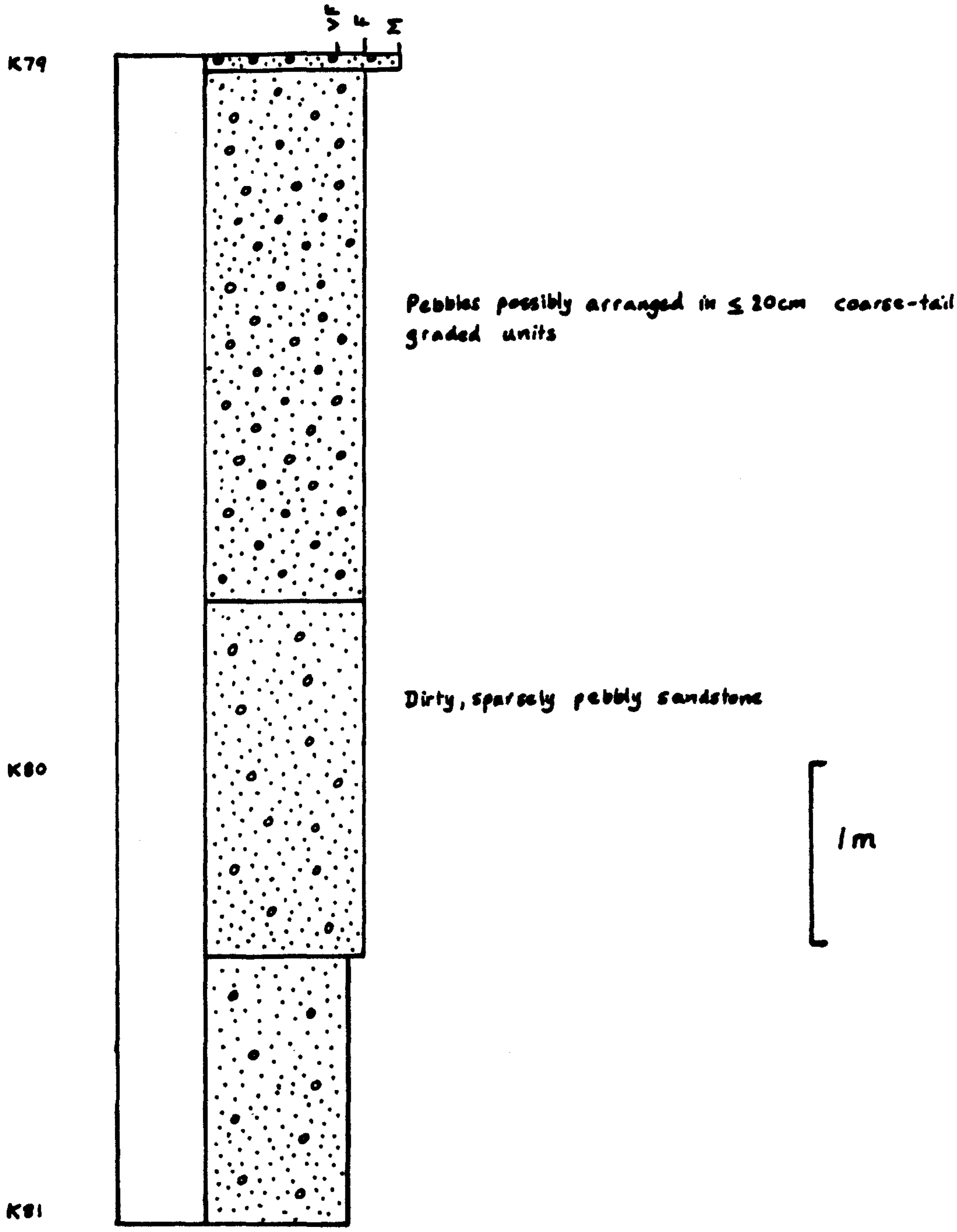


Fig. 6.13 Section B2 Allt na Cùile Sandstone body.

streams which occur in this area, and is approximately 350m south west of the mouth of the Allt na Cùile gorge. The sandstones exposed in the sides of the valley are mostly of the bedded variety like the upper part of section B1, but just inside the mouth of the valley, intercalated in these bedded sandstones, is a 1.5m thick, clast to matrix-supported breccia unit (Plate 6.10b). This band appears to be more-or-less conformable with the bedding (although naturally somewhat more irregular) and consists of a soft, friable sandstone matrix with abundant angular to subangular clasts (70-80cm) of better cemented sandstone. This outcrop is only briefly described in order to demonstrate the fact that the thickness and scale of these intraformational breccia horizons (only ∿ 10cm in B1) apparently increases in a progressive fashion south westerly along the sandstone outcrop. In fact two breccia horizons occur in this particular valley (q.v. Linsley, 1972, p.64).

Locality B4 Allt Choll (NC 935090)

This locality occurs in the sides of the small gorge of the Allt Choll between the road bridge and the cliff line on the northern side of the railway. It represents one of the best exposures of intraformational breccia to be observed in the Lothbeg area. No section can be offered due to the lack of bedding but, allowing for dip, up to about 40m of the breccia may be exposed in the course of the gorge. The best exposed face is shown in Plate 6.11. Little can be added to the combined descriptions of Bailey and Weir (1932) and Linsley (1972), and the following is based on both these sources with only minor additions. The breccia is essentially a larger version of that exposed at locality B3; it varies from clast-rich to clast-supported and consists of a friable sandstone matrix with angular clasts of better cemented, but essentially identical sandstone, which vary from a few centimetres up to 3m in maximum dimension and are often internally parallel or cross-bedded. The structure of the breccia is largely

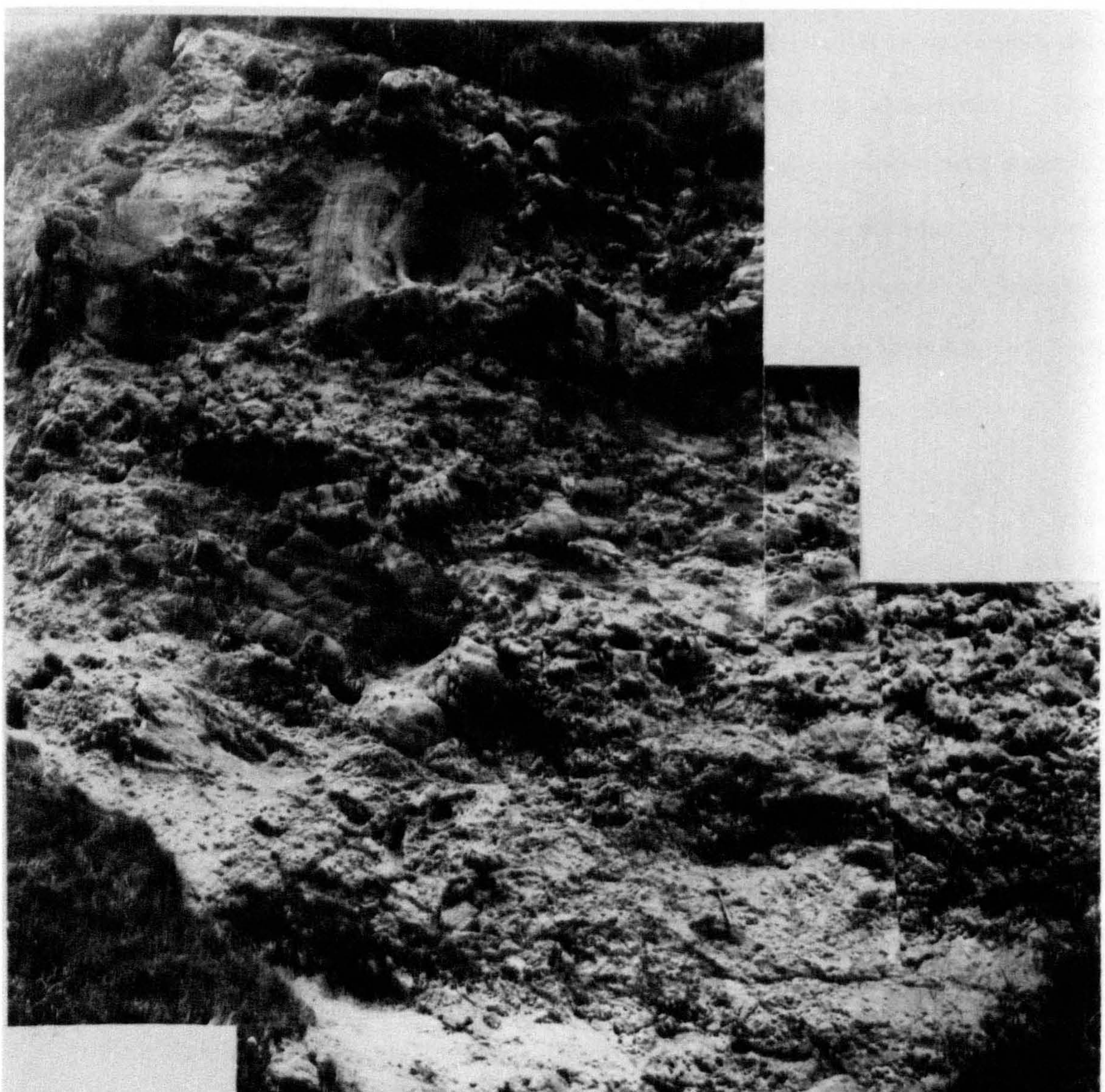


PLATE 6.11 Main exposure of the Allt Choll breccia facies locality B4; hammer near base gives scale. Note crude imbrication.

chaotic, although when viewed from a distance (as in Plate 6.11) some crude imbrication is apparent. Brookfield (1973, p.527 and 1976, p.182) records the presence of rare, exotic angular blocks of Moinian metasediment up to 1m in length, although these were not commented upon by Linsley or Bailey and Weir (1932) and were not conclusively identified during this study. Linsley (1972, p.60-63) has shown that Bailey and Weir's famous 60 foot long horizontal boulder (Bailey and Weir 1932, p.440) located near to the top of the cliff at the base of the gorge of the Allt Choll, is, in all probability, an intercalated in situ lens of bedded sandstone. This presumably reflects the lateral facies change to sequences in which bedded sandstone dominates - as at locality B3. The matrix of the breccia was described by Bailey and Weir (1932, p.440) as containing more large quartz grains than the boulders and as exhibiting casts of belemnites and shells. One feature noted in this study, but not in previous works, was the presence of a light grey waxy clay as matrix between some of the boulders in the breccia exposed in the east cliff at the very mouth of the Allt Choll gorge. This clay is generally inconspicuous, being covered by loose sand, and was only discovered by accident. Subsequent XRD analysis showed a kaolinite-illite composition identical to that of the similar clay described from section B1.

Three other examples of this facies (friable, yellow-orange, quartz sandstones with intraformational boulder-sized clasts) were located and described during this study; one of these is also located in the Lothbeg area, the two others geographically removed (see also the similar facies described from Crackaig Links). These three outcrops are described below.

Localities B5 (NC 944104 and locality B6 (NC 943104) Loth Burn

Sandstone breccia very similar to that described from Allt Choll is well exposed in the banks of Loth Burn between the new and old road bridges (locality B5). This outcrop was first described by Bailey and

Weir (1932, p.441) whose account was reproduced by Linsley (1972) without the addition of further detail. A more complete description was made during the present investigation. The banks of Loth Burn consist of a chaotic sandstone breccia composed of angular blocks of parallel or cross-bedded, pure or variably carbonaceous, yellow to grey, friable to relatively hard quartz sandstone in a matrix which variably consists of sandstone, carbonaceous sandstone or light grey clay. The sandstone clasts are generally between 30 and 50cm (up to $\sim$ 1m) in diameter, although there are also several blocks which are 4-6m thick (measured normal to their internal bedding). Like Linsley (1972, p.66), I could find no sign of the fifty-foot boulder reported by Bailey and Weir (1932, p.441). The internal structures of the sandstone clasts and blocks include parallel bedding (in 5-10 or 3-5cm units), very common planar current bedding and more rarely, small scale slump folds and loading structures. Carbonaceous laminae are often abundant and horizons with clay clasts ($\leq$ 10cm) are also common. No in situ bedded sandstones are present in the river banks. The grey clay matrix,which is relatively common in some parts of the exposures, is identical in character to the clay observed at localities B1 and B4. The breccia is illustrated in Plate 6.12 of the east bank of Loth Burn.

Sandstones are also patchily exposed on the grassy slope between the old A9 and the Glen Loth road on the west side of the river (locality B6). These exposures consist of friable, orange-brown or yellow, fine grained sandstones, occasionally with rare fine bioclastic debris. The exposures are rather small and the sediments highly weathered. Discrepancies in the orientation of feint bedding (?) structures suggests that these outcrops are, at least in part, of breccia rather than in situ sediments. One surface gave a strike and dip value of $330^{\circ}/50^{\circ}$N but it is impossible to say how dependable this measurement was. Several of the sandstone outcrops exhibit clasts of orange-brown clay (generally 2-3cm in diameter), one laminated clay clast measuring

PLATE 6.12 Exposures of breccia facies on the east bank of the Loth River upstream of the new road bridge, locality B5. Plate 6.12B (BOTTOM) shows a large internally bedded mega-clast.

25cm across. This clay also has a kaolinite-illite composition and probably represents a stained and/or oxidised equivalent of the light grey clay observed elsewhere. Small outcrops of the same fine grained sandstone occur at the junction of the A9 and the Glen Loth road and adjacent to the cattle grid on the Glen Loth road a little above the exposures just described, where the sandstone is partly better cemented.

Locality X1 Clyne Burn, Clynekirkton (NC 893063)

Only brief descriptions of this locality have so far appeared in the literature (Judd, 1873, p.118 and Linsley, 1972, p.68). The locality is situated in the valley of Clyne Burne just below the prominent waterfall and gorge formed by the Moinian, and is some 5km south west of Allt Choll. The main exposures occur in the steep sides of a small spur on the east side of the Burn, only about 14m downstream of the Moine metasediments. The north west face of this spur consists of a chaotic, matrix-supported breccia composed of a ferruginous, orange to pale yellow, friable sandstone matrix with common to abundant ($\leq$ 75%) angular to subrounded clasts (< 1cm to 18cm in diameter) of identical, but slightly better cemented sandstone. The friable matrix is veined, dominantly well sorted except for patches of quartz pebbles ($\leq$ 2cm diameter), and is very fine grained, becoming streaky and carbonaceous towards the base of the face. At the change in slope at the base of this face ($\sim$ 1m above the pipeline) there is an irregular but apparently continuous band of light greenish-grey waxy clay up to 18cm in thickness. In places it contains carbonaceous streaks (and possibly also small sandstone clasts) and is clearly laminated in part. It is underlain (i.e. on its upstream side) by sandstone but very little is exposed. If it were assumed that this clay band was a planar, originally horizontal feature the estimated strike and dip would be about $026^{\circ}/70^{\circ}$E. The breccia is also exposed in the south west face which also consists of dirty, carbonaceous, hard sandstone (just above where the pipeline changes its direction). The latter lithology appears to be sharply demarcated and may represent a clast (at least 1m across). The down-stream (south east) face also shows a good exposure of the sandstone breccia. Most of the sandstone clasts measure 20cm or less in diameter but a few are ∿ 40cm in maximum dimension and exhibit signs of internal parallel bedding. The matrix contains occasional small, angular quartzitic pebbles (possibly Moinian?), as seen in the other faces, and also has orange-brown clay clasts. The clays are of a kaolinite-illite composition as recorded elsewhere in this facies.

Downstream of the spur small outcrops occur in the stream for about 75m (see Fig. 6.14). These outcrops consist of massive or wavy to irregularly bedded, friable, calcareous sandstones which vary in colour from pale grey-green to brownish or pink coloured when weathered. They show occasional bedding surfaces rich in medium to coarse grains and sometimes small pale green (clayey?) clasts, and at least two thin (≤ 10cm) lenticular, rubbly, conglomeratic bands were noted. These sediments are clearly different from those exposed in the small spur described above. They are predominantly fine to very fine grained and show normal sparite or poikolitic calcite cements. In thin section the rubbly bands are seen to consist of granules (and pebbles) of very finely crystalline calcite and scattered silt and sand-sized quartz grains in a matrix consisting of partially dolomitised coarse sparite and coarse silt-sized polygonal calcite. These sandstones also appear to contain more feldspar than is typical of most Kimmeridgian sediments examined from this field area. Small outcrops of these calcareous sandstones occasionally occur through the grass cover on the east side of the valley between the spur and the road. In the stream they appear to have a high dip which is consistently in a downstream direction. These sediments are of an unknown age and their precise relationship with those exposed in the spur is also a matter of speculation. The sandstone breccia, however, is so similar to its Kimmeridgian counterparts further eas

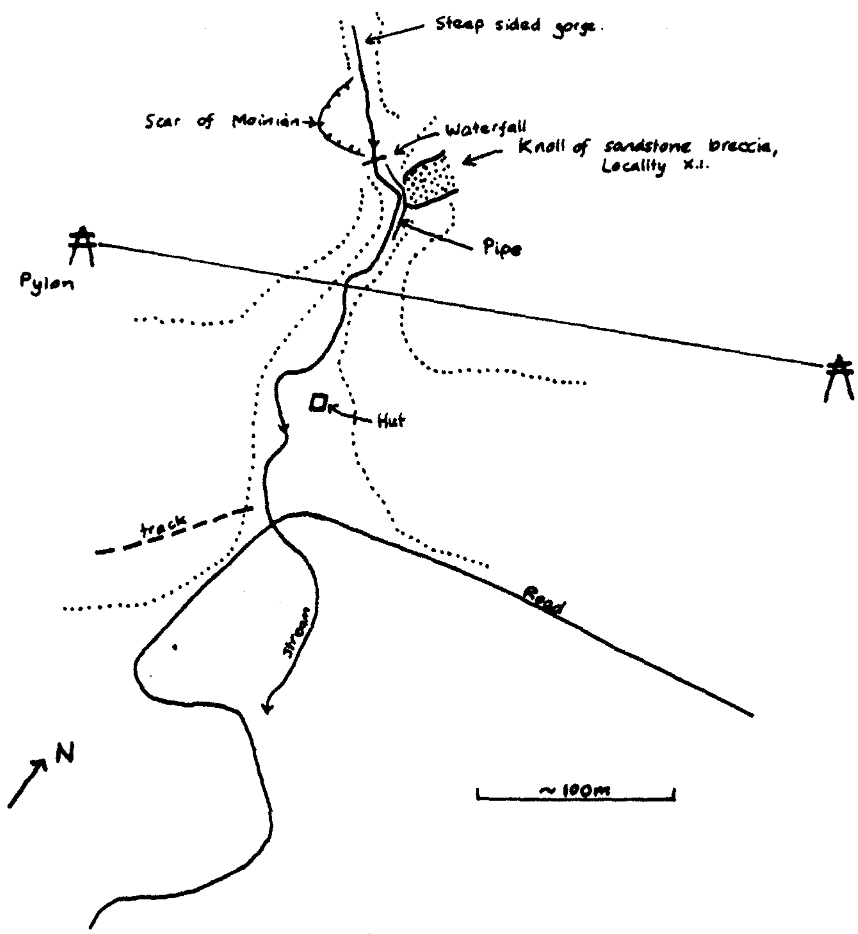


Fig. 6.14 Sketch map of locality X1, Clynekirkton.

there seems little doubt that it is probably more-or-less the same age - even though the only fossil reported from this locality is a non-diagnostic crinoid stem (Lee, 1925). The sandstones exposed in the Clyne Burn itself are problematic and do not appear to be comtemporaneous with the breccia, although the difference in facies may be misleading. The sandstone breccia exposed in the small spur appears to have been caught and down-faulted between (at least) two branches of the Helmsdale Fault which must lie immediately to the north (see Fig. 6.1).

Locality X2 'Sput Burn' (ND 047166)

This locality is not actually situated within Sput Burn, but occurs in the cliff of the raised beach 200m north east of where Sput Burn crosses the cliff line and falls to beach level (about 1km east of Navidale). The locality has been previously described by Lee (1925), Bailey and Weir (1932, p.441) and Linsley (1972, p.56-58). The outcrop consists of a very ferruginous, strongly fractured and indurated quartz sandstone with siliceous (and in part clayey-altered?) veins. The top of this particular small outcrop passes gradually (almost imperceptibly) upwards into iron stained sand of the raised beach - indicated by the presence of granite pebbles. It is part of a thin slice of sandstone sporadically exposed between here and Sput Burn which was evidently caught between two branches of the Helmsdale Fault (q.v. Bailey and Weir, 1932, p.441). Part of this strip consists of bedded sandstone and part of chaotic sandstone breccia. Bailey and Weir (1932, p.441) report the presence of casts of marine fossils and in accordance with Lee (1925, p.109) considered this sandstone to be a probable lateral equivalent of those exposed in the Lothbeg area.

Locality B7 Allt na Cùile (NC 939094).

This locality occurs in the small cliff of an incised meander on the west side of the Allt na Cùile approximately 300m south of the road bridge (about 20m downstream of the small ford unmarked on the 1:25000 sheet). The total exposure is only about 2m high and grades into raised beach sand with granite pebbles at its very top. The outcrop consists of friable orange-brown quartz sandstone (pale brown to white on fresh surfaces). The lower part is apparently bedded (although it could be a large block) and the upper part consists of a matrix-supported sandstone breccia containing angular clasts of better cemented sandstone ($\leq$ 10cm diameter) and quartz pebbles ($\leq$ 3cm). A few very small inaccessible outcrops of similar sandstone occur in the undercut banks of the stream north of this locality.

Locality B8 Lothbeg Farm (NC 942097)

This locality represents the scattered outcrops around the two low hills situated roughly midway between Allt na Cùile and Loth Burn and some 500m SSW of Lothbeg (see Figs. 61 and 6.11). The hills are clearly visible on aerial photographs of the area, standing proud of the 'flat' cultivated fields on either side. The exposure is limited to several small areas projecting through the grass cover. These small outcrops suggest that the hills are composed entirely of veined, light grey, fine grained quartz sandstones. Owing to the small size of the exposures little can be said about the internal organisation of these sandstones other than that they appear to be bedded (with gentle northward dips) and that no breccia horizons were observed.

Locality B9 Lothbeg (NC 941093 to NC 944096)

This locality represents the cliff of the raised beach which forms the railway cutting at NC 941093 (location of section B1), then turns inland before running north eastwards for about 550m until the cliff line is dissected by a small un-named stream at NC 944096. Yellow to orange weathering, thinly bedded, bioturbated sandstones like those in the upper part of section B1 (with which this locality is in lateral continuity), can be followed eastwards along the cliff for about 75-100m. After this the sandstones change in character, becoming more massive and grey or slightly pinkish-brown weathering. From about half way along the cliff the latter sandstones, which are very fine to fine grained, moderate to well sorted and vary from sparsely to strongly veined, contain subordinate interbeds of laminated, highly carbonaceous, shaley, very fine sandstone (30-35cm thick). About 100m from the east end of the cliff (at NC 943095) the major part of the outcrop consists of at least 50% of this carbonaceous shaley sandstone (one individual, uninterupted unit being 3m thick), although the base of the outcrop consists of the usual, more massive, veined, grey-brown variety. Eastwards the carbonaceous layers appear to rapidly die away and at NC 944096, on the west side of the valley of the small stream, the outcrop consists of relatively massive, non-carbonaceous, veined, grey sandstone. Although the exposure along the foot of this cliff is more-or-less continuous with only minor gaps, it is not perfect and is often obscured by vegetation. It is not possible to stand back from the cliff and follow the beds along strike with the eye in order to fully assess the lateral variation in facies. The sandstones at locality B8 are situated stratigraphically above, and to the north of those exposed in this cliff line.

Locality B10 (NC 945096)

This locality is separated from the east end of locality B9 by the small un-named stream (see Fig. 6.11). A total of 4.2m of sediments are exposed in a small outcrop at the top of the cliff (here little more than a steep slope - see Plate 6.13a). The lower 2.7m consist of carbonaceous grey shales with thin sand stringers ($\leq$ 1cm), the remainder of sandstone. The contact between the two appears to be erosional (although the outcrop is too small to confirm this) and the base of the sandstone is rubbly and contains sandstone clasts $\leq$ 30cm in diameter (see Plate 6.13b). The sandstone is pale brown to grey coloured, fine grained and clean, the top passing through a weathered zone into a sandy

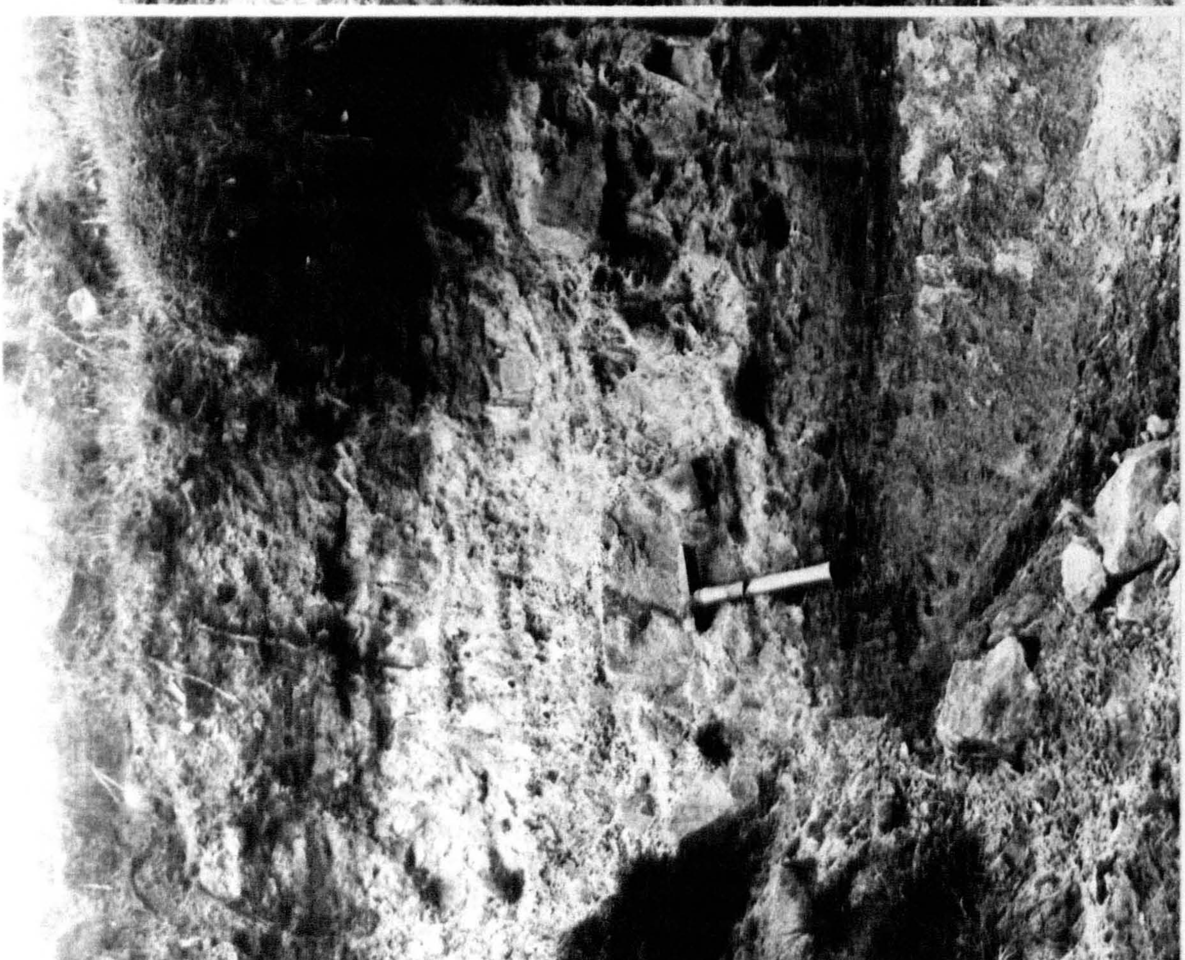


PLATE 6.13 Exposures at locality B10 (see Fig. 6.11)

6.13A Exposure of sandy shales

6.13B Contact of sandy shales and overlying bouldery sandstone (Just out of view at top left of 6.13A) soil (relict beach deposit?). A strike and dip of $267^{O}/14^{O}$N was measured at the base of the shale.

Localities B11 (NC 946096) and B12 (NC 947097) Lothbeg

Locality B11 occurs just on the south west side of the railway cutting where the cliff line rejoins the railway (approximately 30m ESE of locality B10). The outcrop consists of about 5m of massive, grey to light brown, fine grained, veined sandstone with a little scattered carbonaceous material. This lithology can be followed north eastwards in the railway cutting. Locality B12 is located at the north east end of the railway cutting before the cliffline once more swings inland and the outcrop becomes obscured by grass cover (see Fig. 6.11 and Plate 6.14a). The locality has been briefly described by Lee (1925, p.106) and Linsley (1972, p.67). A measured section (equivalent to Lee's) was recorded from the north east end of this outcrop and is shown in Fig. 6.15. The top 3m of the exposure (the prominent face in Plate 6.14a) consists of fine to medium grained, pale brown to grey, veined sandstone with rare bioclastic debris and common rounded to subrounded, white to pink quartz pebbles ($\leq$ 25%) ranging in size up to 1.5cm in diameter. In parts of the outcrop poorly defined, inclined planar bedding is visible but there is little other structure; the pebbles are dispersed in the sandstone matrix and there are no indications of graded bedding. At the very north east end of the outcrop the pebbly sandstone contains dispersed subangular to subrounded clasts of yellow, cross bedded, fine to medium grained sandstone varying from 2-45cm (see Plate 6.14b of the north east face). The clasts do not show any pronounced organisation but there does appear to be a tendency for the largest clasts to occur only at lower levels, possibly suggesting crudely graded units (q.v. Plate 6.14b). One clast of shaley, carbonaceous sandstone was noted. Below this upper 3m unit there is 1.5m of carbonaceous, silty, sandy, shales with thin, uncemented bands of sand ($\leq$ 5cm thick), which is poorly exposed in the grassy slope

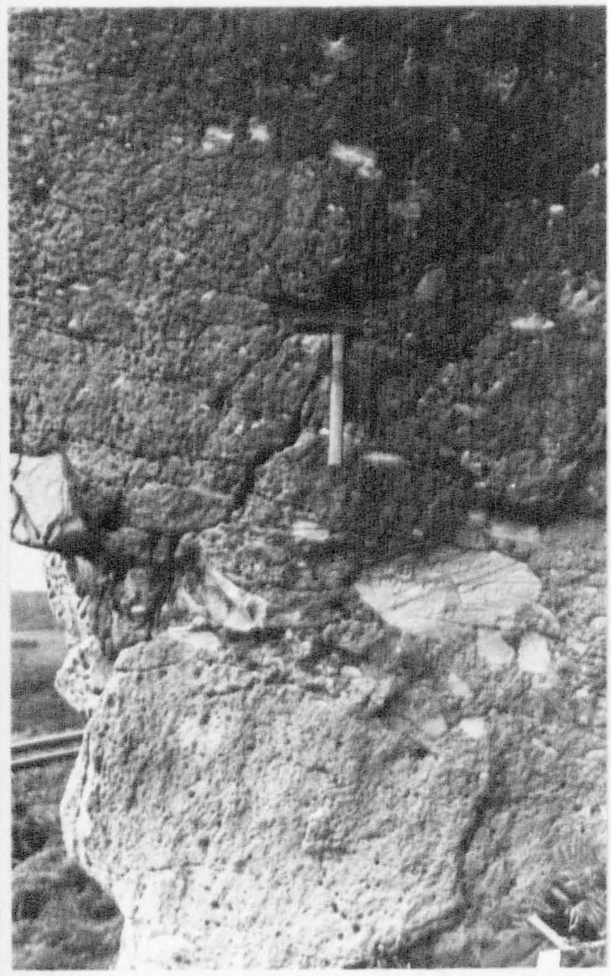


PLATE 6.14 Generally massive sandstone exposed at locality B12.

6.14A General view of outcrop

6.14B Detail of right hand side of outcrop (looking west) showing dispersed clasts and pebbles

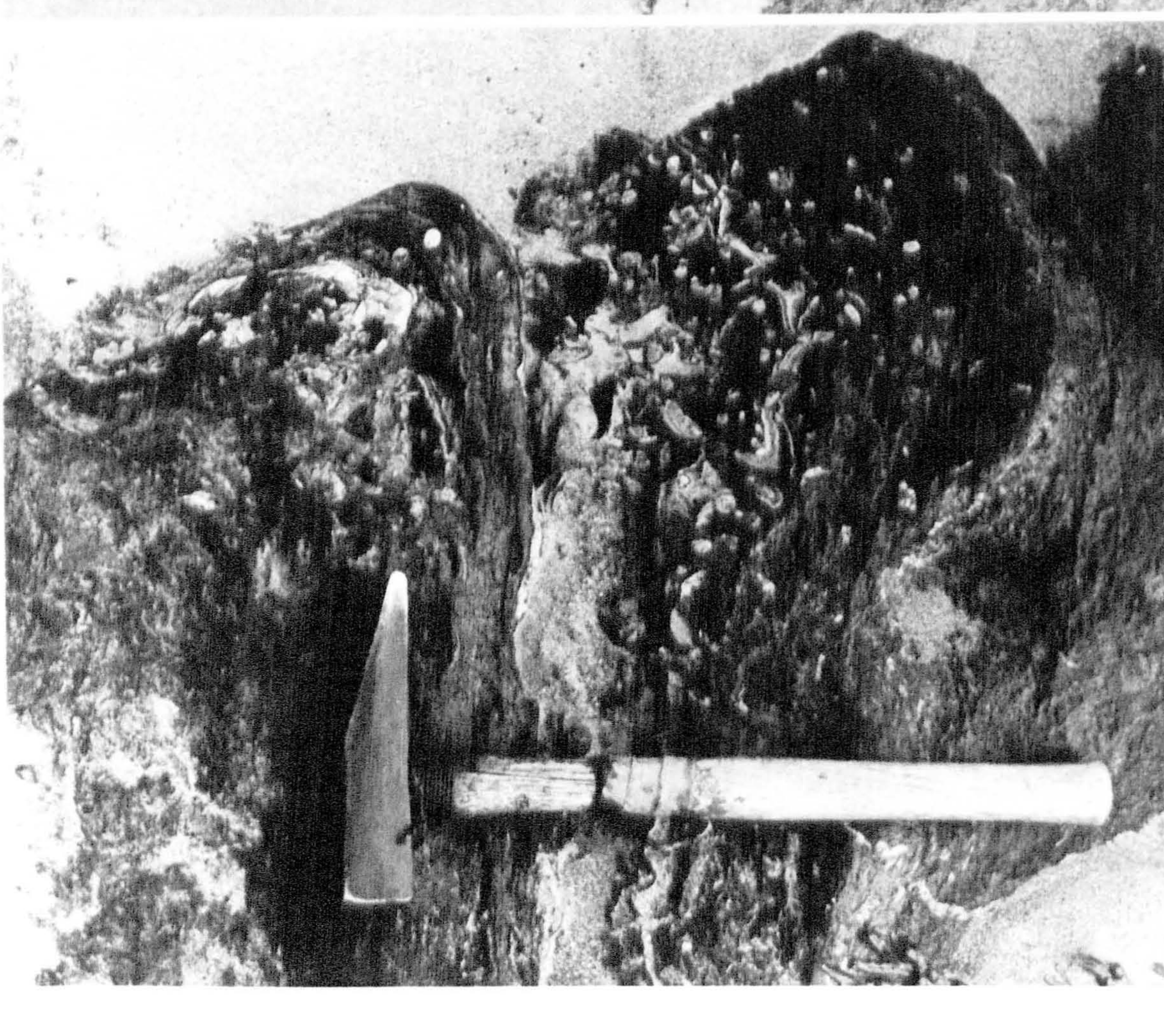



PLATE 6.15 Bioturbated shaley, carbonaceous sandstones exposed on Lothbeg beach, locality B13. Burrows are vertically restricted and apparently of Planolites affinities

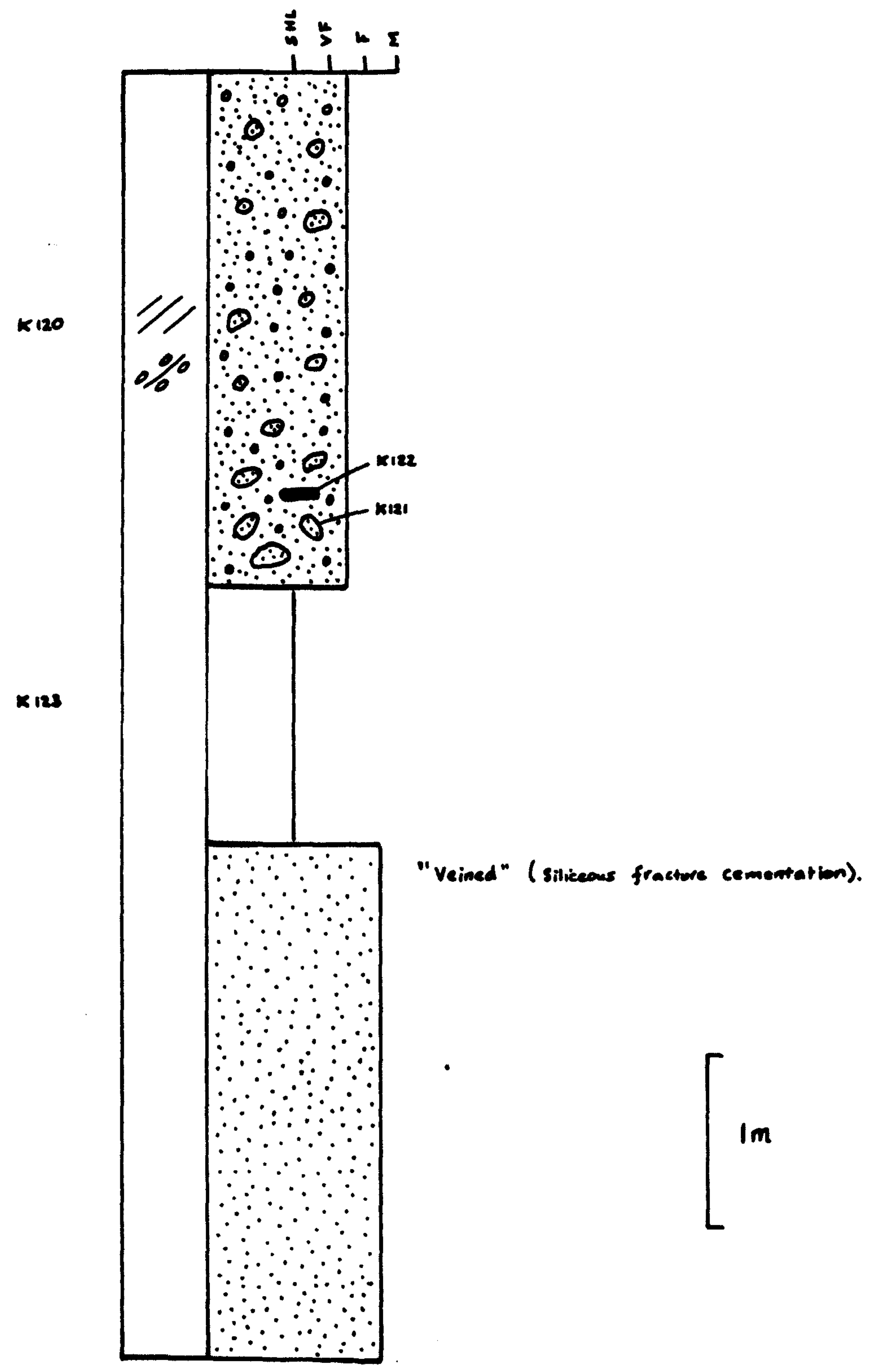


Fig. 6.15 Section B12 Lothbeg.

visible in Plate 6.14a. This is underlain by 3m of pebble-free, fine grained, pale brown, veined quartz sandstone, roughly along strike and at the same level as locality B11 (which occurs about 100m to the south west). The upper pebbly sandstone unit continues north eastwards from B12 as scattered poorly exposed outcrops to about NC 949098 (approximately 200m along the railway).

Locality B13 Lothbeg beach south (NC 946094)

This locality is situated on the west side of the low lying promontory that is exposed at low tide just under 1km west of the mouth of Loth Burn, and about 200m due south of localities B11 and B12. The outcrop consists of some 11.3m of gently dipping highly carbonaceous, shaley, very fine grained, silty, friable sandstones (strike and dip $234^{\circ}/11^{\circ}N$). Although the sandstones are predominantly shaley and carbonaceous, they contain lenses of relatively pure very fine to fine grained sandstone, generally $\leq$ 10cm in thickness, but one prominent band (containing only rare, 1mm laminae of carbonaceous material) measures 30cm. The amount of carbonaceous material shows great lateral variability. Although the sandstones are generally laminated, their top surfaces often show conspicuous ?Planolites bioturbation (Plate 6.15) which is of a very limited vertical extent and often hardly noticeable in side view. This locality occurs at the shoreward base of the promontory; the outer (mostly southerly part) appears to consist mostly of veined, grey quartzitic sandstone.

Locality B14 Lothbeg beach south east (NC 950097)

This locality represents the low-lying, poorly exposed outcrops between the low and high tide mark that occur between the mouth of Loth Burn and the small promontory to the west (i.e. locality B13). The sediments are largely obscured by sand and seaweed and it is impossible to give a measured section. The sequence is dominated by laminated, carbonaceous, shaley, fine grained, friable sandstones with interbeds of pure or sparsely carbonaceous, friable sandstone (up to 50cm thick but mostly ∿ 10cm). The sequence is very like that at locality B13 and is similarly bioturbated. The sequence appears to include at least a few low angle dislocation surfaces like those observed at Kintradwell. See locality B21.

Section B15 Loth Burn railway bridge (NC 952099)

This locality occurs in the steep cliff on the east bank of Loth Burn immediately to the south of the railway bridge. It is the type area of the so-called 'Loth River Shales' and has been briefly described by Lee (1925, p.106), Arkell and Callomon (1963, p.242) and Brookfield (1976, p.182). During the present study a measured section of 21.85m of sandstone (43%) and shale (57%) was recorded as shown in Fig. 6.16; the locality is shown in Plate 6.16a. The shales are highly carbonaceous and contain common laminae and occasional beds ($\leq$ 1-5cm) of uncemented very fine to fine grained sand which often increase in number and thickness toward the bases of the prominent sandstone beds. They are sometimes richly fossiliferous (containing ammonites, belemnites and pelecypods) but are totally decalcified, leaving only impressions rather than true body fossils. The sandstones vary from pure to highly carbonaceous and shaley, are generally strongly veined and are predominantly very pale brown to light grey in colour. At the base of the section they are friable, uncemented and highly porous (with < 53% voids in thin section), but the two upper (1.8m) beds are partly calcite cemented and partly decalcified producing the sphaeroidal 'concentric weathering' pattern noted by Lee (1925, p.106). The sandstones are mainly fine grained rather than medium grained as reported by Brookfield (1976, p.182). Because of the common presence of carbonaceous laminae it can be seen that the ubiquitous 'veining' results from lines of quartz cementation along small fractures and faults which show small displacements of $\leq$ 4cm. No bioclastic debris was noted in the sandstones

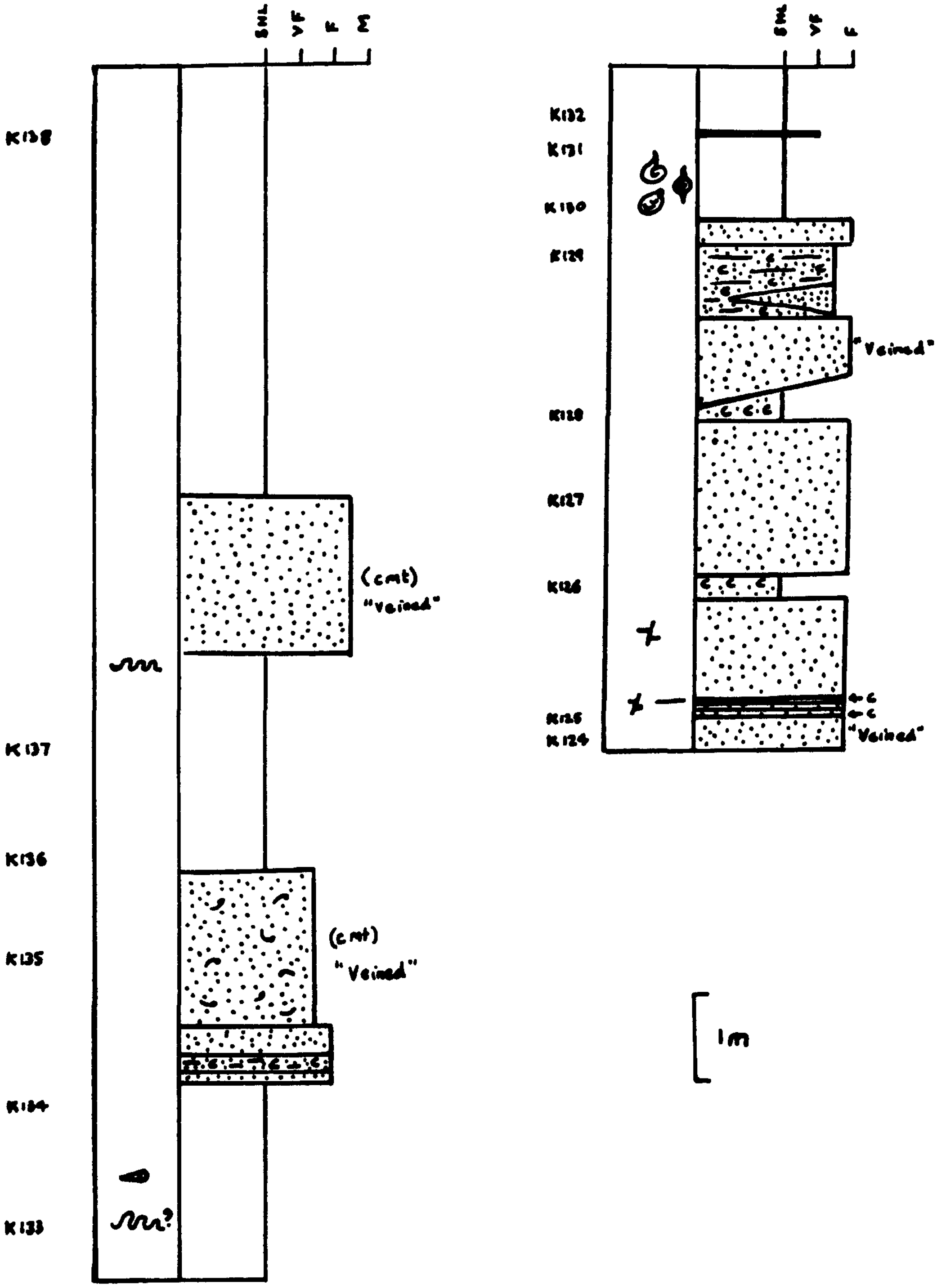
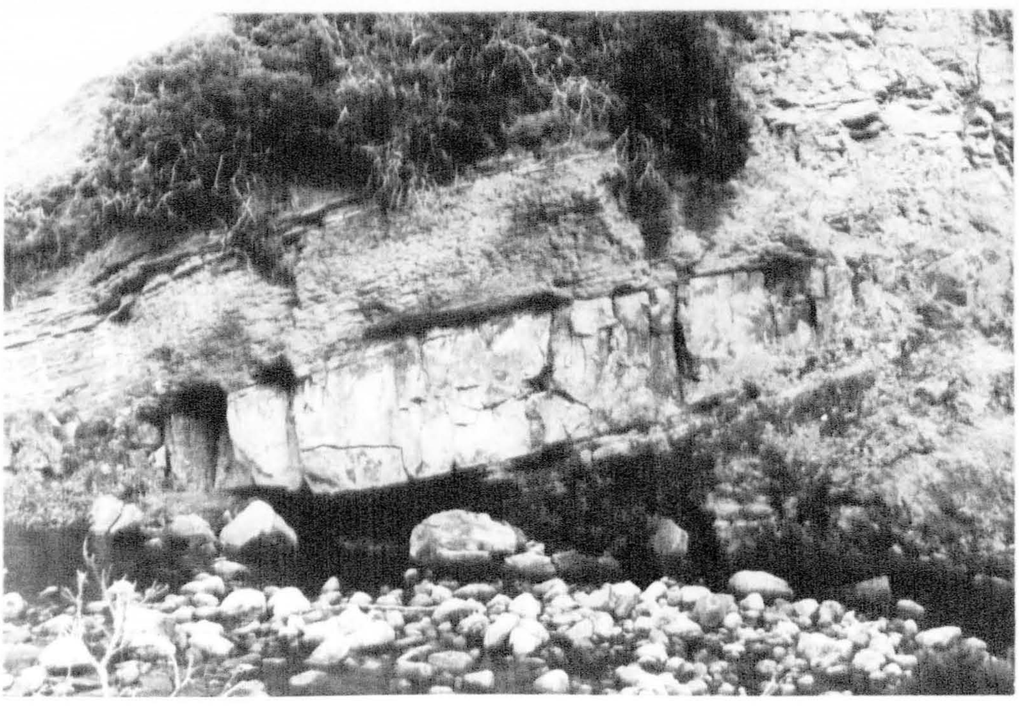


Fig. 6.16 Section B15, Loth Burn railway bridge.

PLATE 6.16 Shales and lenticular sandstones exposed near mouth of Loth River.

6.16A Main exposure on east bank (locality B15)

6.16B Lenticular sandstone (locality B16); hammer for scale (centre)



during the field description, but one thin section of sparite cemented sandstone showed small amount ($\sim$ 5%) of echinoderm, pelecypod and foraminiferal debris. Brookfield's comments that "these beds show extensive synsedimentary deformation, including recumbent folds" (Brookfield, 1976, p.182) is rather misleading. Some rotated shale blocks ($\leq$ 40cm) occur in the third shale unit down but I was not convinced that these were not a result of 'modern' slope failure; the only other structures are common microfaulting, and in the second shale unit down, the presence of small scale sedimentary overfolds picked out by thin sand laminae. This can hardly be described as 'extensive synsedimentary deformation'. Plate 6.16a demonstrates the fact that all the sandstone units are lenticular bodies which have apparent widths of between 11 up to 19+m.

<u>Localities B16-20 Loth Burn, subsidary exposures (Nc 952099 to NC 948100)</u>

Locality B16 occurs on the east bank of Loth Burn on the north side of the railway bridge about 50m upstream of locality B15. None of the sandstone units observed at B15 can be seen in the river cliff, although this section is at least partly laterally equivalent. The outcrop consists predominantly of sandy carbonaceous shales but has a prominent lenticular sandstone body about 2m in maximum thickness whose top 10cm is laminated and rich in woody debris (see Plate 6.16b). All other exposures between here and the road bridge (q.v. localities B5 and B6) occur in the west bank of Loth Burn. Locality B17 is the exposure in the west bank of the river on the south side of the railway bridge directly opposite locality B15. The exposure is much poorer than on the opposite bank but a rough section can be measured. Exposed in the river bank at the base of the section is a 2.5-3m thick unit of veined, light grey, friable, very fine grained sandstone which is overlain by 70cm of shaley, carbonaceous laminated sandstone. This is followed by another 2.5-3m unit of massive, sparsely carbonaceous, friable, very fine grained sandstone. It contains occasional concentrations of carbonaceous debris and a few lenses of laminated, shaley, carbonaceous sandstone suggesting that it may, in part, be an almagamated unit. This sandstone gives way to about a 2m gap in outcrop (probably carbonaceous shales and shaley sandstones) and is then followed by a further 2m of massive, veined, light grey, friable sandstone before the outcrop on the south side of the bridge is lost. The section is shown diagrammatically in Fig. 6.17; it is presumably laterally equivalent to the lower sand part of section B15 in Fig. 6.16.

The only outcrop on the west side of the river immediately north of the railway bridge is a thick sandstone exposed high on the side of the valley. This sandstone descends with the dip and is eventually exposed in a meander cliff at locality B18 (NC 950099) stratigraphically above localities B15-B17. The meander cliff consists mainly of a 2.5m thick, fine to very fine grained, pale yellowish grey, veined sandstone. Although no petrographic samples were taken, the occasional presence of sphaeroidal weathering suggests (by analogy with section B15) that this sandstone is partly calcite cemented and suffering decalcification. It is overlain by about 2m of carbonaceous sandy shale before outcrop is lost below adjacent river deposits. Another, approximately 2m thick, yellow-grey, friable, veined, fine to very fine grained sandstone unit is exposed about 100m further upstream in the river bank below the terrace of the raised beach (locality B19, NC 950100). The final outcrop, locality B20, occurs in a meander cliff at NC 948100, where at least 3m of friable, veined, pale brownish grey, fine grained sandstone is exposed. No other exposures occur in the banks of Loth Burn until locality B5.

<u>Locality B21 Lothbeg Point (NC 955098 to NC 961094)</u>

The outcrops which make up this locality occur between the mouth of Loth Burn and the tip of Lothbeg Point (as exposed at low tide), and between the low tide mark on the south west of Lothbeg Point and the inland side

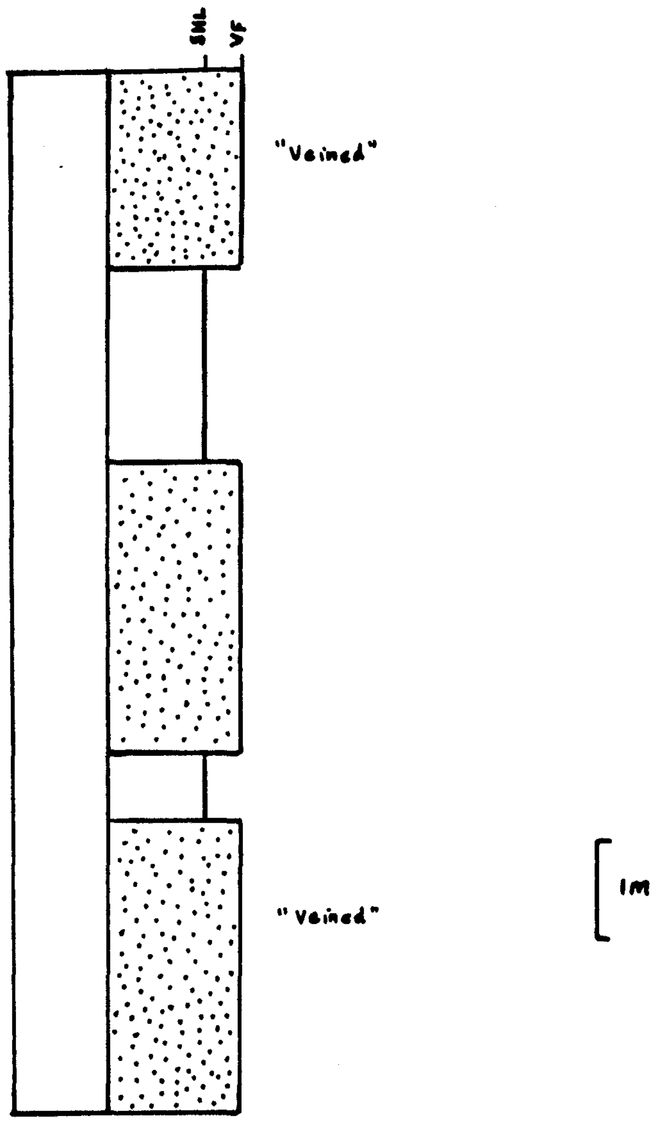


Fig. 6.17 Section B17 Loth Burn.

of the grassy track which runs from the top of the Point to the mouth of Loth Burn (see Fig. 6.11). The outcrops are predominantly sandstone and appear to all represent various parts of a single sandstone body. The outcrops in the vicinity of the Point itself (locality B21a, NC 959096) are of massive, pure, light grey, veined, fine to medium grained, moderately sorted quartz sandstone (see Fig. 6.11). The sandstones are uncemented and very porous (containing up to 30% void space in thin section). The fairly common presence of small shell casts indicates that they have been decalcified. Small amounts of coarse to very coarse grains and rare granules occur scattered throughout. Although any sign of bedding is generally rare at this particular locality, a strike and dip of $113^{\circ}/14^{\circ}$N was recorded on a fairly reliable looking surface on the south west side of the Point. South eastwards from the exposures at the base of the cliff thin, shaley, carbonaceous bands appear in the sandstone (see Fig. 6.11). These bands are generally 2.5cm or less in thickness and become more common south eastwards (along the rocky promontory exposed at low tide) where the sandstone also becomes less massive, better bedded and often laminated. The thin carbonaceous bands are often microfaulted as shown in Plate 6.17a (locality B21b, approximately NC 960095, see Fig. 6.11).

Neves and Selley (1975) have previously described this locality in the following terms: "As seen at Lothbeg Point, for example, the Allt-na-Cùile sandstone occurs in massive graded units about a metre in thickness. The base of each unit is irregularly channelled and it fines upwards, through structureless medium grained sand into fine grained laminated carbonaceous sand" (p. JNNSS/5-11). I cannot reconcile this description with my own observations of this locality. Although I did observe one approximately 1m thick unit of relatively poorly sorted sandstone overlain by a thin ($\sim$ 2cm) carbonaceous horizon (locality B21c on the west side of the Point at NC 959095, see Fig. 6.11), this particular band did not show grading (except with respect to carbonaceous material at the

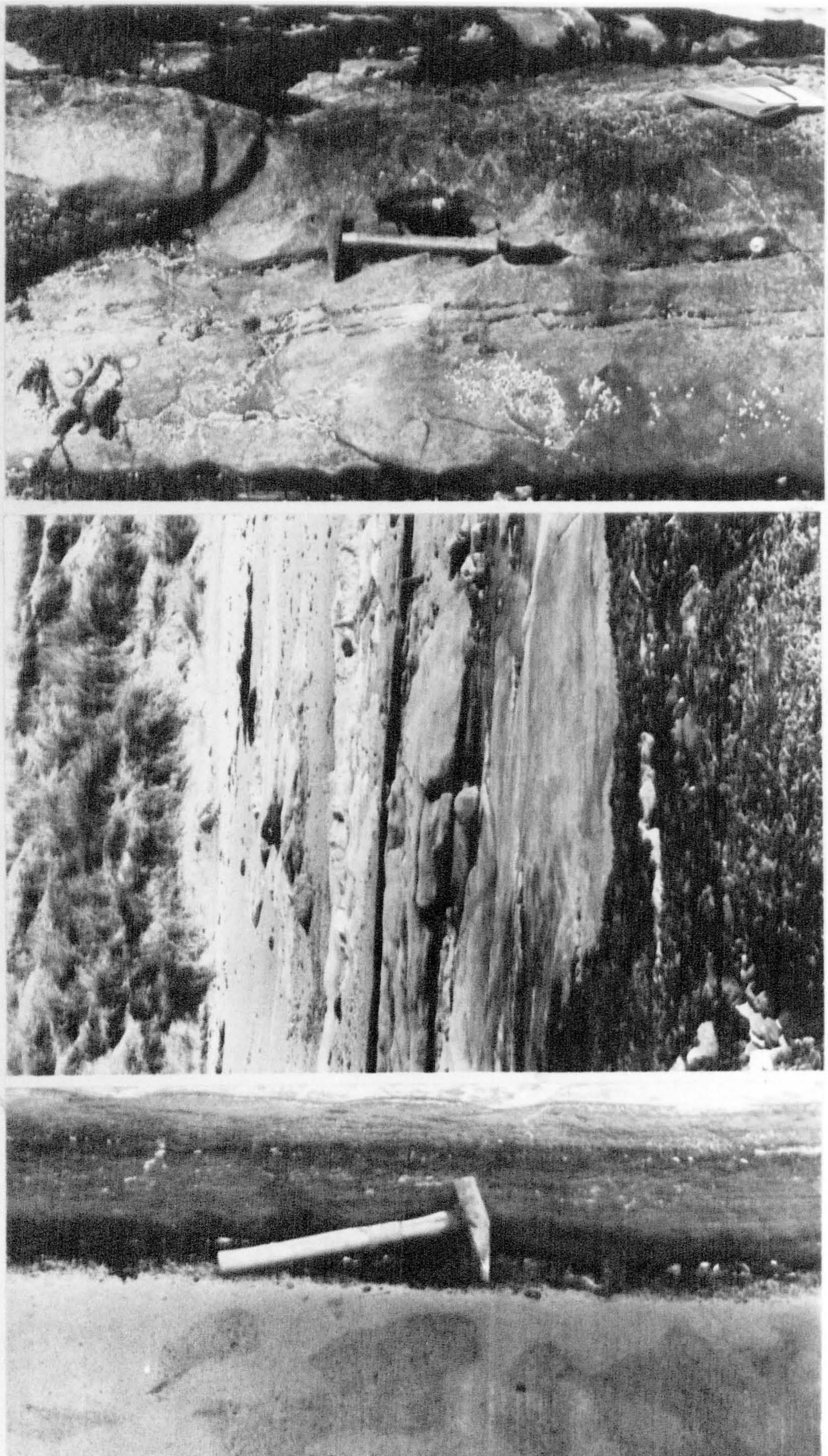


PLATE 6.17 Sedimentological features of the marginal facies of the Lothbeg Point Sandstone body

6.17A Microfaulted carbonaceous laminae in generally massive sandstone (locality B21b)

6.17B Thin interbedded highly carbonaceous shales (locality B21e)

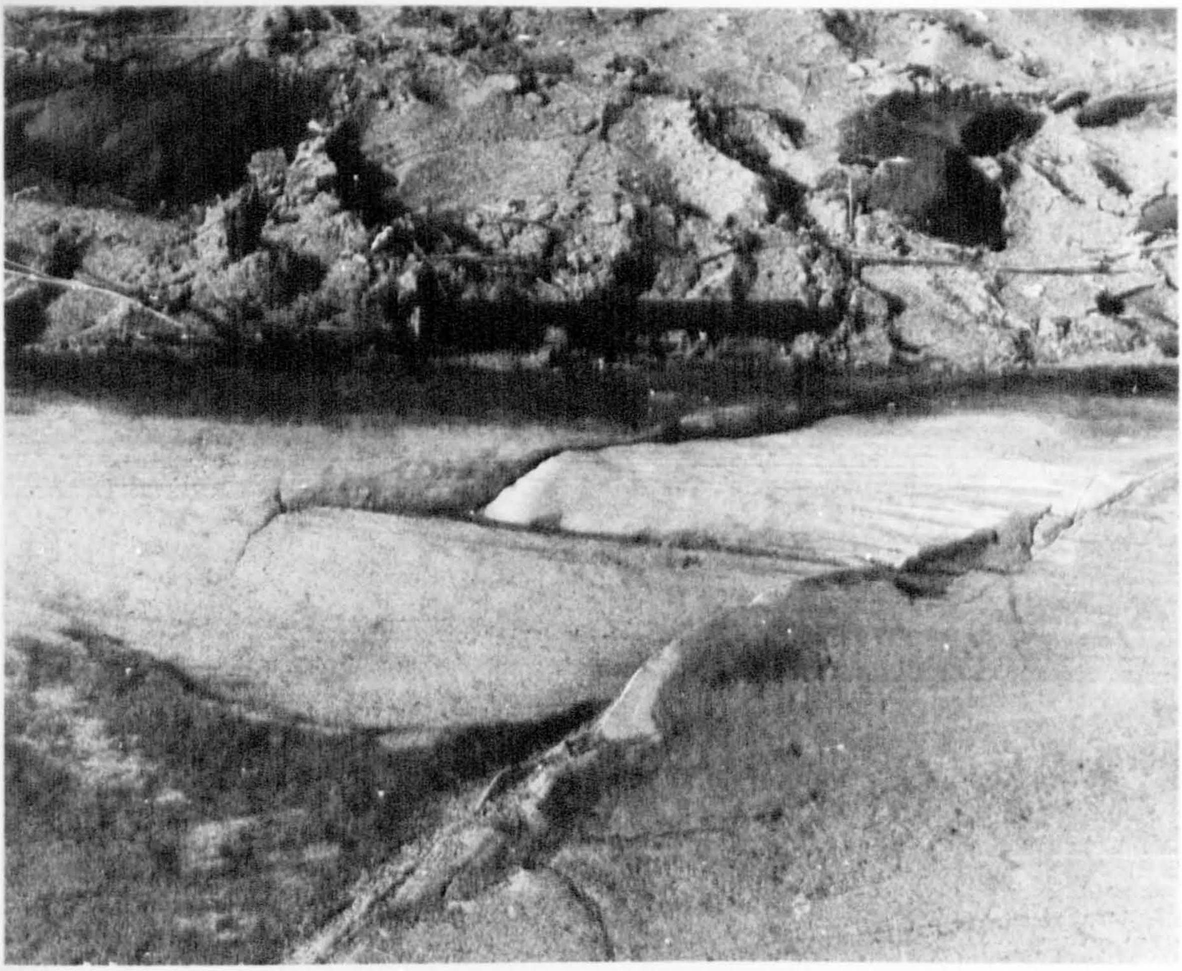
6.17C Load casting of carbonaceous sandstone below thin shale unit (locality B21e) top), nor an erosive base and I could find no other which even vaguely appeared to fit Neves and Selley's description. Although I cannot refute their account I must call to attention the fact that it is a highly misleading representation of the locality as a whole.

The massive, non-carbonaceous variety of this sandstone (as at locality B21a above) forms most of the outcrop along the back of the beach northwest of Lothbeg Point (i.e. at the south west end of Crackaig Links). The sandstone ranges from about 2m below the level of the sandy beach up to the top of the Links (i.e. to about the 50 foot contour) without any apparent interruption. Above the beach, outcrop is sporadic and mainly confined to the sides of the track running between the Point and the mouth of Loth Burn; these outcrops are, however, more-or-less identical to those at beach level. The highest exposure occurs at locality B21d (NC 957097), above and to the west of the large buried bunker, and about 50m south west of the small installation at the top of the Links (these two particular second world war constructions are not marked on the 1:25000 sheet). At this locality the sandstone is light grey, fine-medium grained, moderately sorted and veined. It contains a few 1mm thick, carbonaceous laminae but has no carbonaceous bands as observed elsewhere. Scattered through the outcrop are small ($\sim$ 5mm) casts of shells whose symmetry suggests brachiopods.

It is only in the beach outcrops near the mouth of Loth Burn (at locality B21e, NC 955097) that any significant differences are seen in this sandstone. In this area the sandstone (below the high tide mark) shows more indications of bedding than elsewhere and contains several thin ($\leq$ 10cm), lenticular, carbonaceous sandy shale bands which extend up to about 9m along strike (see Plate 6.17b). Four such bands were observed evenly spaced within about 3m of sandstone. These shales occasionally contain small sand injections ('dykes'), while the adjacent sandstones sometimes show loading structures as in Plate 6.17c. The sandstones in which the shales are intercalated are fine grained, contain scattered carbonaceous material (which often appears to be coarser and more abundant immediately below the shales), and sometimes exhibit feint parallel lamination. The sandstones also show abundant 'veins' resulting from cementation along anastmosing fractures and microfaults; where carbonaceous material has been caught up in these fractures it has occasionally been streaked-out parallel to the slickenslides. Nearer to the low water mark, some of the sandstones show convoluted carbonaceous, laminated horizons, apparently produced by loading effects. Just south east of where the lenticular shales are exposed, another beach outcrop (partly obscured by loose sand) exhibited good low-angle cross lamination (see Plate 6.18). The outcrop appears to represent the cross-laminated infill of a channel structure (about 4m wide and 70-80cm deep) which was cut down into more massive, carbonaceous sand. The latter contains several carbonaceous haloes which appear to be carbonaceous-rimmed burrows (Planolites or Ophiomorpha, see Plate 6.18b). This was the only example of undisputable cross-lamination observed in any of the Lothbeg Point sandstone localities.

Behind the small bunker at locality B21f (NC 955098, on the east side of the mouth of Loth Burn) there is a 4m outcrop of light grey weathering, sulphurous, carbonaceous, laminated sandy shales. They contain common bands ($\leq$ 1.5cm) of very fine sand and are fossiliferous (ammonites, belemnites and small pelecypods - probably Buchia) but totally decalcified. The shale is also highly disturbed and shows common dislocation surfaces and small scale sedimentary overfolds like those noted at locality B15, with which this exposure is more-or-less laterally equivalent. The presence of shales at this point, and of the sandstone at locality B21d to the south east at a topographically higher level, indicates that the Lothbeg Point sandstone body must be lenticular since its base appears to remain approximately at the same level along strike (see Fig. 6.11). This accounts for the somewhat different character of the sandstone at localities B21b and B21e, which must be situated near

PLATE 6.18 Previously undescribed cross-bedding and channeling near base of Lothbeg Point sandstone body (locality B21e). Carbonaceous (dark) haloes visible in 6.18B appear to be original features.




the lateral margins of the body and therefore represent a facies transition into the shales on either side. Owing to the lack of continuous exposure it is possible that the outcrop at locality B21d may be separated from that at the beach by a thin shale unit, but it is also clear that if such a shale exists it is cut out (e.g. by the amalgamation of two sandstone bands) at least 200m north west of Lothbeg Point. The maximum possible dimensions of the Lothbeg Point sandstone body are about 58 x 850m.

The sequence below the Lothbeg Point sandstone body is poorly exposed near the low tide mark on the beach between the mouth of Loth Burn and Lothbeg Point. The outcrop is mostly covered by sand, seaweed or water, and consists of about 3m of carbonaceous shale with thin, uncemented sand laminae and occasional lenticular beds of friable sandstone up to 1m in thickness. The sequence is in part bioturbated and shows common microfaulting and loading and injection structures and is very similar to that exposed on the opposite side of Loth Burn at localities B13 and B14, although the sandstone beds are somewhat thicker.

Section B22 Lothbeg Point north east (NC 961095) mutabilis zone

Section B22 is a thick shale sequence (42.32m, only 4% sandstone) exposed between the high and low tide mark north east of Lothbeg Point (see Fig. 6.11 and Linsley's map XV). There is a small gap in exposure between the top of the sandstone at Lothbeg Point and the first outcrop at the base of this section (representing about 8.5m of missing exposure when compensated for dip). On his map XV Linsley marks this gap as poorly exposed but consisting mainly of sandstone. Although the area in question was buried by sand during my main field season, during a brief visit to the region in April 1978 it had been swept clean by storms and was seen to consist predominantly of sandy shales (the top of the sandstone being quite sharp). Section B22 is shown diagrammatically in Fig. 6.18 and the more significant features of the sequence

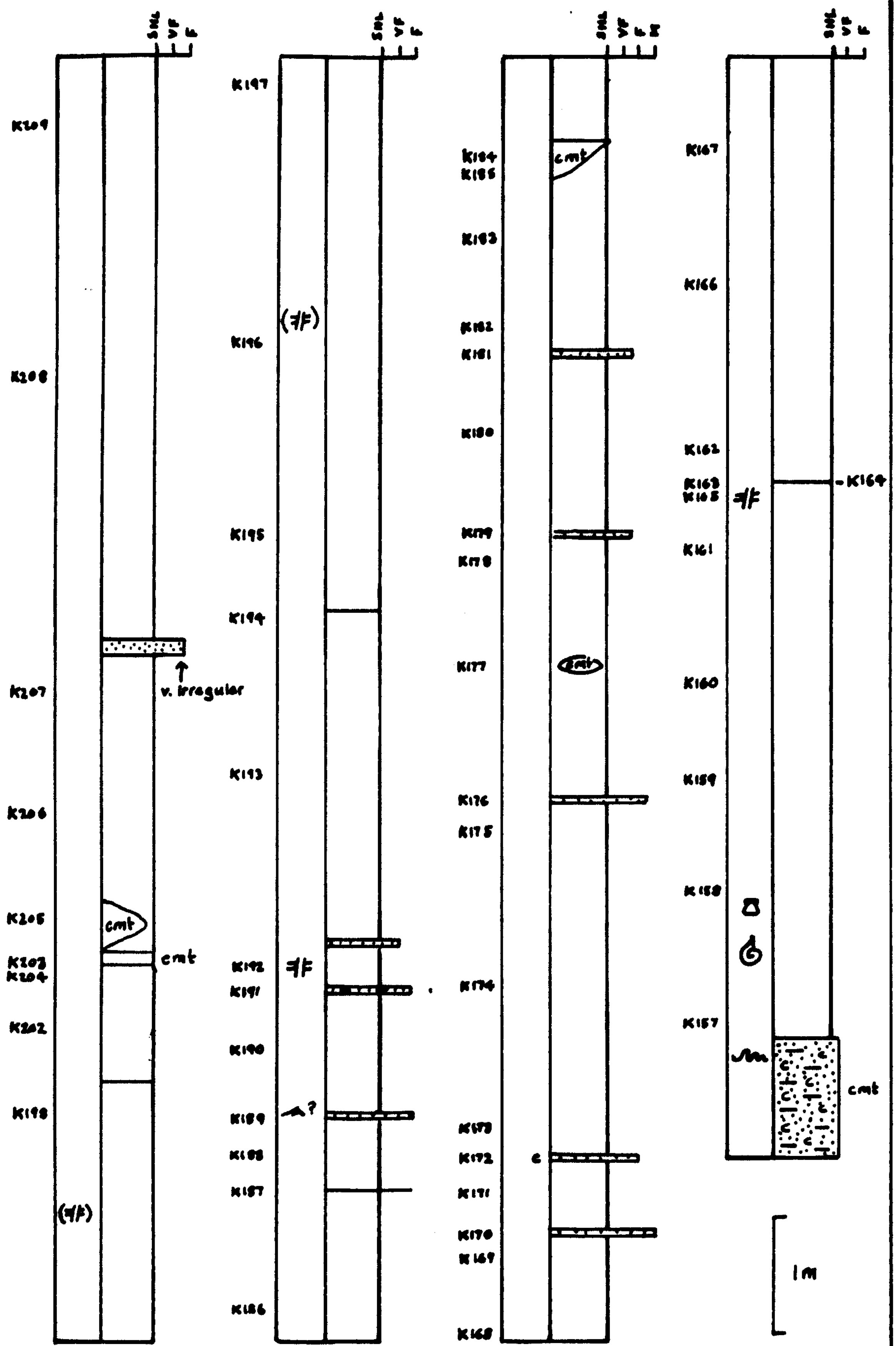


Fig. 6.18 Section B22, north east side of Lothbeg Point.

discussed below.

The shales in this section are dark grey to black, carbonaceous and fairly fossiliferous, containing ammonites, small pelecypods (including Buchia?) and occasional oysters. They vary in character from papery (as at the base) to more blocky, depending upon the degree of calcite cementation. Above the basal 8.7m, thin ($\leq$ 1cm) cemented sandstone 'stringers' are relatively common. One particularly interesting feature is the presence of a thin (2cm) band of waxy, light grey, laminated clay 5.7m from the base of the section. Subsequent XRD analysis indicates a kaolinite-illite composition identical to those clays already described from the sandstone outcrop at localities B1, B4, B5 and X1. The section also contains several lenticular masses (concretions 30-50cm x 2m) of calcite cemented shale which weather out and form the most conspicuous topographic features of this low lying outcrop; scattered smaller nodules are fairly common.

The lowest metre of the recorded interval consists of thinly bedded, calcite-cemented (lustre-mottled), carbonaceous, fine grained sandstone which shows strong internal deformation and overfolding. Above this level there is very little sandstone, only thin beds which are predominantly less than 7cm thick. These thin sandstones are all calcite cemented (about half of those examined being lustre-mottled), fine to medium grained and range from poorly sorted to dominantly moderately sorted and occasionally well sorted. They sometimes show carbonaceous laminae in their upper part but exhibit no indications of grading. Most of these beds show a variable thickness along strike (varying by a factor of 10 or more) and often pass laterally into thin 'stringers' ($\leq$ 1cm). They can only rarely be seen to exhibit slightly irregular, erosive bases. One sandstone near the top of the section shows a rippled top; the ripple crests strike at 060° (assymetric toward the SSE) on a bed with a strike and dip of 128°/32°NE. The sandstones generally contain less than 5% bioclastic debris, although the top 5mm of one band contained $\leq$ 50% shell and echinoderm fragments. The same sandstone bed also exhibited thin, microsparite-cemented siltstone lenticules with incipient cone-in-cone cements at their margins (second sandstone from the top of the sequence). Several small sandstone dykes are present in the section, some appearing as short, stubby masses (∿ 30 x 30cm) when viewed in the direction of the dip. The dykes are often carbonaceous and are all calcite cemented.

Locality B23 Crackaig Links (NC 962099) eudoxus zone

The outcrops exposed at this locality are described alongside those of the Lothbeg area because of their similar facies. Two small 'finger-like' outcrops occur to the north east of section B22 (250m and 290m respectively), from which they are separated by a zone of no exposure (see Figs. 6.11 and 6.19). These outcrops have been described by Linsley (1972, p.55) as "typical of the Allt na Cùile sandstone series" but he gives no details (note the incorrect grid reference). The location of the outcrops is shown on his map XIV of Crackaig Links. The stratigraphically lowest, most south westerly 'finger', consists of ∿ 3.5m of white to pale orange-brown, friable, 'veined', quartz sandstones which are generally fine grained but contain poorly sorted bands with common granules and pebbles of quartz ($\leq$ 6mm diameter). These sandstones are pure and non-carbonaceous while the higher, north eastern 'finger', consists of ∿ 5m of fine to very fine grained, 'veined' friable, pale grey sandstone with common 10-30cm carbonaceous bands. Some of these carbonaceous bands contain shale intraclasts (averaging 5cm but up to 10cm in diameter). The more pure, sparsely carbonaceous sandstones can sometimes be seen to have channelled down into the underlying carbonaceous bands (e.g. see Plate 6.19a). About 40m to the north east of this last outcrop, the next exposure consists of 1.8m of brownish grey, moderately sorted, fine to medium grained sandstones with subordinate carbonaceous shales, overlain by about 2m (seen) of

Fig. 6.19 Foreshore exposure at Crackaig Links (after Linsley, 1972).

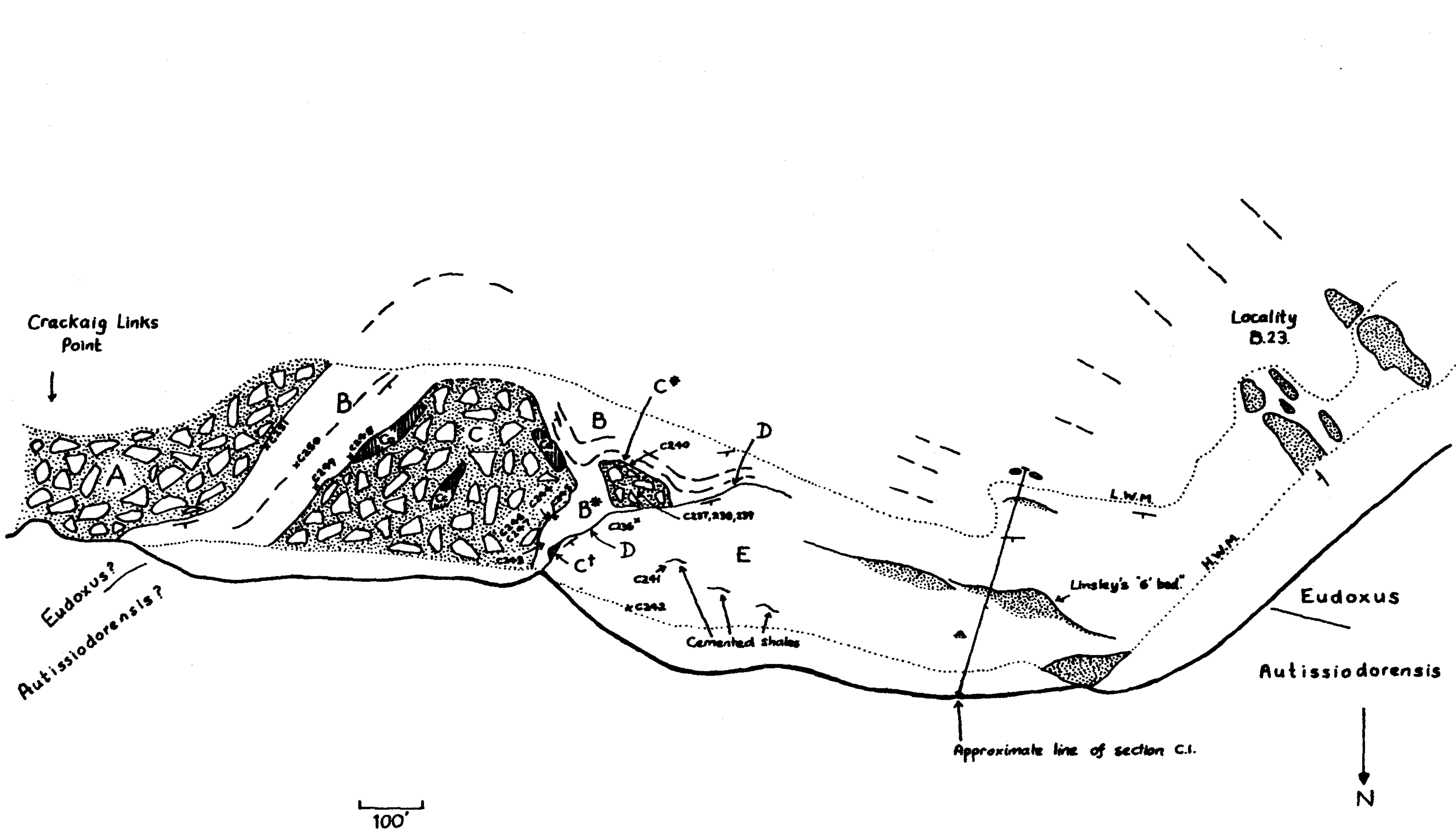
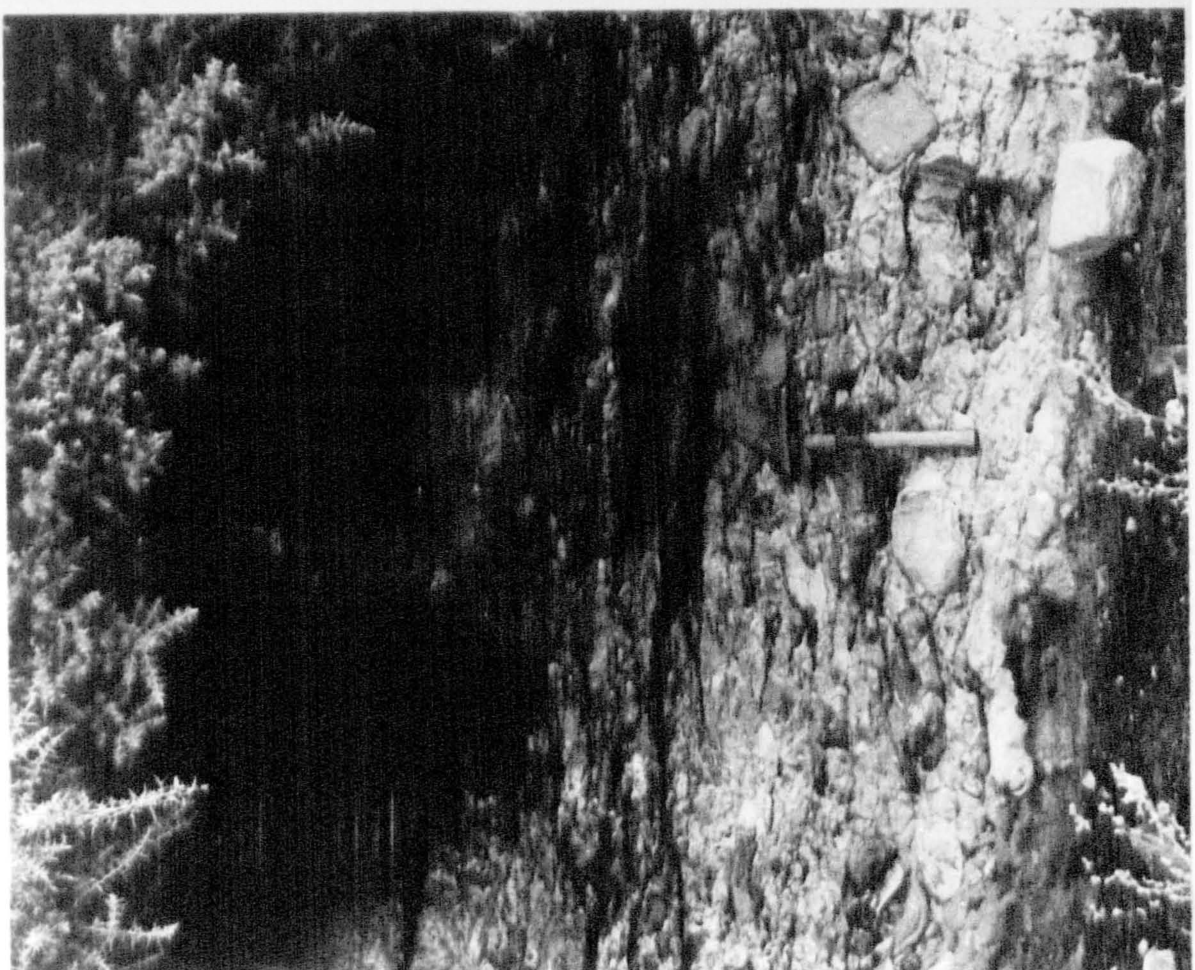


PLATE 6.19 Sedimentological features of lenticular sandstone units at localities B23 and B24 (Crackaig Links and Loth Station)

6.19A Channeling and amalgamation of sandstone beds, locality B23. Sandstones in part pebbly and shale-clast bearing.

6.19B Clast-supported breccia lens at base (?) of sandstone outcrop, locality B24. Note fining upward trend. See also Fig. 6.20.



sandy shales with subordinate, uncemented, very fine sandstone bands in the lower half. The lower (1.8m) sandstones contain common but unrecognisable fossil casts indicating decalcification. A strike and dip of $124^{o}/24^{o}$NE was recorded at this point (at the edge of the main Crackaig Links outcrops - see Linsley's map XIV).

Localities B24 to B27 Loth Station north west (NC 955102)

These outcrops occur in four small hills to the north and north west of the old (disused) Loth railway station (see Fig. 6.11). The general location of the two largest outcrops is indicated by the 50 foot contour on the 1:25000 sheet and all four are clearly visible on aerial photographs of the area. Three of the localities were briefly noted by Linsley (1972, p.67), although the grid references he gives are again incorrect. The largest and most interesting of the four localities (B24) occurs 250m west north west of Loth Station at NC 955102 and consists of 6m of sandstone breccia and pebbly sandstone (see Figs. 6.11 and 6.20). The base of the outcrop is composed of at least 76cm of sandstone breccia-conglomerate. The matrix to this bed is poorly sorted, fine grained sandstone with common to abundant, rounded to well rounded quartz pebbles up to 2cm in diameter. The clasts consist of very fine to medium grained sandstones (rarely pebbly, pebbles $\leq$ 5mm) and often show parallel bedding; their long axes are generally aligned more-or-less parallel to each other (and bedding?). Occasional clay clasts are also present and are like those observed elsewhere but were not submitted for XRD analysis. Overlying this basal unit there is a 30-70cm band which is essentially identical in character but is less coarse and weathers in a more rubbly fashion. The clasts in this band are generally smaller ($\leq$ 15cm long dimension) and more elongate (platey), resulting in a stronger parallel fabric than in the underlying unit. These two breccias can be clearly distinguished in Plate 6.19b; note the grading in clast size and the upward increase in the amount of matrix. The

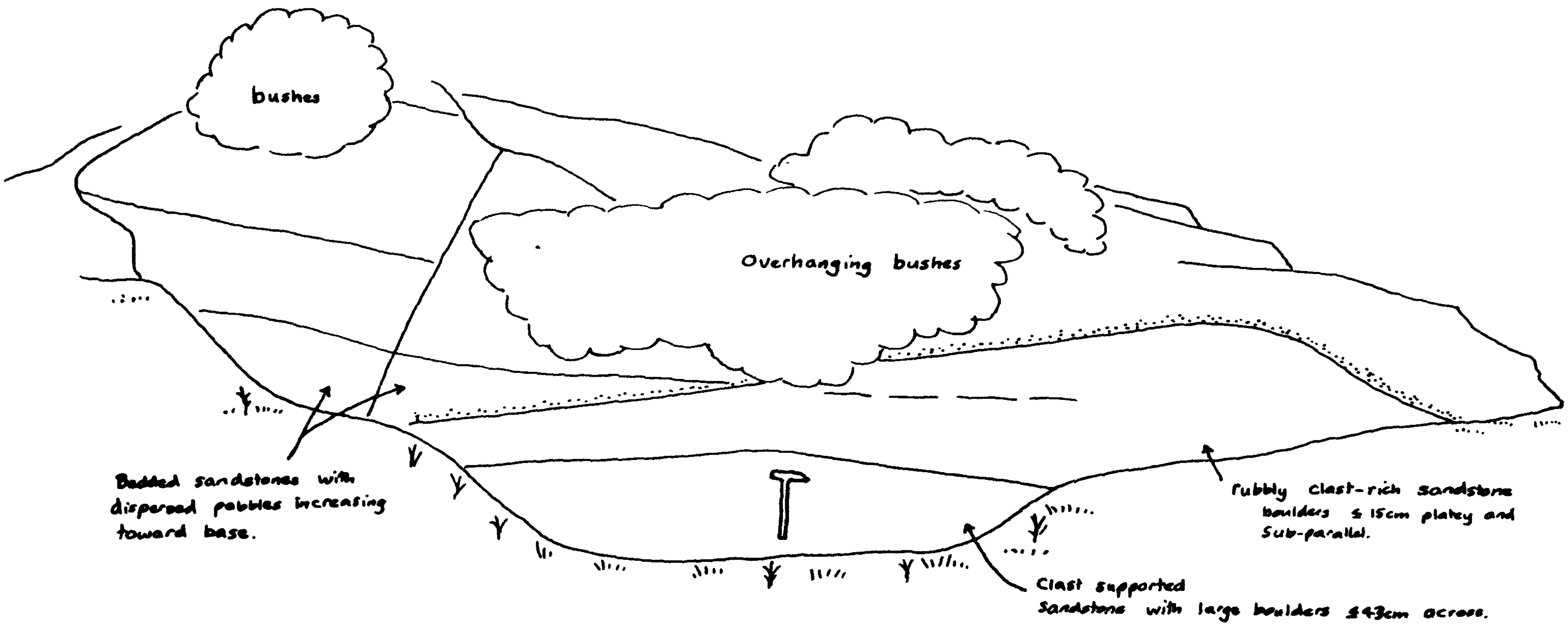


Fig. 6.20 Field sketch of locality B24.

topmost 4.5m exposed at this locality consists of pebbly sandstones bedded in 20-50cm thick units showing internal low angle (foreset) laminations. The sandstones are predominantly fine grained but are graded with respect to their pebble content; the pebbles and granules (which are generally less than 5mm in diameter) are most common in the basal metre but occur scattered through the whole 4.5m. The sandstones are partly calcite-cemented. The lateral variations of the three lithologies (the two breccias and the sandstone) are indicated in Fig. 6.20 (based on a field sketch). Note the diachronous distribution of the breccia facies. A strike and dip measurement of $348^{\circ}/12^{\circ}$E was recorded from the pebbly sandstones.

Locality B25 occurs about 150m ESE of B24 and 125m north west of Loth Station at NC 956101. Approximately 7m of sediment are exposed on the north side of the small hill (see Fig. 6.11). The base of the section consists of a diachronous breccia bed varying in thickness from 0.5-1.0m and containing clasts of orange-brown bedded sandstone which are generally about 5cm in diameter but vary up to 38cm. This is followed by 3.5m of pebbly sandstones like those described at B24 but containing several lenses of coarser material, as in the overlying rubbly bed, which consists of 50cm of bouldery, pebbly sandstone (boulders $\leq$ 10cm). The top of the section is capped by 2m of pebble-free or sparsely pebbly fine to very fine grained sandstone.

The two remaining localities, B26 and B27 are relatively minor. Locality B26 occurs 100m north of Loth Station at NC 957101 and consists of about 2m of slightly pebbly, fine grained sandstone. Locality B27 is situated about 30m WNW of B24, and 300m WNW of Loth Station at NC 954102. Only about 1m of sediment is exposed in a small outcrop (at the south east side of a low grassy mound) which is apparently situated stratigraphically below locality B24. The lower 50cm consists of a rubbly breccia bed (like that at B24) and is overlain by fine grained pebbly sandstones.

DISCUSSION OF THE LOTHBEG REGION

(1) Stratigraphic relationships

Before discussing the facies relationships observed in the Lothbeg region it is essential to briefly review the biostratigraphic data available for the area in order to gain the perspective necessary to distinguish between temporal and spatial facies variations. The quality of the biostratigraphic control in the Lothbeg region is unfortunately poor when it is considered in relation to the importance and significance of this part of the Brora-Helmsdale section. The number of reliable ammonite determinations leaves much to be desired, although is understandable in view of the predominantly unfavourable lithologies, poor preservation, problems of correlation and the vagaries of the beach exposure. Linsley (1972), to whom we owe most of our contemporary knowledge of the biostratigraphy of the Kimmeridgian section, apparently collected no ammonites between east Kintradwell and Lothbeg Point. Perhaps the present work will encourage ammonite workers to re-examine the area.

Previous discussions of the biostratigraphyof the Lothbeg outcrops have been mainly confined to arguments over the age of the "Allt na Cùile sandstone". Judd (1873, p.176) originally assigned the sandstone exposed in the Allt na Cùile to the Kimmeridgian, but we now know that his identifications of _Ammonites eudoxus_ (and _mutabilis_?) were almost certainly in error. Woodward (1911) later mapped these sandstones as Corallian, probably because of their pelecypod fauna (see below) and general similarities with the Brora and Clynelish sandstones. Lee (1925) was clearly more non-committal but was swayed by the Kimmeridgian age given to ammonites by Judd (loc. cit) and Buckman (1932), although he later became more inclined to accept Woodward's Corallian age (Lee and Pringle, 1932). Macgregor et al. (1930) were clearly convinced that the Allt na Cùile sandstones were of Corallian age and this conclusion was also later accepted by Bailey and Weir (1932). Arkell (1933, p.477) was more reluctant to assign a specific age to these beds, pointing out that many of the pelecypod and other invertebrate (non-ammonite) fossils were insufficiently age diagnostic to conclusively place the sandstone in the Corallian. One suspects that, as an ammonite worker himself, he was more impressed by Buckman's reported finds of Rasenia and Pictonia, as he appears to have been quite willing to believe that at least the top of the 'Allt na Cùile sandstone' was Kimmeridgian in age.

Callomon (in Arkell and Callomon, 1963, p.242) recorded a mutabilis zone fauna from the upper part of the "Loth River Shales", which to judge from his written log, came from about 10cm above the base of section B15. Both Lee (1925, p.106) and Macgregor et al. (1930, p.81) have previously recorded cymodoce zone faunas from this locality, but it is not clear whether these were inaccurately assigned or came from lower in the section (the latter would indicate that the cymodoce-mutabilis boundary occurs within the sequence at locality B15). Lee (1925, p.106) noted that the 'Loth River Shales' "seem to replace part of the sandstones exposed in the previous section" (i.e. at localities B1, 9, 10, 11 and 12). This clearly indicates that he believed the top of these shales to interfinger with the top of the "Allt na Cùile sandstone series" and is another probable reason why he initially assigned a Kimmeridgian age to the latter. Linsley (1972) has more recently shown that the shales at Lothbeg Point (i.e. locality B22) are of mutabilis age (apart from a few metres of eudoxus sediments at the very top), which implies that most of the sequence at locality B15 is also probably of mutabilis age as it lies directly along strike.

Linsley (1972, p.55) has examined the ammonite casts collected by Manson from the gorge of the Allt na Cùile (see Lee, 1925, p.105) and has interpreted them as representative of the askepta subzone (the highest of the three subzones of the cymodoce zone proposed by Morris, 1968). Brookfield (1976) notes the presence of the pelecypod Velata anglica in the lower part of the sandstone (his 'unit one', p.182) which, in Dorset at least, is indicative of a pseudocordata zone to baylei zone age. He also reports finds of an ammonite fauna in large loose blocks of sandstone in a field at NC 940095 (above the level of the Allt na Cùile gorge and near locality B7) which he believes represents the upper part of the cymodoce zone. Birkelund et al. (1978) have recently discussed the nature of the British cymodoce faunas and consider that there are at least four "horizons" which can be recognised by their ammonite assemblages. The uppermost of these, their "horizon IV" (p.35-36) is characterised by Rasenia (semirasenia) askepta Ziegler and associated Rasenia (Rasenioides) lepidula Oppel. The latter species is one of those described by Brookfield (1976) from loose sandstone blocks and hence confirms the view of Linsley (1972, p.55) and Brookfield (1976, p.184) that the 'Allt na Cùile sandstone' ranges up to the top of the cymodoce zone in its type area.

Even a brief examination of the type Allt na Cùile localities (B1-B4) indicates plentiful evidence for the likelihood of reworking. The presence of erosional surfaces and scour pockets indicates reworking on a short term, penecontemporaneous time scale, while the sandstone breccias with their similar but more cohesive sandstone clasts, suggest reworking and redeposition involving a rather larger separation in time between clasts and matrix. Under these circumstances the youngest age is most likely to be the more respresentative and it follows that the age of the Allt na Cùile sandstone body should be regarded as at least as young as late cymodoce zone age. Despite Brookfield's find of Velata anglica (whose stratigraphic range in Dorset may well have been ecologically influenced considering the changes between the Black Head and Kimmeridge Members), I believe it is unlikely that the pseudocordata or baylei zones are present in situ in the type area. It would appear that the epifaunal and shallow-infaunal pelecypod faunas recorded from the Allt na Cùile gorge are most characteristic of shallow water sand environments (q.v. Brookfield, 1976, p.183 and Sykes, 1975) and have been derived from the upthrow side of the fault, where one can easily imagine late Oxfordian and early Kimmeridgian shallow water sediments being reworked and redeposited as blocks and unconsolidated sediment. It is also apparent from the field relations that the type "Allt na Cùile sandstone" may range up into the lower mutabilis zone but probably not much higher. Cope's comment that "Linsley believes that the sandstone facies is developed at various points up to the base of the eudoxus zone and is not uniquely found bear the base of the succession (Cope, 1980, p.85), is not "contra Brookfield" (as stated) but results from the confusing double identity of the name "Allt na Cùile sandstone" as both as facies and a semi-stratigraphic term,and is an unfair and unjustified criticism.

With regard to the relationship between the Lothbeg and Kintradwell areas, Linsley (1972, p.75) states that the fossil evidence "suggests an upper cymodoce zone age for the sandstone at Allt na Cùile, whilst the succession at Kintradwell is known to be in the same zone but lower down". From his Table 1 (p.77) and previous comments (p.55) it is apparent that he considers the Kintradwell exposures to be largely representative of Morris's (1968) uralensis (middle cymodoce) subzone. I agree with Linsley's opinion that "part or all of the sandstones exposed between Allt na Cùile and Allt Choll to the south west, are probably the equivalent of the sandstones exposed sporadically along the beach north of Kintradwell" (p.75) and that no major fault has probably intervened between the two. The Kintradwell exposures are, therefore, probably the oldest Kimmeridgian sediments exposed in the Brora-Helmsdale outlier. The Corallian, baylei (and lower cymodoce?) zone ages that have been variously suggested for the "Allt na Cùile sandstone" appear to have largely resulted from the 'optimistic' biostratigraphic usage of non-diagnostic (and possibly reworked) fossils and the misinterpretation of similar facies as lateral equivalents of the same lithostratigraphic unit (see below). The problem of the precise correlation between

Kintradwell and Allt na Cùile is presently at the mercy of the exposure on the beach of Kintradwell Links.

(2) Lithofacies relationships

The major reason for the confused state of the 'lithostratigraphy' of the Lothbeg area lies in the discontinuous nature of the exposure. The sandstone outcrop initially appears to permit two interpretations; the first that the outcrop represents the discontinuous exposure of a continuous lithostratigraphic unit, and the second that the outcrop represents a series of exposures of separate, discrete and smaller units of similar facies. Unfortunately, their previous experiences with 'layer-cake geology', led the early Geological Survey workers to accept the first of these two interpretations, even though a detailed examination of the ground clearly indicates that it is the other which is correct. Lee, for example correlated the exposures of massive, veined sandstone at the mouth of the Allt na Cùile with those at Lothbeg Point and elsewhere (Lee, 1925, p.104, 106). This attempt at simplification irrovocably confused the lithostratigraphy, tended to exaggerate the significance of the 'Loth anticline' and led Lee to seriously underestimate the thickness of the sequence (Lee, 1925, p.103). Since all the existing rock terminology in the Lothbeg area (e.g. "Allt na Cùile Sandstone", "Loth River shale") is based on a misconception of the lithostratigraphy (either implicit in the original definitions or by subsequent (mis-) usage), it is all essentially invalid, although tediously entrenched in the literature.

Lithofacies rather than lithostratigraphy are the real key to understanding the geology of the Lothbeg region. The principal lithofacies found in the Lothbeg outcrops are listed in Table 6.1 and their geographic distribution, based on outcrop data and partly extrapolated from aerial photographs, is shown in Fig. 6.21. I believe the configuration of the outcrops indicates that the Lothbeg region consists

TABLE 6.1

Kimmeridgian lithofacies developed in the Lothbeg region

1. Current bedded to parallel bedded sandstones (B1, B3, X2, B9, B21)

2. Massive, fine-medium grained, veined sandstones (B6, B8, B9, B11, B2, B25)

3. Pebbly sandstones (pebbles dispersed or showing coarse tail grading)

   (a) Structureless (B12, B21, B25, B26, B27)

   (b) Current bedded (B12, B24)

   (c) With carbonaceous horizons (B1, B2)

4. Sandstone breccias

   (a) Thick, massive chaotic, predominantly clast-supported: occasional kaolinitic clay matrix (B4, B5, B6, X1, X2, B7)

   (b) Thinner interbeds, sometimes graded or imbricate, clast to matrix supported (B1, B3, B6, X2, B7, B10, B12, B24, B25, B27)

5. Shales with lenticular sandstones $\leq$ 2m (B10, B13, B14, B15, B16, B17, B18, B19, B20, B21, B23)

6. Shales with only rare sandstones ($\leq$ 30 cm) (B22)

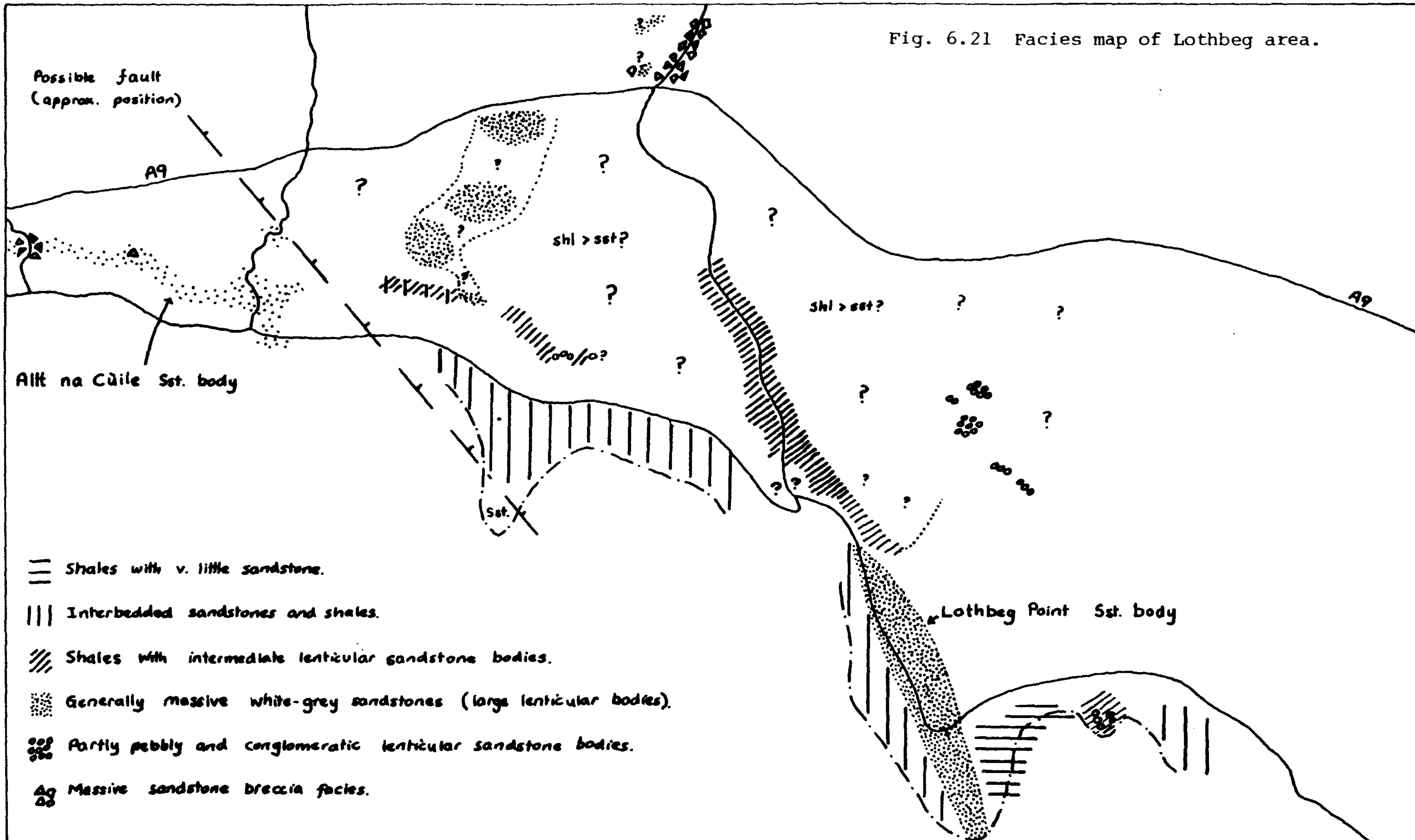


Fig. 6.21 Facies map of Lothbeg area.

mainly of a background sandy shale facies which contains several prominent lenticular sandstone bodies (see Fig. 6.22). This pattern is somewhat complicated by the varying proximity of the outcrops to the Helmsdale fault (where sandstones are more predominant) and may include an element resulting from facies variation in time as well as space. The two main lenticular sandstone bodies are described below.

(i) The Allt na Cùile sandstone body.

The apparent dimensions of this unit lie somewhere in the region of about 1000m x 50-90m. It is the most complicated of the large sandstone bodies observed in the Lothbeg region since it contains at least three major facies transitions; the vertical change from the massive, pebbly basal part to the upper, bedded division (see locality B1), the apparently proximal-distal (and partly lateral) change from breccia-dominated to sandstone dominated (compare localities B1, B3 and B4), and the apparent lateral change with the incoming of carbonaceous sandy shale units observed at locality B9. The proximity to the Helmsdale fault and the presence of the distally-thinning breccias implies that the body (or at least some of its facies components) are fan-shaped. The topography of the area and the relative positions and nature of localities B1, B9 and B13 suggests that the body is at least partly underlain by sandy shale facies (like that at B13) and may be thinning laterally north eastwards along the outcrop. Conclusive evidence is not available. It is difficult to say what happens in the area between locality B9 and the A9 road because of the very poor exposure. The featureless fields lying between the Allt na Cùile and localities B9 and B8 possibly suggest unexposed softer, shaley sediments, while the sandstones exposed in the hills at locality B8, being somewhat different from those typical of the Allt na Cùile outcrops, probably belong to distinctly separate bodies or represent parts of a single unit lying above the Allt na Cùile body. The massive sandstones seen in the hills of locality B8 may be continuous with the very similar sandstone seen

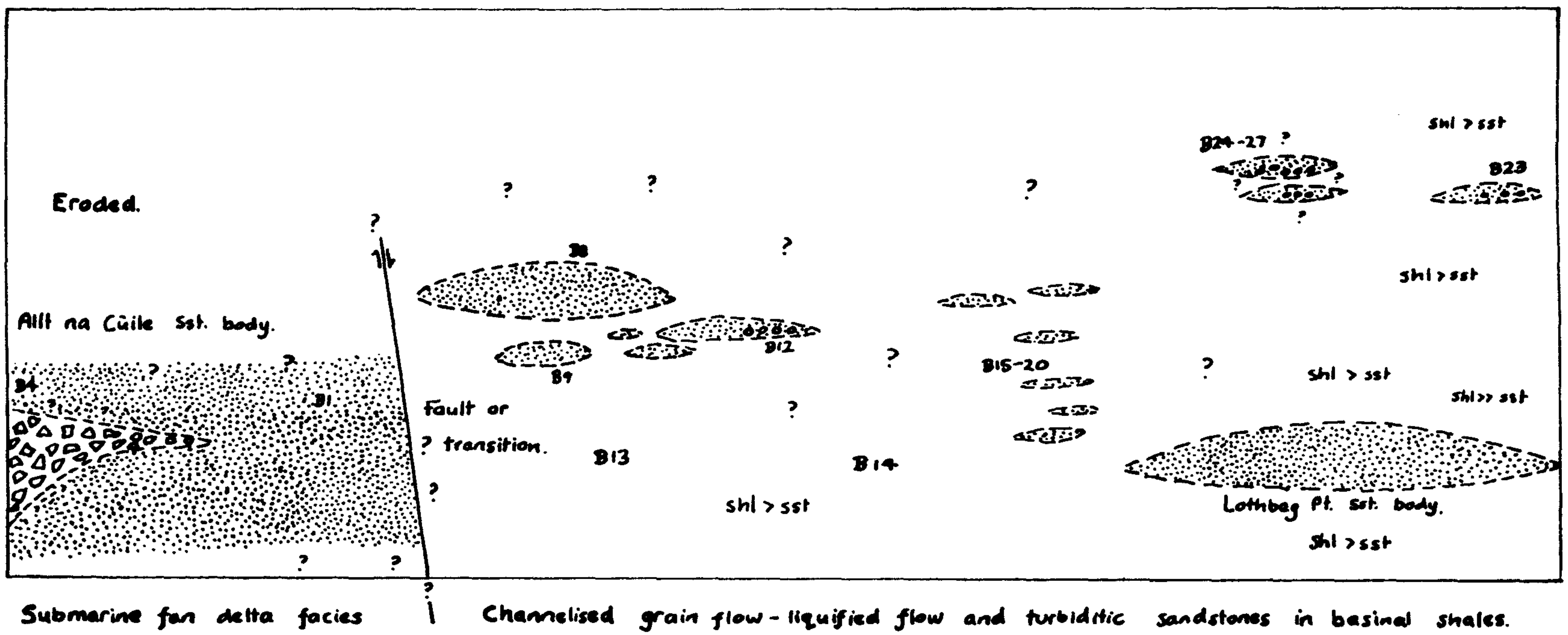


Fig. 6.22 Schematic facies section of Lothbeg area.

at the very north east end of locality B9; these sandstones may together form part of a body (or bodies) which have coalesced and amalgamated with the main Allt na Cùile unit. Considering the proximity of the Helmsdale Fault, it would not be surprising if the sandstones represent a complex of lobate units (a sort of 'submarine bajada') with shaley sediments only deposited in the vicinity of the fault scarp in the areas between active fans or loci of sediment input. The massive development of the sandstone breccias within the Allt na Cùile body (i.e. as at locality B4) appears to be limited to within 150m of the fault line. The interpretation of the facies is discussed at the end of the chapter.

(ii) The Lothbeg Point sandstone body

This is a much better defined unit than the All na Cùile body since its lateral margins and base can be confidently ascertained from the outcrop (see description of locality B21). The body has maximum dimensions in the range of 850 x 58m and consists predominantly of massive, structureless sandstone but contains a better bedded, more carbonaceous transitional facies at its base and lateral margins. It is surrounded by the sandy shale facies on three, and presumably four, sides (see localities B21, B22 and B15) and is clearly of lenticular cross-section (see Fig. 6.21). The thick sandstone unit at the base of section B15 is not a lateral continuation of this sandstone but belongs to a separate body which lies above the north west margin of the Lothbeg Point body. The contrasting facies observed in the Lothbeg and Allt na Cùile bodies may be mainly attributed to their differing proximities to the Helmsdale Fault, although the fact that they are probably of slightly different ages may also be significant.

There are probably several other relatively large lenticular sandstone bodies in the Lothbeg region (e.g. at the base of section B15, the sandstones at locality B23, and the pebbly sandstones at localities B12 and B24-B27) but in each of these cases the exposure is insufficient to allow reconstruction of their likely size and geometry, or to tell whether more than one body is present. The pebbly sandstones with interbedded breccias at localities B24-B27 may, for example, represent the discontinuous exposure of a single body (since no shales were observed in this area) or the exposure of a series of separate, smaller units. Another interesting problem is the relationship of the sandstone breccias at locality B5 to the shales and sandstones exposed around the lower reaches of Loth Burn. There may be an unexposed facies transition from breccias to bedded sandstones (as in the All na Cùile body) and hence to shales with lenticular sandstones (i.e. the facies at localities B15-B20) in the fields in the immediate vicinity of Lothbeg, but it is also conceivable that a fault may occur between localities B5 and B20. Presumably if such a facies transition does not now occur at the surface (covered by drift) it probably does so at depth and probably also occurred at higher levels which have since been eroded. It also seems likely that the Helmsdale Fault may well be more complicated than the general single line shown on the geological map of the area, and that there may be other smaller, intra-Mesozoic faults in the outlier which have not been recognised because of the poor exposure. The sharp change in the strike of the Helmsdale Fault (about 70°) which occurs in the vicinity of the old Loth Burn road bridge, does perhaps suggest the fault may be a little more complicated than shown on the maps.

The afore-going account and discussion of the Lothbeg region makes it quite clear that no simple lithostratigraphic rationalisation of the sequence is possible and that any generalisations made along these lines (e.g. Brookfield, 1976, p.182) are largely meaningless <u>except</u> at specific localities. Although the lithofacies organisation in this region is complex it can be fairly successfully summarised as in Fig. 6.21 and Table 6.1. Remarkably, Linsley (1972, map A) has previously mapped the whole of the Lothbeg area, with the exception of the shales at localities B15 and B22, as consisting entirely of one facies - the "Allt na Cùile facies". While a facies-orientated approach is clearly better than the earlier lithostratigraphic ones, such a sweeping generalisation is meaningless; he might as well have said that there are a lot of veined, decalcified sandstones in this area, since this all such a broad label implies. Indeed, Lee (1925) seems to have been more impressed with the lithological variation in the "Allt na Cùile sandstone series" than Linsley was to be 47 years later. Linsley's vague and naive appreciation of the lithofacies distribution is exemplified by the following comment on the Loth River railway bridge section, that "the shales are thought by the author to represent a period of quiescence during the formation of the Allt na Cùile sandstone" (Linsley, 1972, p.79 - see also his Map A).

(3) Recommendations on rock unit nomenclature

(i) The name "Allt na Cùile sandstone" should be discontinued; the name has no lithostratigraphic significance and is too general for usage in an informal facies sense.

(ii) The name of the Allt na Cùile should be used only with regard to the type locality (e.g. 'the sandstones exposed in the gorge of the Allt na Cùile'),or in a general sense to indicate the lateral extent of the sandstones exposed in this stream and the Allt Choll (e.g. 'the Allt na Cùile sandstone body'). The name should not be applied to any unit which is not identical to, and which cannot be clearly seen to be continuous with, those sediments exposed in the type area (sections B1-B4).

(iii) The name "Lothbeg River Shales" should be discontinued as it has no lithostratigraphic significance and represents only the background sandy shale facies.

(iv) The informal term "the Lothbeg Point sandstone body" may be used for the large lenticular sandstone unit exposed along the shore at the south west end of Crackaig Links between the mouth of Loth Burn and

Lothbeg Point. This term should only be used to indicate the locality of this unit and not its lithological character.

(v) Owing to the facies complexity in this area, all sandstone units should be described with regard to their facies rather than just named (e.g. see Table 6.1).

The facies presently exposed in the Lothbeg region are not necessarily typical of coeval but unexposed sediments elsewhere within the Brora outlier or of the unexposed or now eroded older and younger sediments in this particular locality. For further comments see the interpretive section at the end of the chapter.

THE CRACKAIG LINKS AREA

Section C1 Central Crackaig Links (NC 964100) eudoxus-autissoidorensis zone

The location of this section is shown on Fig. 6.19 based on Linsley's Maps XIV and XIII. The 27m of sediments consists mainly of shale but includes about 30% (by thickness) of lenticular, usually amalgamated,fine to medium grained, calcite cemented sandstone beds (see Fig. 6.23). The most conspicuous of these amalgamated sandstone units is Linsley's "six foot thick bed" whose position is marked on his map XIV, and the equivalent 2.76m unit indicated on Fig. 6.19. Linsley's "fairly well developed load casts" at the base of this composite unit (weathering as one massive bed) appear to be a misinterpretation of the lower shale intraclast-rich sandstone; only very minor loading is in fact present. The individual beds within this composite unit are often strongly lenticular and this characteristic is also seen in the associated, smaller discrete sandstone bodies (e.g. Plate 6.20a). Linsley (1972, p.29) also comments on the rapid lateral changes in bed thickness.

The most interesting feature of the amalgamated sandstone units in the central Crackaig Links section is the repeated, predictable sequence of features exhibited by the individual component beds. Although the

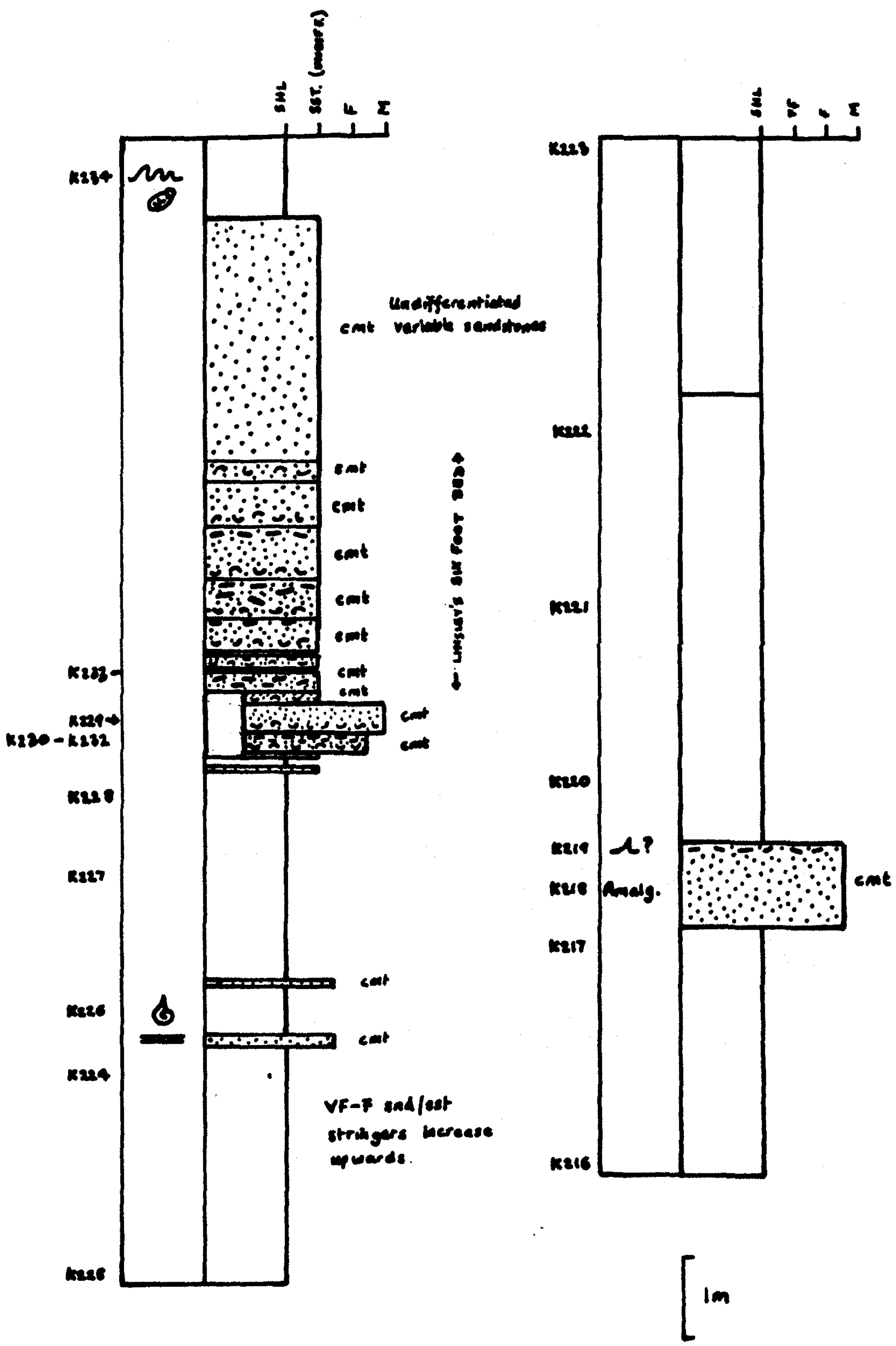
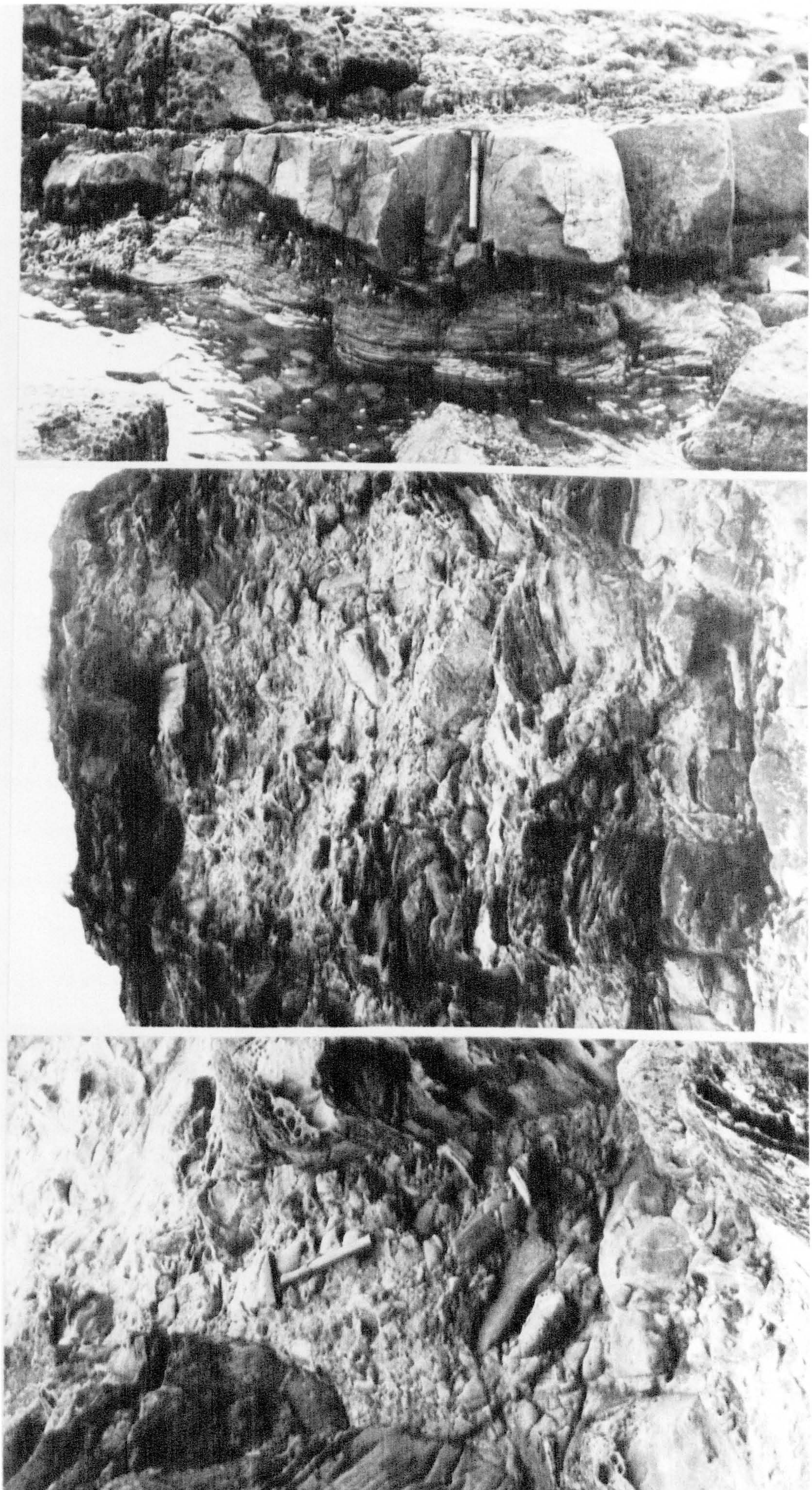


Fig. 6.23 Section C1 Central Crackaig Links.

PLATE 6.20 Sedimentological features of Crackaig Links exposure.

6.20A Channelised sandstone near base of section C1.

6.20B,C Clast-supported breccia of Old Red Sandstone blocks at Crackaig Links Point (see Fig. 6.19); hammer for scale.



sandstones show little indications of grading according to grain size, the basal parts of the beds are conspicuously richer in bioclastic debris (mainly shell fragments) which decrease in abundance (and size?) upwards. This sequence is topped by a division rich in sub-parallel shale clasts (2-10cm) which is continuous with the sandstone below. No shale or other clasts were observed in the basal parts of these beds. It is clear that these individual sequences represent discrete depositional events, i.e. they were each formed during one episode. The upper division (rich in shale clasts) is usually overlain by an erosional contact with the base of the succeeding bed and has often been much reduced in thickness or completely eroded. In other parts of the section this shale-clast rich unit can be seen to pass sharply upward into shale. These sequences vary in thickness from about 10cm (e.g. in the subordinate sandstones in the shales near the base of the section) to 66cm in Linsley's "six foot thick bed" (see Fig. 6.19). The sandstones are mainly calcite cemented (with poikolitic cements) although the thinner bands within the shales at the base of the section are often uncemented. The uppermost 3m of undifferentiated sandstones is rather poorly exposed, appears to have been partly deformed and slightly folded, and consists of massive and finely laminated alternations and sequences like those described above. A strike and dip of 090/20°N was recorded near the base of section C1. The lateral relationships and the remainder of the Central Crackaigs Links outcrops are discussed below after the description of section C2.

Section C2 Crackaig Links east (NC 97101) autissiodorensis zone?

This section is located approximately 130m east of the small headland at the east end of Crackaig Links (the feature termed "Crackaig Links Point" by Linsley and shown on Fig. 6.19). The section consists of 13.5m of sediments including 13% sandstones and 11% bouldery sandstones. One of the sandstone beds shows the sequence of features observed in many of the sandstones in section C1, namely a bioclast-rich base and a shale clast-rich top. Only one (1.5m thick) "boulder bed" unit is present; it is matrix supported (>50%) but clast-rich, containing clast of sandstone up to 2m across (generally 10-50cm) in a shaley or argillaceous sandstone matrix. No other particularly noteworthy features are present; the section is shown in Fig. 6.24.

DISCUSSION OF THE CRACKAIG LINKS AREA

With the possible exception of the sandstones at locality B23, the stratigraphically lowest unit exposed on the shore of Crackaig Links is the massive boulder bed unit which forms the short cliff line at 'Crackaig Links Point' (Bailey and Weir's "Boulder Bed A"; see Fig. 6.19). The full thickness of this unit is not exposed even at low tide, but assuming it is not much greater than that part which is exposed, it may be estimated to be around 30m. The boulder bed is very dense, predominantly clast supported, and has a strong superficial resemblance to the Allt Choll intraformational breccias (see Plate 6.20). It is composed of angular to subangular clasts of very fine to fine grained, orange-brown or orange-speckled white sandstone in a fine to medium grained, similarly coloured sandstone matrix. The sandstone clasts are almost all parallel-bedded (in flaggy 5mm to 2cm units) or show low angle planar cross-bedding. Some of the clasts occasionally show laminations (water escape structures and deformation?) and ripple bedding; a few also contain scattered quartz pebbles ($\leq$1cm) and small siltstone or mudstone clasts. The boulder bed is essentially monomict; the only other lithology present is a pale blue-green mudstone which occurs as small (<1cm) rare clasts in the matrix. There seems little doubt that the boulders are predominantly of Old Red Sandstone (John O'Groats sandstone facies). Some of the clasts show a typical red-brown colouration but in most the iron has been leached and redistributed and iron-pan weathering is common. Thin sections of the boulder bed matrix show partial sparite

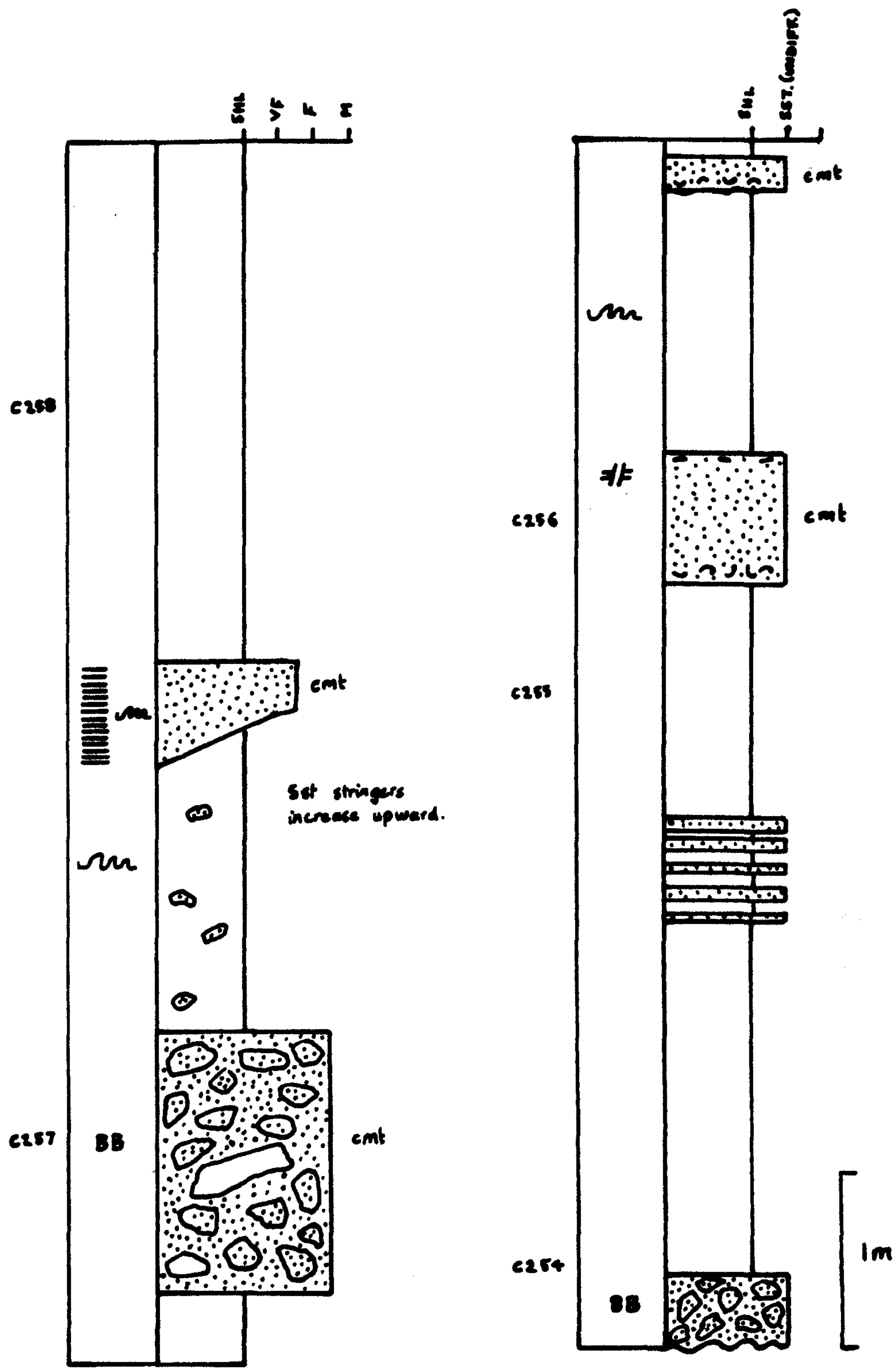


Fig. 6.24 Section C2, Crackaig Links east.

cements, relatively common kaolinite pore-fillings and trace amounts of glauconite and bioclastic debris. As noted by Linsley (1972, p.38), to the west of "Crackaig Links Point" the top of this boulder bed has been capped and smoothed by an immediately overlying sandstone bed; to the east, however, at the base of section C2, this sandstone is absent and the boulder bed is topped by a shaley argillaceous sandstone which has been contorted by compaction around the irregular top of the boulder bed.

Overlying the boulder bed on the west side of "Crackaig Links Point" and at least partly equivalent to section C2, is a shale unit whose thickness is estimated at between 18 to 23m (Bailey and Weir's "B Shale"). The centre of this shale band contains two prominent sandstone beds separated by a thin (56cm) shale. The upper of the two sandstone bands is of variable thickness ($\leq$34cm) and is a composite, amalgamated bed which has a base rich in bioclastic debris and contains shale clasts in its upper part. The lower band is 20cm thick, more uniform in character and laminated. Convolute laminations are present in parts of these sandstones and occasional small, short sandstone dykes occur in the shale between them. Throughout the whole of the "B Shale" there are numerous stringers of uncemented very fine sand; only thicker bands appear to have been calcite-cemented.

Overlying the "B shale" is Bailey and Weir's "Boulder Bed C". This latter unit has attracted more attention than most because of its interesting field relations (see Bailey and Weir, 1932; p.449-450 and Linsley, 1972, p.36-39, 80). The boulder bed and its lateral relationships are shown in Fig. 6.19 which is labelled in accordance with Bailey and Weir's original description (Bailey and Weir, 1932, Fig. 6, p.449). It is very similar in character to "Boulder Bed A" and consists predominantly of angular clasts of white to pale grey, orange-brown speckled, very fine to fine grained, parallel or cross-bedded sandstone in a fine to medium grained sandstone matrix. Much less common clast lithologies include pale blue-green mudstones and siltstones, alternations of which form the large ($\leq$44.2m long) conspicuous clasts whose positions are shown in Fig. 6.19. The bedded, flaggy, relatively micaceous sandstone clasts and the pale blue green mudstone siltstone blocks are almost certainly of Old Red Sandstone age and clasts of Jurassic lithologies appear to be very rare. Thin sections of the clast lithologies show that they have been variably attacked and replaced by calcite-cements; some samples show minor sparite cementations with most of the closed pores filled with kaolinite aggregates, in others replacement has advanced further and the cements are extensive and poikolitic (lustre-mottled). Some of the siltstones and mudstones show incipient metamorphic fabrics, with mica flakes aligned in a criss-cross fashion (approximately 120° and 60°). Towards the base of the boulder bed (as seen at the margins) the matrix sandstone is more argillaceous, dirty, carbonaceous and poorly sorted and contains small amounts of bioclastic (including phosphatic) debris.

Bailey and Weir (1932, p.450) considered "Boulder Bed C" to have been a "landslip" which had dug into the underlying shales (B). This seems a fairly reasonable interpretation, but that subsequently proposed by Linsley (1972) verges on the realms of the fantastic and must be recounted in his own words in order to demonstrate that it is not a product of my own misunderstanding. Linsley (1972, p.38) considered that the base of this particular unit was "unlike that of any other boulder bed", a fact in itself which is not even true, and that the "once sharp base has been smeared out of recognition". His interpretation proposes that the boulder bed, rather than being simply a submarine landslip or similar phenomenon, was also "injected sideways during folding" (Linsley, 1972, p.80). He considered that "the contortion of the (underlying) shale was probably the result of post-depositional movement, during which the boulder bed C was injected along the strike towards the south in a semi-plastic state, causing the deformation of basal contact with the shales". His account continues ... "the evidence of sub-parallelism seen in some of the boulders (i.e. the large mudstone blocks) close to the contact is offered in support of this theory. The movement, which was not excessive, resulted in the deformation of the shales above the boulder bed, the capping sandstone bed D forming an upper limit, beneath which the semi-consolidated shales B* were squeezed and contorted" (Linsley, 1972, p.39). Such a complicated and unlikely interpretation is not necessitated by the facts.

Linsley's interpretation places great emphasis on the nature of the basal contact of the boulder bed and the characteristic features of this contact are discussed below. The base of the boulder bed is everywhere rather irregular and the underlying shale tends to be rather disturbed and structureless at the contact with a transition from shale proper to structureless argillaceous sandy matrix to clean sandstone matrix across the boundary. The vertical extent of this transitional base is variable, not least because of the differing dips on the two margins of the boulder bed, and is often only a few centimetres in thickness. Deformation of 'the shale' is most pronounced on the west side of the boulder bed in the area marked B*. In this particular region 'the shale' has been totally disrupted, is largely structureless and the outcrop essentially a matrix-supported argillaceous boulder bed. The matrix consists of largely homogenised sandy shale and contains $\leq$25% clasts (up to 10cm in diameter) which are composed of argillaceous sandstone, shelly bioclastic sandstones, shale clast-rich sandstones, flaggy sandstones and pale blue-grey and blue-green mudstones and siltstones. The clasts are predominantly angular (some are subrounded), most are parallel-bedded and there is clearly a mixture of Jurassic and Old Red Sandstone lithologies. As noted by Bailey and Weir (1932, p.450) this argillaceous boulder bed (B*) at least partly overlaps the "Boulder Bed C" and therefore appears to be younger or so situated because the sandy massive boulder bed (C) dug down through the underlying shale during its emplacement. Adjacent to the relatively clast poor argillaceous boulder bed (B*) there are at least two more dense boulder bed tongues which project from under sandstone bed D; these are marked C* and $C^{\dagger}$ on Fig. 6.19. These tongues vary from matrix-supported to clast-supported (>80% clasts) and contain clasts which are generally 70cm or less in diameter (but which range up to 2m across) in a matrix which consists not of shale but of poorly sorted argillaceous sandstone (see Plate 6.24a for similar boulder bed).

One of the confusing aspects of the outcrop of boulder bed C is the relationship between the structure and topography of the area. Boulder bed C lies at the 'core' of a strongly assymetric, northerly plunging syncline which has dips of 30-36^{o} on its east side and $\sim$15^{o} on its west side. This change in dip means that the outcrop is largely two-dimensional on the east side and becomes more three-dimensional in its west side. The large clasts "ca" and "Cd" are oriented parallel to the base of the unit, while the clast "Cc" is probably situated near to its top. If the thickness of the boulder bed outcrop (as measured along its approximate fold axis from its nose) is not much greater than that which is at present visible, the thickness of the boulder bed is probably about 15-20 m. The orientation of the large clasts does not indicate sense of movement as supposed by Linsley, but illustrates crude bedding-parallel 'imbrication', which is hardly surprising given the size of these clasts. Given this structural situation it is apparent that the argillaceous boulder bed (B*) does not so much lie over the top of boulder bed C as overlap against its sides, and is therefore at least partly laterally equivalent rather than being stratigraphically above or below. The present interpretation is that B*, C* and $C^{\dagger}$ represent argillaceous boulder bed units (one or several amalgamated) which partly overlap laterally and lie along strike from boulder bed C; B* is not a lateral equivalent of shale unit B, which occurs on the east side of C, but is in fact a younger horizon (as indicated by the dips). The westward continuation of unit B is the shale which occurs to the south of C* and B* and to the southwest of C. The true basal contact of boulder bed C is probably structureless and homogenised because of liquifaction, shearing and drag which occurred during emplacement; there seems little reason to propose subsequent post-depositional tectonic deformation. The small outcrop of boulder bed (C?) on the east side of "Crackaig Links Point" may be a lateral equivalent of C; if so the boulder bed is diachronous and the thickness of B has been much reduced.

One of the most interesting features of the central Crackaig Links area is the unconformable relationship between the prominent sandstone bed D and the underlying units C*, $C^{\dagger}$ and B*. Sandstone D is 50 to 70cm thick and consists of several amalgamated beds; the top surface is rippled, the upper part of the bed exhibits water escape type structures (convolutions, etc.) and there are several horizons rich in shale clasts. Its base is sharp and flat. Bailey and Weir (1932, p.450) believed that the top of the underlying units had been planed-off by erosion due to "tunamis" prior to the deposition of this sandstone. However, Linsley (1972, p.38) notes that "the apparent erosion surface at its base is not continuous westwards from the area of the contorted shale (B*), i.e. it is not an especially eroded horizon when seen elsewhere". This is suggested on Fig. 6.19 where to the west of C* it can be seen that the strike of D and E becomes parallel with the underlying shales (the westward continuation of B). The impression that I received on examining this area was that an horizon of minor decollement may have occurred at the base of sandstone D and that this band and the overlying shales (E) were subsequently 'thrust' over the top of the underlying argillaceous boulder beds. The extent of the translation is probably small. Units D and E (including much of section Cl?) probably formed a detached sheet which slid (or crept) down the palaeoslope. To the west of C* this sheet probably slid across a shale surface (hence the parallelism of strikes above and below, and the absence of a visible disconformity) but where it encountered the boulder beds it was probably forced up and over them creating a noticeable disconformity. It is difficult to say if this decollement and 'thrusting' was at all effected by the subsequent folding or whether it might have even been produced by this tectonic activity as a result of the differing competence of the lithologies involved.

The E shale (lateral equivalent of section C1) contains three prominent large calcareous concretions (see Fig. 6.19) like those in section B22. They are calcite veined, calcite-cemented shales which contain sandstone lenticules and stringers which are partly load-casted, pseudo-noduled and folded; redeposited bioclast-rich layers (≤5cm) are also present (with oysters and ammonites) and some surfaces show aligned woody debris. The E shale also contains several sandstone bands (≤20cm thick) and occasional, isolated subangular sandstone clasts up to 10cm in diameter. From Fig. 6.19 it is clear that there is a significant change in strike between locality B23 and section C1; the significance of this is unclear and a fault may intervene between the two areas or there may be further such complications as horizons of decollement, etc.

It should be noted that Linsley (1972) only obtained ammonites (indicative of the <u>autissiodorensis</u> zone) from the top of shale B (on the west side of "Crackaig Links Point") and that most of the shale B, boulder bed A and section C2 are of unknown age; the <u>eudoxus</u> age shown by Linsley is only conjecture.

THE CULGOWER AREA

<u>Section D1 Culgower (NC 986115)</u> <u>eudoxus</u> or <u>mutabilis</u> zone

This section is located on the shoreward side of a low-lying promontory situated south-south-west of Culgower and approximately 650m west of Srồn Rubha na Gaoithe. The area was not mapped by Linsley (1972) but the promontory is shown on the Ordnance Survey Map (1:25,000) and its position with respect to the other Culgower outcrops is indicated in the panoramic view in Fig. 6.25 (redrawn from photographs). Because of

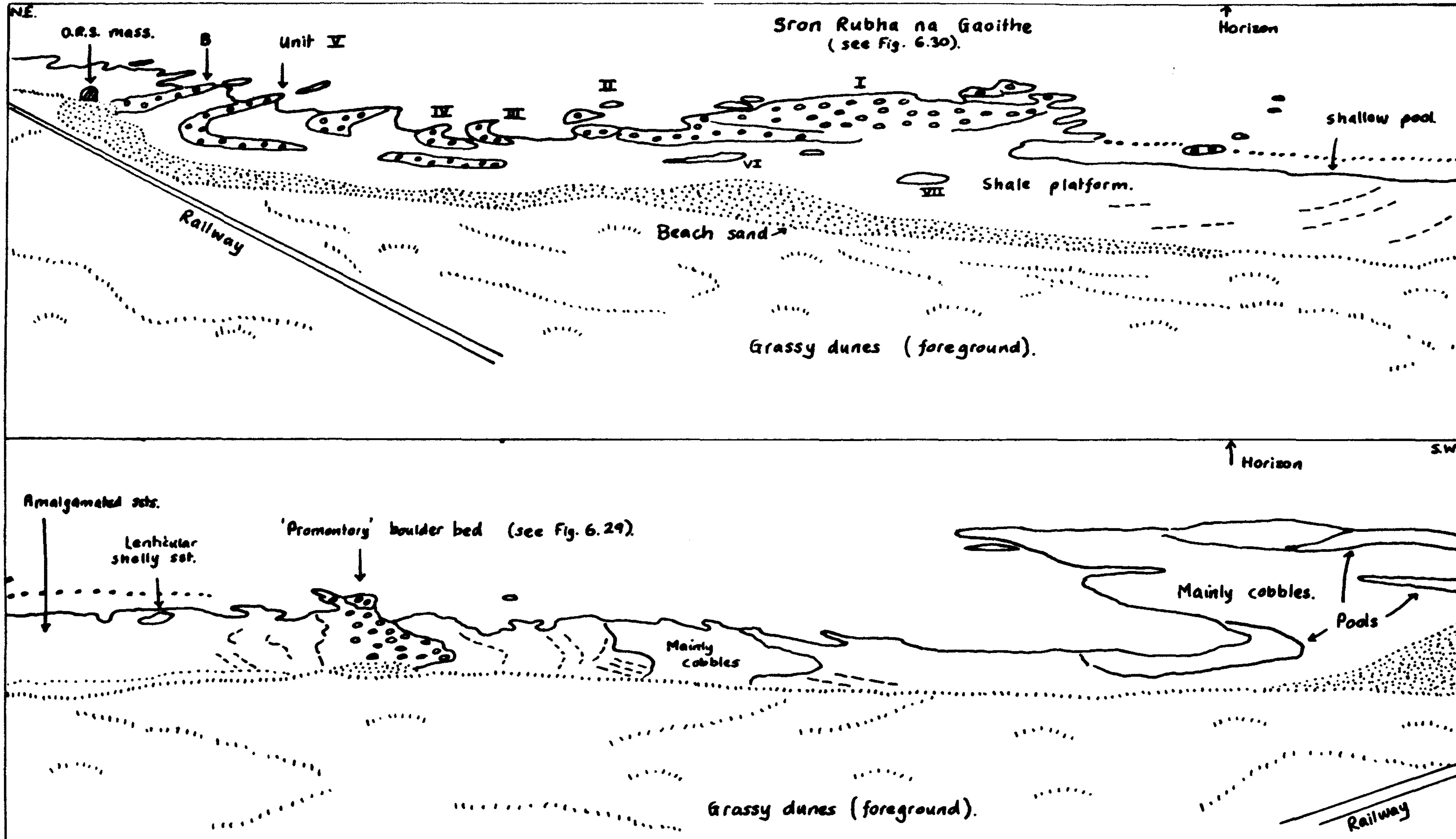


Fig. 6.25 Panoramic view of Culgower area redrawn from photographs.

the poor exposure in this area no precise correlations between the two sections measured on this promontory (D1 and D2) can be given except that they probably overlap and the base of section D1 is older than that of D2. No outcrops occur between this promontory and Crackaig Links to the west; the promontory shown on the maps at NC 982111 is formed by a bouldery raised beach deposit (with granite boulders) and is not Jurassic in age.

Section D1 comprises 8.92m of sediments of which ∿50% is represented by sandstone beds. The sequence consists essentially of a rapid alternation of thin sandstones (≤7cm) and shales with only a few thick sandstone beds (i.e. greater than 10cm in thickness; see Fig. 6.26). The prominent sandstone beds often show bioclast-rich bases and occasionally shale-clast containing tops (i.e. like the Crackaig Links sandstones). The lower part of the uppermost sandstone also contains sandstone clasts up to 5cm in diameter in a very shelly matrix; on its upper surface occasional clasts up to 10cm in diamater are visible and this may suggest reverse grading or may be a result of fortuitous outcrop. The bases of the sandstone beds are more irregular than their tops and appear to be slightly erosional. The sandstones are fine (to medium) grained, calcite-cemented (often lustre-mottled) and contain trace amounts of glauconite and common pelecypod, brachiopod, gastropod, echinoderm, ammonite and foraminiferal debris. Only slight grain size grading was observed (most variation within Wentworth size classes) and vertical changes were most pronounced in terms of composition (amount of bioclastic and carbonaceous material) and sorting. A few beds showed common belemnites oriented in an east-west direction. A strike and dip of $154^{\circ}/20^{\circ}$E was recorded on this section.

<u>Section D2 Culgower (NC 978114)</u> <u>eudoxus</u> or <u>mutabilis</u> zone

The line of section D2 is situated approximately 50m to the south of D1 on the same low-lying promontory. The section comprises 16.72m

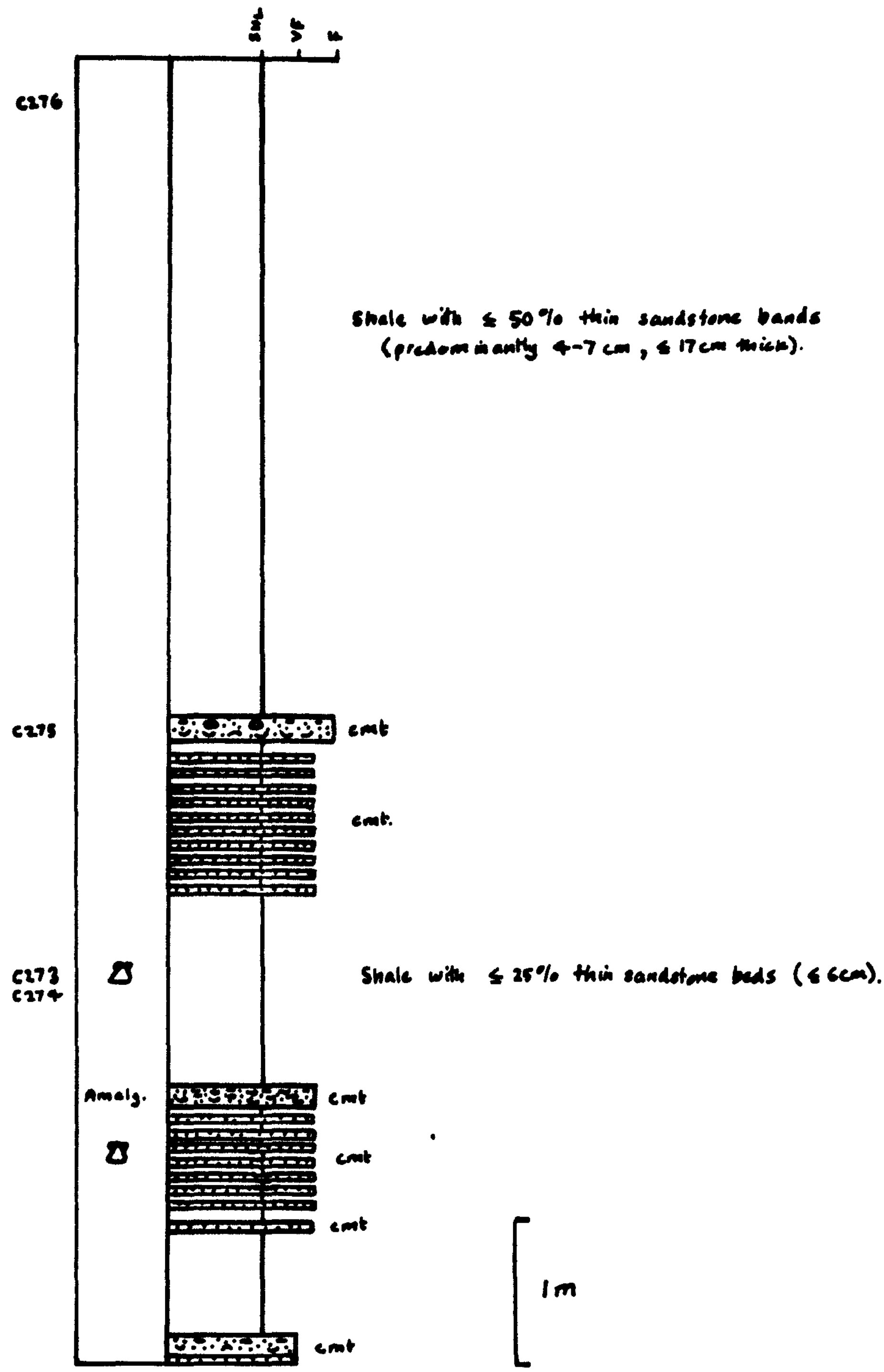


Fig. 6.26 Section D1 Culgower.

of alternating sandstones and shales and is essentially of an identical facies to that in section D1 but includes two boulder bed tongues (see Fig. 6.27. There are only a few thick (>10 cm) sandstone beds but the shale units contain common thin sandstones which are generally 4-6cm thick but range up to 8cm; the lowest 3m shale unit contains uncemented fine to very fine grained sands up to 10cm in thickness.

The sandstone beds (and particularly the thicker ones) usually have a basal layer (up to two-thirds of the total thickness) which is rich in bioclastic debris. Shale clasts were only observed in the upper part of one of these sandstones. The sandstones are predominantly fine grained; most are calcite-cemented (lustre-mottled with up to 75% calcite in the very shelly bands) and their tops often show carbonaceous laminae. The most interesting features of this section occur above the 4.2m shale unit (see Fig. 6.27). A lower 16cm shelly sandstone is partly separated by a largely eroded 0-10cm thick shale from an upper 27cm shelly sandstone, the beds having amalgamated in the absence of the thin shale. Along strike the lower 16cm of the upper shelly sandstone locally passes rapidly into a matrix-supported boulder bed containing sandstone clasts up to 35cm (mostly ≤10cm) in diameter in an extremely shelly sandstone matrix. The overlying 3m shale contains thin uncemented sand bands, is highly contorted in places and contains slumped and rotated blocks of shale. The succeeding 30cm sandstone is of particular interest because it exhibits the closest approximation to a Bouma sequence observed during the whole of my investigations on the Brora-Helmsdale section (see Plate 6.21a). The sandstone is generally pure and very fine to fine grained- it has an erosive base, a relatively massive lower part with a few shale clasts, a cross-laminated central part (see photograph) and a parallel-bedded top with occasional carbonaceous laminae. At one point along its strike this bed is down-warped by the presence of a boulder bed tongue in the overlying shale. This boulder bed has a highly weathered, friable, dirty, sandy matrix which contains some clasts of Old Red

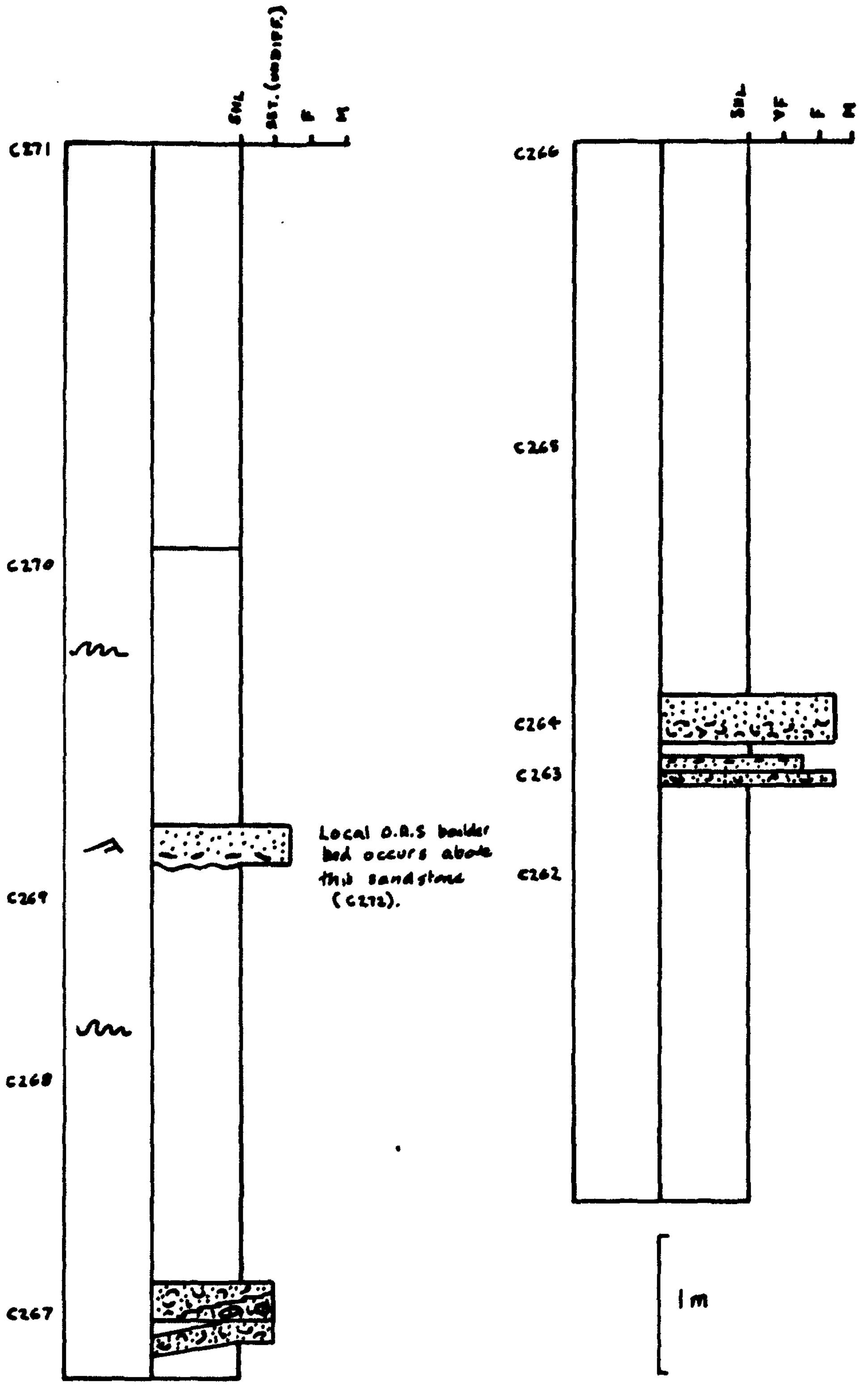


Fig. 6.27 Section D2 Culgower.

PLATE 6.21 Sandstone units from Culgower area

6.21A Sandstone bed exhibiting Bouma sequence from section D2.

6.21B Amalgamated cemented sandstones (not clast bearing basal intervals); locality D3iii.

Sandstone ($\leq$50cm) but consists mostly of smaller ($\leq$10cm) clasts of silty shale and sandstone which have clearly been derived from the adjacent Kimmeridgian shale facies. The 2m of sandy shale which overlies the 30cm sandstone contains common thin ($\leq$10cm) sandstones which have been slump-folded and distorted. Higher up, a poorly exposed boulder bed occurs on the east edge of the promontory (see Fig. 6.25).

<u>Locality D3 Culgower (NC 989115)</u> <u>eudoxus</u> zone

This locality is situated about 400m to the west of Sron Rubha na Gaoithe and approximately 200m east of the low-lying promontory of the previous sections, from which it is separated by a gap in exposure. At low tide the prominent boulder bed at the top of the section (Fig. 6.28) forms a short but distinct promontory (NC 990115) which is shown on Linsley's Map XII. It should be noted that when Linsley made this map the area was poorly exposed (Linsley, 1972, p.85). A more useful map (made under better conditions,like those existing when I examined the area) is that shown by Bailey and Weir (1932, Fig. 15, p.454) which, although somewhat stylised, illustrates the main features of the field relations in this vicinity, and is largely in agreement with my own observations. It is remarkable to note that although these two maps were made of the same locality and the degree of exposure was not that drastically different, Linsley (1972, p.85) was not sure of this fact. The position of the main 'promontory boulder bed' is also indicated on Fig. 6.25 (panoramic view) and on Fig. 6.29 (based on previous maps and my photographs and field sketches). Because of the low-lying and poorly exposed nature of the outcrop and the apparent inconsistency of the strike, it was found impossible to construct a single continuous stratigraphic section and hence the locality is described in terms of its component areas (in order of decreasing age: D3iii, D3ii and D3i; see Figs. 6.29 and 6.28).

The most westerly and stratigraphically lowest area (D3iii) consists

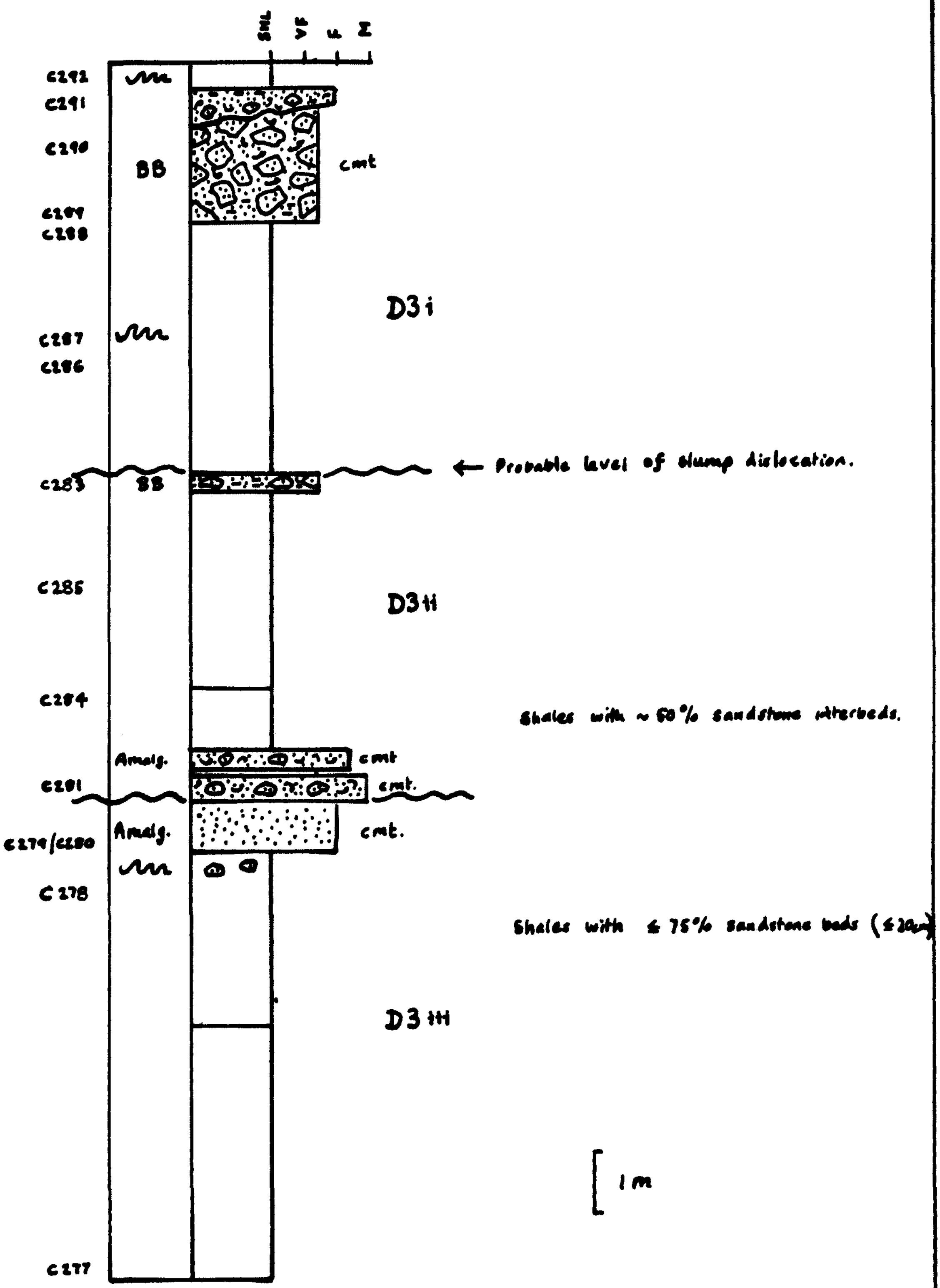


Fig. 6.28 Section D3 Culgower

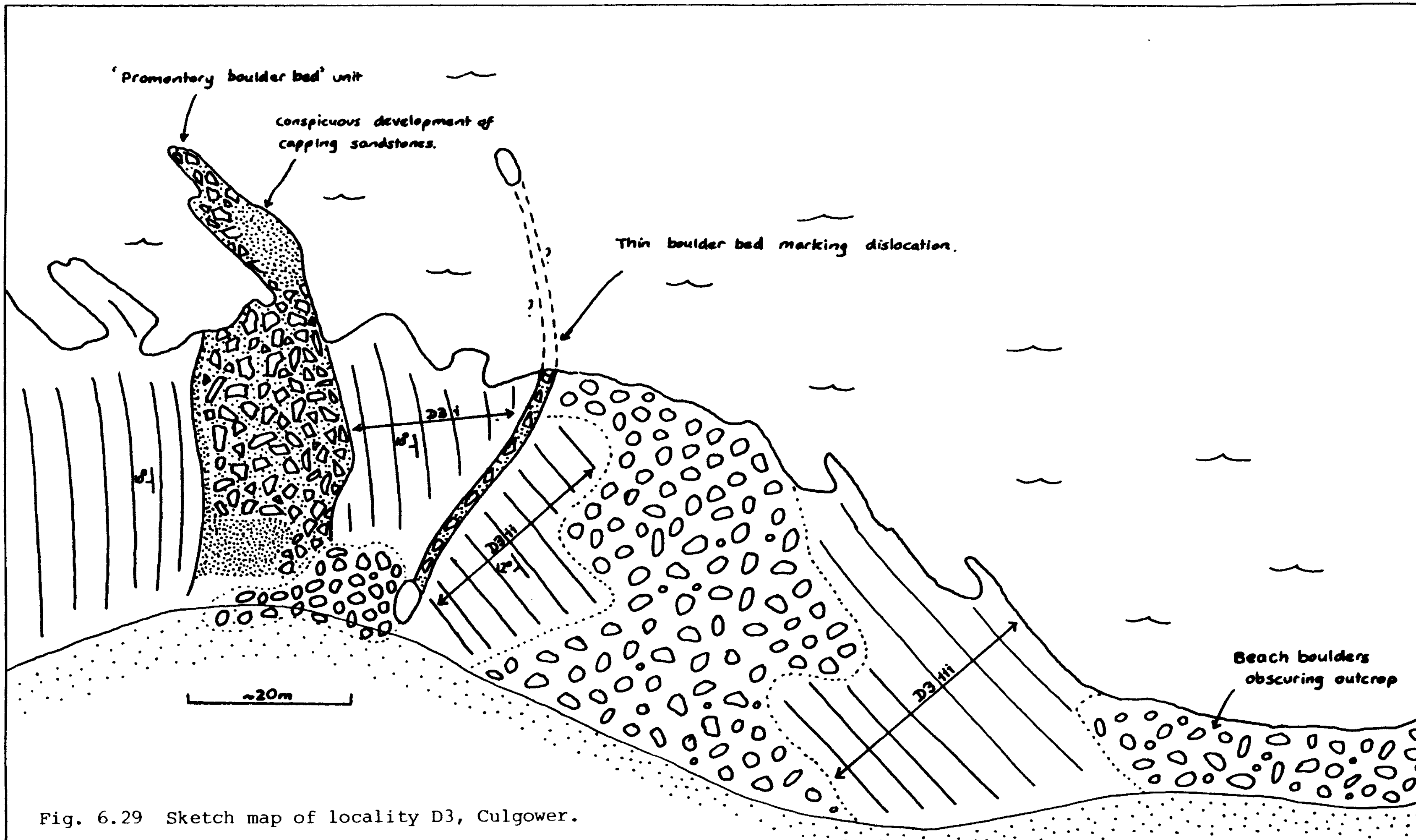


Fig. 6.29 Sketch map of locality D3, Culgower.

of an estimated 7.5m of alternating sandstones and shales. Although exposure is discontinuous, sandstones appear to represent up to 75% of the interval (especially in the upper 3.5m). The majority of the sandstone beds range from 5-10cm in thickness ($\leq$20cm) but in one part of the outcrop (near high water, approximately at the top of the interval) there is an 83cm thick amalgamated unit (see Plate 6.21b). This unit contains several sandstone beds with shale clast-rich tops, and the thicker bands usually contain scattered sandstone (ORS) clasts which are generally $\leq$5cm in diameter but range up to 30cm across. The shale below the amalgamated sandstones is interesting in that it is contorted and contains scattered sandstone clasts (generally 5-10cm, up to 30cm in diameter). The sandstones in this area are predominantly fine grained, are all calcite-cemented (partly lustre-mottled) and are often very shelly, particularly in the lower half of their thickness.

There appears to be only a slight change in strike between areas D3iii and D3ii. The central area, defined at its base by a gap in exposure and at its top by the base of a thin boulder bed, also consists of a sandstone shale sequence which is estimated to be up to 5m in thickness. A clear discordance in strike occurs between this sequence and the base of the overlying boulder bed (as indicated by Bailey and Weir, 1932), such that the thickness of sediments in the D3ii area increases shoreward (mainly shales and thin sandstones) and decreases seaward (shales and a few thick sandstones $\leq$40cm). The sandstone bands are fine to medium grained and sparite cemented; they vary in thickness along strike and where thickest often have very shelly bases containing sandstone clasts (5-20cm in diameter).

The matrix-supported boulder bed which divides areas D3ii and D3i (see Fig. 6.29) is approximately 30cm thick and consists of a very fine to fine grained, dirty, argillaceous sandstone matrix containing angular blocks of bedded Old Red Sandstone which are generally 10-30cm in diameter, but range up to 110cm across. The discordance between this 'bed' and the underlying shale is most pronounced at the shoreward end, although owing to sand cover at the edge of the beach, I was not able to verify the sharp cross-cutting relationship shown by Bailey and Weir (1932, Fig. 15, p.454). Two views of this thin boulder bed (partially obscured by loose boulders) are shown in Plate 6.22. Between this unit and the main boulder bed which forms the promontory, is up to 4m of shale with thin (≤10cm) lenticular sandstones. The sandstones are fine to medium grained, calcite-cemented (lustre-mottled) and often show carbonaceous laminae, are frequently rich in shale clasts and commonly contorted and folded. The shale also contains several 'slumped, slid and rotated' shale blocks and is somewhat undulose. Bailey and Weir (1932, p.453-454) interpreted the thin boulder bed (forming the base of D3i) as a "sill like injection". While I wholly agree that it shows a discordant contact I am extremely reticent to agree with their interpretation and think it much more likely that there is a dislocation surface resulting from 'syn-sedimentary' slumping or 'tectonic' decollement at the base (and perhaps also the top) of the boulder bed. The overlying shale certainly yields plenty of evidence for slumping and sliding and a 'syn-sedimentary' mechanism may be the more likely.

The boulder bed which forms the top of the section and eastern edge of area D3i is an excellent example of this particular lithofacies. It is a chaotic, clast-supported boulder bed 150cm in thickness whose matrix consists of very fine to fine grained sandstone which is dirty and argillaceous in the lower 50cm,and cleaner, variably shelly, sparite cemented and calcite veined in the remainder of the bed. It contains angular clasts of Old Red sandstone (including parallel-bedded, relatively micaceous sandstones, convolute bedded siltstones and interbedded sandstone-siltstones) which are mainly 5-40cm in diameter,but more rarely up to 2.25m in their longest dimension. In a few places the underlying shale has apparently penetrated into the base of the boulder bed (during compaction?). Overlying the boulder bed is a classic example of a

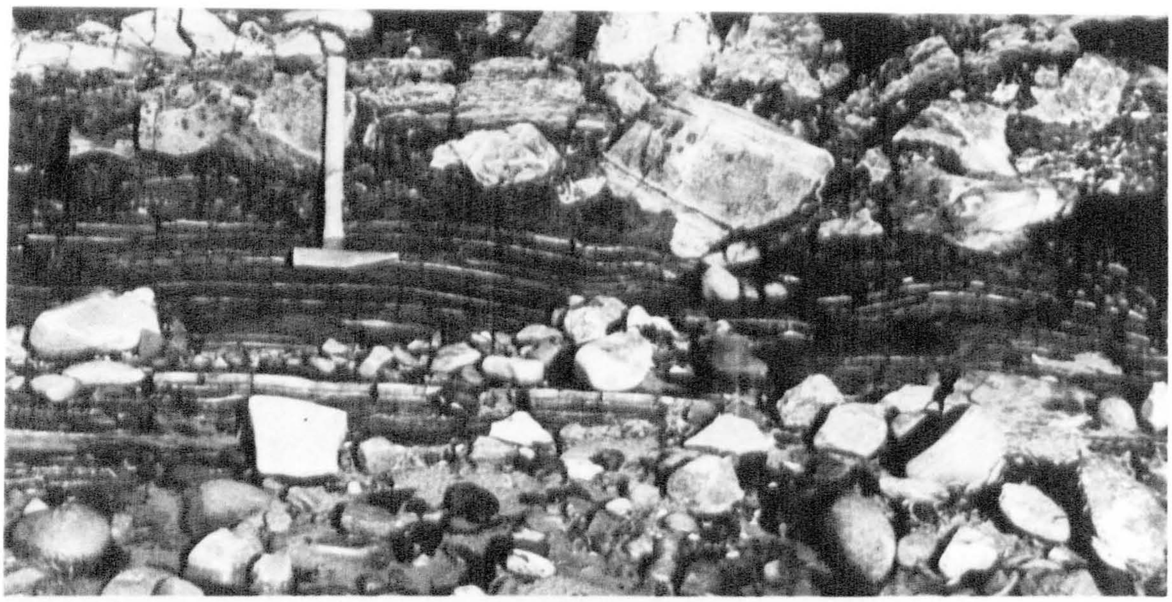


PLATE 6.22 Argillaceous boulder bed unit from section D3 (see Fig. 6.29).

capping sandstone, now mostly removed by erosion, which largely in-filled the irregular topography of the underlying bed and produced a flat top (see Plate 6.23a). The maximum thickness measured for this unit was 60cm, but is variable, possibly amalgamated, and may be thicker in places. The sandstone is fine grained, sparite cemented, moderately to poorly sorted and varies from exceptionally to moderately shelly (especially in the vertical direction). It has a base rich in sandstone clasts which are mainly 6cm or less in diameter (rarely $\leq$30cm); larger clasts projecting through the flat top of the bed probably belong to the underlying boulder bed. The flat capping sandstone is well developed on the eastern flank and shoreward end of the boulder bed where the surface of the sandstone exhibits numerous bedding parallel, platey shale clasts up to 10cm or more across (see Plate 6.23b). Projecting through the beach sand just below high water mark at the landward end of the promontory are several large blocks of Old Red sandstone several meters in diameter. On his map XII Linsley (1972) states that these blocks represent the "locality of the fault suggested by Bailey and Weir (1932)". This is a misconception; nothing (not even the blocks) is indicated on their detailed map of the vicinity (Fig. 15, p.454). The fault they show in the Culgower vicinity on their inadequately large-scale general locality map (Bailey and Weir, 1932, Fig. 2, p.430) was probably copied directly from the geological survey sheet (see Linsley, 1972, p.85). The ORS blocks are probably large clasts in the boulder bed (or loose blocks?).

<u>Locality D4 Srðn Rubha na Gaoithe (NC 992116)</u> <u>eudoxus</u> zone

This locality represents all the outcrop between D3 in the west and the mouth of Westgarty Burn in the east, and includes the low tide promontory of Srðn Rubha na Gaoithe. The area can be seen on the panoramic view in Fig. 6.25 and most of it is also shown on Fig. 6.30 based on Linsley's map XI. Structurally, the locality represents the

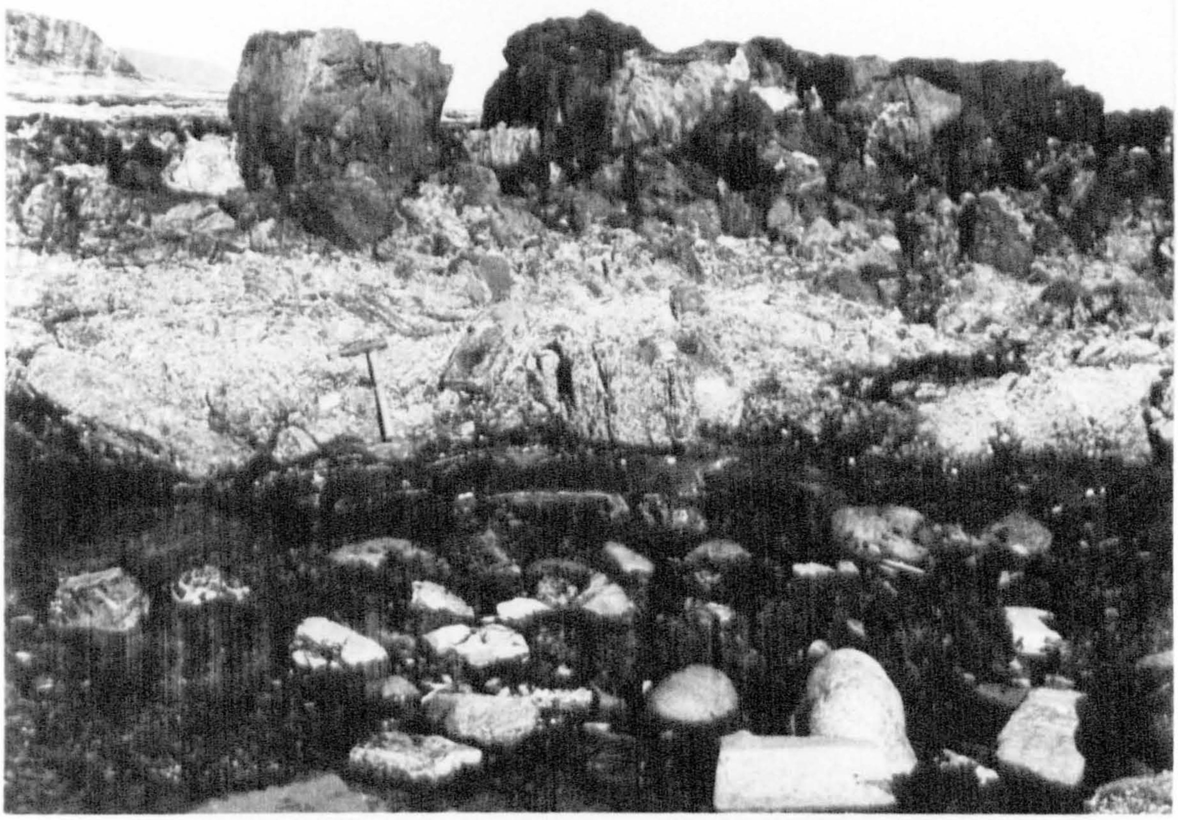
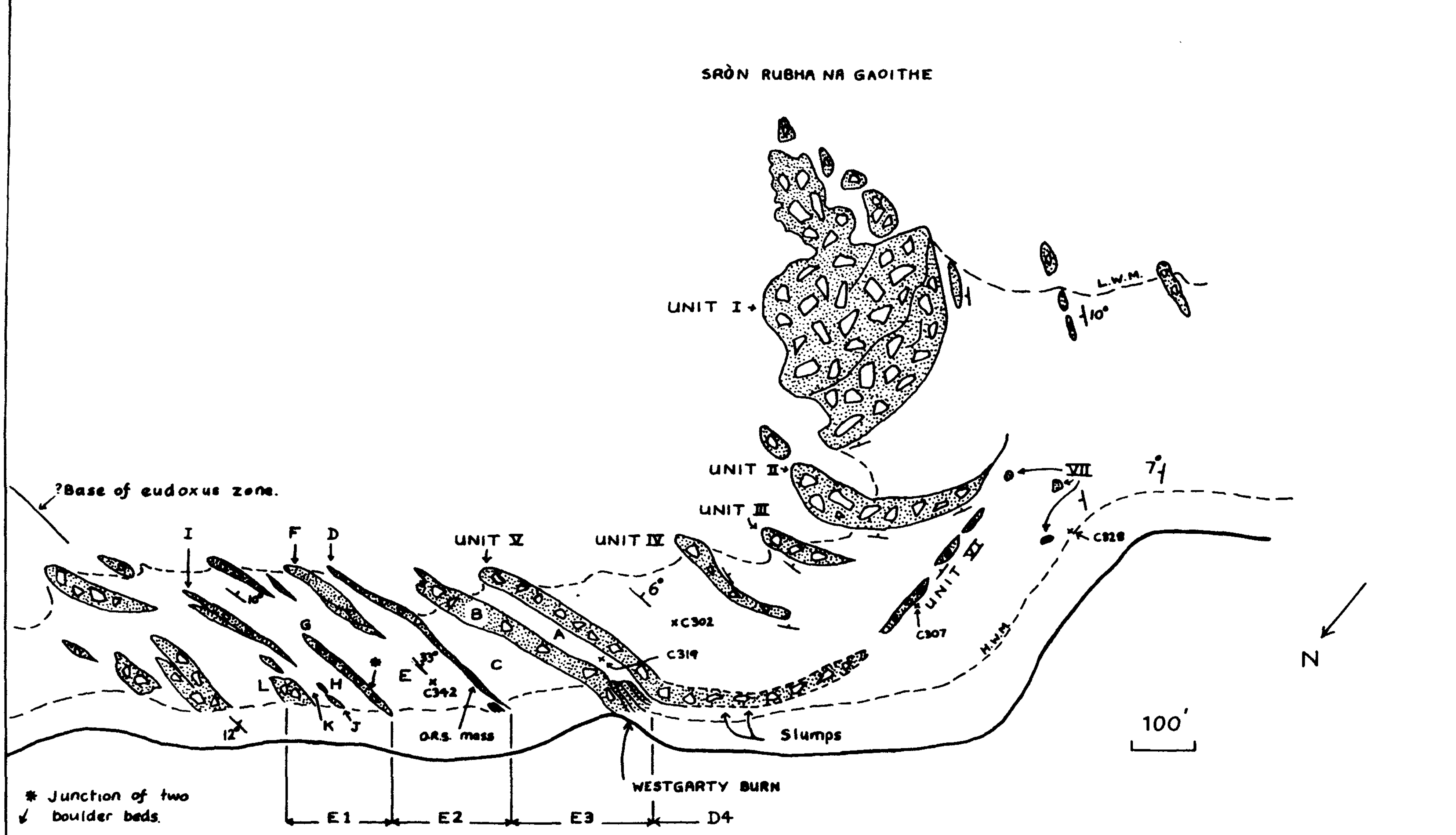


PLATE 6.23 Promontory boulder bed at top of section D3 (see Figs. 6.28, 6.29)

6.23A General view of boulder bed and overling capping sandstone

6.23B Shale intraclast-rich top of capping sandstone

Fig. 6.30 Sketch map of foreshore exposure on Culgower-West Garty beach (after Linsley, 1972)



core of a south-easterly plunging syncline and is very flat-lying with dip values generally less than 10° in the vicinity of the 'headland'. Because of the low-lying nature of the outcrop and the generally discontinuous nature of the sandstone and boulder bed units the area was described rather than logged.

The 150m long strip of outcrop between D3i and the west side of the Sròn Rubha na Gaoithe boulder bed mass consists predominantly of shale. The most westerly 50m of this shale outcrop contains up to 30% of thin (≤10cm) sandstone beds, many of which are highly deformed and there is common evidence of slumping. The most prominent coarse grained units include a slumped, lenticular (channelled?) shelly sandstone (6m x ≤58cm) and a 1m thick amalgamated sandstone unit consisting of at least ten beds of shelly sandstones with common shale clasts (some of which exhibit folded sandstone stringers) and a few sandstone boulders which occasionally reach up to 50cm in diameter. The locations of these units are indicated on Fig. 6.25. The 80cm of shale below the latter amalgamated sandstone unit is partly calcite-cemented, much less compacted and weathers out as lenticular masses measuring 4 to 5m across (along strike). These cemented sandy shales exhibit internal deformation and contain shelly and plant rich layers. The most easterly 100m of this strip of outcrop contains only thin sandstone stringers (≤1cm) and a few thin matrix-supported argillaceous boulder beds such as form the small islands shown on Figs. 6.25 and 6.30.

The remainder of the description of locality D4 will utilise the numbering of the outcrops as shown on Fig. 6.30, with each facies described in turn.

<u>Boulder bed units</u>

<u>Unit 1</u>: this is the flat lying boulder bed mass which forms Sròn Rubha na Gaoithe. It is of unknown thickness and varies from clast to matrix-supported with angular blocks of Old Red sandstone in a sandstone matrix.

The clasts are mainly 10-80cm in diameter but include quite common boulders measuring up to 2 x 4m. The clast lithologies include fine grained sandstones (predominant), red, grey or pale blue-green laminated argillaceous, calcareous siltstones and silty calcilutites and also occasional micaceous metaquartzites. The clasts often consist of alternations of sandstone and siltstone, and commonly exhibit convolute bedding, flame structures, load casts, pseudo-nodules and subaqueous shrinkage cracks (q.v. Donovan and Foster, 1972) but cross-bedding is comparatively rare. The A, B, C and D lithological associations of the Middle Devonian Caithness Flagstones are probably all represented (see Donovan, 1980). In places the boulder bed shows a locally developed argillaceous matrix where the underlying shale has intruded between the boulders; this is particularly true of the western margin of the unit which may represent the 'back' of the bed.

Unit II: this unit is approximately 1m thick, tends to be clast-supported but is highly variable and consists of angular clasts of Old Red sandstone (sandstones and thinly bedded green, grey or brown siltstones up to 3m or more in diameter) in a generally argillaceous sandstone matrix. See above for details of clast types.

Unit IV: this outcrop is an approximately 1.5m thick boulder bed containing up to 50% angular Old Red sandstone clasts in a medium to coarse grained, calcite-cemented sandstone matrix. The clasts are generally less than 50cm in diameter but range up to 2.4m across (see above for lithological details). The boulder bed is underlain by about 80cm of alternating sandstones and shales which appear to have slid with the bed.

Unit V: the outcrop of this unit consists of two parts, one striking NE-SW and the other E-W; only the latter part was shown on Linsley's map XI (perhaps due to poor exposure) and this ommission has been subsequently corrected on Fig. 6.30. The boulder bed is evidently folded and swings around from E-W to lie on strike with outcrop VI, suggesting a dislocation between outcrops II-IV and V-VI. The approximately E-W striking part of the unit consists of up to 1.6m of clast-supported boulder bed with angular clasts of Old Red sandstone in a sandstone matrix. The clasts are generally less than 40cm in diameter but their size is very variable and ranges up to 1.6m across. The boulder bed apparently thickens seaward and is underlain by slumped masses of thinly bedded sandstone up to 1.3m thick. The NE-SW striking part is a much lower lying, broader outcrop with a very low dip. Here the boulder bed is estimated to be between 1-1.5m thick and has a matrix-rich to matrix-supported base with angular Old Red sandstone clasts in a dirty, argillaceous, very fine to fine grained, sparite cemented sandstone matrix. The clasts are mostly between 2-40cm but range up to 3.5m in their longest dimension and include sandstones and siltstones, some of which show convolute and ripple-bedding (see Plate 6.24). The upper part of the boulder bed is clast-supported and the matrix less argillaceous. Approximately 50m south west of Westgarty Burn three tongues of slumped material measuring 2-4m across and trending at approximately 130^{o}, project from below the base of the unit.

Other outcrops

Unit VI: this unit is represented in a conspicuously linear chain of outcrops along strike with the Unit V boulder bed (see Fig. 6.30). It consists of shale with abundant sparite cemented sand laminae overlain by a boulder bed unit which is up to 70cm thick and passes laterally into a rubbly, shelly, calcite cemented sandstone with abundant grey and green mudstone-siltstone clasts. The underlying shales show small sandstone injections.

Outcrop VII: this exposure consists of calcite cemented shales and sandstone stringers but also includes one 10-20cm thick matrix-supported boulder bed with sandstone clasts up to 8cm in diameter (see Plate 6.24b). In parts of the outcrop these cemented shales, which have clearly

PLATE 6.24 Sedimentological features of Culgower beach exposure.

6.24A Argillaceous boulder bed with flaggy ORS clasts; Unit V (see Fig. 6.30)

6.24B Cemented shale mass (with subordinate coarse grained units) exhibiting radial calcite veining; Unit VII (see Fig. 6.30).

suffered much less compaction, can be clearly seen to pass into the ordinary surrounding shales. Two parts of the exposure exhibit remarkable radial and concentric calcite veining (see Plate 6.24c) as previously reported by Macgregor et al. (1930, p.80). In petrographic sections the shales can be seen to be strongly microsparite cemented and also show incipient cone-in-cone cement formation (larger prismatic crystals growing on the margins of sand grains and bioclasts). Linsley (1972, annotation on map XI) apparently thought this outcrop to be merely very finely bedded sandstone.

DISCUSSION OF THE CULGOWER AREA

There are only a few points worthy of discussion in the Culgower area which have not been covered in the locality descriptions above. One of these is the structural relationship between the Culgower syncline and the Crackaig Links outcrops. The most easterly outcrops on Crackaig Links are of lower autissiodorensis or upper eudoxus zone age and dip towards the north-west; the next outcrop (i.e. in the Culgower area) is of lower eudoxus zone or upper mutabilis zone age and shows dips towards the Culgower syncline. Linsley (1972, p.82) says of this situation that "the author was unable to devise a folded structure between the two localities which satisfied the palaeontological evidence, and therefore there must be a fault between them". Although there may be a fault or faults between these areas (the gap in exposure being 1.7km measured NE-SW), there is a possible fold structure which can explain the observed data (Fig. 6.31). This structure, which is no more probable than a fault solution, would predict a structural col to exist between paired anticlines and synclines somewhere to the north west of Sròn Rubha na Gaoithe (although terminated by the Helmsdale Fault).

The other point worthy of brief mention is the apparent decrease in sandstone frequency up through the eudoxus zone (compare locality D4 with D1-D3 and with the West Garty area). Note also the difference in the

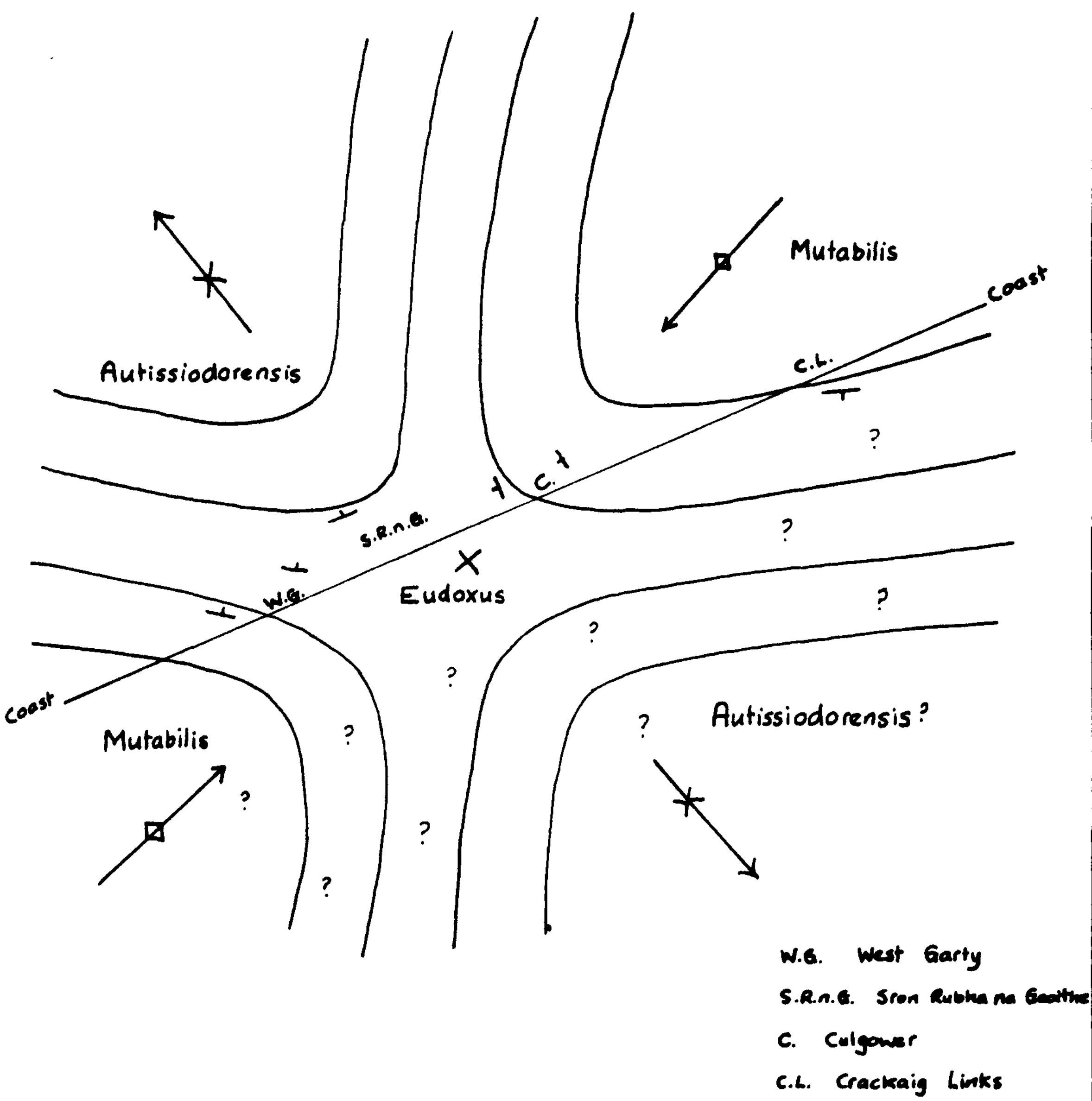


Fig. 6.31 Possible structural relationship of areas C-E.

facies of the Old Red sandstone clasts observed in the boulder beds at locality D4 and those on Crackaig Links.

THE WEST GARTY AREA

For convenience the boundary between the West Garty and Culgower areas was taken at Westgarty Burn; this division is essentially artificial although there are some differences in the facies composition of the two areas (particularly in the proportion of sandstones). The West Garty section respresents the north east limb of the Culgower syncline (here dipping at up to 30° to the south west) and is therefore laterally equivalent (but not lithostratigraphically correlateable) with the sediments exposed in most of the Culgower area. Owing to the strong lateral variation and lenticularity of the coarser grained units (sandstones and boulder beds) the measured sections are by necessity strongly tied to Linsley's map of the West Garty outcrops (map XI, reproduced with slight modification in Fig. 6.30; the individual outcrops have been labelled for simplification and clarity). It should be stressed again that the logging was only undertaken in order to describe the overall nature and proportions of the lithologies and in so doing the general character of the facies. Three continuous sections are presented in order of decreasing age (E1 to E3); all are of <u>eudoxus</u> zone age according to Linsley's map, but it is apparent from the distribution of his ammonite specimens that all of section E1 and at least half of section E2 <u>could</u> be of <u>mutabilis</u> zone age. Boulder beds are here defined as containing >10% clasts.

<u>Section E1 West Garty (NC 993119)</u> <u>eudoxus</u> (or <u>mutabilis</u>?) zone

The base of section E1 occurs in outcrop L approximately 150m to the north east of the mouth of Westgarty Burn (see Fig. 6.30). For details of the section see Fig. 6.32; only specific points are discussed below. The lowest bed of the section is particularly interesting in that it shows a very well developed erosional (partly channelled)

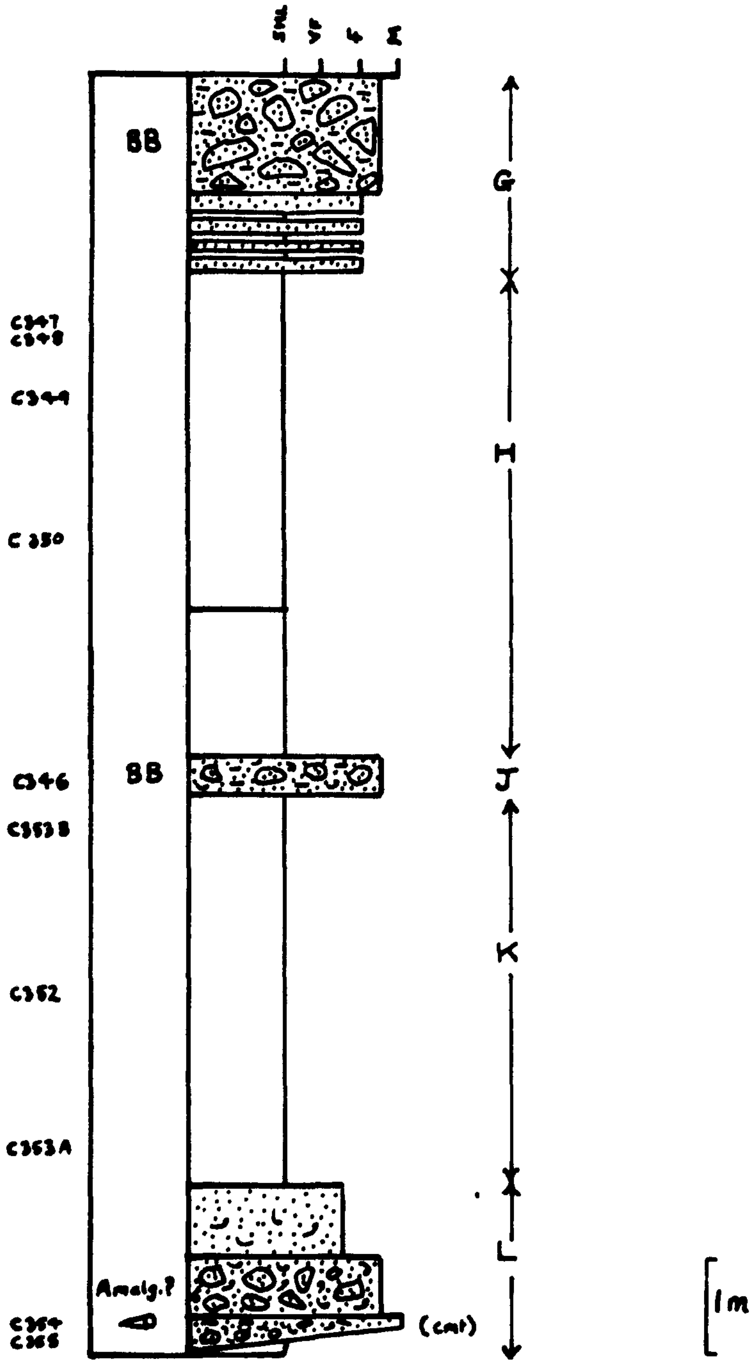


Fig. 6.32 Section El West Garty.

base (see Plate 6.25a). The irregular sand laminae visible beneath the bed are uncemented and the sandstone itself shows variable cementation, passing in a few centimetres from calcite-cemented (lustre-mottled) to friable, uncemented sand with widespread authigenic kaolinite in the pores and common quartz overgrowths (absent at grain contacts and therefore formed in situ). The top of this composite ($\leq$40cm) sandstone bed is sometimes rich in shale clasts and its upper surface contains relatively common belemnites oriented at about 140°-150° (not structurally corrected). It is overlain by a shelly boulder bed and more sandstone (see log). Separating the upper and lower shale intervals in section E1 is a small, lenticular (4m x $\leq$40cm), matrix-supported Old Red Sandstone boulder bed (outcrop J, not previously shown by Linsley, 1972), which is laterally equivalent to a composite sandstone unit (outcrop I). The composite unit consists predominantly of very shelly, calcite cemented sandstones which contain up to 10% floating Old Red sandstone clasts ($\leq$40cm diameter) and show a crude layering due to the subparallel orientation of the bioclastic debris. Subordinate lithologies include about 15% intercalated shale, a few true thin ($\leq$30cm) boulder bed lenses and common shell-free sandstones with carbonaceous and/or shale clast-rich laminae. These sandstones are also partly equivalent to the shale and sandstone sequence in the lower 1.5m of shale outcrop H. The boulder bed at the top of section E1 (outcrop G) is, along with those in outcrop D, one of the few clast-supported, true boulder beds in the West Garty sequence. Its clasts are generally 50cm or less in diameter, but one individual block exposed at the beach end of the outcrop measured 1.5m x 2.3m x 0.36m. In detail outcrop G can be seen to be a composite unit formed by (at least) two overlapping boulder beds; the point where these beds meet is indicated on Fig. 6.30.

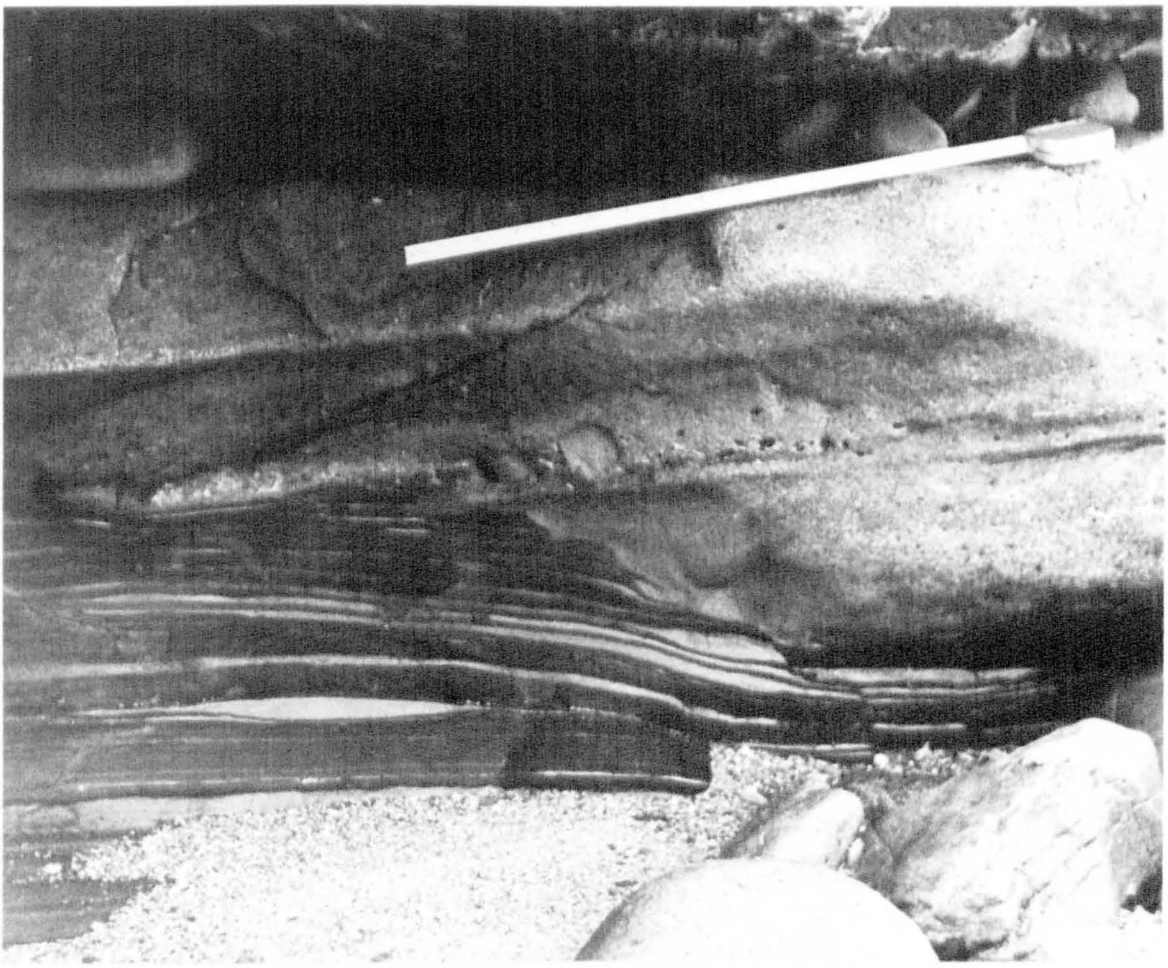
Section E2 West Garty (NC 993119) *eudoxus* zone

Section E2 follows on directly above the top of outcrop G in section

PLATE 6.25 Sedimentological features from West Garty exposures

6.25A Marked channeling at base of section E1 (Unit L, see Fig. 6.30)

6.25B River cliff exposure of shales cut by sandstone dykes, section E4.



El (see Fig. 6.30). The bulk of this section is represented by shale with common thin ($\leq$1cm) sand or sandstone stringers (outcrop E) but only a few pure or sparsely shelly, thin ($\leq$5cm) calcite-cemented sandstones. The most interesting parts of the section are outcrops F and D. Outcrop F is a lenticular sandstone unit which occurs about 3m below the top of the shale outcrop E and is somewhat unusual in that the sandstones are predominantly pure or only sparsely shelly and are free of clasts. Although there are a few intercalated boulder beds in the middle of the outcrop most of the unit is composed of thinly bedded (1mm-3cm) pure, calcite-cemented,fine grained sandstones with abundant shale clast and/ or wood debris-rich laminae.

Outcrop D consists of a series of complicated, amalgamated boulder bed and sandstone units with intercalated shales. The measured section presented in Fig. 6.33 was taken on the shoreward side of the large prominent Old Red sandstone mass whose position is indicated on Fig. 6.30; this mass measures 5.8m parallel to, and 4m normal to <u>its</u> bedding. The boulder beds usually consist of a shelly sandstone matrix containing up to 75% Old Red Sandstone clasts (sandstones, siltstones and pale grey, slightly silty, microsparitic limestones); relatively clast free and sometimes shale clast rich 'capping sandstones' are common. Two instances of the complex lateral and vertical relations of the boulder beds are shown in Fig. 6.34; note the erosive bases, sandstone caps, laterally variable lithologies and amalgamation. The interbedded shales in outcrop D contain lenticular sandstones ($\leq$10cm) and localised lenticular boulder beds (to bouldery sandstones) and slumped blocks of Old Red siltstones; one such block in the upper shale measures 3m x 1.7m x 0.5m. The sandstones (including the boulder bed matrices) are all fine grained (occasionally fine to medium grained) but often contain very coarse grained bioclastic debris and granule grade clasts.

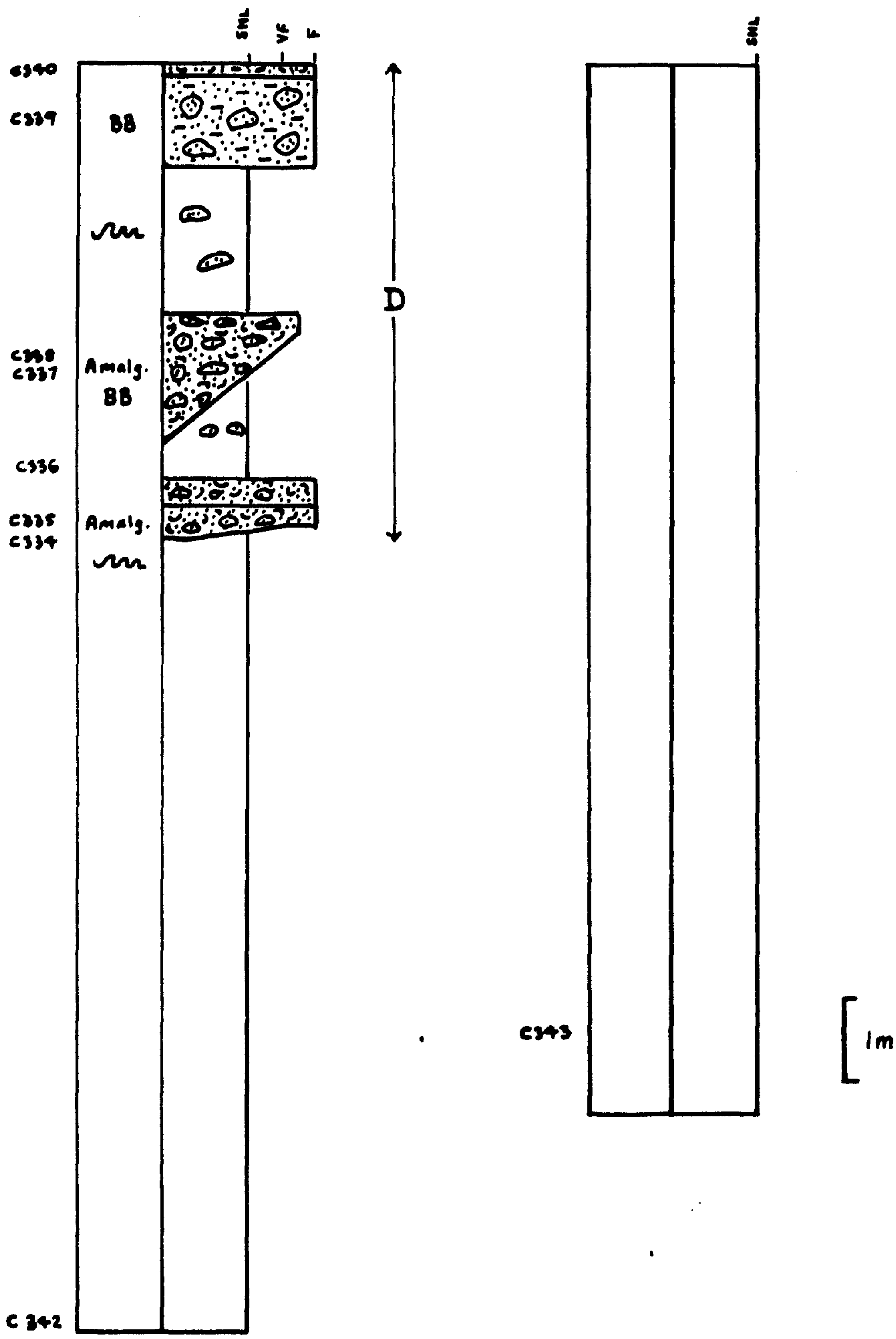


Fig. 6.33 Section E2 West Garty.

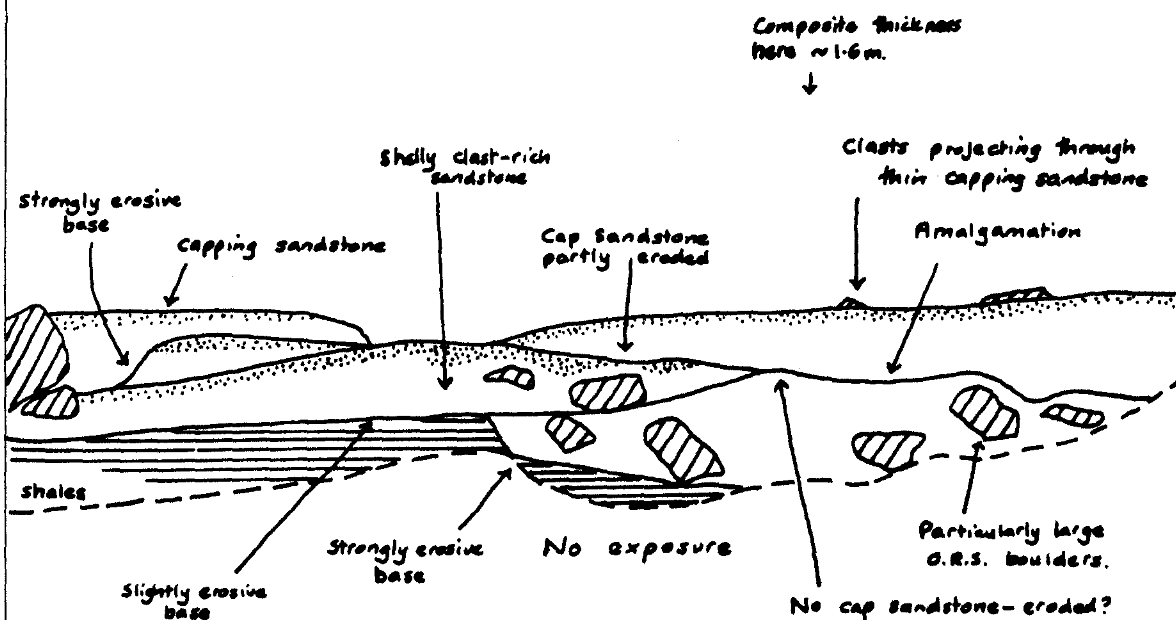
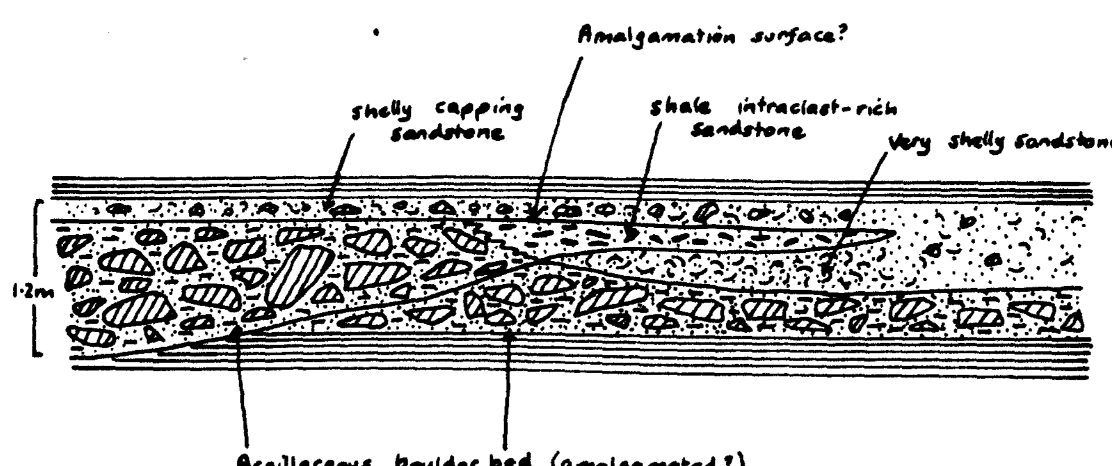


Fig. 6.34 Field sketches of lateral variability in units of section E2.



Section E3 West Garty (NC 992119) *eudoxus* zone

The lower part of the section shown in Fig. 6.35 consists of shales with thin sandstones, and the upper part an alternation of interbedded shale, 'normal' sandstone, clast-rich sandstone ($\leq$10% clasts) and boulder beds ($\leq$25% clasts). The sandstone beds in the lower shaley part of the section are generally 4cm or less in thickness and vary from shelly to pure (the latter with or without shelly basal layers) and often contain plant and shale clast rich laminae. The lower shale division also contains several slumped and partly rotated blocks of shale and shale-sandstone alternations up to 3m in diameter, and occasional isolated blocks of Old Red Sandstone lithologies ranging from a few centimetres to 70-100cm in their longest dimension. The sandstones and boulder bed matrices of the upper part of the section are all extremely shelly, calcite cemented and often calcite-veined; total calcite contents are often in the range 50-75% of the rock. The Old Red Sandstone clasts in both the sandstones and boulder beds consist predominantly of parallel, convolute or cross-bedded sandstones and are mainly less than 20cm in diameter. Many of the coarse grained beds are clearly composite, amalgamated units since they often include thin remnant shale horizons and show irregular clast distributions. Three sandstone dykes measuring up to 18cm in diameter were observed to cut a lenticular shelly sandstone bed near the top of the section. Approximately 3-7m of shale (outcrop A) occur between the sandstones of outcrop B and the base of the unit V boulder bed of locality D4.

Locality E4 Westgarty Burn (NC 991120) *eudoxus* zone?

A small exposure of shales and sandstones occurs on the east side of the mouth of the Westgarty Burn valley (see Plate 6.25b). The shales contain abundant lenticular and irregular (continuous) uncemented fine grained sand laminae which are generally 1-2cm thick and exhibit small scale overfolds, boudinage and injection structures (note sandstone

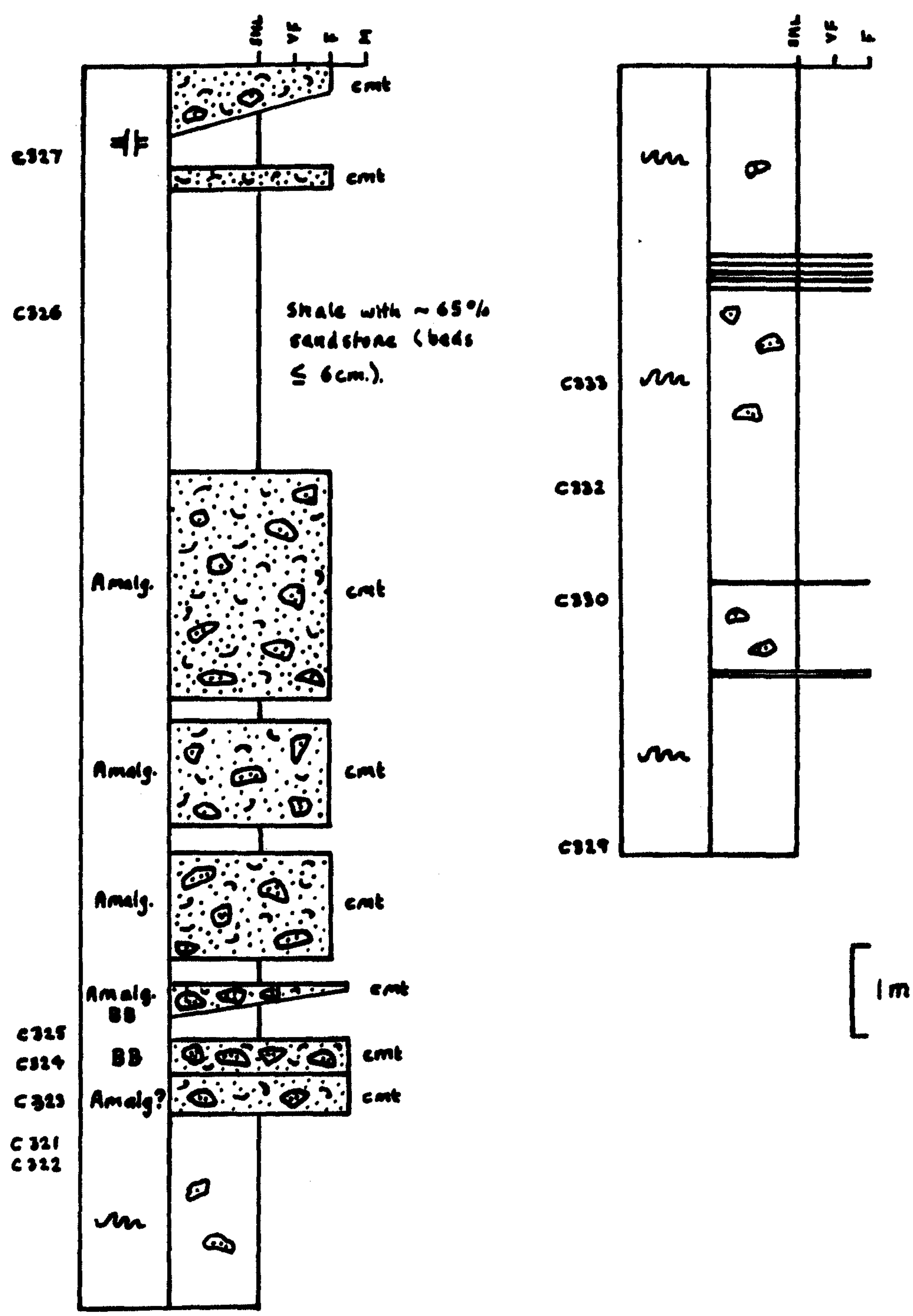


Fig. 6.35 Section E3 West Garty.

dyke in centre of the photograph). At the top of the exposure there are several thicker bands (5-20cm) of extremely shelly, calcite cemented, fine to medium grained sandstones with small ORS pebbles; the skeletal component consists mainly of oyster (and brachiopod?) debris with lesser amounts of echinoderm (echinoid) fragments, and one piece of coral was observed in thin section. Further upstream (near the end of the stone wall which runs up the centre of the valley) there are a few, poor, river cliff exposures of boulder beds.

No discussion is offered for the West Garty area.

THE SEQUENCE BETWEEN WEST GARTY AND ORD POINT

Except for brief reconnaisance visits the sequence north east of the West Garty sections was not examined during this present study. The following comments are,therefore,based largeley on existing accounts (principally Bailey and Weir, 1932 and Linsley, 1972). Apart from his detailed maps (which bear only rather superficial annotations) Linsley's thesis contains little in the way of specific sedimentological observations that could not already have been gathered from earlier works and so only a very general account can be offered here. The chief localities, their maps references, zonal ages and the numbers of corresponding detailed maps made by Linsley (1972) are listed in Table 6.2 (see Fig. 6.1 for map of localities).

As the width of the Brora outlier decreases north eastwards from West Garty the most noticeable change in the sequence is the increasingly conspicuous presence of thick, lenticular, clast-rich boulder bed units and the increase in the size range of the boulders. Thick boulder bed masses are particularly well developed between Helmsdale and Ord Point, but three such units occur in the Portgower area (the thickest measuring $\leq$24.5m) and one at Midgarty (see below). With the exception of parts of the Helmsdale, Navidale and Dun Glas areas, boulder beds remain subordinate to the sandstone-shale sequence and the facies development is essentially the same as that in the West Garty and Culgower sections: shales with

TABLE 6.2 Kimmeridgian-Volgian sections north east of West Garty, Brora Helmsdale outlier

Area	Map references	Ammonite zones*	Corresponding maps in Linsley, 1972
Midgarty	NC994119 - NC999124	mutabilis	X - IX
Portgower	NC999124 - ND010136	mutabilis - autissiodorensis	IX - VII
Gartymore	ND010136 - ND019142	autissiodorensis - elegans	VII - VI
Green Park (W. Helmsdale)	ND020144 - ND022146	wheatleyensis	VI
Helmsdale	ND031151 - ND042156	hudlestoni - ?pallasioides	V - IV
Navidale	ND042156 - ND047164	?pallasioides - glaucolithus	IV - II
Sput Burn - Ord Point	ND047164 - ND062174	?pallasioides - glaucolithus?	II - I

* Linsley (1972) and Riley (1980)

thin sandstones, shelly bouldery sandstones and thin ($\leq$1-2m) boulder beds. Another notable feature of the upper part of the sequence (especially north-eastwards from northern Gartymore) is the presence of rounded coral masses ($\leq$45cm; principally Isastraea) amongst the Old Red Sandstone blocks in the boulder beds. Particular features and aspects of some of the localities are discussed below.

Midgarty

The central part of the Midgarty exposure (opposite the end of the small valley leading down to the beach from West Garty Farm) contains a prominent 38m thick boulder bed unit. Linsley (1972, p.27) notes that the bedding in this unit is defined by variations in clast size wich indicate "at least four periods of deposition". Clast sizes range from a few centimetres upward to a maximum of 36m (longest dimension) and blocks measuring 1.5 to 6m in diameter are common in some parts of the unit. The top of the boulder bed contains the largest clasts,which nearly all consist of light brown sandstone similar to that observed in boulder beds A and C at Crackaig Links (Linsley, 1972, p.84) and resemble the John O'Groats sandstone facies of the middle Old Red Sandstone. The boundary between the Midgarty and south westerly part of the Portgower area is marked by a fault.

Portgower to Gartymore

Bailey and Weir (1932, p.445-6) have presented a measured section for the strip of outcrop occurring between the fault at NC 999124 and ND 019142. This section consists of 310m of shales (318.5m according to Linsley, 1972, p.72), including 22% of boulder beds and 7.2% of shelly sandstones. It is apparent from Bailey and Weir's Figure 4 (p.445) that their section must terminate some 125m to the west of the 'fallen stack' boulder bed at a level just above the lowest thick boulder bed unit in the south east part of the Portgower area. However, Bailey and Weir (1932, p.445) state that their section "starts ... with 20 feet of soft, well-bedded sandstone", apparently corresponding to Linsley's statement (1972, p.85) that "the lowest 60 feet of the succession is formed by a yellow-brown to blue-grey sandstone with thin argillaeous laminae", which, however, refers to the sediment immediately to the north of the fault at NC 999124 and as much as 60m below the apparent base of Bailey and Weir's section! No similar sandstone occurs above the lower thick boulder bed unit (Linsley, 1972, map IX).

Three main thick boulder beds occur in the Portgower area. The lowest (mentioned above) is about 20m thick, clast-rich with "most blocks <10'" in diameter and is clearly composite, with thin shale beds picking out the bedding (Linsley, 1972, p.87 and map IX). Proceeding north eastwards the next thick boulder bed unit is that containing the famous 'fallen stack' block. This bed is up to 13.7m thick, but is clearly lenticular and thins shoreward. The remarkable Old Red Sandstone boulder, the so-called 'fallen stack' or 'Portgower giant', consists of an alternation of shrinkage cracked, rippled and convolute bedded "soft red sandstones, grey, green and red mudstones and bands of cream-coloured limestone" (Macgregor, 1916, p.81). Various authors have given its dimensions; the maximum width has been consistently given as 27.4m, although less agreement is evident concerning the maximum length and height: 45.7m (L) x 9.1m (H) according to Macgregor (1916), 30.5m (L) x ~6.1m (H) according to Bailey and Weir (1932), and 33.5m (L) x 4.6+m (H) according to Linsley (1972). The large value given by Macgregor (1916) for the length of the 'fallen stack' probably includes the small island which lies directly offshore and along strike from the main mass, to which it is, or was, probably connected. Remarkably, this formidable mass only disturbs the underlying shale to a depth averaging 15-20cm (Linsley, 1972, p.87). The final thick boulder bed unit occurs 400m further along the shore and consists of 24.4m of amalgamated boulder beds with a few thin shale bands; the clasts are mainly flaggy sandstones, the matrix is rich in coarse bioclastic debris and capping sandstones are present (Bailey and Weir, 1932; p.446; Linsley, 1972; p.89, map VIII).

Bailey and Weir (1932, p.446) note that the thick boulder beds described above are "quite exceptional" and that most of the Portgower-Gartymore boulder beds are about 1m thick, with about 25% of the occurrence being only 30cm in thickness (although not with necessarily smaller clasts). The presence of sandstone caps (pure or less shelly, clast-free or clast-poor) gives the boulder beds an overall fining upward pattern, although much of the grading is compostional rather than textural. Many of the interbedded sandstones exhibit clearly erosional, clast-rich bases, are commonly amalgamated and are frequently so rich in bioclastic debris and cement that they become sandy skeletal grainstones (limestones). At the north east end of the Gartymore outcrops the shales become very sandy and sometimes pass into uncemented, bluish-brown sandstones with carbonaceous shale partings (see Linsley's maps VII and VI).

Helmsdale

The Helmsdale area consists of extensive boulder bed units and clast-rich shelly sandstones which together exceed the thickness of shale (although this may not have been true prior to compaction). Bailey and Weir (1932, p.450) note that the Helmsdale outcrops show a "spreading out of individual boulder deposits into extensive beds" which at least appear to extend up to 380m or more along strike according to their representation on Linsley's map V. Bailey and Weir (1932, p.450) also report that: "In the great majority of the Helmsdale boulder beds the top is levelled down with rubble and shelly sand, and presents a flat surface to the succeeding shale. By contrast, the basal part tends to contain larger boulders and to be irregular and transgressive". Linsley's map annotations indicate that the thick boulder bed units ($\leq$31m) often contain thin shale bands which pick out the bedding (and define cyclic sequences), and that the shales are often very sandy, decalcified and have common small-scale injection structures.

Navidale

Bailey and Weir (1932, p.450) note that "smooth rubble tops" are particularly well-developed in the Navidale area, and the large boulder bed masses are formed of rhythmic repetitions of thin shales, boulder beds and capping shelly sandstones. Many of the boulder bed units are relatively laterally extensive (e.g. in the Navidale Bay anticline and the anticline south of Sput Burn) but there are also several "excessively bulbous lenses" which Linsley (1972, p.94) believes must have been deposited in channels within the shales. In several parts of the outcrop the boulder beds can be seen to pass laterally into very sandy shales (e.g. core of Navidale Bay anticline). The sandstones and boulder bed matrices are extremely rich in bioclastic debris; large pieces of Coral (Isastraea) are common in places and Macgregor et al. (1930, p.79) note the presence of sandstones composed largely of Nanogyra shells. It was from the Navidale Bay exposures that Riley (1980) recently obtained palynological samples indicating an early Portlandian albani-glaucolithus zone age, appreciably younger than Linsley's (1972) previous estimation of a pallasioides zone age for the top of the succession.

Sput Burn to Dun Glas

From about 150m east of Sput Burn to the Dun Glas headland the sequence consists almost exclusively of amalgamated, clast-rich boulder beds. In many parts of the exposures there are no, or only laterally impersistent, thin shale horizons, and bedding can only sometimes be ascertained by the presence of the relatively clast free capping sandstones. The boulder bed is particularly massive and structureless in the immediate vicinity and 200m to the west of Dun Glas, where according to Macgregor (1916, p.80) the clasts consist of light grey sun-cracked sandstones, dark flaggy sandstones and light-coloured sandy limestones. Macgregor's account continues: "One hundred yards or so further south along the shore (from Dun Glas) the blocks are mainly composed of flaggy red and grey mottled mudstones often as large as 10 x 6 x 4 feet. Further south again this type is replaced by yellow and brown marly sandstones. The blocks along this section of the coast are very numerous, and lie heaped upon one another without the least order or arrangement. While practically all angular, they vary enormously in size and shape. Only a few among the smaller ones show an approach to roundness". Between Allt Briste and Dun Glas the Helmsdale Fault crosses the beach and then runs parallel to the shore just above the high water mark. It is therefore difficult to say how much the apparent proximal shift in facies (increasing proportion and density of the boulder beds) is due to variation in time and how much to variation in space.

Total thickness of the Kintradwell-Ord Point sequence

Linsley (1972, p.77) estimates the total maximum thickness for the whole succession to be something in excess of 973m. Bailey and Weir (1932, p.445) gave a minimum thickness of 462m and considered that the aggregate movement of the Helmsdale Fault has "much exceeded" 610m (p.457).

PETROGRAPHIC OBSERVATIONS ON THE KINTRADWELL - WEST GARTY SEQUENCE

No systematic detailed petrographic studies were carried out during my work on these sediments, but general observations were made on some 171 thin sections, with particular emphasis placed on the sandstones; some of the findings are summarised in Table 6.3 (see also Plates 6.26 and 6.27).

Grain Size

There is only a comparatively small range of modal grain sizes in the Kintradwell - West Garty section. Some 87% of all samples (i.e. thin sections) show a modal grain size lying between 0.1 and 0.3mm, and half of this total lies between 0.125 and 0.25mm (fine grained on the Wentworth scale). Only 6% of the sample fell in the range 0.3 to 1.0mm (medium to coarse grained) and the only modal siltstones observed were clasts of Old Red Sandstone lithologies. Since the distribution of coarse bioclastic debris and granule and pebble grade lithoclasts clearly indicates the periodic presence of currents which were strong enough to have transported coarser sand grains had they been present, this limited range of grain sizes presumably reflects the provenance of the sediments rather than sorting.

Sorting

Sandstone samples show variable sorting with a slight peak in the poor to moderately sorted range (see Table 6.3). Bouldery sandstones and boulder bed sand matrices are, not surprisingly, much less well sorted than the 'normal' sandstones and just under three-quarters of the thin sections in these lithologies showed very poor to poor sorting. It is interesting to speculate on the extent to which the replacive calcite cements have effected the apparent degree of sorting; the cements have often clearly corroded and embayed the edges of the quartz grains and it is probable that silt and even very fine sand grade material could have been totally replaced. The shales vary from pure to very silty and often

TABLE 6.3 Petrographic data

Lithological distribution of slides, cementation and sorting

			Calcite cemented samples		Sorting		
	No.	%	%	No.	VP - P	P - M	M - G
Sandstones	109	64	76	83	27%	42%	24%
Boulder bed matrices	24	15	80	19	71%	25%	4.0%
Shales	22	13	55	12*	-	-	-
Boulder bed clasts	16	9	88	14	–	–	–
	171			128			

* 33% of cemented shales show incipient cone-in-cone cements

Modal sandstone and boulder bed matrix grain size:

0.1 - 0.3 mm	(1.75 - 3.25 ϕ)	VF to F - M	87% of slides
0.125 - 0.25 mm	(2.0 - 3.0 ϕ)	F	43% of slides
0.3 - 1.0 mm	(1.75 - 0.0 ϕ)	M - C	6.0% of slides

Modal siltstone grain sizes occur only as Old Red Sandstone clasts

Calcite cementation type in sandstones and boulder bed matrices:

	% all cemented samples	% cemented boulder bed matrices	% cemented sandstones
Lustre-mottled (poikiloptic) cements	53	10	70
Granular sparite cements	34	53	27.2
Microsparite cements	12	37	2.8

Incipient cone-in-cone cements occur in seven samples: shales (K109, K241B, C314, C318), sandstones (K203) and boulder beds (C310, K237)

TABLE 6.3 (CONT) Petrographic data

Sandstone composition based on 20 point counted slides:

Sandstone type	%	No.	Mean % cement or porosity	Mean % bioclasts
Calcareous quartz arenites	5	1	33	0
Calcareous sublitharenites	40	8	29.4	6.3 (0 - 66%)
sublitharenites	35	7	20.9	0
Calcareous lithic arenites	20	4	39	0.8

N.B. Boulder bed matrices are lithic arenites

Mean carbonate contents of cemented sandstones and boulder bed matrices:

	Visual assessment	Point counts
Mean % cement	39%	35%
Mean % bioclasts	16%	7%
Mean total calcite	56%	43%

Comparisons of cemented sandstone and boulder bed matrix carbonate contents for areas A - C and D - E:

	Kintradwell - Crackaig	Culgower - West Garty
Mean total calcite*	50%	60%
Mean bioclasts*	7%	24%
Mean cement*	45%	34%

* Figures based on visual assessments utilising visual aids

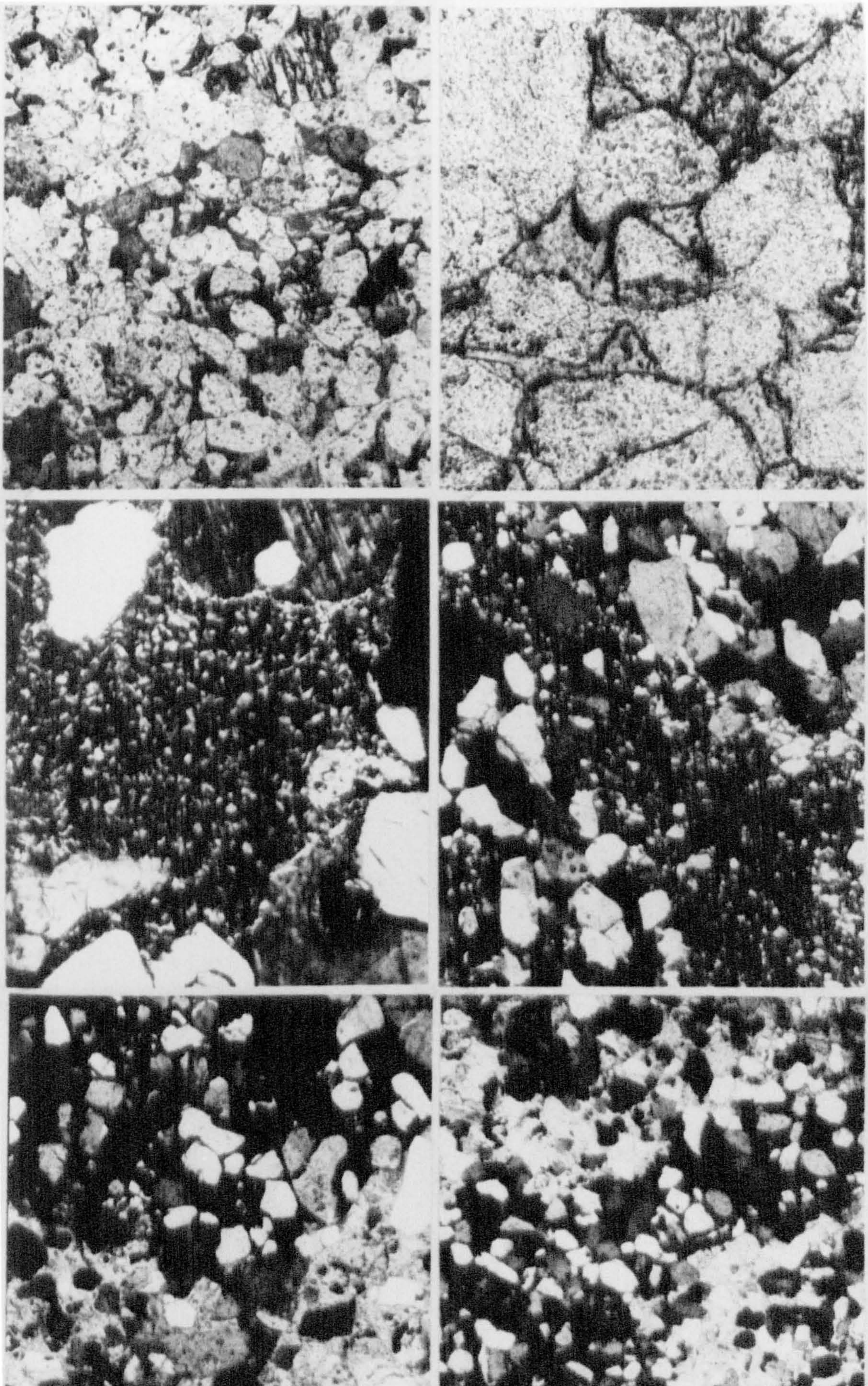
PLATES 6.26 and 6.27. Petrographic features of Kimmeridgian sediments from the Brora-Helmsdale outlier. Figures in parentheses denote widths (long dimensions) of photographs in millimetres.

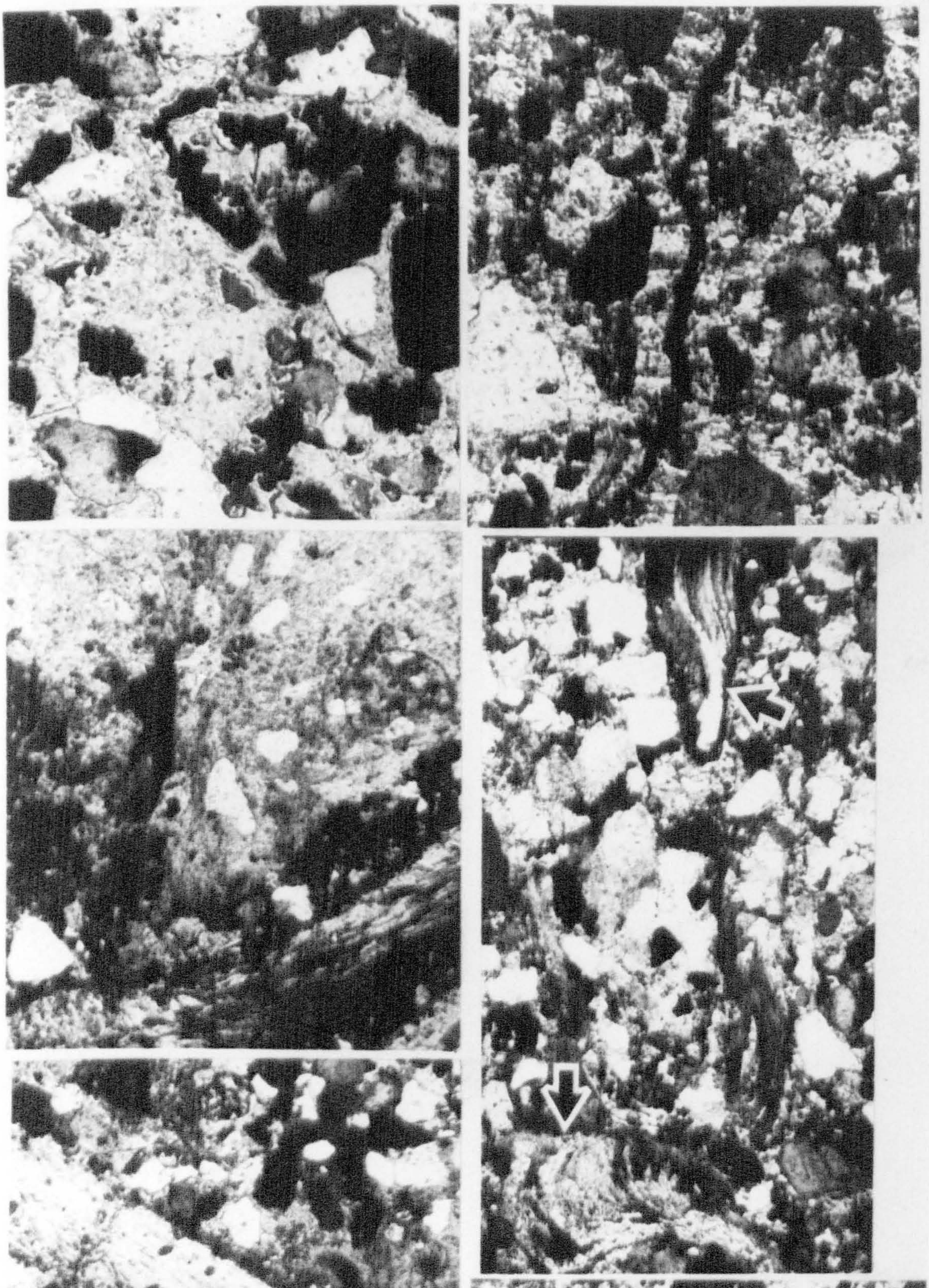
PLATE 6.26

Top Left:	Porous sandstone; darker colour is stained impregnating medium. Sample K 78 (2.2)
Top Right:	Porous sandstone showing slight pressure solution; darker rims are ferruginous 'cement'. Sample K 79 (0.6)
Centre Left:	Kaolinite pore filling aggregate in illite lined pore. Sample K 76 (0.6)
Centre Right:	Fracture cementation ("veining") in Lothbeg sandstone. Sample K 85B (2.2)
Bottom Left:	Bleb of opaque cement in calcite cemented sandstone. Sample K 28 (2.2)
Bottom Right:	Poikoloptic (lustre mottled) calcite cement. Sample C 324 (2.2)

PLATE 6.27

Top Left:	Highly corrosive carbonate cement. Sample K 64 (0.6)
Top Right:	Incipient cone-in-cone fabric/cement. Sample K 203 (0.6).
Centre Left:	Coral debris in shelly sandstone. Sample C 346 (2.2)
Centre Right:	Quartz- replacement of shell debris (conformable with shell microstructure; lighter areas in arrowed bioclasts are quartz). Sample C 324 (3.2)
Bottom Left:	Algal-bored shell debris. Sample C 334. (2.2)
Bottom Right:	Chalcedonised ?calcisphere; note also cone-in-cone cement fabric at rim. Sample C 314 (0.4)





contain numerous thin laminae of very fine to fine grained sand; the matrix silt or sand fraction is often comparatively well sorted.

Sandstone composition

The composition of the sandstones was mainly judged by eye with the use of visual comparison charts for the assessment of percentages; this semiquantitative analysis was supplemented by point counting a sample of 20 sandstones (see Table 6.4). The sandstones were named according to the classification of Pettijohn (1975, Fig. 7-6, p.211). They are predominantly sublitharenites (∿75%), but lithic arenites are common (∿20%) and rare quartz arenites are also present. Boulder bed matrices are mainly lithic arenites. The lithic grains in the sandstones are predominantly composite quartz grains (quartzites and minor schistose and gneissose rock fragments) but in the boulder bed matrices there is usually also a significant component of granule grade 'Old Red' siltstones. The proportion of lithic grains increases as grain size increases and sorting decreases, and hence the coarser and more poorly sorted sandstones and the boulder bed matrices tend to be lithic arenites. This relationship is evident in one of the graded beds from section A1. The base of the bed is medium grained, contains 28% lithic quartz and is a lithic arenite, while the upper part of the bed is fine grained, better sorted, contains 33% <u>less</u> lithic grains and is a sublitharenite (see samples K65B and K65T, Table 6.4). The average feldspar content of the sandstones is 2-5%; it tends to be slightly higher in the lithic arenites which sometimes contain small amounts of quartzofeldspathic lithic grains. Trace amounts of glauconite are usually present.

Since some 76% of the sandstones examined in thin section are calcite cemented, the prefix "calcareous" is usually appropriate; uncemented sandstones are only common in the Lothbeg area. Many of the calcite-cemented beds contain significant amounts of bioclastic debris, while the uncemented units tend to have been decalcified and leached.

TABLE 6.4 Summary of point count data (see Appendix 6F). Sandstone terminology after Pettijohn (1975) based on normalised quartzo-feldspathic grain composition

Sample	(Section)	Grain Composition (%) Quartz	Lithics	Feldspar	Grain Size	Name	% Voids	% Cement	% Bioclasts
K 65B	(A1)	67	28	6	M	Lithic arenite	0	40	0.2
K 65T	(A1)	75	18	7	F	Sublithic arenite	0	37	0.4
K 11	(A3)	95	3	2	F	Quartz arenite	0	33	0
K 41	(A4)	76	18	6	F	Sublithic arenite	0	39	1.4
K 21	(A5)	92	8	1	F	Sublithic arenite	0	36	0.2
K 83	(B1)	90	8	2	F	Sublithic arenite	12	0	0
K 79	(B2)	81	16	3	F-M	Sublithic arenite	22	0	0
K 78	(B6)	86	10	4	F	Sublithic arenite	10	0	0.1
K127	(B15)	91	8	1	F	Sublithic arenite	23	0	0
K135	(B15)	92	7	1	F	Sublithic arenite	0	36	0.6
K199	(B21c)	93	7	0	F	Sublithic arenite	18	0	0
K152	(B21e)	92	7	0	F-M	Sublithic arenite	19	0	0
K189A	(B22)	92	7	1	F-M	Sublithic arenite	0	34	0
K218	(C1)	68	27	6	F	Lithic arenite	0	39	0.3
C232	(C1)	88	12	1	M	Sublithic arenite	0	28	2.3
C263	(D2)	66	27	7	F-M	Lithic arenite	0	39	0.5
C286A	(D2)	58	37	5	F-M	Lithic arenite	0	44	3.1
C335T	(E2)	86	12	2	F-M	Sublithic arenite	0	42	24.1
C326	(E3)	82	12	6	F	Sublithic arenite	0	17	66.1
C356	(X1)	80	10	8	F	Sublithic arenite	13	0	0
C300	(D4)	52	37	11	M-C	Lithic arenite (boulder bed matrix)	0	32	0

Bioclastic skeletal debris varies from trace amounts up to 60-70% of the rock, and is particularly abundant in the Culgower and West Garty sections (where pure sandstones are the exception) and in some parts of the Kintradwell sequence. The average content of bioclastic material in the Culgower and West Garty sections is 25% and hence the sandstones here are mainly calcareous, shelly sublitharenites and lithic arenites, but in some beds they grade into arenaceous biosparites (sandy skeletal grainstones) with up to 66% bioclastic debris (q.v. sample C362, Table 6.4).

Bioclastic skeletal components

Shell debris is the dominant skeletal component of the sandstones and boulder bed matrices. Both brachiopod and pelecypod fragments are present in significant amounts, with the latter consisting mainly of 'oyster'. Echinoderm debris occasionally comprises up to 75% of the bioclastic materials observed in thin section and appears to be exclusively represented by echinoid plates and spines. The sandstones of the Culgower and West Garty sections also contain small amounts of coral and bryozoan debris which were not noted lower in the sequence. A little gastropod debris occurs in some beds and a few percent of foraminifera and phosphatic fish debris are present in most of the thin sections. The shell debris is quite commonly bored and often contains numerous small pyrite crystals concentrated in particular layers or near the rims of the shells and sometimes disseminated throughout. Most of the shell debris has retained its original microstructure (fibrous and prismatic layers, punctae, etc.) but recrystallisation to blocky sparite is sometimes fairly common, particularly in the thinner and smaller fragments. Initially the shell debris retains sharp outlines but where recrystallisation is most advanced the bioclasts remain only as 'ghosts' in the sparite cement. Recrystallised shell debris and echinoderm fragements are often syntaxial with the cements. In many of the thin sections the fairly marked indented contacts between the siliceous grains and the bioclastic material is a clear indicator of early pressure solution of the primary carbonate content. Although cephalopod debris is a conspicuous macroscopic component of some sandstone beds, the size of this material (often whole fossils) is such that it rarely appears in thin section. The only other significant fossil components are the small unidentified "calcispheres' which can often represent up to 5% of the whole rock. These particles are 0.09 to 0.17mm in diameter (mostly 0.12-0.17mm) and are almost invariably recrystallied to sparite (ghosts in the sparite cements) or chalcedony with the total loss of all structure. Only very rarely, where recrystallisation has been less extreme or there are contrasts in the clarity of the calcite, is any microstructure apparent. In such cases there is a three-layered concentric structure in which the outer layer is the thinnest (0.01-0.02mm) and best defined. Sections through these particles are 'all' circular or slightly oblate (sometimes slightly indented on one side) and this strongly suggests a 'spherical' shape. There is no evidence to suggest that they are partly crinoid ossicles, ooliths or foraminifera, and although they are probably polygenetic (see below) most of them may represent recrystallised calcisphaerulids and calcitised radiolaria.

Silica diagenesis of bioclastic components

Very rare chalcedonic replacement of calcite is present in many of the thin sections examined but was only significant in about five samples. The most common fossil components to be replaced are the 'calcispheres' described above, but there are also rare ovoid to elongate bioclasts which are also chalcedonised. The latter are 0.1 to 0.17mm in shortest and up to 0.8mm in longest dimension and appear to represent sections of tubular particles, end views of which may account for some of the calcispheres (note similar dimensions). These elongate sections often exhibit a very thin central 'canal' which has quite often been the path for subsequent calcite replacement, but exhibit no other structure. Like the 'calcispheres' they are of unknown origins but may represent sponge spicules or echinoid spinelets. In the few thin sections where chalcedonisation has been more significant, shell debris (and in one case coral fragments) have also been affected. Here the chalcedony (or quartz) occurs mainly as small discontinuous, elongate patches within the fibrous or lamellar microstructure,and only exceptionally constitutes more than 10% of the bioclast. The chalcedony often appears to have lost its original texture and crystal structure and to have recrystallised to larger (sometimes single) less complex, but often highly irregularly shaped crystals.

Calcite cements

Three quarters of all the thin sections examined were found to have calcite cements; the lithological distribution of the calcite cementation is recorded in Table 6.3. Three main cement fabrics were observed; in order of decreasing abundance (reflecting the proportions of lithologies): poikolitic (lustre-mottled) cements, granular sparite cements and granular microsparite cements (see Table 6.3). Seven samples also contained some partial incipient cone-in-cone cementation. Although the name 'calcite' is used in the following descriptions, no attempt was made to verify the mineralogy of the cements by staining or other techniques. The type of cement present is clearly influenced by the original porosity (and therefore sorting, etc.) of the host lithology. Since the more poorly sorted argillaceous or shelly lithologies are those which had the least primary porosity, the moderately sorted, sparsely to moderately shelly sandstones contain poikolitic cements, the poorly sorted shelly sandstones and boulder bed matrices contain granular sparite cements,and the argillaceous sandstones, argillaceous boulder bed matrices (and the shales) all exhibit granular microsparite cements. The 'normal' sparite and poikolitic cements are clearly corrosive and replacive; quartz grains often show corroded and embayed margins (although to very variable extents) and the cements have often penetrated and partly replaced Moinian and Old Red sandstone lithoclasts. Some 70% of the calcite-cemented sandstones are lustre-mottled, with the poikolitic cement usually representing 35 to 40% of the rock. Since the great majority of bioclasts are unrecrystallised or are still visible as 'ghosts', the large volume of these cements must reflect the high original porosity of the sandstones and to a lesser extent their replacive nature. The poikolitic cements are generally very coarse to extremely coarsely crystalline (1.0 - >4.0mm).

A comparison between the cemented sandstones of areas A, B, C and D and E indicates that although the average amount of cement falls by about 10% (from 44 to 34%), the total calcite content of the rocks increases by about 10% (from 50 to 60%), the difference being more than made up for by a 20% increase in the bioclastic content in the higher sections. The increase in the bioclastic component and poorer sorting apparently reduced the primary porosity to an extent where granular sparite cements formed preferentially to poikolitic ones (because of the lack of space). In addition the more shelly sandstones may have contained more 'nuclei' for cement growth with a consequent reduction in the individual crystal sizes. Echinoderm debris is often syntaxial with the cements. Argillaceous sandstones are invariably microsparitic, presumably because of even lower primary porosities. Cementation in the shales is mainly restricted to sparite cementation of the sand stringers and only rarely effects the matrix resulting in large lenticular, microsparite cemented masses, e.g. see sections B22, D4 (outcrop VII) and the E shale unit on Crackaig Links. Incipient cone-in-cone fabrics (q.v. Füchtbauer, 1971) were observed only in the cemented shales and argillaceous sandstones and boulder bed matrices.

In many of the thin sections of cemented sandstones there are small patches of opaque cement (limonite and/or pyrite) which are of irregular or rounded outline and 0.5-9.0mm across. In one sample in particular

(K28, section A5) these patches form distinct, rounded (often spherical) blebs in amongst the normal calcite cement. Pettijohn (1975, p.240) has described identical features from 'spotted sandstones' as being due to the oxidation of siderite cements to limonite. No interpretation of these opaque blebs is offered here in lieu of a more detailed study.

Diagenesis of the non-calcite - cemented sandstones

Most of these sandstones are only weakly cemented as a result of slight pressure solution,which is indicated by slightly concavo-convex grain contacts. In addition most of the sandstones of the Allt na Cùile body,and several other scattered samples,are also variably iron cemented. Authigenic quartz overgrowths are rare in the non-calcite-cemented sandstones and when present are usually confined to the smaller, more enclosed pore spaces. Authigenic clay minerals are mainly limited to patchy discontinuous illite grain coatings. Kaolinite pore fills (book and vermicular habits) were common in only four of the sandstone thin sections (15% of the non-calcite-cemented sample) and two sandstone clast thin sections; illite pore fillings were only observed in one sample (K78). Total porosity reduction due to authigenic clay minerals and iron cements is usually in the range 5-20%, but exceptionally reaches up to 60% reduction, although even here some 10% porosity usually remains. The lack of illite grain coatings beneath the authigenic quartz overgrowths and the presence of kaolinite aggregates in illite-coated pores suggests that overgrowths formed first and were followed by authigenic illite and then kaolinite. Iron cementation may be comparatively recent in geological terms and relate more to weathering than diagenesis in its usual sense. Other diagentically significant events in the history of the sandstones include decalcification (by early pressure solution and leaching) and the formation of the so-called 'veins' in the Allt na Cùile and Lothbeg sandstone bodies. These features are not veins in the normal sense, but anastmosing fractures and microfaults. In thin section they appear as 'crush zones' of tightly-packed (presumably quartz cemented) silt-sized angular quartz grains and their formation is probably related to the compaction history of the sandstones. Such features were not observed in the thinner sandstone beds intercalated in the shale sequences. The presence of these 'veins' which often form quite dense networks (particularly in parts of the Lothbeg sandstone body) presumably has a significant effect on the net permeability of these units and is therefore of some economic interest. The 'veins' are sharply defined and the intervening sandstones perfectly normal.

PALYNOFACIES OF THE KINTRADWELL - WEST GARTY SEQUENCE

Organic carbon data

Seventy four organic carbon analyses were carried out on my behalf by Roberston Research. The organic carbon values show a clear relationship to the sample lithology (see Table 6.5) with the shales being almost twice as rich as the more carbonaceous of the sandstones. The light grey kaolinite-illite clays (as described under sections B1, B4, B22, X1 etc) are the least organic-rich lithology (0.0-0.9% org.c.), which is not surprising considering their colour and that all but one are intercalated in sandstone units. Table 6.6 shows the distribution of organic carbon by area and section. The only visible trend is that the shales in areas C, D and to a lesser extent E, are slightly richer (by about 2% org. c.) than those in areas A and B. The mean organic carbon content of the shales is 7.5%, with only 13% of the shales analysed yielding values over 10.0% org.c. (maximum 17.5%).

Kerogen composition

The kerogen composition of the Kintradwell-West Garty sequence is also clearly related to lithology (see Table 6.7). The kerogen assemblages are dominated by woody material with the autochthonous component only averaging at about 25% even in the shales. The woody

TABLE 6.5 Lithological variation of organic carbon and pyrolysis data (analyses by Robertson Research : see Appendic 6E)

Lithology	Organic carbon wt. % No. analyses	$\bar{x}$	δ	Rock Eval. Pyrolysis No. analyses	Hydrogen Index $\bar{x}$	Hydrogen Index δ	Oxygen Index $\bar{x}$	Oxygen Index δ
Shales	62	7.5	3.1*	24	147.5	78.1	18.1	11.2
Carbonaceous shaley sands	7	4.2	2.6	2	48.0	-	31.5	-
Argillaceous sandstone	1	2.6	-	0	-	-	-	-
Shaley boulder bed	1	3.8	-	0	-	-	-	-
Kaolinitic clay	4	0.3	0.4	3	0	-	-	-

* Only 13% of shale samples show values above 10.0%

$\bar{x}$ = mean δ = standard deviation

TABLE 6.6 Distribution of geochemical data by section
(analyses by Robertson Research : see Appendix 6E)

Section	Org.C $\bar{x}$	Hydrogen Index $\bar{x}$	Oxygen Index $\bar{x}$	No. analyses Org.C	No. analyses Pyrolysis	
A1	3.9	-	-	2	-	
A2	7.0	66.8	16.4	13	4	*
A3	6.8	173.0	27.0	2	1	
A4	6.9	93.0	19.0	3	1	
A5	4.4	92.0	49.0	2	1	
A6	6.5	187.0	14.5	2	2	
B1 snd	3.0	0	14.0	1	1	
B13	6.0	42.0	12.0	1	1	
B14 snd	3.4	33.0	17.0	2	1	
B15 shl	6.7	122.0	21.0	2	1	*
B15 snd	7.0	63.0	46.0	2	1	
B18	6.2	102.0	34.0	1	1	
B21e	12.8	131.0	10.7	3	3	
B21f	2.9	-	-	1	-	
B22	7.8	130.0	10.0	8	1	*
C1	10.0	215.0	14.0	3	1	
C(B)	8.5	152.0	13.0	2	1	
C2	8.5	-	-	1	-	*
D2	9.0	182.0	9.0	3	1	*
D3i	7.9	266.0	13.0	2	1	
D3ii	9.2	-	-	1	-	
D3iii	11.3	-	-	1	-	
D4	6.0	232.0	14.0	3	1	**
E1	8.9	-	-	1	-	**
E2	5.6	248.0	23.0	1	1	**
E3	6.2	300.0	10.0	4	1	**
E4	6.9	-	-	1	-	
X1 snd	0.6	-	-	1	-	
X1 cly	0	0	0	1	1	

* denotes section whose samples show strongest A.O.M. fluorescence under ultra-violet

** denotes section whose samples show strongest and most consistent A.O.M. fluorescence

TABLE 6.7 Lithological variation of main kerogen components (%)

Lithology	No.	Palynomorphs		A.O.M		Phytoclasts*		Inertinite	
		$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ
Shale	175	8.4	5.4	28.6	11.9	57.8	12.7	9.6	5.1
Argillaceous sandstone	8	9.9	4.9	18.0	10.9	64.8	11.7	12.3	5.8
Argillaceous boulder bed	16	8.6	3.8	21.6	12.8	62.1	13.6	9.8	4.1
Argillaceous clasts	5	7.8	4.6	19.1	10.8	68.6	13.1	14.5	12.6
Carbonaceous shaley sand	21	2.9	3.2	2.4	1.8	87.6	10.0	20.9	7.1
Kaolinitic clays	3	3.3	2.0	12.5	5.4	80.8	8.0	19.1	2.5

* N.B. Phytoclast category includes inertinite

$\bar{x}$ = mean δ = standard deviation debris is proprotionately most abundant in the carbonaceous sandstones (88% mean particle abundance) and least proportionately abundant in the shales (58%), although in absolute terms the reverse is probably true (see Table 6.5 and Plate 6.28). The relative abundances of woody materials observed in kerogen preparations are a function of three main factors: sorting (hydrodynamic equivalence), redox conditions (since oxidation of A.O.M. increases the percentage of terrestial materials), and the rate of supply. Sorting and redox conditions are probably especially important controls on the kerogen assemblages of sandstones and redox and supply most important for shales. Within the argillaceous sediments the wood fraction tends to be higher in the more allochthonous sediments (boulder bed matrices and shaley clasts) which is partly a result of their higher sand contents; the kaolinite-illite clays only show high proportions of wood because of the oxidation and destruction of most other materials (note the very low organic carbon values). The proportion of inertinite in the wood fraction is significantly (2-3 times) higher in the carbonaceous sandstones and kaolinite-illite clays than in the shales and this probably also reflects the oxidation potential of their relative depositional and diagenetic environments. Although oxidation may be higher in the carbonaceous sandstones, the ratios of wood of a fresh appearance to that of a degraded appearance is in fact lower in the sandstones (about 1:2) than the argillaceous lithologies ($\sim$1:3-4). This suggests that physical degradation (breakage, splintering, etc.) predominates in the sandstones while biological degradation (by bacteria, fungi, detritus feeders) characterises the shales.

The redox and hydrodynamic equivalence relationships are such that the distribution of amorphous organic matter tends to be inversely related to that of woody material. Amorphous particles are proportionately most abundant in the shales (28.6% mean particle abundance) and the other argillaceous facies and lithologies, with a marked minimum of 2.2% in the carbonaceous sandstones. Palynomorphs are relatively most abundant

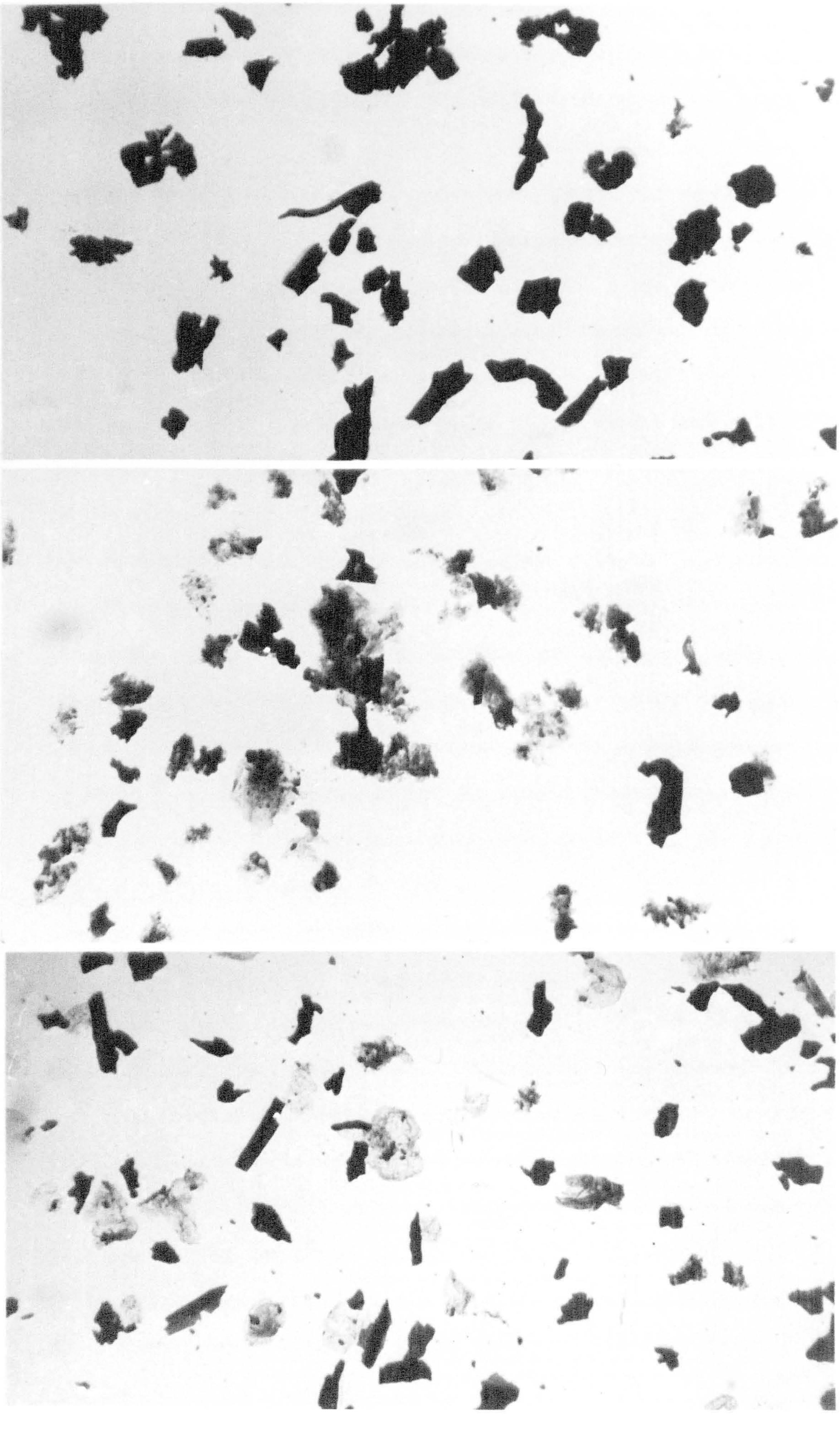


PLATE 6.28. Palynofacies variants from Kimmeridgian sediments of the Brora-Helmsdale outlier. Figures in parentheses are the widths of the photographs in millimetres.

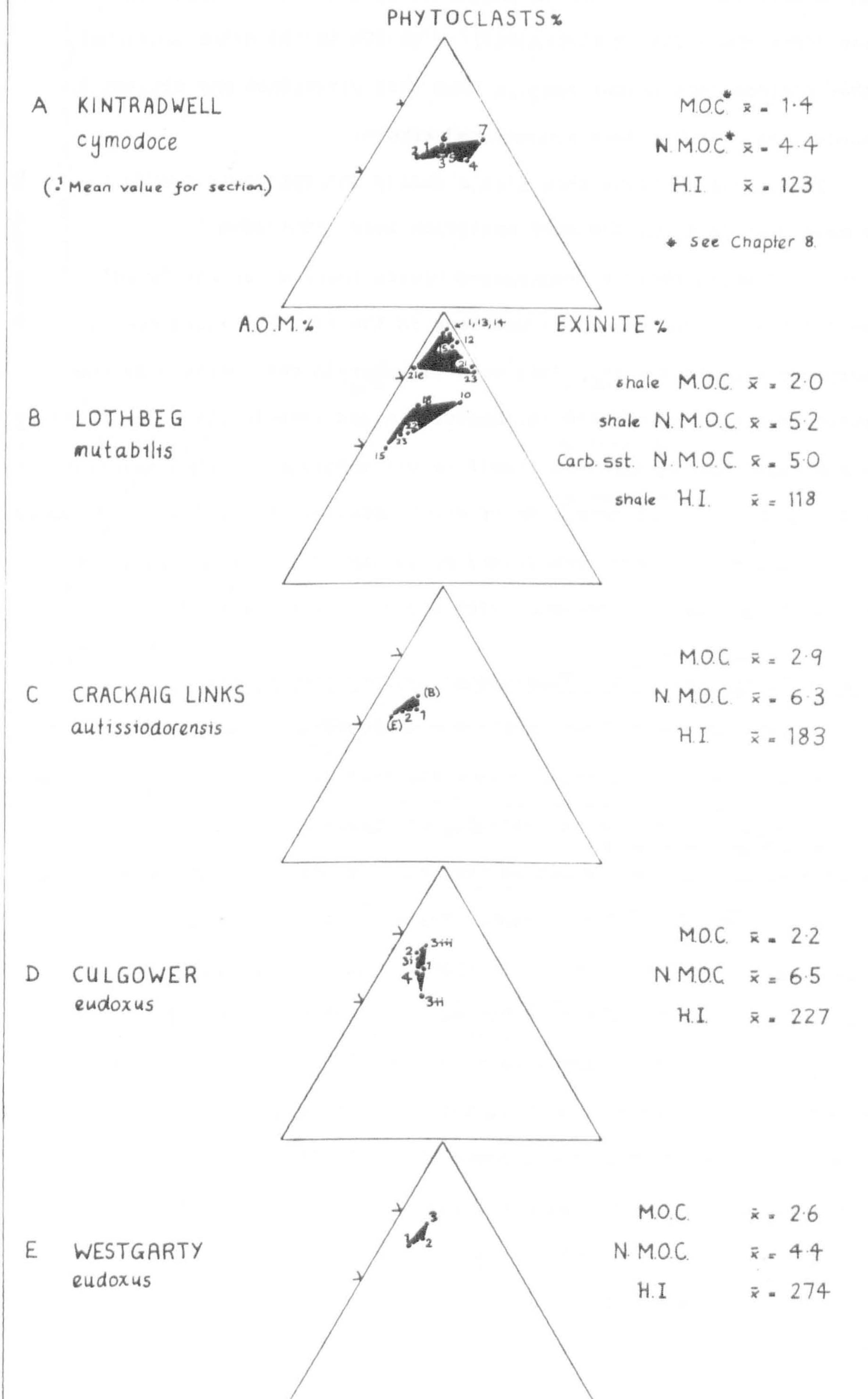
Top: Assemblage rich in dark, oxidised plant debris ("black wood" - probably dominantly semi-fusainite or inertinite). Carbonaceous shaley sandstone. Sample K 144 (0.8)

Centre: More usual mixed phytoclast and A.O.M. assemblage typical of shales. Sample K 122 (0.8)

Bottom: An assemblage unusually rich in palynomorphs and palynodebris. Sample K 36 (0.8) in the argillaceous lithologies (mean values 6.4 to 10.0%) and again show minima in the carbonaceous sandstones (1.9%) and those kaolinite-illite clays interbedded within sandstone sequences. Sample K164 from section B22 is a kaolinite-illite clay interbedded in a shale sequence; it has a somewhat greater illite content (as judged by peak intensity from X.R.D.) and an A.O.M. and palynomorph content which is much more like the usual shale than the other similar clays (K77, K86, C358). This indicates that it is the setting of this lithology, rather than anything inherent in the sediment type, which determines its kerogen characteristics. Sorting is the principle control on palynomorph abundances.

Figure 6.36 shows the distribution of the main kerogen components (phytoclasts, A.O.M. and palynomorphs) by area and section. The only facts which are immediately obvious from this diagram are the inverse relationships of phytoclastic wood materials and A.O.M., the strong influence of lithology and the general similarity of the shales in all areas despite the general high standard deviations in individual sections. On closer inspection it can be seen that the kerogen distributions in the shaley sediments of area B (Lothbeg) fall into two groups: those with high wood particle abundances which are from sections where the shales are interbedded within predominantly sandstone sequences, and those with wood particle abundances like those in areas A, C, D and E which are from predominantly shale sections (B18, 15, 22 and 23). This difference is probably due to the fact that the shales of the former group are much sandier than those of the latter (being transitional to carbonaceous sandstones) and may also indicate that the influx of sands was associated with an active input of woody material. Clearly, both sorting and supply factors are effectively inseparable. Although there is significant overlap in the range of values from individual sections the relative abundance of amorphous particles appears to be higher in the mutabilis-autissiodorensis shales (areas C,

Fig. 6.36 Geochemical trends in Kimmeridgian sediments



* See Chapter 8.

D, E and shale dominated sections in B) than in the cymodoce deposits at Kintradwell (area A). This is indicated by a 25-75% increase in the area mean (from about 20% at Kintradwell to 25-35% in the other sections). Other factors (see below) suggest that this difference may reflect a genuine trend rather than accumulated errors.

Individual sections show little easily interpretable trends in kerogen distribution, the only exception being Kintradwell section A2 which shows a progressive upward increase in the 'wood' component and a corresponding decrease in the relative abundance of amorphous organic matter. This trend may herald the increase in the sandstone shale ratio which is observed in,and immediately below,sections A3 and A4,although the trend itself is not continued in these sections or in A6 and A7. The distribution of woody material in section B15 suggests a correlation with sand content and proximity to sandstone bodies (see Fig. 6.37) and is in accordance with the general patterns discussed above.

Rock Eval. pyrolysis and ultra-violet fluorescence microscopy

Twenty eight Rock Eval. pyrolysis analyses were carried out on my behalf by Robertson Research. Since the relative particle abundances of woody materials observed in palynological preparations normally lies between 66 and 75%, it is not surprising that all the analyses indicated a Type III composition (see Chapter Three), but some lithological and stratigraphic variations are also apparent (see Figs. 6.36, 6.37 and Table 6.5). From Table 6.5 it can be seen that the mean hydrogen index value for the shales is approximately three times greater and the mean oxygen index roughly half of that for the carbonaceous sands. This lithological trend is in total agreement (qualitatively if not quantitatively) with the palynological (visual) kerogen data presented in Table 6.7. Despite some overlap in the range of values, Fig. 6.36 shows that the hydrogen indices of shale samples appear to be higher in areas D and E (and to a lesser extent C) than those in areas A and B, a

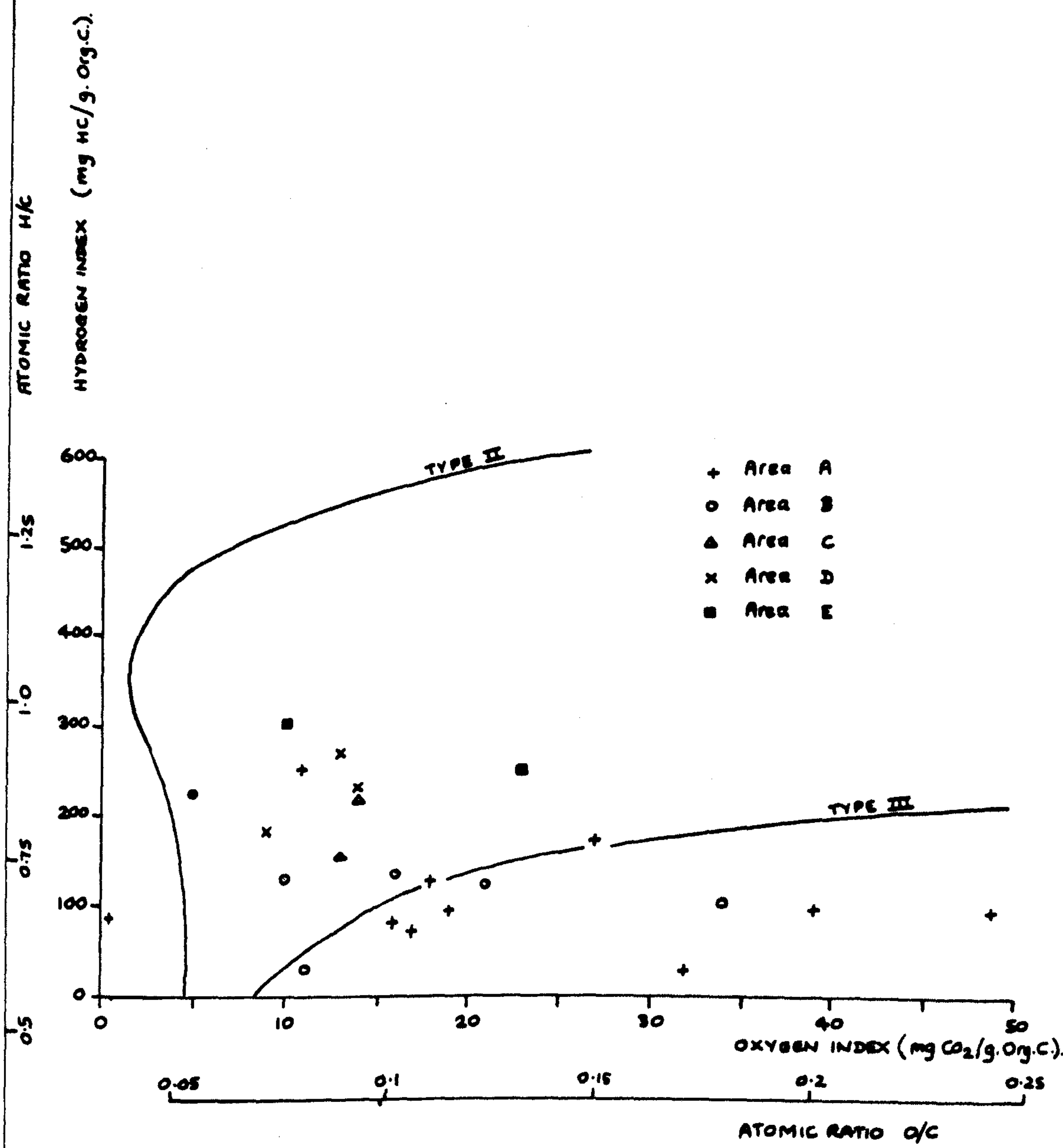


Fig. 6.37 Pyrolysis data for the Lower Kimmeridgian of the Brora-Helmsdale outlier.

similar trend to that observed for organic carbon values.

Under ultra-violet fluorescence the particles of amorphous organic matter are seen to consist of a dull matrix (with various woody fragments, pyrite, palynomorphs, etc.) which contains varying amounts of small (<20μ), more highly fluorescent fragments and debris of what may be exinite alginitic material derived from blue-green algae (cyanobacteria). The overall appearance under ultra-violet is very like that observed in the examined kerogenous shales of the Swalland Member in Dorset (see Chapter Five). The abundance and fluorescence of these alginitic inclusions is most marked in the shale dominated sections and is most consistent and best developed in the shales from sections C2, D2, D4, and E1-E3, where they are probably responsible for the higher hydrogen indices.

Palynofacies characteristics

Since only kerogen (un-oxidised) palynological preparation were made available for this study, the low relative abundance of palynomorphs in the Kintradwell-West Garty sequence placed considerable constraints on the potential information to be had by this technique. Only a quarter of all the slides examined (59 out of 234) contained greater than ten percent total palynomorphs and 60% of this number were from sections in the Kintradwell area. Excluding the atypical Kintradwell samples, the modal palynomorph abundance (representing half of the total number of slides) was found to lie between three and eight percent. The masking effect of the amorphous organic matter is a formidable problem in some of the slides, but there is no reason to suppose that any one group of palynomorphs is especially more susceptible to masking than any other, and the most important effect is probably that the total palynomorph abundance is somewhat underestimated. Forty-five slides (19% of the total) were selected on the basis of suitable palynomorph densities to represent the different lithologies and sections and the nature of their palynomorph assemblages determined by counting. In all cases the assemblages are dominated by sporomorphs and are of a relatively distal aspect with the most abundant groups being the undifferentiated ('light') sphaeromorphs (∿40-60%), bisaccates (∿10-20%) and unornamented ('light') deltoid trilete spores (∿10-20%). Other sporomorph groups including Callialasporites, Perinopollenites and Cerebropollenites were present in most samples but in low numbers and together usually total less than 5-7%.

Although sample coverage was admittedly rather poor, few logical correlations were observed between the nature of the palynomorph assemblage and the lithology and stratigraphic position of the samples (see Tables 6.8 and 6.9). The relatively minor differences in the abundance rankings of the differentiated sporomorph groups may reflect only the granulometric composition of the samples or an unpredictable combination of sorting and supply factors. The most conspicuous contrasts in the assemblages are those seen in the total palynomorph abundances (see Fig. 6.36) and the distribution of plankton (principally dinocysts and acritarchs). The sparcity of plankton in the Kintradwell-West Garty sequence was immediately obvious even from the general kerogen analysis. Subsequent detailed counts demonstrated that the total plankton abundance was usually less than 5% of the total palynomorph content. Unlike the sporomorphs, some correlation appears to exist between the percentage of plankton and the sample lithology (see Table 6.8). It is interesting to note that plankton is most abundant in the boulder bed matrices and shale clasts, both of which may be considered to be more allochthonous than the shales. Dinocysts either dominate or are equally as abundant as acritarchs in 70% (32) of the counted samples and chorate forms are the single largest or equal largest dinocyst group in 70% of the counted slides. The only other significant component of the plankton is the relatively long-spined acanthomorph acritarch Micrhystridium fragile which is the single most abundant element of the plankton in 36% (16) of the counted samples. No stratigraphic trends in the distribution of the

TABLE 6.8 Lithological variation of palynomorph assemblages

% PALYNOMORPH GROUP		LITHOLOGY	Shale (32)	Argillaceous sandstone (3)	Argillaceous boulder bed (4)	Argillaceous clasts (2)	Kaolinitic clays (2)	Carbonaceous shaley sand (2)
DINOCYSTS	Undifferentiated	I	0.7	0.9	1.0	1.2	0.6	0.9
	Proximate	A	0.1	0.2	0.2	0.4	1.1	0.1
	Cavate	B	0.1	0	0.2	0	0.1	0
	Proximo-chorate	C	0.2	0.1	0.4	0.2	0	0.2
	Chorate	D	0.5	0.7	1.4	1.2	0.4	0.5
	Peridinean	E	0	0	0.1	0	0.1	0
ACRITARCHS	Short spined acanthomorphs	F	0	0	0.2	0.1	0.1	0.1
	Long spined acanthomorphs	G	0.5	0.6	1.5	0.6	0.1	0.3
PRASINOPHYCEAE		H	0.1	0	0.1	0.1	0.3	0
	TOTAL PLANKTON		2.2	2.5	5.1	3.8	2.8	2.1
	Bisaccates	J	15.5	8.0	11.9	14.2	14.0	12.9
POLLEN	Perinopollenites	K	1.9	1.0	2.4	0.7	1.9	0.6
and	Light, unornamented	L	13.4	17.0	15.0	14.7	11.6	20.0
SPORES	Zonate	M	0.8	0.7	0.5	0.6	0.6	0.3
	Heavy, ornamented*	N	14.0	15.6	9.5	12.4	12.4	8.4
UNDIFFERENTIATED PALYNOMORPHS		O	52.5	55.0	55.6	53.5	57.0	56.0

* erroneously includes ≤ 2.0% Cerebropollenites

TABLE 6.9 Distribution of mean palynofacies assemblages by section (argillaceous lithologies only) (See Table 6.8 for key)

Section	No.	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	ΣA - I
A1	3	0.1	-	-	0.1	-	-	0.2	-	0.3	15.4	1.0	13.7	0.4	10.9	57.1	1.0*
A2	5	0.1	-	0.1	0.5	-	-	0.4	-	0.9	19.2	1.8	12.1	1.2	17.3	46.4	1.1
A3	4	0.2	0.1	0.2	0.7	0.1	-	0.2	0.3	0.5	22.5	2.4	4.6	0.7	14.1	50.8	1.5
A3/A5	2	-	0.1	0.1	1.2	-	-	0.8	0.1	0.9	13.1	2.1	14.4	0.2	7.9	59.8	2.2
A6	3	0.1	-	-	0.6	-	0.1	0.8	-	0.6	8.8	1.2	12.8	1.0	13.7	57.1	1.5
B6(clast)	1	0.4	-	-	-	-	-	-	-	0.4	12.3	1.5	13.8	0.4	12.3	59.0	0.4
B15	3	0.1	0.3	0.1	1.4	0.1	-	0.4	-	0.9	17.4	3.0	9.7	1.0	6.4	59.3	2.3
B22	8	0.3	0.1	0.3	0.3	0.1	-	0.5	0.1	0.9	14.5	2.4	15.2	0.7	12.5	51.9	1.7
C1	4	0.1	-	0.1	0.2	0.1	0.1	0.7	0.1	0.9	9.9	1.1	18.2	0.7	22.8	45.2	1.2**
C2	1	-	-	-	-	-	-	0.9	-	0.2	8.0	1.4	13.0	0.7	9.3	66.6	0.9
D1	1	-	-	0.5	0.2	-	-	0.9	-	0.7	7.3	0.7	13.1	-	17.9	58.7	1.6
D4	1	0.2	-	0.5	0.7	-	-	0.7	-	-	16.2	1.9	18.0	-	9.4	52.5	2.1
E3	1	-	-	0.2	0.5	-	-	1.4	-	1.4	23.6	0.2	11.3	0.5	15.0	46.0	2.1

* 0.5 without sample K66 (clast)

** 0.7 without sample K230 (clast)

total phytoplankton is apparent from Table 6.9, but the counts made during the general kerogen analyses strongly suggest that the plankton (like the total palynomorphs) is relatively more abundant in the Kintradwell area (see data in Appendix). Differences in the plankton counts derived from the kerogen and the palynomorph counts probably reflect the differing degrees of accuracy employed in the two sets of analyses (e.g. see Chapter Three). The greater abundance of plankton in the Kintradwell sections is probably not apparent in Table 6.9 because of the atypical nature of the samples from areas B to E which were selected for counting on the basis of their above average palynomorph contents. The relative apparent abundance of the plankton could be related to the degree of masking by A.O.M., since such material is somewhat less abundant in the Kintradwell sections than it is elsewhere. Prasinophycean algae only very rarely exceed 0.2% of the total palynomorph count.

Conclusions and discussion

(1) The high relative abundance of woody material in all the Kintradwell-West Garty sediments results in Type III kerogens and suggests deposition relatively close to a major source of terrigenous material. The sandier, more wood-rich shales in the vicinity of the Allt na Cùile and Lothbeg sandstone bodies may imply that the Lothbeg area was a locus of terrigenous organic matter supply.

(2) Although woody debris is abundant the sporomorph assemblages are of a distal aspect and their general uniformity indicates that the depositional conditions were relatively stable, and that most of the sorting of the palynomorphs occurred up slope in unsampled environments.

(3) The abundance of amorphous organic matter (20-30% relative particle abundance) and the presence of relatively undegraded liptinitic inclusions suggests that bottom conditions were reducing; the R.P.D. probably occurred very close to the sediment water interface and dysaerobic

conditions may have predominated immediately above the bottom. High accumulation rates may have assisted the preservation of this autochthonous organic matter. The maintenance of these reducing conditions in the majority of the sections (with exceptions in the Lothbeg area) tends to suggest relatively uniform bottom water conditions, i.e. a sub-mixed layer environment.

(4) The paucity of dinocysts in the sediments also suggests stable stratified conditions, but may also party be due to ecological exclusion by blue-green algae (whose presence may be indicated by the 'alginitic' inclusions in the A.O.M.). The dinocyst component present has probably been redeposited from up-slope mixed layer environments, with selective transportation resulting in the dominance by chorate morphotypes. This is tentatively supported by higher dinocyst densities in shale clasts and boulder bed matrices. The up-slope environment may in part be salinity stressed but there is insufficient data to determine this on the basis of the plankton. The dinocyst assemblages are strongly dominated by gonyaulacean forms (see also Lam and Porter, 1977).

(5) The somewhat higher organic carbon values, hydrogen indices, relative abundances of A.O.M. and stronger and more consistent ultra-violet fluorescence of samples from areas C, D and E (primarily eudoxus and autissiodorensis zones) suggests that bottom conditions became more strongly reducing after the cymodoce and mutabilis zones. This could reflect the similar trend observed between the Black Head and Kimmeridge Members in the Dorset type section (see Chapter Five).

(6) Kerogen compositions are strongly and predictably influenced by lithology. Phytoclastic debris is positively correlated with the sand content as a consequence of hydrodynamic sorting and redox factors, but palynomorph assemblages show only relatively small lithological contrasts. The kerogen character is strongly influenced by the overall facies character and context of the sections (particularly with regard to the dominant lithology) as well as by the lithology of the individual units sampled.

The only relevant published palynological works on the Upper Jurassic of the Brora-Helmsdale outlier are those by Lam and Porter (1977) and Riley (1980), both of which are essentially biostratigraphy papers. Lam and Porter (1977, p.49) note that "Throughout the section microplankton abundance is low compared with that of the pollen and spores" and their Fig. 5 (p.51) shows plankton abundances which range between 10 and 50% with a mean value at about 20-25% for that part of the sequence studied here. Although I clearly agree that plankton is poorly represented, my own investigations indicate that it does not exceed 5%, let alone 20-25%! Although I am not a palynologist as such, and I was using kerogen rather than oxidised palynostratigraphic preparations, I still believe my figure to be more representative than Lam and Porter's. My own analysis was based on far more samples (234 as apposed to 23) and even if I excluded the largest single group in my counts (the undifferentiated sphaeromorphs - which undoubtedly does include some misidentification) the percentage of plankton (in the 45 samples) still does not exceed 10%!

Considering that redox trends observed in the Dorset Kimmeridge Clay Formation may be reflected in the Brora-Helmsdale sequence (point 5 above), it is interesting to speculate on the likely palynofacies trends to be expected in the higher, upper Kimmeridgian-Portlandian (or Volgian) parts of the succession. Lam and Porter (1977) record that pteridophyte spores decline after the lower Kimmeridgian but that bisaccate pollen remain abundant and this suggests a distal shift in the palynofacies. In addition they note that "species of Crassosphaera, Pterospermopsis and Tasmanites are commonly found in the Kimmeridgian deposits from the mutabilis zone onwards" (p.54) and this is echoed by Riley (1980, p.30) who records "large numbers of Pterospermopsis" in the lower Portlandian (middle middle Volgian) sediments at Navidale. In the Viking and Central Graben areas of the North Sea the increased abundance of prasinophyceae (particularly Pterospermopsis) is characteristic of the Volgian (especially from the middle Volgian) to the Ryazanian (Berriasian) interval of the Kimmeridge Clay Formation (pers. comm. L.A. Riley; see also Chapters 3 and 4). This 'prasinophycean facies' generally correlates with a marked distal facies shift (e.g. see Chapter Four) and an increase in the relative abundance of A.O.M., and a similar trend may be anticipated for the Volgian part of the Brora-Helmsdale sequence. Although the nature of the coarse grained lithofacies suggests a proximal facies shift with regard to the position of the active fault scarp, this does not necessarily imply an increase in the terrigenous influence in the Midgarty-Dun Glas succession. The presence of displaced corals in this upper interval argues against there being a major fluvial input of terrigenous organic matter, although 'drift wood' is often encountered in the boulder beds.

During the present study a single sample of shaley mudstone of cymodoce zone age was collected from the small lower Kimmeridgian outcrop at Eathie, Sutherland (for description see Waterston, 1950, 1951, with additional comments in Bellamy, 1979, p.97). No kerogen or palynofacies counts were performed on this sample, but it was observed to be rich in both degraded A.O.M. (dull under U.V.) and palynomorphs, and to contain only a little woody material. The sporomorphs are dominated by undifferentiated sphaeromorphs, bisaccates and Classopollis with rare Cerebropollenites and a little or no ornamented heavy spores. Plankton is much better represented than in the coeval sediments of Kintradwell and the dinocysts are strongly dominated by chorate morphotypes. The overall assemblage is very like that observed in the normal mudstone shales in the type Kimmeridge Clay (Chapter Five) and the sediment was probably formed under similar conditions. The Eathie sediments appear to have been deposited in a distal lower mixed layer environment with generally oxygenated bottom conditions, and hence in shallower conditions than at Kintradwell. The latter area was undoubtedly influenced by greater subsidence adjacent to the Brora-Helmsdale Fault, while Eathie may have occurred closer to the 'hinge' of the half-graben.

SYNTHESIS, SUMMARY AND DISCUSSION

(1) FACIES INTERPRETATION AND DEPOSITIONAL PROCESSES

Introduction

Bailey and Weir (1932), Crowell (1960) and Linsley (1972) have satisfactorily established that the Brora-Helmsdale sequence represents a fault scarp facies association consisting of redeposited sandstones and boulder beds intercalated within basinal marine shales. No further consideration is given to the other sedimentary models proposed by earlier workers, but sceptics who might scoff at Macgregor's (1916) fossil cliff interpretation (the most cogently argued of the early models) are recommended to examine the Torridonian cliffed shoreline deposits of north west Scotland (e.g. see Lawson, 1976). The best single published work on marine fault scarp facies associations is thatby Surlyk (1978) which describes the Late Jurassic-Early Cretaceous submarine fan deposits of East Greenland. Since the latter sediments are partly coeval and partly of similar facies to those of the Brora-Helmsdale succession, Surlyk's excellent paper has been used as the principal analogue for the interpretation of the sequences studied here. It should be appreciated that Surlyk's models of fault scarp-fan sedimentation were constructed from the examination of much more extensive and better exposed sequences than were available in this study, and therefore represent an invaluable guide to facies interpretation. Other relevant sequences and analogues occur in the Upper Jurassic of the Magnus, Brae and Toni-Thelma oilfields of the Northern North Sea (De'Ath and Schuyleman, 1981; Ziegler, 1980; Harms et al. 1981). The Toni-Thelma and Brae sequences are described and discussed in more detail in Chapter Seven.

A general review of the submarine fan and turbidite facies association is not presented here but the key information used in the present interpretations may be found in Carter (1975), Middleton and Hampton (1976), Lowe (1976, 1979), Kelling and Stanley (1976), Walker (1975, 1978) Nardin et al. (1979) and Stow and Shanmugam (1980). In the following account each of the major facies types of the Brora-Helmsdale sequence is examined and interpreted in turn.

MAJOR SANDSTONE BODIES

(a) The Allt na Cùile body (sections B1, 2, 3, 4, 9 and ?7)

The reliability of the facies interpretation of this unit is strongly constrained by the character of the exposure (see earlier comments in the discussion of the Lothbeg area). Most of the outcrop consists of vertical and inaccessible cliffs whose bases are mainly surrounded by dense vegetation; the irregular weathering effects and lack of clear marker horizons effectively prevent interpretation from a distance. What exposure there is indicates that the Allt na Cùile body includes at least four different facies: the intraformational breccias, the crudely bedded pebbly sandstones, the thinly bedded, current-bedded sandstores and the carbonaceous shaley sandstone-sandy shales.

(i) Crudely bedded, pebbly sandstones

This facies appears to be most characteristic of the basal part of the Allt na Cùile exposures (see sections B1 and B2). The sandstones are mainly massive or thick bedded and structureless except for occasional poorly defined layering or lamination of carbonaceous debris and dispersed pebbles and granules. Possible pebble-rich channel fills occur in the cliff exposures immediately adjacent to the railway line between the Allt na Cùile and section B3. Indistinct bioturbation (mottling) is sometimes present. Subordinate intercalations of this facies may occur higher in the sandstone.

(ii) Thinly bedded, current bedded sandstones

The thinly bedded facies immediately overlies (i) and appears to form most of the exposed part of the Allt na Cùile body. It consists of pebble free, thinly bedded (1-5cm) units of sandstone arranged in sets with planar or slightly concave upward binding surfaces. Individual beds may be highly bioturbated giving them a somewhat irregular appearance (enhanced by weathering and irregular cementation), but the burrows only rarely cross bed boundaries. The burrows are at least partly of *Skolithos* ichnofacies type.

(iii) Intraformational breccias

This facies appears to be exclusively interbedded within facies (ii). The thickness and scale of the breccias apparently increases both towards the Helmsdale Fault and laterally from section B1 to B4. They are predominantly clast supported (sometimes only clast rich) and are either chaotic or show rather poorly developed clast imbrication (subparallel arrangement with long axes dipping in an offshore direction). No internal bedding structures or discontinuities are apparent in the breccias in the Allt Chollgorge. At locality B3 the tops of the breccia bands are infilled and then overlain by thinly bedded sandstones. The basal contact of the breccias is not well exposed; it is presumed to be at least slightly erosional (but see later discussion).

(iv) Carbonaceous shaley sandstones and sandy shales

This facies is a lateral equivalent of facies (ii), occurring only to the east of section B1. The individual sandstone beds are thicker bedded or massive but the orientation of dispersed carbonaceous debris sometimes suggests a laminated fabric. The proportions of clean and carbonaceous sandstone/sandy shale are highly variable and the sandstones may show variable thickness. The carbonaceous units are much more organic-rich than those in facies (i) and are free of pebbles; they are sometimes bioturbated. Similar sediments are probably also laterally equivalent to (i) representing a transition from sandstone dominated to shale dominated sections.

The first question to be asked in the facies interpretation is whether the Allt na Cùile sandstone body represents a 'shallow' or 'deep' water facies association (i.e. mixed layer or bottom water environment). I favour at least a relatively deep water interpretation (lower mixed layer or bottom water environment), as the available ammonite data and field relations suggest an upper *cymodoce* to *mutabilis* zone age for the

Allt na Cùile body, placing it above what are obviously deep water deposits at Kintradwell. Other sediments in the Lothbeg area which are clearly of mutabilis age are also of deep (bottom water) facies. The overall geometry of the Allt na Cùile body is unclear - mostly because of the poor exposure, but also because of probable amalgamation with other units. The interpretation of the individual facies is discussed below.

The lower pebbly sandstone facies (i) is thought to represent deposition from a combination of gravity flow and traction (bottom current) processes. Pebbles are often dispersed rather than concentrated as basal lag deposits and this, along with the general homogeneity of some of the beds, suggests deposition was at least partly from grain-flows and liquified flows and possibly also turbidity currents. No liquifaction structures such as dish or pillar structures were observed, but this may merely reflect the inadequacy of the exposure. The facies may be loosely "equivalent" to a combination of facies A3, A4 and B of Walker and Mutti (1973) and facies 6 and 9 of Surlyk (1978).

If the Allt na Cùile body is interpreted as a bottom water facies, the interpretation of the thinly bedded, current bedded sandstones (ii), becomes rather problematical. The literature on submarine fan and turbidite facies contains very few references to well bedded, non-pebbly sandstones intercalated with massive breccias. The only recorded sediments which appear to be at all similar are those of the "arenaceous facies" (facies B) of Mutti and Ricci-Lucchi (1972). These are described as medium coarse to fine sand and as being better sorted, thinner bedded and exhibiting greater parallelism, more extensive strata and more pelitic layers than in the sandstone-conglomerate facies (facies A). They have thick and parallel laminae which are broadly undulating (with wavelengths of several to several tens of metres) or flat and discordant at very low angles (wedge shaped cross stratification). This lamination is described as being similar to that produced in the foreshore swash zone of beach sands. The "arenaceous facies" also exhibits dish structures and shows all types of gradation with conglomeratic sequences. Mutti and Ricci-Lucchi (1972) interpret it as the coarse grained fill of large submarine valleys which was deposited from a combination of grain flow and traction; the 'undulating' laminations were interpreted by these authors as the product of upper flow regime antidune bedforms. It is interesting to note that this sandstone facies was not included in the subsequent standard facies scheme of Walker and Mutti (1973).

Although I have not made any specific measurements, the current bedding in facies (ii) of the Allt na Cùile sandstone appears to be in a broadly unimodal 'offshore' direction. The current bedding was probably formed by the migration of straight-crested megaripples under lower flow regime conditions; the uniformity of the thinly bedded sandstones implies a stability and continuity of the sedimentary regime. Antidune bedding (like that in facies B of Walker and Mutti, 1973) has not been observed. The intensive bioturbation of some of the sandstone beds suggests that the rate of sedimentation was relatively slow or somewhat intermittent, but the rarity of argillaceous and carbonaceous materials indicates that even residual currents were sufficient to prevent the accumulation of finer materials. The pattern of bioturbation suggests that the organisms may have been mining relatively organic-rich horizons within the sand.

Skolithos-type trace fossils like those in facies (ii) are normally considered indicators of unstable shallow water sedimentary regimes (e.g. see Chapter Two). However, in a study of recent sediments from the floor of the Hueneme submarine canyon, Scott and Birdsall (1978, p.60) noted the abundance of vertical, lined dwelling tubes of suspension feeding polychaetes (burrows 0.5-1.0cm in diameter). They recognised that these burrows were a "recent analogue of the trace fossil Skolithos that, where abundant, is considered indicative of a littoral or shallow sublittoral relatively high energy environment" and that "this observation indicates that the distribution of endobenthic organisms within the canyon is not strictly controlled by water depth or sediment type and that organisms respond to other ... stimuli, such as food suply and energy conditions, produced by down-canyon currents" (Scott and Birdsall, 1978, p.60). Contours of benthic stability may be expected to show pronounced seaward deflections in the vicinity of active submarine canyons, channels and fans and this does not contradict the general concepts discussed in Chapter Two. Frey et al. (1978, p.218-9) have also noted that Ophiomorpha has been recorded from environments characterised by Zoophycus and Nereites trace fossil assemblages.

The breccia facies of the Allt na Cùile body probably represents deposition by submarine rock avalanche deposits variably remobilised as debris flows. The sandstone clasts are assumed to be more-or-less contemporaneous (i.e. intraformational) but there is no precise evidence for their exact age; they were clearly consolidated prior to their erosion. The maximum clast size appears to diminish away from the fault line (contrast B5, B4, B3) and breccia dominated sections are probably restricted to within about 250m of the Helmsdale Fault. It is possible that latest Oxfordian or early Kimmeridgian equivalents of the Brora and Clynelish Quarry Sandstone (q.v. Chapter Four) may be being reworked on the upthrow side of the fault. Sykes (1975, p.189) notes that the Allt na Cùile macrofauna (which is presumably reworked) consists of a shallow water assemblage of pectinid bivalves and rhynconellids like that which occurs in the Oxfordian Brora Sandstone (see also Brookfield, 1976). The crude imbrication of the clasts in the Allt Chol breccias might reflect original depositional slope; the imbrication does not dip 'upstream' as is typical for clast-supported conglomerates of the turbidite association (q.v. Walker, 1975). This general interpretation of the breccia facies also holds true for the similar developments at localities B5 and X1.

Analgous breccias have been described from eastern Greenland (facies 10 of Surlyk, 1978). The latter have little or no matrix, are structureless or contain only vaque horizontal bedding and occur only in close proximity to the major faults. Surlyk (1978, p.67) notes that "breccias interfingering with conglomerates show a feint bedding and always rest with a gentle even or slightly undulating surface on the underlying sandstones or conglomerates; they are rarely found as fills in channels or other erosional structures". The breccias occur either as "very thick completely chaotic beds or form more continuous beds only a few metres thick, which seem to have been deposited as more-or-less horizontal units" (Surlyk, 1978, p.76-77). The former mode of occurrence sounds very like that observed in the gorge of the Allt Chol (locality B4), and the latter like that found at locality B3.

The carbonaceous shaley sandstone-sandy shale facies is clearly a transitional sediment type developed at the margins of the Allt na Cùile body. The sandstones were probably deposited by a variety of processes including gravity flows, turbidity currents and traction currents, while the shales represent lower energy background conditions (hemipelagic settling and fine grained turbidites). This facies may correspond to a combination of inter-channel, inter-fan, overbank and levee deposits as defined in the normal submarine fan models (e.g. see Walker, 1978).

Linsley (1972) gives only a very general interpretation of the depositional environment of the "Allt na Cùile sandstones' (remember that by his definition this includes all the sediments in sections B1 to B27, with the exception of B15-20 and B22). He suggests that "whereas at other localities the fault scarp may have been notched with steep canyons, it is envisaged that south of Loth, a wider, more gentle deepening valley, or number of closely set valleys existed. For clarity only one valley is considered here. The head of the valley was near a supply of sandy detritus on the shallow shelf to the east (sic), perhaps near a river mouth or channel along which sediment was normally transported" (Linsley, 1972, p.107). Submarine currents were considered to have channeled this sand and deposited "rather structureless sandstones" (!)

on the downthrow side of the fault while transporting finer material eastwards. "The sediments were affected by the same seismic disturbances that affected or rather caused the formation of the boulder beds. Their effect was to cause the movement of some of the sediment on the shelf and on the valley flanks down the wide submarine valleys redepositing it eastwards of the fault line. There is no evidence that the movement of this sediment ever resulted in turbidity flow. As some of the sediment being redeposited was partly lithified, the result was the formation of intraformational breccias" (Linsley, 1972, p.108). In a rather similar vein, Brookfield (1973, p.528) interprets the absence of the boulder bed facies at Loth as an indication that no fault scarp existed and that sand derived from shallow water areas interdigitated with deep water shales. The interpretation of the "Allt na Cùile sandstone" given in Neves and Selley (1975) refers only to Lothbeg Point exposures and therefore applies to the Lothbeg Point sandstone body (B21a-e) and not the Allt na Cùile body as defined in this work.

Many of the features of the Allt na Cùile body (particularly of facies i, ii, and iv) are very similar to those described from the Raukelv Formation of eastern Greenland by Surlyk et al. (1973) and Surlyk (1975). The Raukelv Formation of Jameson Land is Middle to Late Volgian in age and consists of "cyclically alternating, thick, massive or large scale cross bedded sandstone units and shaley siltstones" (Surlyk et al. 1973, p.49). It lies conformably above the Hareelv Formation which consists of laminated, black, basinal mudstones and intercalated submarine channel sands. The Raukelv sandstones are poorly sorted, coarse or granule grade, variably glauconitic and carbonaceous quartz sands containing common quartz pebbles. They occur as massive units in the form of 0.1-0.3m thick, large irregular sheets and as large scale cross bedded units with sigmoidal, or more commonly, tangential foresets in tabular 0.3-10m form sets. The foreset bedding forms large fans exhibiting very uniform offshore palaeocurrent directions. The massive sandstone units are highly fossiliferous and include "pavements of large pectinids" (Surlyk et al. 1973, p.53) while the cross bedded sandstones contain only very few body fossils (including oysters) but have characteristic Diplocraterion and Monocraterion trace fossils. Cyclicity occurs in the form of repeated coarsening or fining upward sequences consisting of cross-stratified sandstones at the base, overlain by massive sandstones then sandy or silty shales. The basal cross bedded units comprise one or two 5-10m thick sets and have erosional upper surfaces penetrated by numerous vertical burrows and bearing a ferruginous crust. The succeeding poorly sorted massive sandstone unit is 5-10m thick and has an identical erosive, burrowed, ferruginised top which is overlain by silty or fine sandy, parallel-laminated, but often intensely bioturbated, shale with rare ammonites and bivalves. Rootlet horizons occur at certain horizons within the formation and plant debris is present throughout.

In the present context, the main feature of interest is one of the varieties of cross stratified sandstone (facies 7 of Surlyk, 1975). Surlyk notes that "Interbedded with the cross bedded facies there commonly occur very thick ($\leq$60m) units of very coarse grained, even, non-parallel bedded sandstones. The bedding is always of a relatively irregular nature and is either perfectly horizontal or forms a very low angle giant scale cross bedding with a foreset dip of about 5-10°. The facies is bioturbated by Diplocraterion, Monocraterion and Thalassinoides or Ophiomorpha. It contains ammonites and bivalves, the latter deposited as pavements ... The facies was probably laid down in the upper flow regime or in the transition between the lower and upper flow regime" (Surlyk, 1975, p.JNNSS/7-23). The latter facies clearly sounds very like the thinly bedded, current bedded sandstones of the Allt na Cùile body, while the massive sandstones described above (facies 8 of Surlyk, 1975) can be equated with the crudely bedded, pebbly sandstone facies, and the silty shales (facies 1 and 2 of Surlyk, 1975) with facies

(iv) of the Allt na Cùile body. Given the poor accessibility of most of the Allt na Cùile outcrops it is also quite possible that it may contain equivalents of Surlyk's facies 5 and 6, the mega-cross-bedded sandstones of the Raukelv Formation.

Surlyk et al. (1973) consider the change from the Hareelv Formation to the Raukelv Formation to have been produced by tectonic rejuvenation of the coarse sediment supply; the contact between the formations is apparently conformable but the wheatleyensis to pectinatus zones have not yet been proved in either unit and a hiatus may be present. Surlyk et al. (1973, p.72) interpret much of the Raukelv Formation to have been "deposited as huge cross-bedded fans, presumably in the marine part of a delta characterised by great differences in niveau and torrential sediment supply". Surlyk (1975, p.JNNSS/7-25) is more specific and interprets these sediments as "prograding coarsening upward sequences of a braided river deltaic compex with Gilbertian foresets built out in a shallow shelf"; he considers the fining-upward sequences to be distributary deposits and the dark shaley mudstones to be a combination of interdistributary bay, open bay and normal shelf sediments. Surlyk believes the Raukelv Formation to be predominantly marine on the basis of fossils and glauconite, shallow water on the basis of grain size and the scale of the cross bedding, periodically emergent because of the presence of rootlet horizons (water depths 0-100m), and fluvially dominated with only minimal marine reworking because of the consistency of the palaeocurrent directions. The only unequivocal evidence for shallow water lies in the rootlet horizons which implies that the prograding fans occasionally built up toward sea level. The fauna of the Raukelv Formation does not seem to offer much support for a fluvially dominated depositional environment; the massive sandstones which overlie the cross-bedded units contain diverse faunas with cephalopods, bivalves and crinoids, and ammonites and bivalves appear to occur in all the facies. Significant shallowing appears to be implied for the Hareelv-Raukelv transition, possibly related to local tectonics and perhaps associated with a hiatus or period of condensing.

Sykes (1975, p.262) has cited Surlyk (personal communication) as considering the "Allt na Cùile sandstone" to have a strong resemblence to the late Volgian submarine fan delta facies of Wollaston Foreland. This similarity is given general support by my own observations. However, the Allt na Cùile body differs in several important respects:

(1) It consists of predominantly finer grained sandstones.

(2) The grain size and sedimentary structures suggest deposition under lower energy conditions.

(3) Giant foreset (25°) cross-bedding has not yet been recorded (but may well be present).

(4) It does not appear to contain any evidence of emergence.

(5) Its deposition was clearly affected by synsedimentary tectonics (breccia formation).

(6) Sedimentation was probably of a less extensive and smaller scale.

It is interesting to note that both the Allt na Cùile body and the Raukelv Formation overlie what are almost certainly deep water deposits. It is possible that the higher palynomorph abundances, and in particular the higher phytoplankton abundances in the Kintradwell shales, may indicate that their depositional environment was a shallower part of the bottomwater zone than elsewhere in the Kintradwell-West Garty section. The upward increase in phytoclastic debris observed in section A2, may in part reflect progradation in more marginal areas that could have subsequently led to the deposition of the Allt na Cùile body, but the latter is still considered to be a bottom water, or at least a lower mixed layer facies. The contrasts between the Allt na Cùile body and the Rauklev Formation, indicate that the former is a relatively more distal, but related type of deposit.

Summary of facies interpretation, Allt na Cùile sandstone body

The Allt na Cùile sandstone body is a composite unit consisting of at least four main lithofacies. The most 'proximal' lithofacies consists of sandstone breccias which are probably intraformational and occur only in the vicinity of the Helmsdale Fault. The geometry of the breccias will depend on the geometry of the fault line, the distribution of tectonic activity in time and space, the submarine topography at the time of deposition and the mode of emplacement. They probably occur in the form of several discrete or coalesced fans or as a strike-parallel wedge with a relatively flat base and irregular top. The sandstone clasts may have been derived from the uplift of partially lithified sediment in a fault scarp, and/or eroded or slumped from the sides of a submarine valley-canyon. The clasts are probably early Kimmeridgian in age (possibly latest Oxfordian) and were deposited in chaotic masses by rock fall avalanches and debris flows.

The thinly bedded sandstone lithofacies (ii) is considered to have been deposited in distal submarine fan-deltas by lower flow regime traction currents (flowing down or issuing from submarine valleys or canyons?). Such sedimentation was periodically disturbed by fault activity with the relatively thin intercalated breccia horizons which occur in facies (ii) representing debris flows which have moved downslope from the main breccia mass(es) adjacent to the fault. The pebbly sandstones (i) were probably deposited by a combination of gravity flow and traction processes; gravity flows may have been triggered by seismic activity, resulting in the redeposition of loose sand from the upthrow side of the fault (prior to and coincident with breccia formation), by the transfer of inertial energy from slumps and rockfall avalanches, and the over-steepening and instability of fan delta profiles. The shaley lithofacies (iv) occurs as a lateral transition into the background deep water shale environment. The overall facies association suggests a fault controlled, or at least tectonically modified, relatively deep water, submarine fan delta formed by progradation and/or channelling and redeposition of shallow shelf sand.

(b) The Lothbeg Point Sandstone Body

The greater part of the Lothbeg Point sandstone unit consists of massive, structureless sandstone and the only interesting features are those which occur in the transitional basal and lateral facies (see localities B21a, b, c and e). The lateral facies transition is characterised by a change from pure massive sandstones to well bedded, partly laminated sandstones with increasingly more frequent thin carbonaceous intercalations. The base of the body is seen at locality B21e. Thin, lenticular carbonaceous shale beds seem to be the most conspicuous feature of the basal transition and appear to have been deposited in shallow channels or other slight topographic hollows. The presence of irregular loading effects in some parts of the basal sandstone testifies to the sand having been in a liquified state during or immediately following deposition. Rare exposures through the beach cover also indicate shallow erosive channels in the sandstone and the presence of low angle planar cross lamination as proof of the presence of erosive and competent traction currents.

The only previous published interpretation of the Lothbeg Point sandstone is that given by Neves and Selley (1975, p. JNNSS/5-11). They interpreted the sandstone as a grain flow deposit, noting that "the clean well sorted nature of the sediment and its lack of cross-bedding are features found in other sediments interpreted as grain flow deposits". Their report of "massive graded units" appears to be a misleading and unrepresentative description of the transitional facies described above (see locality B21). The predominant massive lithofacies of the Lothbeg Point body is featureless except for "veining" and small amounts of scattered coarse to granule grade grains and rare decalcified comminuted shell debris; my search for dish structures proved unsuccessful. The greater part of the Lothbeg Point sandstone body is here interpreted as consisting of liquified flows to grain flows (probably deposited in a submarine channel) but traction current and turbidity flow processes were probably of increasing importance towards the base and margins. This sandstone facies is 'equated' with the 'massive sandstone' category of Walker (1978), facies B2 of Walker and Mutti (1973) and facies 6 of Surlyk (1978). Its character implies that it is more distal than the facies developed in the Allt na Cùile body; this is also supported by its relative proximity to the Helmsdale Fault.

(c) Un-named sandstone units (localities B11, 12, 24-27 and B8)

In their general appearance, if not in their lithological details, the other major sandstone outcrops show a greater similarity to those at Lothbeg Point than they do with those of the Allt na Cùile body. Like the Lothbeg Point sandstones they are predominantly white, relatively uniform in character and tend to be massive, but they differ in that they contain more conspicuous amounts of pebble and granule grade material and sandstone clasts, and are more commonly bedded. Sparsely to moderately "pebbly" (pebble to granule bearing) sandstones are developed at localities B12, 24 and 25. These sandstones are either massive or exhibit poorly defined horizontal and/or inclined planar bedding that is sometimes picked out by the imbrication and alignment of elongate coarser grains. They appear to be transitional between non-pebbly massive sandstones and clast-bearing, matrix and clast supported conglomerates and breccias. Graded and possibly imbricated, clast supported breccias were only observed at localities B24 and B27, but scattered clasts also occur within the pebbly sandstones at locality B12.

The exposure of the large un-named sandstone units is not sufficient to observe their geometry but it seems likely that they may have been deposited in channels. The clast bearing breccia-conglomerates probably represent proximal turbidity current and debris flow deposits and may be equated with the graded conglomerates of Walker (1975, 1978), facies A1 and A2 conglomerates of Walker and Mutti (1973) and facies 8d, 8e and 8f conglomerates of Surlyk (1978). They represent a subordinate part of the un-named sandstone outcrops and pass into the pebbly sandstones which were probably deposited by a combination of traction processes and liquified and grain flows; the latter may be equated with the pebbly sandstones of Walker (1978), facies A3 and A4 of Walker and Mutti (1973) and facies 9 of Surlyk (1978). The pebbly sandstones in turn pass into non-pebbly massive sandstones like those forming the bulk of the Lothbeg Point body. The overall nature of the un-named sandstone units indicates a facies which (in relative proximal-distal terms) would be situated between those of the Allt na Cùile and Lothbeg Point bodies, as is perhaps suggested by their intermediate proximity to the Helmsdale Fault. The very poorly exposed sandstones of locality B8 probably comprise a very similar lithofacies.

MINOR SANDSTONE UNITS

(a) Loth River section (localities B15-20)

The sandstones of the Loth River sections are 1-2m in thickness and range from 11m to over 19m in lateral extent. They are of lenticular cross section or show flat, more-or-less concordant bases and concordant or irregular tops, and are fine grained and massive but may contain planar carbonaceous laminae (particularly near their margins). The presence of 'false bedding' reported by Callomon (in Arkell and Callomon, 1963) was not confirmed during this study. Although the upper and lower contacts of the sandstones are usually relatively sharp, there is often a progressive increase in the abundance and thickness of sand laminae in the shales immediately underlying the beds. The sands are moderate to well sorted and there is no conspicuous grading. The only previous interpretation of thse sandstone units is that given by Sykes (1975, p.163) who was impressed by their similarity with the massive, submarine channel grain flow sands intercalated in the Hareelv Formation of East Greenland (see Sykes, 1975, p.191-192; Surlyk et al. 1973, Sykes and Surlyk, 1976). The sandstone of the Loth River sections are here interpreted as having been deposited by liquified flows to grain flows with possible subordinate contributions from turbidity current and traction processes. They were probably deposited in shallow channels but may include liquified/grain flow 'tongues' deposited over the shale surface rather than in erosional hollows. They can be equated with finer grained, better sorted versions of Surlyk's "structureless and non-graded sandstones" (facies 6, Surlyk, 1978). Since these thinner sandstone units overlie the Lothbeg Point body, they may form part of a 'waning sequence' resulting from reduced sediment supply or a reduction in the efficiency of redepositional processes.

(b) Crackaig Links, locality B23

The two sandstone outcrops which occur at this locality are clearly composite units consisting of massive and banded sandstones which are variably carbonaceous and contain pebbly bands and horizons with shale clasts. In some parts of the exposure there is clear evidence of channelling. These sandstones are here interpreted as possible channel fills deposited by a combination of grain flows, liquified flows and turbidity currents; the presence of erosional features suggests turbidity currents may have been more important than in (a) above.

(c) Turbiditic sandstones

Relatively thin-bedded (0.5-70cm) sand or sandstone units are common throughout the shale sequence in the Kintradwell-Navidale section. They are predominantly structureless and featureless (particularly when only a few centimetres thick) and only extremely rarely exhibit anything approaching a Bouma sequence (for notable exceptions see sections A1 and D2). Grading is also very rare except where it involves the distribution of clasts and bioclastic debris. The thin bedded sandstones are often more-or-less parallel sided or show only slightly irregular bases;

strongly erosive bases and channelling do occur but they are the exception rather than the rule in most of the sections studied. Amalgamation, the abundance of clasts and proportion of bioclastic material all noticeably increase in the Culgower and West Garty areas. The main lithological variants are discussed below.

(i) Sand or sandstone stringers: thin beds up to 1.5cm in thickness, parallel-sided or irregularly thickening and thinning along outcrop; they can sometimes be observed to pass into other (thicker) sandstone types along strike. These units are interpreted as fine grained turbidites (e.g. T_2-T_4 sequences of Stow and Shanmugam, 1980).

(ii) Parallel sided, structureless, 'ungraded' beds: usually less than 5cm thick, upper part may contain horizontal and very rarely rippled carbonaceous laminae; thicker beds may show sparsely shelly, more irregular bases. These beds are common in all sections and are interpreted as low energy turbidity current deposits.

(iii) 'ungraded' structureless beds with shelly bases and shale clast rich tops: generally between 5 and 70cm thick with erosive bases and planar tops (unless amalgamated). No pronounced grading of sandstone matrix, but lower third to two thirds of bed noticeably richer in scattered (occasionally sub-parallel) shell debris, and the uppermost portion rich in subparallel platey shale clasts. These beds are interpreted as turbidites deposited in situations without a coarse sediment supply; the upper shale-clast rich division appears to indicate the deposition of shale clasts ripped up in front of, and settling out behind, the head of a turbidity current after most of the bed was deposited. The tendency for shale clasts to occur only at the tops of the beds might also result partly from their relative buoyancy within the sandstone matrix. This type of turbidite is best developed in section C1 but occurs throughout the Crackaig, Culgower and West Garty areas. The source of the shale clasts was clearly well consolidated prior to erosion and redeposition.

(iv) 'ungraded' structureless beds with extremely shelly and clast-bearing bases: generally between 8 and 70cm thick with erosive bases and planar tops (unless amalgamated). No pronounced grading of matrix is present except for that produced by very coarse, granule, pebble and cobble grade sandstone lithoclasts. The base of the beds is often extremely shelly (e.g. sandy bioclastic grainstone or sandy biosparite) and contains scattered angular lithoclasts, while the top of the beds are pure or sparsely shelly and clast free, although occasionally with shale-clast rich tops as in (iii). This type of bed grades through bouldery sandstones into clast-supported sandy boulder beds, and is interpreted as a relatively more proximal turbidite (turbidity current transitional to liquified flow?). These sandstones are developed only in the Culgower and West Garty areas (and higher parts of the succession not studied here). Essentially a 'more proximal' version of (iii).

(v) Graded beds approaching Bouma sequences: usually less than 50cm thick and with flat tops and variably erosive bases. May be graded from medium to fine or very fine grained and contain occasional sandstone clasts in their lower part. The lower portion may also exhibit irregular internal folding, the central part poorly defined current bedding,and the upper part convolute bedding and water escape structures or parallel (or rarely rippled) carbonaceous laminae. This type of sandstone appears to represent more-or-less classical turbidites (a-e and b-e Bouma sequences) with various degrees of penecontemporaneous deformation (liquifaction and fluidisation - especially at Kintradwell). They are very rare (only observed in sections A1, A5, D2); this may partly reflect a limited range of grain sizes in the sediment source area.

The sequence (i) to (iv) probably occurs as a lateral transition within a single bed, principally along strike but also possible down dip. Convincing sole structures are not common, but this partly reflects the lie of the outcrop which is not entirely suitable for examining the bases of the beds.

OLD RED SANDSTONE BOULDER BEDS

At least four types of ORS-clast bearing boulder beds are developed in the Kintradwell-West Garty succession. All transitions occur between these 'end-members' and individual beds often show a high degree of lateral variability in the proportion of clasts, matrix type, degree of amalgamation and bed thickness (e.g. see descriptions of sections D2 and E2). The boulder beds are all principally chaotic but elongate clasts are quite often oriented parallel to bedding and crude 'statistical' grading of clasts (distribution and modal size) and rare internal deformation (q.v. Crowell, 1960, p.209) may also be present. No convincing imbrication or reverse grading was observed. The main boulder bed types are described and discussed below.

(a) Thick bedded, chaotic, clast-supported boulder beds

This category applies exclusively to the A and C units on Crackaig Links (see discussion of Crackaig Links area) and represents a lithofacies unique in the sequence studied. These boulder beds are 15 to 30m thick, clast supported and consist of a relatively clean sandstone matrix with angular blocks of ORS of the John O'Groats sandstone facies which often exceed 1m in diameter. They are essentially conformable on the underlying shales and exhibit a homogeneous basal transition layer, where the matrix is argillaceous and clast-poor, and which was probably formed by liquification, shearing and drag during emplacement. Their most similar counterparts elsewhere in the sequence are the Allt Chollbreccia facies of the Allt na Cùile body, but unlike the latter they were apparently deposited much further from the line of the fault and consist almost entirely of exotic clasts. They do not contain any obvious evidence of internal bedding or discontinuities and therefore appear to have been deposited more-or-less en masse. Their conformable basal contacts seem to argue against any high energy mechanism for their <u>final</u> emplacement, but it is possible that they have slid downslope following their initial deposition adjacent to the fault scarp. They are interpreted as rock fall avalanche deposits derived from an ORS fault scarp, variably remobilised as sandy debris flows and possibly finally slumped into position. They are analogous to facies 10 of Surlyk (1978) which consists of chaotic, thick bedded, clast-supported proximal breccias. Surlyk (1978, p.67) notes that such breccias usually have gentle, even or slightly undulose bases and are rarely found in channels or other erosive structures. The 38m boulder bed at Midgarty may be a composite example of this lithofacies.

(b) 'Thinly bedded', clast-rich to clast supported bouldery sandstones

This category includes a whole spectrum of sediments which are transitional between dense, clast-supported boulder beds and clast-bearing (type iv) turbiditic sandstones. Individual beds often demonstrate a high degree of lateral variability in all characteristics. They are generally about 30cm to 1m thick and usually contain 10-25% of ORS clasts which are generally less than 20cm (1-100cm) in diameter, and are scattered throughout or tend to be more abundant toward the base of the bed. The thicker beds are generally composite and contain erosional surfaces and sometimes intercalated shale remnants or bands of relatively clast free or shale-clast rich sandstone. The matrix is usually clean and extremely shelly, but its very base may be argillaceous if the bed overlies shale, and the uppermost part may be relatively shell and clast free. The clasts are usually angular to subangular (more rarely subrounded) and where elongate may be oriented parallel to bedding: they are usually contained entirely within the beds and do not project beyond the sandstone matrix. This type of boulder bed is particularly well seen in outcrop B of section E3, but occurs throughout the section except in the Lothbeg area. These sediments are interpreted as a combination of proximal turbidite, sandy debris flow, density modified grain flow and liquified flow deposits. They can be roughly equated with facies 8a, 8b, 8d and 8g of Surlyk (1978).

(c) Chaotic, clast supported sandy boulder beds

This division represents the classical boulder beds of the Brora-Helmsdale sequence. For that part of the sequence studied here, they are generally 1-2m thick, but for the whole upper Jurassic section they can range from 30cm to greater than 25m, with the thickest units generally composite. The best examples examined during this study were the main boulder bed tongue at east Kintradwell (see sections A3-A5), units I-IV of section D4, and the top of bed section D3i. The following composite description was obtained from the literature.

(i) "The breccias consist of a chaotic jumble of brown sandstone blocks... set within a sparse muddy and sandy matrix which here and there is conspicuously calcareous" (Crowell, 1960, p.203, 205).

(ii) Many of the "occurrences are only one foot thick and in some cases are too thin to include the boulders which belong to them" (Bailey and Weir, 1932, p.446).

(iii) "As the blocks are distributed without any arrangement no stratification is visible within the beds" (Lee, 1925, p.108).

(iv) "Minor folds are revealed within the thick breccia beds by the size sorting and shapes of the blocks" (Crowell, 1960, p.209). "The folds are clearly overturned and so give the sense of movement during emplacement" (loc. cit. p.207). Assymetric subadjacent minor folds occur beneath the boulder beds; they are strongest at the contact and die out quickly and were formed during emplacement (Crowell, 1960, p.209-210). "Recumbent and isoclinal folds ... are usually best developed a few inches to a foot below the breccia base. At places the folds are compressed to form small thrusts" (loc. cit. p.210).

(v) "Although the base of a boulder bed is usually sharp, there is normally a progressive and very noticeable increase in the thickness and concentration of the silt laminae in the shales immediately below the bed" (Linsley, 1972, p.28).

(vi) "The Old Red sandstone blocks are not necessarily restricted to the boulder beds; isolated examples are occasionally embedded or rather sunk into the Kimmeridgian shales, the laminae of which are contorted and follow their contours, thus showing that the blocks must have fallen into the still plastic sediment" (Lee, 1925, p.108).

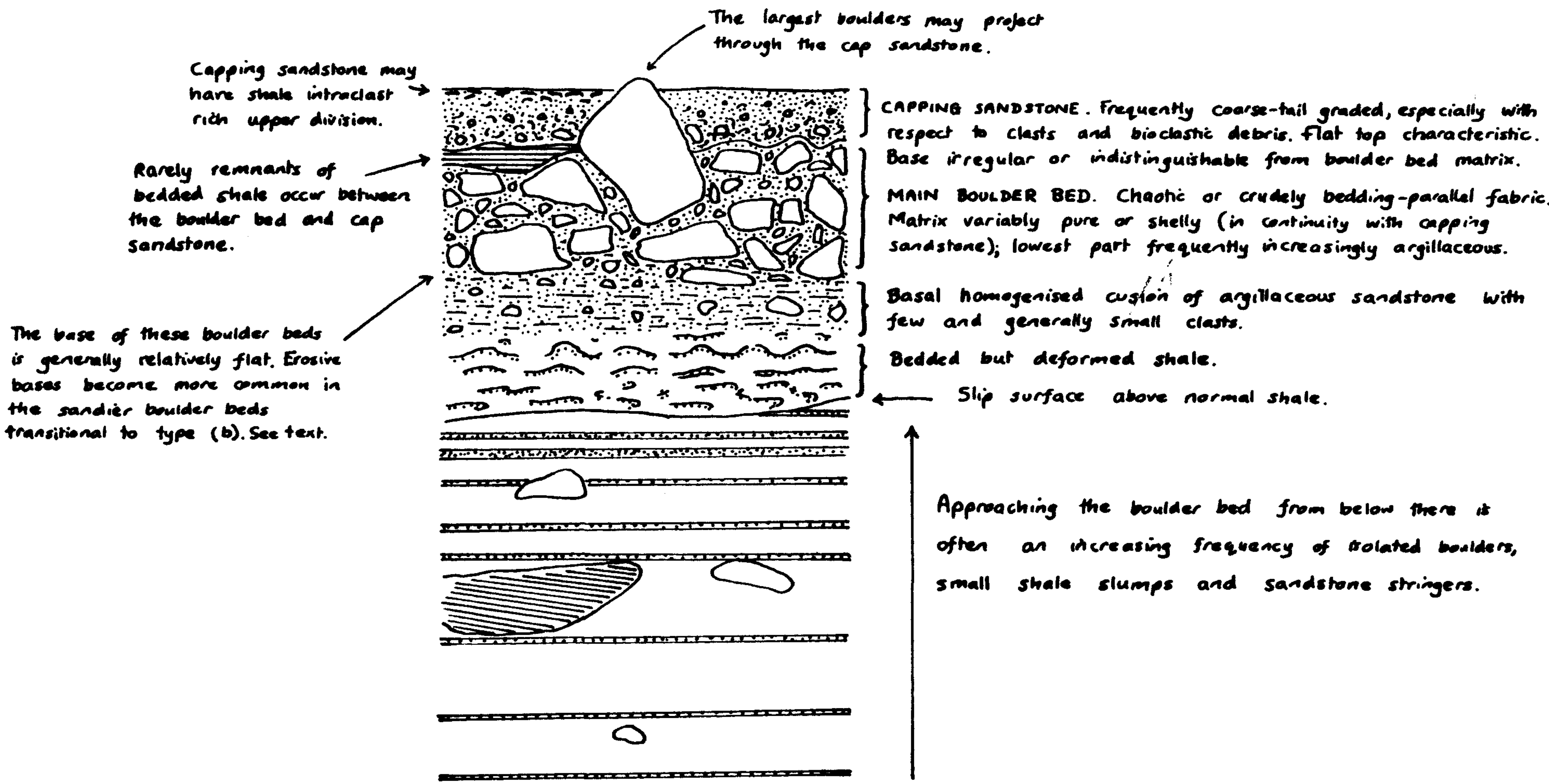
(vii) "The majority of the isolated blocks are less than 3 feet in maximum dimension and are most common ... immediately beneath a boulder bed" (Linsley, 1972, p28).

(viii) "The boulder beds themselves are usually conformable with the underlying shale and disturb the sediments surprisingly little" (Brookfield, 1972, p.529).

The general features of this variety of boulder bed are summarised in Fig. 6.38. The matrix sandstone to the beds is often argillaceous near its base and more rarely throughout the whole bed. Although the matrix can be shelly, the most bioclast-rich portions of the beds are the capping sandstones where these are present (e.g. see section D3 and description of the Helmsdale and Navidale outcrops). These sandstones are generally type (iii) to (v) turbiditic sandstones or type (b) boulder beds. The infilling of the irregular upper topography of the boulder beds by sandstone and or shale often exaggerates the apparent amount of the matrix, but there is often evidence to indicate that the boulder beds and the capping sandstones are distinctly different units (e.g. remnants of shale beds overlying the boulder bed, differences in matrix composition, grading, etc.).

A quick comparison between the size of the included clasts and yet the relative conformity of these boulder beds clearly indicates that they were not deposited from high energy transport mechanisms. They are here interpreted as debris flows (mudflows) sensu Lowe (1979). They represent masses of ORS boulders which have moved down slope under gravity resulting in the distortion and deformation of the underlying sediment but very little if any erosion. Deformation appears to have resulted from shearing and drag, sediment loading and compaction around the

Fig. 6.38 Summary diagram showing principal features of type C boulder beds.



irregular bed boundaries. The boulders may have moved downslope in channels, sheet-like masses or tongues of debris whose geometry was determined only by the supply of boulders. The ultimate source of the boulders must have been a 'deep' submarine fault scarp; the clasts are predominantly angular and unworked and there are no reports of an upper Jurassic epifauna. The supply of boulders may have been strongly episodic and related to periodic movement on the fault,and likewise the downslope movement of boulders accumulated at the foot of the fault scarp may have depended on periodic seismic triggering. Rare isolated boulders within the shales do, however, indicate that small amounts of boulders were continually being displaced down the palaeoslope, albeit less efficiently than when large amounts of debris had accumulated. It is possible that the accumulation of boulders at the foot of the scarp and the resulting loading on the underlying sediments eventually resulted in loss of shear strength and flow and/or the formation of minor horizons of decollement. Such sediment failure would clearly have been facilitated by regular seismic activity. Kelling and Stanley (1976) note that debris flows may generate turbidity currents by the transfer of kinetic energy and by mixing with water at their margins even though their forward motion is normally sluggish. "Thus is it possible for the deposits of a debris flow to be followed directly upward by a turbidite of proximal aspect. Such "bipartite" units are encountered in the fills of some ancient submarine canyons" (Kelling and Stanley, 1976, p.415). While some of the debris flow and capping sandstone couplets may be of such an origin, in other cases the cap sandstone appears to have been a distinctly different event formed by a turbidity current which may have flowed down the same submarine channel or originated from the same point on the fault scarp.

The present interpretation is essentially identical to that of Crowell (1960). "The slow sliding of debris across mud and sand rather than sudden emplacement, was probably the principal process. Here and there isolated blocks lying in the midst of shale and sandstone, evidently rolled and slid into position. They perhaps tumbled down a submarine scree slope as individuals whereas the main breccia layers formed when large masses of the scree became unstable and slid seaward" (.217). "Although the slope of the depositional surface must have been gentle, as judged from the evenness and thinness of bedding, locally slump balls and folds are present in the shales. The palaeogeographic picture ... is of a submarine fault scarp, notched with steep canyons down which inshore material could pass and spread outward to form the sea floor. Between these notches, piles of scree lay along the base of the scarp and locally bordered by debris flows" (Crowell, 1960, p.218). It should be noted that one of the early geological survey workers, Woodward had considered the boulder beds to represent "embedded cliff talus or scree materials" (Lee, 1925, p.107; Macgregor, 1916, p.78), although he probably considered the cliff subaerial rather than submarine. No equivalent lithofacies appears to be developed in the east Greenland fault scarp-fan sequences.

THE BACKGROUND SHALE ENVIRONMENT

The palynofacies data and the nature of the coarse grained interbeds indicates that the Kintradwell to Ord Point shales were deposited in a deep (bottom water) environment. The only indications of current activity are the rare occurrence of current bedding in sandstone beds and the irregularity of the thin sandstone stringers within the shales. Both of these features may be attributed to the action of downslope turbidity currents. Palynofacies evidence indicates that conditions were reducing at least within the sediments if not immediately above. Benthos appears to be very rare within most of the shales and where present often appears to have been redeposited. Bioturbation is similarly very rare and was only conspicuous in the Lothbeg area with and adjacent to the main sandstone interbeds,whose deposition may have been associated with better oxygenation. The sea floor environment may have therefore been predominantly dysaerobic to anaerobic. In the lower part of the sequence the clay grade material may have been derived from the upthrow side of the fault, but higher up in the sequence the presence of redeposited corals argues against any major fluvial source of clay (and plant debris) on the upthrow side, and this material may have entered the basin from a different direction.

The presence of slumps and debris flows is normally taken as an indication that the slope of the sea floor was in excess of 3°, although slumping does occur on slopes as low as 1° (Carter, 1975). In this case the bottom gradients may have been quite slight since the likelihood of regular seismic shocks and the organic-rich nature of the sediments (which would probably have significantly lowered their shear strengths) may have made them more prone to remobilisation than normal. Besides the normal debris flow, it is apparent that the downslope translation of slumped sediment also resulted from the formation of horizons of decollement within the sediments. This largely bedding-parallel slumping may have been analagous to that observed in recent sediments in the 'block -glide terrane' of the Alaskan continental margin described by Grantz et al. (1981, p.478-81). This block-glide terrane consists of slabs of sediment (up to 140m thick) which are sliding downslope over a gently dipping (1°) glide plane. This downslope movement opens up fissures which are 7-17m deep and become infilled from above. Linsley (1972, p.46) has noted that sandstone dykes within the Helmsdale section have horizontally bedded fills (unlike the dykes at Kintradwell) and he likened them to infilled 'bergschrund' type fissures produced by slumping-clearly analagous with the recent features described by Grantz et al. (1981).

SUMMARY OF THE UPPER JURASSIC GEOLOGICAL HISTORY OF THE BRORA OUTLIER

During the Callovian to Oxfordian interval the Brora outlier was characterised by shallow water (upper and lower mixed layer) sedimentation (see Chapter Four for brief reviews). Part of the Oxfordian sandstone sequence contain occasional clasts of Moine and Old Red sandstone (Sykes, 1975, p.25) indicating that Devonian and Precambrian rocks outcropped in the hinterland. During the latest Oxfordian to earliest Kimmeridgian activation of of the Helsmdale Fault resulted in a sharp differentiation between a shallow water area to the north west and a deep, rapidly subsiding basin to the south east, although there was probably no prominent fault scarp developed at this time. During the cymodoce zone the shallow water sands on the upthrow side of the fault were redeposited seawards in prograding submarine fan deltas (e.g. the Allt na Cùile body), while between these fans shales with turbiditic sandstones and occasional fault scarp-derived Old Red sandsone debris flows were deposited. Submarine fan delta sedimentation was periodically interrupted by faulting resulting in the deposition of breccia boulder beds by rock fall avalanches and debris flows. The kaolinite-illite clays in the Allt na Cùile body and other units, were probably formed by the weathering of Middle Devonian lacustrine mudstones exposed somewhere in the hinterland. The background cymodoce shales appear to be bottom water deposits.

The difference in facies between the Allt na Cùile body and the Lothbeg body (and the other similar, predominantly massive sandstones) strongly suggests that the latter are slightly younger in age. This difference in facies appears to reflect a distal shift in the depositional environment with the submarine fan deltas being replaced by channelised grain flow-liquified flow sands. This change occurred during the mutabilis zone when the background shale environment also became apparently more reducing and oxygen deficient. The latter trend correlates with that observed in the type Kimmeridge Clay Formation and may be of secular origin (resulting from a rise in sea level?). The supply of sand from the upthrow side of the fault diminished rapidly in the eudoxus zone and the basinal shale facies became overwhelmingly dominant with only relatively minor intercalations of turbidites and debris flows. This change may have resulted from either the early Kimmeridgian transgression and/or the fact that the shallow water block had been essentially stripped of its loose sediment cover. The eudoxus zone was also the first time that a significant fault scarp appeared on the sea floor, perhaps because there was no longer any large sand influx to swamp it. Clasts consisting predominantly of the John O'Groats sandstone facies of the Old Red Sandstone were derived from this fault scarp at Crackaig Links during eudoxus zone times. The underlying lacustrine mudstone facies of the Old Red Sandstone did not become exposed in the fault scarp in significant quantities until the autissiodorensis and eudoxus zone intervals, after which it is often the predominant contributor to the Old Red sandstone debris flows.

As the coarse sediment supply waned during the eudoxus zone the upthrow side of the fault apparently underwent a significant change in environment. From having been relatively fluvially influenced, it apparently became a warm, euphotic, turbulent shallow water platform where conditions were ideal for the proliferation of heavily calcified, stenohaline benthic organisms including corals and large brachiopods. This change is reflected in the character of the redeposited sediments which become extremely shelly calcareous sandstones (e.g. Culgower, West Garty). The low fluvial influence and the limited supply of terrigenous clastic materials presumably reflects the size of the 'Sutherland' platform area, which must have been small since the black shale facies is also widespread in the West Shetland Basin (see Ridd, 1981). The basinal conditions may have become even more reducing from the wheatleyensis onward. The facies in the upper (unstudied) part of the sequence reflects a proximal shift in the position of the outcrop with respect to the Helmsdale Fault.

The overall character of the sequence is very like that in the coeval sediments of East Greenland (as described by Sykes, 1975 and Surlyk 1978), where Kimmeridgian faulting resulted in the redeposition of shallow water sandstones (e.g. the Pecten Sandstone) in submarine fan deltas (Raukelv Formation) and submarine grain flows (Hareelv Formation). The main contrast appears to result from the fact that following the eudoxus zone the Scottish basin was comparatively starved of sand grade material, while in East Greenland the supply continued throughout the upper Jurassic and lower Cretaceous. The Magnus Member of the Kimmeridge Clay Formation in the Northern North Sea (see De'Ath and Schuyleman, 1981) appears to be equivalent in character to the Lothbeg Point sandstone body and its associated lithofacies.

APPENDIX 6A: LIST OF MEASURED SECTIONS AND LOCALITIES

Section/ locality	Location	Map reference	Thickness logged	% sandstone	% Boulder bed*
A1	Kintradwell Burn	NC921070	6.8m	25	-
A2	Kintradwell beach west	NC926074	37.0m	16	-
A3	"	NC927075	13.5m	21	-
A4	"	"	12.7m	10	-
A5	"	"	9.4m	20	-
A6	Allt Garbh-Chlais	NC925076	5.2m	9	-
A7	Kintradwell bluff	NC926077	5.5m	13	-
B1	Allt na Cùile	NC941093	10.0m	100	-
B2	Allt na Cùile beach	NC941092	5.6m	100	-
B3	Allt na Cùile south west	NC937091	-		
B4	Allt Choll	NC934090	-		
B5	Loth Burn	NC944104	-		
B6	"	NC943104	-		
B7	Allt na Cùile	NC939094	-		
B8	Lothbeg Farm	NC942097	-		
B9	Lothbeg	NC941093-NC944096	-		
B10	"	NC945096	-		
B11	"	NC946096	-		
B12	"	NC947097	7.5m	80	-
B13	Lothbeg beach south	NC946094	11.3m		
B14	Lothbeg beach south east	NC950097	-		
B15	Loth Burn railway bridge	NC952099	21.9m	43	-
B16	Loth Burn	NC950009	-		
B17	"	NC952099	10.7m	75	-
B18	"	NC950099	-		
B19	"	NC950100	-		
B20	"	NC948100	-		
B21a	Lothbeg Point east	NC959096	-		
B21b	"	NC960095	-		
B21c	"	NC959095	-		
B21d	"	NC957P97	-		

APPENDIX 6A (CONT)

Section/ locality	Location	Map reference	Thickness logged	% sandstone	% Boulder bed*
B21e	Lothbeg Point	NC955097	-		
B21f	"	NC955098	-		
B22	Lothbeg Point north east	NC961095	42.3m	5	-
B23	Crackaig Links	NC962099			
B24	Loth Station north west	NC955102	6.0m	100	-
B25	"	NC956101	7.0m	100	-
B26	"	NC957101	-		
B27	"	NC954102	-		
C1	Central Crackaig Links	NC964100	27.0m	29	-
C2	Crackaig Links east	NC971101	13.5m	13	11
D1	Culgower	NC986115	8.9m	11	-
D2	"	NC987114	16.7m	8	-
D3i	"	NC989115	6.1m		
D3ii	"	NC989115	5.0m		
D3iii	"	NC989115	7.5m		
E3	West Garty	NC992119	22.8m	25	3
E2	"	NC993119	27.4m		
E1	"	NC993119	13.2m		
X1	Clyne Burn, Clynekirkton	NC893063	-		
X2	Sput Burn	ND047166	-		

* NB Boulder beds defined as units containing over 10% clasts.

APPENDIX 6B: KEROGEN DATA

Sample	Locality/ Section	Lithology	% Particle Abundance Wd	Wu	I	C	UP	P	M	F	A	Pd	Counts
K63	A1	Shl	34.8	13.6	7.0	0	4.4	0.2	2.6	1.2	23.2	13.0	500
K66		Clast	39.0	17.8	7.0	0.2	8.2	1.4	5.2	1.8	14.4	5.0	"
K68		Shl	35.6	6.6	8.0	0.2	4.0	0.6	0.8	0.4	34.8	9.0	"
K69		Shl	35.4	9.6	7.2	0	3.6	0.8	3.0	0.6	27.4	12.4	"
K70		A.sst	37.0	6.8	4.8	0.8	7.0	1.4	2.4	1.4	27.6	10.8	"
K71		Shl	39.4	16.4	17.0	0	1.4	0.2	3.8	0.2	17.8	3.8	"
K72		Shl	30.8	10.0	6.6	0	5.0	0	2.4	1.0	38.6	5.6	"
K74		Shl	38.8	12.0	12.0	0	4.8	0	1.8	1.2	24.6	4.8	"
K94	A2	Shl	13.6	2.4	4.6	0.2	6.4	0.2	1.8	0.8	53.6	10.4	"
K95		Shl	24.8	4.6	5.0	0	6.4	0.6	2.8	1.2	47.8	6.8	"
K96		Shl	33.0	5.2	6.4	0	4.4	0	1.8	0.2	34.6	14.4	"
K97		Shl	27.6	7.2	7.8	0	6.2	0	1.4	1.0	44.0	4.8	"
K98A		Shl	31.4	8.2	13.4	0	4.8	0	0.6	0	35.8	5.8	"
K98B		Shl	38.8	11.0	10.8	0	7.0	0	1.2	0.4	24.8	6.0	"
K100		Shl	24.2	7.2	11.6	0	8.6	0.4	1.2	1.4	41.2	4.2	"
K101		Shl	27.4	9.6	9.8	0	13.0	0.6	5.2	0.4	31.6	2.4	"
K102		Shl	39.4	11.2	21.6	0.2	3.0	0.2	0.6	0.2	19.0	4.8	"
K103		Shl	42.4	7.2	11.8	0	7.6	0.4	2.0	0.2	24.0	4.4	"
K104		Shl	33.8	8.2	10.2	0	7.2	0.4	1.8	0	34.0	4.2	"
K105		Shl	49.2	10.6	20.2	0	6.4	0	2.4	0.2	6.6	4.4	"
K106		A.sst	32.8	8.0	13.6	0	5.8	0.2	1.4	1.4	31.4	5.4	"
K107		Shl	34.2	9.2	6.4	0	5.6	0	0.6	1.0	37.4	5.4	"
K108		Shl	49.2	25.0	14.4	0	3.2	0.2	2.8	0.2	1.4	3.6	"
K109		Shl	47.8	16.6	12.0	0	3.6	0.2	2.2	0.4	8.8	8.4	"
K111		Shl	47.8	11.2	14.0	0	4.8	0.2	3.6	1.0	11.8	5.6	"
K2B	A3	Shl	38.6	21.0	7.8	0.2	3.6	1.4	7.6	1.4	9.6	8.8	"
K3		Shl	24.6	16.0	5.0	0	5.6	2.2	5.8	2.2	30.2	8.4	"
K4		Clast	39.2	19.6	10.0	0	3.2	0.6	6.0	1.4	15.4	4.6	"
K5		Shl	30.2	15.6	5.2	0.6	4.2	0.8	10.0	3.8	26.6	3.0	"
K9		Shl	38.2	30.8	5.8	0.2	3.0	1.2	8.8	1.8	7.4	2.8	"
K10		Shl	33.8	15.2	9.2	0.4	1.4	0.6	9.0	1.8	26.0	2.6	"

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
K13	A3	Shl	31.4	15.6	4.8	0.2	3.2	1.0	4.6	2.4	34.8	2.0	500
K15		Shl	33.0	13.4	9.6	0.2	3.4	1.4	6.8	3.0	29.4	1.8	"
K26		Shl	25.0	13.8	4.0	0.2	13.8	1.6	5.2	2.4	27.2	6.8	"
K27		Shl	21.8	14.0	5.8	0	10.8	1.4	11.2	1.4	27.0	6.6	"
K22	A3/A5	BB	30.6	18.2	7.6	0.4	8.2	1.4	5.2	2.6	20.0	5.6	"
K29		Shl	32.6	18.6	12.2	0	6.8	2.2	8.6	1.2	11.6	6.2	"
K37	A4	Shl	30.6	18.0	7.6	0.2	7.0	3.2	7.6	3.0	14.4	8.4	"
K38		Shl	33.4	11.8	4.6	0	10.6	4.0	9.8	1.6	19.0	5.2	"
K39		Shl	25.8	8.4	5.0	0	8.6	2.4	8.6	1.2	32.4	7.6	"
K42		Shl	40.4	15.8	7.8	0	6.0	1.8	6.8	1.4	9.8	10.2	"
K43		Shl	32.6	23.2	9.4	0	5.6	2.0	9.0	0.4	9.8	8.0	"
K44		Shl	33.6	21.6	12.8	0	5.2	1.4	4.8	0.8	15.0	4.8	"
K45		Shl	31.0	18.8	7.2	0.4	11.0	3.2	12.4	2.2	7.2	6.6	"
K46		Shl	31.6	11.8	8.2	0	8.4	1.6	7.2	1.2	23.2	6.8	"
K48		Shl	36.6	12.4	10.6	0.4	10.4	1.6	8.8	1.4	10.4	7.4	"
K16	A5	Shl	32.8	11.2	5.0	0	2.2	1.0	6.2	1.4	35.2	5.0	"
K18		Shl	32.6	7.2	7.8	0	4.4	0.6	7.6	1.4	26.0	12.4	"
K23		Shl	33.2	17.2	16.8	0.6	7.0	0.2	4.8	1.0	13.6	5.6	"
K25		Shl	36.2	12.0	3.0	0.2	9.2	1.0	9.4	0.8	20.4	7.8	"
K49	A6	Shl	25.6	9.6	10.2	0	10.0	0.2	4.6	2.2	32.0	6.2	"
K50		A.sst	35.4	18.2	19.4	0	5.2	1.0	6.0	0.6	8.0	5.2	"
K51		Shl	33.8	19.0	9.2	0	7.4	2.0	6.8	1.0	13.6	7.2	"
K52		A.sst	39.2	6.6	11.0	0	4.4	0.4	2.0	0.6	18.6	17.2	"
K53		Shl	39.8	10.2	7.0	0	5.8	0.2	1.8	0.8	21.3	13.0	"
K54		Shl	36.4	14.6	8.6	0.2	6.2	0.4	7.2	0.4	17.2	8.8	"
K32	A7	Shl	30.2	17.8	7.0	0.2	11.0	1.0	3.8	0	13.8	15.2	"
K33		Shl	30.2	31.4	6.8	0.2	11.6	1.4	4.4	0.2	7.4	6.4	"
K34		A.sst	37.2	25.0	12.8	0.2	8.8	1.2	9.8	0	0	5.0	"
K36		Shl	14.2	52.4	4.0	0	21.4	1.4	14.2	0	1.8	10.6	"
K82	B1	Snd	33.6	33.6	22.2	0.4	1.2	0	0.6	0	3.4	5.0	"
K86		Cly	53.8	14.6	17.0	0	1.8	0	0.8	0	10.2	1.8	"

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
K87	B1	Snd	38.5	25.8	27.3	0	1.2	0	0	0	4.2	3.0	260
K88		Snd	54.2	29.4	13.0	0.2	0.2	0	1.0	0	0.4	1.8	500
K77	B6	Cly	34.4	9.3	17.4	0	3.5	0	2.4	0.6	36.8	6.6	334
K93	B9	Shl	27.2	44.6	24.4	0	0	0	0.6	0	0	3.2	500
K90	B10	Shl	42.6	16.8	15.8	0.2	8.2	0	1.2	0.4	7.6	7.2	"
K91		Shl	22.0	17.8	17.2	0.2	9.0	0.6	4.6	0	14.4	14.2	"
K122	B12	Clast	39.6	14.4	39.6	0.4	0.8	0	0	0	3.2	2.0	250
K123		Shl	41.2	16.4	26.6	0.6	0.8	0.6	0.8	0	1.6	11.4	500
K114	B13	Snd	56.4	12.2	25.4	0	1.2	0.6	1.0	0	0.4	2.8	"
K115		Snd	46.8	13.4	26.2	0	2.2	0.2	1.6	0.4	5.0	4.2	"
K116		Snd	57.0	13.6	23.4	0	0.4	0	0.6	0.2	2.2	2.6	"
K117		Snd	51.0	13.4	32.2	0	0.6	0	0.4	0	1.0	1.4	"
K118		Snd	53.8	16.2	18.4	0	2.0	0	2.2	0	4.4	3.0	"
K119		Snd	54.0	11.6	25.4	0	1.2	0.2	2.2	0	1.8	1.0	"
K143	B14	Snd	59.6	10.6	24.2	0	0.6	0.2	0.4	0	2.0	2.4	"
K144		Snd	76.0	8.0	9.7	0	0.7	0.2	0.2	0	2.8	2.5	600
K145		Snd	64.2	8.2	18.2	0	0.6	0.6	0.8	0	3.4	4.0	500
K146		Snd	59.3	12.5	18.7	0	1.5	0	0.5	0.2	4.3	3.0	600
K125	B15	Snd	32.4	27.0	32.0	0.6	0.2	0	0.2	0	0.2	7.4	500
K126		Shl	31.2	5.6	5.0	0	5.4	0.4	0.6	0	47.2	4.6	"
K128		Snd	41.2	35.2	17.0	0.2	1.2	0	0.8	0	0.6	3.8	"
K129		Snd	46.3	23.0	10.3	0	3.3	0	1.0	0	7.0	9.1	700
K130		Shl	33.4	2.6	6.0	0	2.2	0	0.8	0.4	50.2	4.4	500
K131		Shl	39.2	4.8	11.2	0	4.0	0.4	0.2	0.4	35.8	4.0	"
K132		Shl	15.6	5.4	1.8	0	8.8	0.2	0.6	0.2	56.6	10.8	"
K133		Shl	29.6	5.6	6.4	0	2.8	0	0.6	0.8	49.0	5.2	"
K134		Shl	42.4	9.0	11.8	0	2.8	0.2	0.4	0.4	30.2	2.8	"
K136		Shl	42.0	11.2	8.0	0	1.2	0	0.6	0.4	32.6	4.0	"
K137		Shl	35.0	3.6	5.8	0	3.0	0.2	0.4	0.8	46.6	4.6	"
K138		Shl	34.2	5.6	9.0	0	2.4	0.8	0.2	0.8	34.8	12.2	"
K201	B18	Shl	42.2	10.7	11.5	0	3.0	0	0.8	0.7	25.5	5.7	600

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
K200	B21c	Snd	31.0	10.8	33.8	0.2	0.8	0	0.8	0	1.0	1.6	500
K147	B21e	Shl	63.2	13.4	19.2	0	0.4	0	0.2	0	2.8	0.8	"
K148		Shl	51.2	11.6	24.8	0	0.6	0.2	0.2	0	10.2	1.2	"
K149		Shl	40.2	7.6	8.2	0	2.0	0	0.4	1.0	36.2	4.4	"
K150		Shl	51.4	10.2	31.8	0	0.2	0	0.4	0	3.6	2.4	"
K153	B21f	Shl	45.8	20.2	25.4	0	1.4	0.2	0.4	0	1.6	5.0	"
K157	B22	Shl	32.8	8.8	9.6	0	2.6	0.2	1.0	0.2	40.6	4.2	"
K158		Shl	33.8	10.8	9.0	0	3.6	0	1.0	1.0	36.4	4.4	"
K159		Shl	24.4	3.8	5.4	0	4.8	1.2	0.6	1.4	52.2	6.2	"
K160		Shl	33.8	4.8	10.0	0	6.6	0.2	1.8	0.6	38.0	4.2	"
K161		Shl	27.2	6.0	8.4	0	5.8	0.8	0.8	1.6	43.0	6.4	"
K162		Shl	31.2	5.4	8.6	0	6.6	0	1.2	1.0	41.6	4.4	"
K164		Shl/Cly	39.4	12.4	17.8	0	4.2	0.2	1.6	0.6	20.0	3.8	"
K166		Shl	33.8	9.8	8.8	0	9.0	0.4	2.0	0.4	33.0	2.8	"
K167		Shl	27.8	9.0	6.6	0	10.8	0.6	5.2	1.4	35.6	3.0	"
K168		Shl	32.8	16.0	10.6	0	11.4	0.4	9.4	0.2	34.6	4.4	600
K169		Shl	31.2	7.8	8.0	0	5.6	0	1.2	0.6	41.2	4.4	500
K171		Shl	33.0	12.4	15.4	0	6.0	0	1.2	1.2	27.2	3.6	"
K173		Shl	22.4	2.8	11.2	0	3.0	0	1.2	1.0	54.2	4.2	"
K174		Shl	31.0	7.2	14.6	0	6.0	0.2	0.8	1.2	38.0	1.0	"
K175		Shl	45.0	9.4	17.2	0	4.6	0.2	1.2	0.4	21.0	1.0	"
K177		Shl	37.6	5.6	17.6	0	4.6	0.6	2.2	0.2	26.2	5.4	"
K178		Shl	35.2	7.4	13.6	0	4.0	0.4	2.8	0.2	34.0	2.2	"
K180		Shl	44.0	7.8	16.6	0	5.8	0.2	4.4	0.2	17.4	3.6	"
K181		A.sst	54.4	9.6	22.0	0.2	1.8	0	0.6	0	4.8	2.6	"
K182		Shl	38.4	8.2	14.6	0	6.2	0.2	1.4	0.2	28.2	2.6	"
K183		Shl	37.4	3.8	13.8	0	6.6	0.2	4.2	1.0	31.6	1.4	"
K184		Shl	36.2	6.4	11.0	0	7.6	0.2	1.0	1.2	34.0	2.4	"
K186		Shl	44.4	8.4	13.0	0	3.4	0.2	3.4	0.8	25.4	1.0	"
K188		Shl	38.2	5.6	16.8	0	3.6	0.2	1.4	0.4	31.6	2.2	"
K190		Shl	39.0	6.4	9.6	0	2.8	0	1.6	0.6	37.6	2.4	"

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance Wd	Wu	I	C	UP	P	M	F	A	Pd	Counts
K192	B22	Shl	30.8	7.4	11.0	0	1.8	0.2	0.2	0	43.8	4.8	500
K193		Shl	28.4	6.8	13.4	0.2	3.0	0.6	0.8	0.6	41.2	5.2	"
K194		Shl	33.5	8.0	18.7	0	4.3	0.2	1.8	0.5	30.2	2.8	600
K195		Shl	40.6	7.2	12.2	0	5.2	0.2	2.0	0.2	29.6	2.8	500
K196		Shl	40.3	4.7	10.5	0.6	7.2	0	2.5	0.8	29.7	4.2	600
K197		Shl	32.6	6.0	8.0	0	6.8	0.4	3.2	0.2	39.8	3.2	500
K198		Shl	37.5	7.8	11.8	0	7.5	0.6	5.2	0.7	26.2	3.2	600
K202		Shl	32.0	7.6	4.4	0.2	4.6	0	0.4	0	47.6	3.2	500
K204		Shl	34.2	3.2	6.6	0	2.0	0	0.8	0.2	45.8	7.2	"
K205		Shl	30.6	9.6	5.6	0	7.2	0.4	1.0	1.2	41.8	2.6	"
K206		Shl	40.8	13.4	9.0	0	8.8	0.2	2.2	1.0	21.8	2.8	"
K207		Shl	40.4	7.6	6.0	0	5.6	0	1.6	0.4	35.6	2.8	"
K208		Shl	31.6	5.2	4.6	0	7.2	0	2.0	1.0	43.6	4.8	"
K210	B23	Snd	34.6	12.0	9.8	0.2	0.6	0	0	0	1.6	1.2	"
K211		Snd	50.3	27.8	19.0	0	0	0	0.3	0	1.3	1.5	400
K212		Clast	39.8	11.2	7.4	0	4.6	0	2.8	0.4	31.6	2.2	500
K214		Shl	38.2	6.4	6.8	0	4.4	0	0.8	1.0	39.4	3.0	"
K154	Cx	Shl	19.0	23.4	13.4	0.2	3.4	0.6	1.4	0	29.2	9.4	"
K216	Cl	Shl	43.2	6.3	4.7	0	7.8	0	4.2	0.2	28.0	5.7	600
K217		Shl	37.8	1.6	2.6	0.4	5.0	0	1.6	0.4	46.2	4.4	500
K220		Shl	45.8	14.4	11.6	0	5.6	0	2.4	0.2	16.4	3.6	"
K221		Shl	39.8	2.2	2.6	0	5.0	0	2.6	1.6	4.3	3.4	"
K222		Shl	37.0	8.0	5.6	0	5.2	0	1.8	0.8	37.4	4.2	"
K223		Shl	44.2	11.0	5.2	0	4.6	0	3.2	0.6	26.6	4.6	"
K224		Shl	40.4	10.2	6.6	0	3.0	0	2.6	0.8	30.8	5.6	"
K226		Shl	44.8	11.8	6.8	0	3.2	0.2	2.8	1.6	20.8	8.0	"
K227		Shl	35.4	8.4	5.0	0	8.6	0	5.2	0.6	24.6	12.2	"
K228		Shl	58.6	5.6	4.6	0	2.8	0	1.0	0.4	23.0	4.0	"
K229		Shl	41.2	6.6	4.2	0	7.6	0.2	4.4	1.0	27.6	7.2	"
K230		Clast	41.8	8.2	8.4	0	4.0	0	2.2	0.4	30.8	4.2	"
K233		Shl	35.4	7.8	9.0	0	3.2	0	1.4	0.2	37.4	5.6	"

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
K234	C1	Shl	36.4	8.0	8.6	0.2	4.8	0.2	3.4	0.6	33.4	4.4	500
C254	C2	Shl	46.5	11.2	8.2	0	4.2	0	1.2	1.2	25.7	2.0	600
C255		Shl	28.2	9.2	3.4	0	3.6	0	0.4	1.6	49.8	3.8	500
C257		BB	42.7	12.0	7.3	0	6.2	0.2	4.0	1.0	21.7	5.0	600
C258		Shl	35.0	6.8	5.8	0	1.2	0	1.4	0.2	44.6	5.0	500
C240	C(B)	Shl	47.8	4.8	4.4	0	2.8	0	2.6	0.4	32.4	4.8	"
C248		Shl	46.2	17.8	4.7	0	5.5	0	2.2	0.8	17.5	5.3	600
C250		Shl	49.8	15.4	9.8	0	4.4	0	1.0	1.0	16.0	2.6	500
C251		Shl	38.0	6.6	5.8	0	2.6	0.2	1.4	1.6	41.4	2.4	"
C252		BB	29.2	4.4	6.8	0	16.0	0.4	1.2	0	34.4	7.6	250
B243	C(B*)	Shl	39.5	4.0	0.8	0	8.2	0	3.5	0.3	37.3	7.3	500
C245	C(C)	BB	58.2	10.8	9.2	0	3.8	0	1.4	1.0	10.8	4.8	"
C237	C(C*)	BB	42.3	8.0	5.5	0.3	8.3	0.2	1.7	1.3	26.5	5.8	"
K235	C(E)	Shl	33.6	6.2	2.6	0	5.0	0	3.8	1.4	42.6	4.8	"
C236A		Shl	37.0	9.4	3.2	0.2	4.8	0	1.8	1.2	37.8	4.6	"
C236B		Shl	36.4	9.4	3.4	0	6.4	0	0.4	0.6	37.8	5.6	"
C242		Shl	37.4	6.0	4.6	0	3.8	0	0.8	1.2	42.2	4.0	"
C274	D1	Shl	45.4	15.6	4.0	0	4.6	0	1.6	0.4	26.0	2.4	"
C276		Shl	39.9	16.1	3.7	0	7.0	0	6.3	0.6	23.6	2.9	700
C266	D2	Shl	46.4	13.8	5.8	0	3.2	0	0.4	0.6	25.4	4.4	500
C268		Shl	46.8	13.0	5.0	0	1.6	0	1.6	0.2	27.0	4.8	"
C269		Shl	39.2	22.7	10.8	0	2.7	0	2.2	0.3	18.8	3.3	600
C270		Shl	53.4	7.6	6.4	0	1.8	0.2	0.8	0.6	27.4	1.8	500
C271		Shl	44.6	8.6	6.8	0	2.8	0	1.6	0.8	30.0	4.8	"
C272		BB	50.0	15.2	10.2	0	2.8	0.8	1.7	0	17.0	2.3	600
C286	D3i	A.sst	45.7	9.2	4.5	0	3.7	0.3	3.3	0.5	30.7	2.2	"
C287		Shl	48.8	11.9	4.7	0	3.2	0	2.6	0.6	25.2	3.0	725
C288		Shl	46.6	12.3	6.3	0	2.3	0.1	2.4	0.3	26.1	3.6	700
C289		BB	51.9	11.6	6.3	0	2.7	0	2.7	0.3	21.1	3.4	"
C292		Shl	53.4	14.3	9.0	0.1	3.1	0.3	2.9	0.6	14.9	1.4	"
C283	D3ii	BB	42.8	1.3	3.8	0.2	5.0	0	2.7	1.3	27.0	3.8	600
C284		Shl	37.6	10.8	5.8	0	3.4	0	2.6	0.2	34.6	5.0	500

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
C285	D3ii	Shl	41.5	12.5	4.0	0.2	7.8	0.2	3.0	0.7	26.2	4.0	600
C277	D3iii	Shl	51.2	13.8	7.6	0	2.6	0	1.6	0	22.0	1.2	500
C278		Shl	45.9	15.4	8.6	0	5.4	0	1.6	0.4	17.9	4.9	700
C259	D4	Shl	37.8	5.2	3.4	0	6.6	0	1.6	0	42.0	3.4	500
C298		Shl	39.4	12.6	8.0	0	7.0	0.6	5.6	1.4	21.4	4.0	"
C299		Shl	45.0	6.6	11.6	0	2.3	0	0.6	0.6	29.8	3.4	"
C300		BB	38.2	10.8	16.2	0	5.6	0	2.0	0.2	22.2	4.8	"
C301		Shl	35.6	7.0	11.6	0	3.2	0.2	1.4	1.0	35.8	4.2	"
C302		Shl	43.8	9.0	11.4	0	6.0	0	1.2	0.8	23.4	4.4	"
C303		BB	44.4	10.0	8.0	0.4	8.2	0.6	4.6	0.6	16.0	7.2	"
C305		Shl	43.2	6.6	13.2	0	3.4	0.2	1.2	1.0	29.2	2.0	"
C306		BB	18.8	4.0	6.8	0	1.2	0.4	1.2	0.2	62.8	4.6	"
C307		Shl	43.8	11.0	18.2	0	4.8	0.2	1.8	0.4	18.2	1.6	"
C313		Shl	44.4	7.4	13.2	0	2.8	0.2	1.8	0	27.4	2.8	"
C314		Shl	42.2	3.8	7.6	0	4.0	0	3.6	1.0	34.8	3.0	"
C316		Shl	46.0	9.8	13.2	0	5.0	0.4	1.8	0.4	19.2	4.2	"
C317		BB	48.4	9.2	12.8	0.2	5.2	0.2	2.6	0.8	16.4	4.2	"
C328		Shl	40.0	3.4	10.6	0.2	5.2	1.8	2.2	0.2	31.2	5.2	"
C346	E1	BB	43.2	14.0	9.0	0	4.0	1.2	1.8	0	23.0	3.8	"
C349		Shl	26.6	4.0	8.6	0	1.1	0.3	0.3	0.9	52.3	6.0	350
C350		Shl	38.3	11.8	15.8	0	2.5	0.5	1.3	0.8	27.8	1.5	400
C353A		Shl	43.0	8.8	13.4	0	2.4	0	2.2	0.2	27.2	2.8	500
C353B		Shl	48.8	15.2	11.8	0	5.0	0	1.0	0.4	13.4	4.4	"
C336	E2	Shl	31.6	7.6	12.2	0	3.6	0.4	1.8	0.4	39.2	3.2	"
C338		A.sst	32.8	20.8	10.4	0.2	9.8	1.8	1.2	0.8	17.0	5.2	"
C339		BB	47.8	12.2	15.6	0.2	4.4	0.6	3.0	0.6	10.2	5.4	"
C342		Shl	39.8	9.0	9.0	0	4.4	0.2	2.2	0.4	30.2	4.8	"
C343		Shl	51.6	8.0	13.6	0	1.0	0.2	0.8	0	23.8	1.0	"
C308	E3	Shl	51.8	12.4	14.0	0.4	4.2	0.2	1.2	0.4	12.8	2.6	"
C309		BB	48.0	13.2	17.0	0	5.2	0.2	1.8	1.0	9.4	4.2	"
C319		Shl	43.0	9.0	10.4	0	2.0	0	2.0	0.8	28.8	4.0	"

APPENDIX 6B (CONT)

Sample	Locality/ Section	Lithology	% Particle Abundance										Counts
			Wd	Wu	I	C	UP	P	M	F	A	Pd	
C322	E3	Shl	48.8	10.2	13.8	0	2.8	0.2	1.0	0	20.2	3.0	500
C324		BB	56.6	13.0	15.0	0	4.2	0.6	1.3	0	6.6	2.6	"
C325		Shl	44.8	5.4	11.0	0.2	2.8	0.2	1.6	0.2	31.2	2.6	"
C327		Shl	50.0	7.2	5.0	0.2	4.6	0.8	4.6	0.4	23.6	3.6	"
C329		Shl	51.8	11.2	12.0	0	3.2	0.2	1.4	0	17.0	3.2	"
C330		Shl	42.2	13.4	19.0	0	2.2	0.2	0.8	0.2	18.8	3.2	"
C332		Shl	37.0	10.6	10.4	0.2	4.4	1.0	0.8	0.4	30.8	4.4	"
C333		Shl	39.6	8.2	12.2	0	3.0	2.0	1.0	0.2	29.8	4.0	"
C293	E4	Shl	42.8	9.8	8.8	0	2.4	0.2	1.6	1.0	30.8	2.6	"
C357	X1	Snd	54.0	12.8	13.8	0	7.4	0.4	2.6	0	1.6	7.4	"
C358		Cly	53.1	11.7	22.6	0	0.6	0	0	0	7.4	4.6	350
C359		Snd	41.5	8.2	18.5	0	10.3	0.7	2.7	0	1.5	16.7	600

Lithological codes:

Shl	Shale
A.sst	Argillaceous sandstone
Snd	Carbonaceous Shaley Sand/Sandy shale
Cly	Kaolinite-illite clay
BB	Boulder bed matrix
Clast	Argillaceous clast (within sandstone or boulder bed)

APPENDIX 6C: KEROGEN DATA vs. LOCALITY/SECTION/AREA

Locality/	Palynomorphs		A.O.M.		Phytoclasts		Inertinite		No.
Section	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	Samples
A1	8.1	3.0	26.1	7.5	56.7	8.4	8.7	3.7	8
A2	8.4	3.2	28.7	14.6	56.0	17.2	11.4	4.6	17
A3	13.9	4.4	23.4	8.8	55.8	10.9	6.7	2.1	10
A4	18.8	4.4	15.7	7.6	54.3	9.1	8.1	2.4	9
A5	13.4	3.8	23.8	7.9	53.8	7.9	8.2	5.3	4
A6	11.9	3.5	18.5	7.4	59.1	8.4	10.9	4.0	6
A7	22.5	8.5	5.8	5.4	62.3	9.9	7.7	3.2	4
ALL A	13.9	4.9	20.3	7.2	56.9	2.7	8.8	1.6	(7)
B1-cly	1.7		4.5		90.8		19.9		1
B1-snd	1.4	0.3	2.6	1.7	92.5	3.0	20.8	5.1	3
B6-cly	4.0		17.2		40.8		11.6		1
B9	0.6		0		96.2		24.4		1
B10	11.8		11.0		66.1		16.5		2
B12	1.5		2.4		88.9		33.1		2
B13-snd	2.8	1.3	2.5	1.7	91.7	3.5	25.2	4.1	6
B14-snd	1.6	0.4	3.1	0.8	92.4	1.8	17.7	5.2	4
B15-snd	2.2	1.6	2.6	3.1	88.1	6.1	19.8	9.1	3
B15	4.4	2.2	42.6	8.8	49.8	13.9	7.2	3.0	9
B18	3.8		25.5		64.4		11.5		1
B21c-snd	1.6		1.0		75.6		33.8		1
B21e	1.2	0.7	21.3	15.0	83.2	16.0	21.0	8.6	4
B21f	2.0		1.6		91.4		25.4		1
B22	7.8	3.7	34.3	9.9	54.1	10.4	11.4	4.3	38
B23-snd	0.5		1.5		76.8		14.4		2
B23	6.3		35.5		54.9		7.1		2
ALL B shl	4.4	3.5	19.4	15.3	72.1	16.9	17.5	8.5	(9)
cly	2.9		10.9		65.8		15.6		(2)
snd	1.7	0.7	2.2	0.7	86.2	7.2	22.0	6.2	(6)
C1	7.8	2.8	30.4	8.1	55.5	8.2	6.1	2.5	14
C(E)	6.7	1.5	40.1	2.3	47.3	2.8	3.5	0.7	4
C(B)	8.1	4.9	28.3	9.9	58.7	12.8	6.3	1.9	5
C2	5.6	2.9	35.5	12.0	54.1	10.3	6.2	1.8	4
ALL C	7.1	1.0	33.6	4.6	53.9	4.2	5.5	1.2	(4)
D1	9.8		24.8		62.4		3.9		2
D2	4.0	0.9	24.3	4.7	67.7	5.1	7.5	2.2	6
D3i	5.9	0.8	23.6	5.3	67.3	5.7	6.2	1.6	5
D3ii	8.2	2.1	29.3	3.8	53.4	4.2	4.5	0.9	3
D3iii	5.6		20.0		71.3		8.1		2
D4-BB	8.0	3.8	29.4	19.5	50.9	16.0	11.0	3.8	4
D4	6.9	2.7	27.0	5.9	60.9	6.6	11.9	2.8	10
ALL D	6.9	1.8	25.5	3.1	62.8	5.9	7.6	2.8	(7)
E1	4.7	1.8	28.7	12.9	62.5	12.3	11.7	2.7	5
E2	7.0	3.4	24.1	10.1	64.4	9.1	12.2	2.3	5
E3	5.6	1.8	20.8	8.4	69.2	8.0	12.7	3.6	11
ALL E	5.8	0.9	24.5	3.2	65.4	2.8	12.2	0.4	(3)
X1-cly	0.6		7.4		87.4		22.6		1
X1-snd	12.1		1.6		74.4		16.2		2

APPENDIX 6D: PALYNOMORPH DATA (Key on page 176-177).

Sample	Lith.	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	No.	Microf.
							Palynomorph Groups (%)											
K 66	Shl.clast	0.2	-	0.2	1.8	-	-	-	-	0.4	16.8	0.9	11.6	0.2	10.7	56.5	457	62
K 69	Shl.	-	-	-	-	-	-	-	-	-	19.8	1.2	13.9	-	11.1	54.0	404	52
K 70	A.sst	0.2	-	-	0.5	-	-	0.7	-	0.7	7.7	1.0	20.6	1.2	15.2	52.0	402	82
K 72	Shl.	0.2	-	-	0.2	-	-	0.5	-	0.5	9.6	0.7	15.7	0.9	11.0	60.7	427	37
K 95	Shl.	-	0.2	-	0.2	-	-	-	-	1.2	11.7	2.1	17.5	1.4	11.7	54.0	428	42
K101	Shl.	-	-	-	0.5	-	-	-	-	0.5	28.3	1.5	10.1	0.7	10.8	46.8	406	13
K103	Shl.	-	-	0.4	0.6	-	-	0.6	-	0.8	16.7	1.3	10.4	0.8	14.0	54.2	480	29
K106	A.sst.	0.5	-	0.2	1.5	-	-	1.0	-	1.0	12.2	1.0	12.2	0.7	10.7	58.9	401	38
K108	Shl.	0.2	-	-	0.2	-	-	1.3	-	2.0	16.6	2.0	11.2	1.1	28.7	36.8	457	20
K111	Shl.	0.2	-	-	0.9	-	-	-	-	0.2	22.6	1.9	11.1	1.9	21.1	40.2	470	44
K 2B	Shl.	0.5	0.2	0.5	2.2	0.2	-	0.2	1.0	1.2	15.1	4.6	8.4	1.4	18.5	45.8	417	
K 3	Shl.	-	-	-	-	-	-	0.2	-	-	41.1	1.0	6.9	1.0	12.1	37.1	404	
K 9	Shl.	0.2	-	0.2	-	-	-	0.2	-	0.6	15.6	2.4	16.8	-	9.9	54.0	463	17
K 27	Shl.	-	-	-	0.5	-	-	0.2	-	-	18.3	1.7	6.2	0.5	15.8	66.4	405	21
K 18	Shl.	-	0.2	0.2	0.7	-	-	1.0	0.2	1.4	11.8	1.9	17.8	0.2	7.7	56.4	415	39
K 22	BB.	-	-	-	2.9	-	-	1.7	-	1.4	3.1	0.2	6.0	-	10.1	73.7	415	66
K 29	Shl.	-	-	-	1.6	-	-	0.5	-	0.4	14.4	2.2	11.0	0.2	8.0	62.2	410	28
K 49	Shl	-	-	-	1.0	-	-	0.5	-	0.5	12.5	0.5	13.0	1.4	13.7	56.9	415	40
K 50	A.sst.	-	-	0.2	0.2	-	-	-	-	0.9	4.2	1.1	18.3	0.2	20.9	54.0	454	21
K 51	Shl.	0.2	-	-	0.6	-	0.2	1.9	-	1.1	16.0	1.9	11.3	0.6	12.4	53.5	469	37
K 54	Shl.	-	-	-	0.2	-	-	-	-	0.2	7.8	1.1	14.2	0.9	14.9	61.0	436	21
K 77	Cly.clast	0.4	-	-	-	-	-	-	-	0.4	12.3	1.5	13.8	0.4	12.3	59.0	261	16
K126	Shl.	0.2	-	-	1.2	-	-	0.7	-	0.7	13.5	4.0	9.5	0.5	6.7	62.9	421	24
K131	Shl.	-	-	0.2	1.2	-	-	0.2	-	0.2	20.0	1.2	5.3	0.5	6.7	64.4	419	30
K132	Shl.	0.2	0.8	-	1.7	0.2	-	0.4	-	1.7	18.7	3.8	14.3	1.9	5.7	50.5	475	10
K160	Shl.	0.2	-	0.9	0.2	-	-	0.7	0.2	0.9	13.2	1.4	11.6	0.7	6.4	63.1	423	26
K164	Cly.	1.7	0.2	-	0.7	0.2	0.2	0.2	0.5	0.7	15.6	2.2	9.4	0.7	12.4	55.0	404	33
K167	Shl.	-	-	0.8	0.2	-	-	0.4	-	1.5	7.8	2.7	19.7	0.6	14.7	51.6	477	32
K171	Shl.	-	-	-	1.0	-	-	1.2	0.2	1.0	26.6	3.0	7.6	1.2	7.9	50.2	406	61
K180	Shl.	-	-	0.2	-	0.5	-	0.2	-	0.5	12.0	2.3	22.4	1.4	14.7	45.9	442	14
K183	Shl	0.2	-	0.6	0.2	-	-	0.2	-	0.8	10.8	2.0	15.6	0.4	10.8	58.4	493	33
K198	Shl.	-	-	-	-	-	-	1.2	-	0.9	15.7	4.9	17.4	0.2	19.4	40.3	432	22

APPENDIX 6D (CONT)

Sample	Lith.	Palynomorph Groups (%)																
		A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	No.	Microf.
K206	Shl.	0.2	0.2	-	0.2	-	-	0.2	-	0.9	14.3	0.9	17.5	0.7	13.4	51.0	441	26
K216	Shl.	-	-	-	-	-	-	0.4	-	0.6	7.5	0.4	16.9	0.6	36.0	37.5	467	3
K227	Shl.	-	-	-	-	-	-	0.5	0.2	0.7	10.2	0.9	18.2	1.1	24.8	43.4	440	3
K229	Shl.	-	-	-	0.2	0.2	-	0.5	-	0.5	10.3	2.4	19.8	0.2	16.4	49.4	409	23
K230	Shl.clast	0.5	-	0.2	0.5	-	0.2	1.2	0.2	1.9	11.6	0.5	17.8	1.0	14.0	50.4	415	33
C254	Shl.	-	-	-	-	-	-	0.9	-	0.2	8.0	1.4	13.0	0.7	9.3	66.6	440	50
C257	BB.	0.4	0.7	1.1	1.1	0.2	-	0.9	-	1.6	11.6	6.3	26.8	0.4	5.1	43.6	447	37
C276	Shl.	-	-	0.5	0.2	-	-	0.9	-	0.7	7.3	0.7	13.1	-	17.9	58.7	436	25
C303	BB/shl.	0.2	-	0.5	0.7	-	-	0.7	-	-	16.2	1.9	18.0	-	9.4	52.5	427	21
C327	Shl.	-	-	0.2	0.5	-	-	1.4	-	1.4	23.6	0.2	11.3	0.5	15.0	46.0	441	20
C346	BB	0.3	-	-	0.7	-	0.7	2.7	0.3	1.0	16.7	1.0	9.2	1.4	13.2	52.7	294	65
C357	Snd.	-	-	0.2	0.6	-	-	0.4	-	1.1	11.6	0.4	23.2	0.2	9.2	53.0	466	0
C359	Snd.	0.1	-	0.1	0.3	-	0.1	0.1	-	0.6	14.2	0.8	16.8	0.4	7.5	58.9	720	0

APPENDIX 6E: GEOCHEMICAL DATA (Analyses by Robertson Research)

Sample	Section/ Locality	Lithology	Org.C(%)	Hydrogen Index	Oxygen Index
K 71	A1	Shl.	4.9		
K 74	A1	Shl.	3.0		
K 94	A2	Shl.	12.5		
K 96	A2	Shl.	5.0		
K 98	A2	Shl.	5.9	75	17
K100	A2	Shl.	7.3		
K101	A2	Shl.	7.2		
K102	A2	Shl.	10.8	84	0.5
K103	A2	Shl.	6.5		
K104	A2	Shl.	5.2		
K105	A2	Shl.	8.3	79	16
K106	A2	Arg.sst.	2.6		
K107	A2	Shl	5.3		
K108	A2	Shl.	8.0		
K109	A2	Shl.	4.1	29	32
K111	A2	Shl.	5.2		
K 3	A3	Shl.	6.4	173	27
K 5	A3	Shl.	7.2		
K 37	A4	Shl.	6.5	93	19
K 38	A4	Shl.	6.6		
K 46	A4	Shl.	7.7		
K 16	A5	Shl.	5.2	97	39
K 25	A5	Shl.	3.7	92	49
K 53	A6	Shl.	5.6	251	11
K 54	A6	Shl.	7.4	123	18
K 88	B1	Carb.snd.	3.0	-	14
K155	B6	Cly	0.1	-	-
K142	B4	Cly	0.1	-	1278
K115	B13	Carb.snd.	6.0	42	12
K144	B14	Carb.snd.	3.8	33	17
K145	B14	Carb.snd.	3.0		
K126	B15	Carb.snd.	8.0	63	46
K129	B15	Carb.snd.	5.6		
K130	B15	Shl.	7.6	122	21
K133	B15	Shl.	5.9		
K201	B18	Shl.	6.2	102	34
K147	B21e	Shl.	7.7	33	11
K148	B21e	Shl.	16.7	135	16
K149	B21e	Shl.	14.0	225	5
K153	B21f	Shl.	2.9		
K159	B22	Shl.	4.9		
K164	B22	Cly.	0.9	-	49
K167	B22	Shl.	4.9	130	10
K174	B22	Shl.	6.0		
K178	B22	Shl.	6.5		
K180	B22	Shl.	6.6		
K190	B22	Shl.	8.7		
K206	B22	Shl.	7.0		
K209	B22	Shl.	17.5		
K217	C1	Shl.	10.9		
K221	C1	Shl.	9.0		
K228	C1	Shl.	10.1	215	14
K255	C2	Shl.	8.5		
K248	C(B)	Shl.	9.3	152	13
K250	C(B)	Shl.	7.7		
K262	D2	Shl.	8.0		
K266	D2	Shl.	9.4	182	9

APPENDIX 6E (CONT)

Sample	Section/ Locality	Lithology	Org.C(%)	Hydrogen Index	Oxygen Index
K271	D2	Shl.	9.5		
K287	D3i	Shl.	8.2	266	13
K292	D3i	Shl.	7.6		
K284	D3ii	Shl.	9.2		
K278	D3iii	Shl.	11.3		
K299	D4	Shl.	6.9	232	14
K303	D4	BB/shl.	3.8		
K307	D4	Shl.	5.2		
K316	D4	Shl.	5.9		
K349	E1	Shl.	8.9		
K342	E2	Shl.	5.6	248	23
K322	E3	Shl.	7.3	300	10
K327	E3	Shl.	4.0		
K329	E3	Shl.	6.4		
K332	E3	Shl.	6.9		
K357	X1	Carb.snd.	0.6		
K358	X1	Cly.	0.0		

APPENDIX 6F: POINT COUNT PERCENTAGE DATA (see also Table 6.4)

Sample	Counts	Voids	Qtz.	Lithic	F.plg.	F.orth.	F.micro.	P.C.M.	A.C.M.	Kaol.	Bcl.	Sp.cmt.	Opq.	Fe.	Wd.
K65B	1007	-	38.7	16.0	-	2.8	0.4	-	-	-	0.2	40.3	-	-	-
K65T	1004	-	46.7	11.4	-	3.0	1.3	-	-	-	0.4	36.8	-	-	0.2
K11	1005	-	62.9	2.0	0.2	1.1	0.1	-	-	-	-	33.1	0.5	-	-
K41	624	-	44.6	10.4	-	2.9	0.6	0.2	-	-	1.4	38.5	-	-	1.1
K21	1000	-	58.8	4.9	0.2	0.2	-	-	-	-	0.2	35.7	-	-	-
K83	675	11.6	71.7	6.4	0.1	1.3	0.3	0.3	2.8	0.7	-	-	0.3	4.1	-
K79	1020	21.5	62.4	12.1	-	1.9	0.3	0.1	-	-	-	-	-	1.9	-
K78	1207	9.8	65.8	7.4	0.1	1.6	1.6	-	6.3	7.0	-	-	-	0.5	-
K127	1000	23.1	69.9	6.1	0.2	0.4	-	0.1	0.2	-	-	-	-	-	-
K135	680	-	58.2	4.7	0.1	0.3	0.1	-	-	-	0.6	36.0	-	-	-
K199	1000	18.1	75.9	6.0	-	-	-	-	-	-	-	-	-	-	-
K152	1001	19.2	74.6	5.8	-	-	-	-	0.4	-	-	-	0.1	-	-
K189A	1001	-	60.4	4.9	0.1	0.4	0.1	-	-	-	-	34.1	0.1	-	-
K218	1003	-	41.0	16.1	0.1	2.7	0.8	0.2	0.2	-	0.3	38.7	-	-	-
C232	1002	-	61.3	8.0	0.1	0.4	-	0.1	-	-	2.3	27.8	0.2	-	-
C263	1003	-	39.6	16.4	-	3.0	0.9	0.1	0.1	-	0.5	39.4	-	-	-
C286A	1012	-	29.8	18.9	-	2.4	0.4	-	0.1	-	3.1	44.3	0.3	-	-
C335T	1003	-	28.9	3.9	-	0.6	0.1	-	-	-	24.1	42.4	0.3	-	-
C326	652	-	13.8	2.1	-	0.8	0.2	-	-	-	66.1	16.6	0.2	-	-
C356	1069	12.8	65.9	8.3	-	4.9	1.8	0.1	0.2	0.7	-	-	-	5.1	-
C300	1028	-	32.1	23.0	-	5.2	1.3	0.5	4.2	1.4	-	32.2	-	-	-

KEY

Qtz.	Quartz
Lithic	Lithic grains (predominantly metaquartzite)
F.plg.	Plagioclase feldspar
F.orth.	Orthoclase feldspar
F.micro.	Microcline feldspar
P.C.M.	Primary clay mineral
A.C.M.	Authigenic clay mineral
Kaol.	Kaolinite pore clay (not included in A.C.M.)
Bcl.	Bioclasts
Sp. cmt.	Sparitic calcite cement
Opq.	Opaque material (mostly pyrite, possibly some wood)
Fe.	Ferruginous cement
Wd.	Wood

CHAPTER SEVEN

Sedimentological and palynofacies observations on Phillips wells 16/17-4, 16/17-1 and 16/29-2x from the northern North Sea

INTRODUCTION

This chapter is based on the results of a study arising from six weeks spent logging and sampling approximately 1040 feet of three inch diameter cores from the Toni-Thelma and Maureen oil fields (see Fig. 7.1). Some 160 palynofacies slides and over 50 thin sections were examined, in addition to which Robertson Research carried out 71 organic carbon determinations and 33 pyrolysis assays on my behalf.

The initial aim of this part of the study was to examine some at least potentially more distal facies of the "Kimmeridge Clay" than those available at the onshore localities. The palynofacies part of the investigation took precedence over what, time permitting, would have been a more detailed sedimentological study and the main effort was expended on the cored sequence from well 16/17-4A which is both the most interesting and the most complete of the cored intervals.

To date nothing has been published on any of these wells with the exception that Ziegler (1980) has described the Toni-Thelma fields as occurring in Late Jurassic "proximal fan deposits" and that virtually identical facies to the latter have recently been described from the nearby Brae oil field by Harms et al. (1981) and Stow et al. (1982).

A COMBINED SEDIMENTOLOGICAL-PALYNOFACIES STUDY OF CORES 1-20, WELL 16/17-4A

Lithological description

The cores from well 16/17-4A are composed of 895 feet of sandstones, shales and "boulder beds" ranging in age from Oxfordian to Middle Volgian (biostratigraphy based on dinocysts and supplied by Dr. L.A. Riley; see Fig. 7.2). The cores consist of six lithological units formed by two very similar sequences each composed of three distinct facies arranged in a crudely fining upward pattern. The three principal facies are described below.

(1) Facies A.

This facies corresponds to units 1 and 4 and consists of a thinly bedded alternation of brown-grey or white sandstones and black or dark

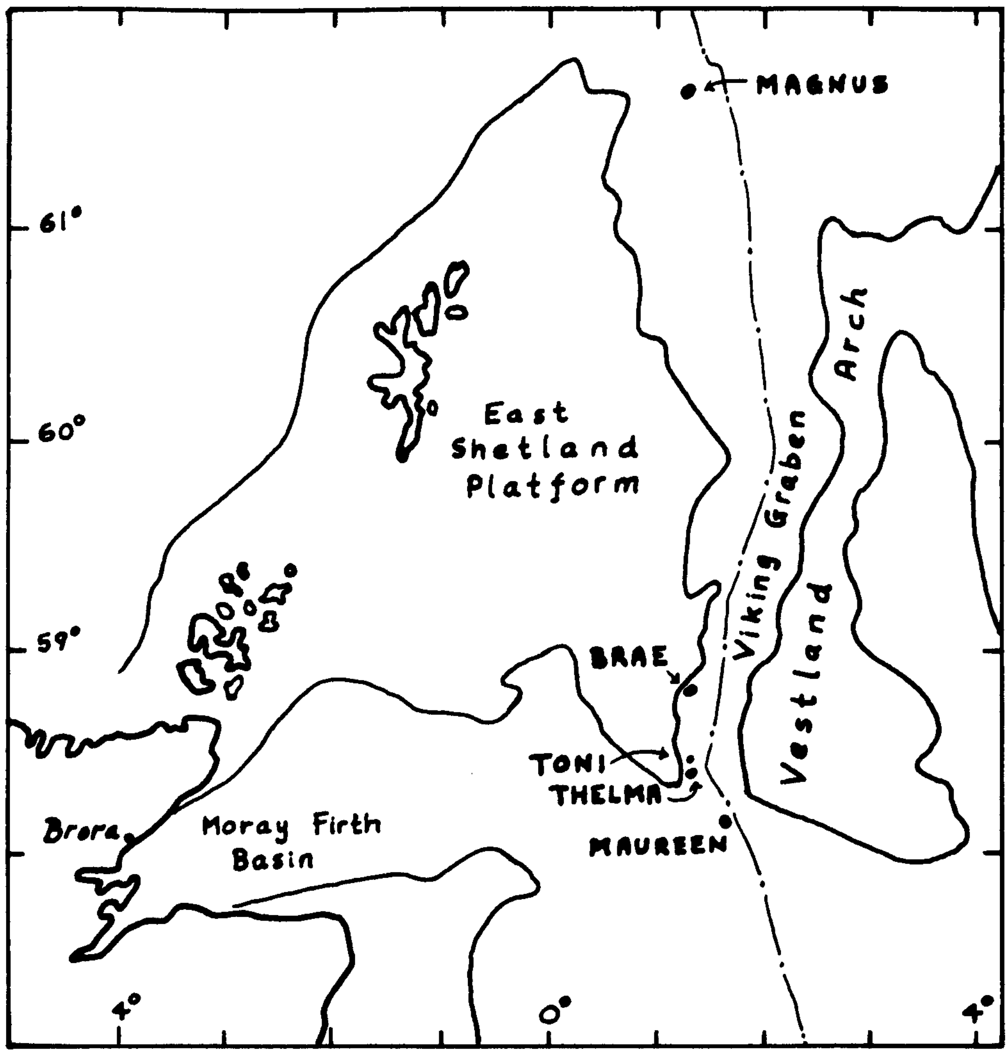


Fig. 7.1 Location map for Toni-Thelma and other oilfields discussed in text.

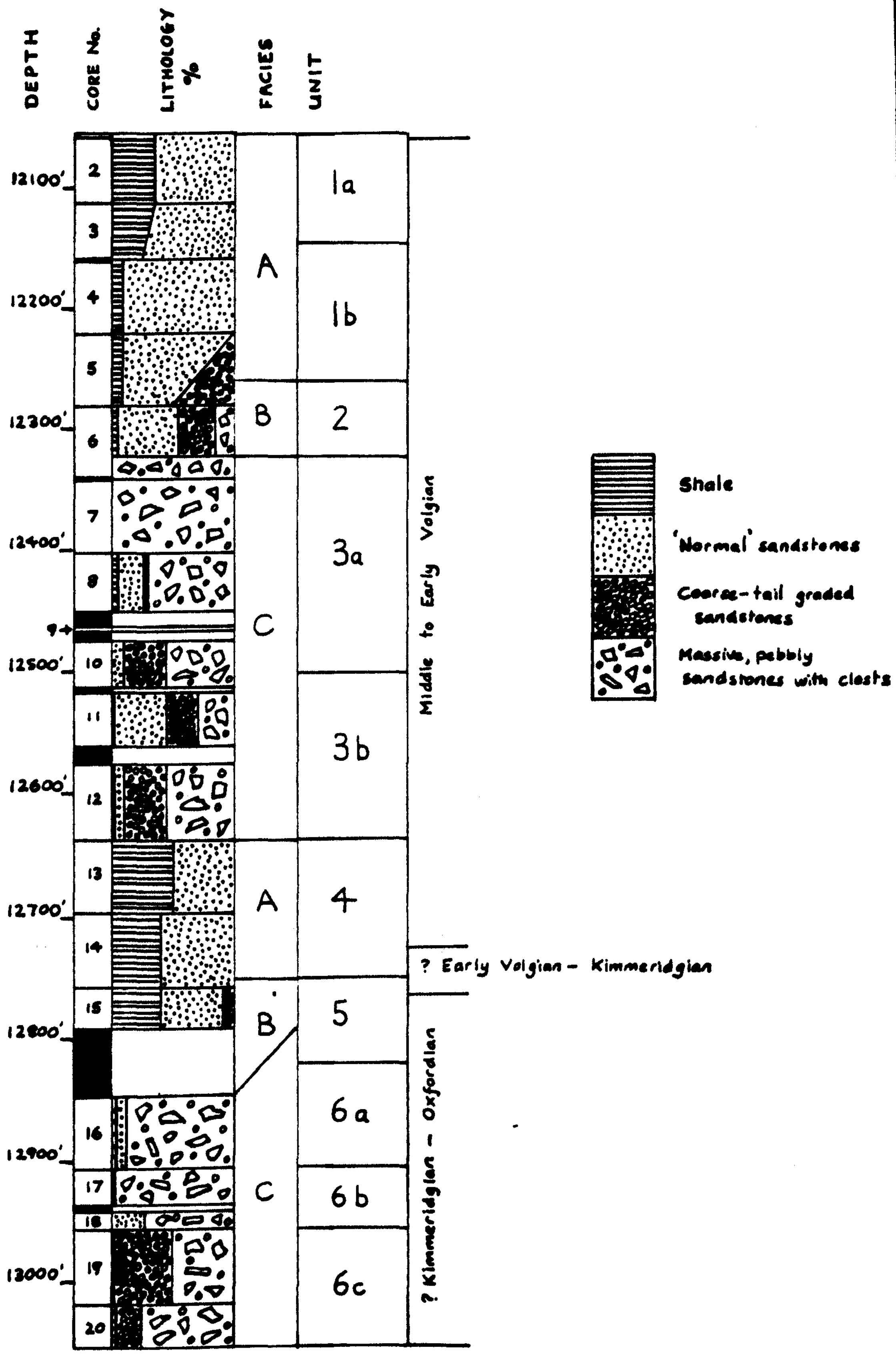


Fig. 7.2 Lithological composition of cored interval from Well 16/17-4A (see also Appendix 7.1)

grey laminated shales. The proportion of shale in individual cores varies up to 50%; in unit 1 it decreases downward from 33% to 6-10% at the base, while in unit 4 it is more variable but averages about 40-50%. Individual un-interrupted shale beds are usually less than 5 cm thick but in unit 1 one 30 cm and two 20 cm thick beds are present. The interbedded sandstones are predominantly fine grained (in the range fine to very-fine - fine to medium grained) except for the top 9 feet of unit 1 which is medium grained. Sorting decreases as modal grain size increases; fine to medium grained sandstones are very poor to poorly sorted and fine to very fine grained sandstones poor to moderately well sorted. Individual sandstone beds generally measure about 5 cm in thickness up to 10-15 cm maximum, but thin lenticular sandstone 'stringers' ($\leq$ 2 cm) are common to abundant throughout the shales.

Grain size variations and/or the presence of thin micaceous, shaley or carbonaceous partings, and laminae rich in platey lath-like shale or carbonaceous clasts ($\leq$ 2 cm long), indicate that the sandstones are predominantly planar-laminated. Larger irregularly bedded shale clasts or disrupted shale laminae are also common in some of the sandstones (e.g. see Fig. 7.3). The most common structure in facies A is microfaulting which is especially abundant where the proportions of shale are highest, but low-relief loading features are also common at the base of the sandstone beds and occasional small sandstone dykes or more irregular liquifaction structures can also be observed. The lenticularity of the thinner sandstones is very suggestive of small scale ripples. Very rare intercalated slumps or otherwise clearly redeposited coarser units are present but are usually only 2-3 cms thick, although one exceptional argillaceous sandstone bed with granules, pebbles and clasts at 12113' measures about 20 cm in thickness. The upper parts of both units 1 and 4 show calcite(?) cementation and the presence of small amounts of bioclastic debris (bivalves and belemnites). No macrobenthos or bioturbation was observed in the shales.

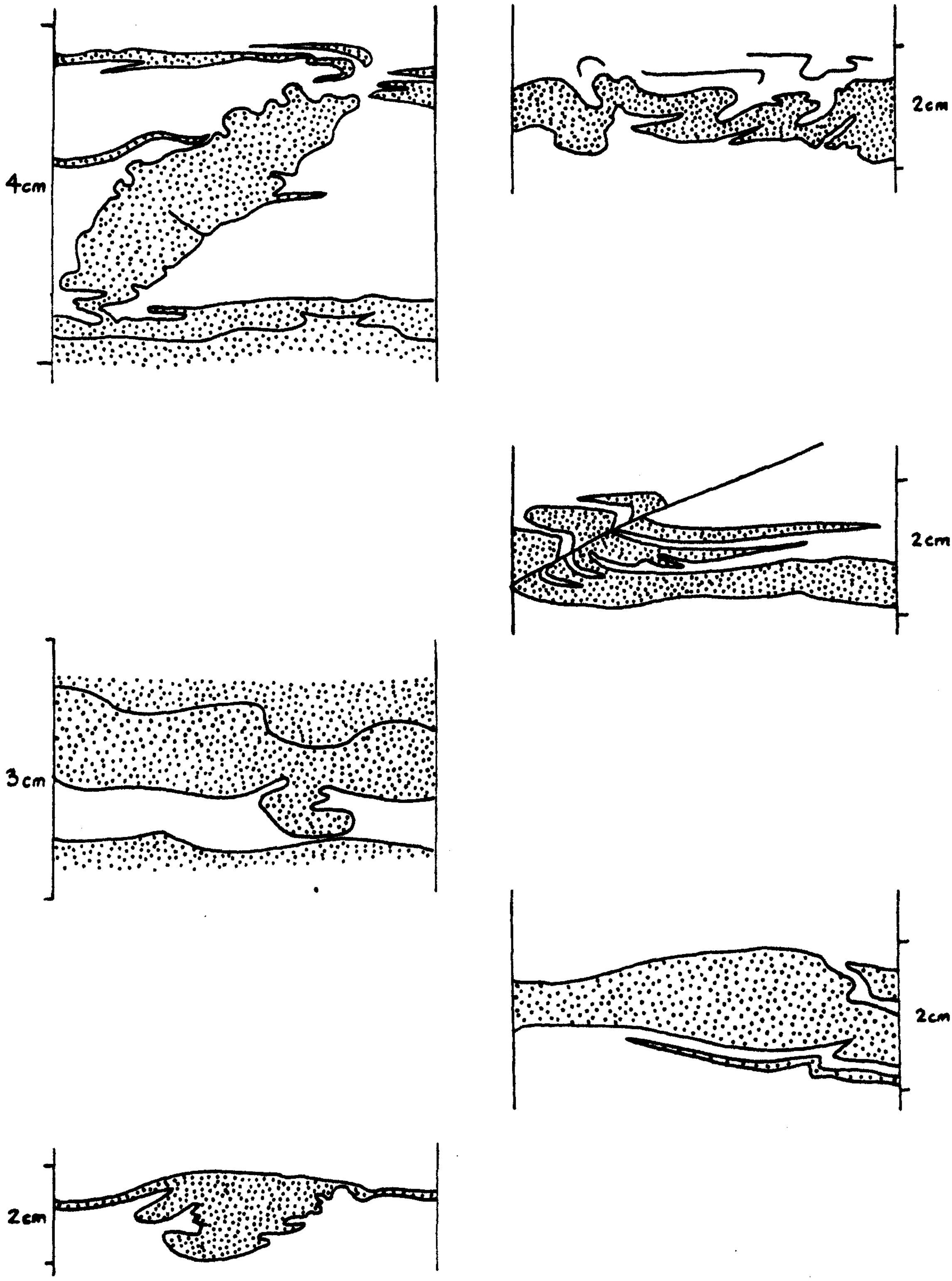


Fig. 7.3 Examples of minor sedimentary structures observed in facies A.

(2) Facies B

This facies forms units 2 and 5 and is almost identical to facies A with the exception that it contains a total of 14-56% (by thickness) of graded or structureless, matrix-supported pebbly sandstones. The pebbly sandstones occur principally as coarse tail graded units of variable thickness (see Table 7.1, Appendix 7.1) characterised by a downward deterioration of matrix sorting and increasingly more coarse and larger amounts of dispersed white or grey (often greenish or pinkish) subrounded to rounded, quartzitic pebbles (predominantly <2.0 cm in diamater). The tops of these graded units (generally the upper 20-50%) are usually pebble-free, sometimes laminated, and may contain carbonaceous laminae or subparallel, platey, carbonaceous clasts. The basal parts of the units only occasionally contain clasts which, when present, are predominantly of wood, shale or claystone, but more rarely of grey-green or light brown sandstones. No bioclastic debris or carbonate cements were macroscopically visible in this facies.

(3) Facies C

This facies is represented by units 3 and 6 and is characterised by its very low shale content (see Appendix 7.1) and an abundance of sandstone clasts which range up to 46 cm in diameter (see Table 7.1). The facies contains a subordinate proportion of clast and pebble-free, fine to medium grained sandstones, and the distribution and presence of recognisable coarse tail graded units (like those in facies B but with clasts) has been used to subdivide units 3 and 6 (Appendix 7.1). The majority of the facies, however, consists of predominantly matrix-supported, poor to very poorly sorted, medium grained, pebbly, clast-bearing sandstones which generally appear to be un-organised except for a variably developed, crude, bedding-parallel clast imbrication. Short intervals of clast-supported sandstone occur locally throughout the facies and are most conspicuous in unit 6 (especially cores 16 and 17). The matrix sandstone is also locally argillaceous (e.g. in parts of

Table 7.1 Statistical trends in thickness of coarse-tail graded (C.T.G.) sandstone units, and mean maximum clast and pebble dimensions

Unit	Facies	No. C.T.G. beds	Thickness (cm) $\bar{x}$	Thickness (cm) S.D.
1	A	1	20.3	-
2	B	19	40.1	12.5
3	C	19	77.7	21.1
4	A	0	-	-
5	B	5	26.4	9.2
6a - c	C	14	79.5	19.9
6a	C	9	74.2	
6c	C	5	88.9	

Unit	Facies	No. Obs.	Max. clast size (cm) $\bar{x}$	Max. clast size (cm) S.D.	No. Obs.	Max. Pebble Size (cm) $\bar{x}$	Max. Pebble Size (cm) S.D.
2	B	5	11.0	3.9	25	1.5	0.9
3	C	25	12.8	11.5	28	2.2	1.4
5	B	2	3.0	0	3	0.7	0.2
6	C	18	10.5	7.3	23	2.5	0.8

cores 8, 12 and 18). The clasts consist predominantly of shale, grey-green, micaceous, fine grained (up to medium grained) laminated sandstones, dark grey-green, harder, cemented, fine to medium grained, quartzite-sandstones, and light brown (to yellowish or orange coloured), fine to medium grained sandstones. The sandstone clasts are nearly all angular except the larger of the dark quartzite clasts which are frequently subrounded to rounded. The clasts are either massive, parallel- or cross-bedded; no diagnostic sedimentary structures were observed.

Petrographic comments on Facies A to C

No systematic detailed petrographic study was carried out during my work on these sediments but general and semi-quantitative observations were made on seventy thin sections (mainly sandstones and principally from units 1 to 3). The sandstones in well 16/17-4A show the same relationship between modal grain size, sorting and composition as observed in Chapter Six, namely the more coarse the sediment the less well sorted it tends to be, and the more lithic the composition. The sandstones are predominantly composed of quartz (as single grains and polycrystalline lithic quartz fragments) and range from sublith-arenites to lithic arenites (sensu Pettijohn, 1975). Allowing for subsequent diagenetic alteration (see below) many of the (non-penecontemporaneous?) sandstone lithoclasts appear to have been more feldspathic in composition than the matrix sediments and were probably subarkoses, although little of the feldspar now survives. Small amounts of bioclastic debris (exceptionally up to 10%) including bivalve, brachiopod and echinoid fragments were observed in sandstones from units 1 and 4, and trace amounts of glauconite were noted sporadically through the section.

Porosity is variably developed in the sandstones, averaging about 10% when present, although about half of the porous samples have values of 5% or less. Higher packing densities and more extensive authigenic silica cements apparently effectively restrict any significant porosity to units 1 to 3a (although the porosity of the lower units may be underestimated due to sampling bias). Carbonate cements in the upper portions of facies A also prevent or restrict porosity in units 1 and 4. The sandstone lithoclasts in facies B and C are usually tight (due to higher packing, silica and/or carbonate cements and pore-filling clay minerals) and only 25% of the clasts examined showed any porosity. Most of the porosity appears to be secondary after the dissolution of k-feldspar, as indicated by corroded or highly altered feldspar remnants and the characteristic angular shapes of most of the pores. Some of these angular pores could, however, result from authigenic quartz overgrowths which grew into original primary pore spaces. Porosity values can be over-estimated in thin section due to loss of pore-filling authigenic clay minerals by plucking during section preparation; a blue stained mounting medium was used to alleviate this problem.

Carbonate cementation is best developed in the upper parts of units 1a and 4 and appears to be correlated with the original distribution of bioclastic debris. The cement varies from a few percent of patchy, medium-coarsely crystalline, replacive sparite, up to 40-50% very coarsely crystalline lustre-mottling, which gives the whole rock a white colour. Where present bioclasts invariably show heavily indented margins indicating that pressure solution occurred prior to the precipitation of the intergranular and replacive cements. Although their maxima correspond, the total distribution of carbonate cements is much greater than that of recognisable bioclasts, which is presumably due to dissolution and complete recrystallisation. Small amounts (2-3%) of patchy sparite is quite common throughout the section and stronger colouration, higher relief and more marked rhombohedral cleavage indicate that a least some of this patchy carbonate is probably dolomite. Sandstone lithoclasts are occasionally preferentially carbonate cemented, which may have occurred either before or after their inclusion in the matrix (the latter due to differential porosity or solution?). The replacive, corrosive nature of the carbonate cements is frequently well illustrated by the way they have attacked the sandstone (and rare micaceous meta-quartzite) lithoclasts. Small amounts of grains, granules and clasts of silty microsparitic limestone are present in the sandstones of facies B and C and are sometimes difficult to distinguish from cements.

It is not uncommon for the sandstones to contain up to 10-20% 'matrix' consisting primarily of authigenic clay minerals. The most common authigenic clay mineral is illite which occurs mainly as concentric (rarely radial) grain coatings and more rarely as pore-filling aggregates. Although a few percent of fresh K-feldspar often remains most of it has been highly altered (sericitised) and transformed into illite (occasionally preserving 'ghosts' of the original cleavage, etc.) or entirely dissolved to leave empty,or once empty pores. Perthite grains exhibit the preferential attack of the more alkaline exsolution lamellae. While illite (as grain coatings and pore-fillings) is much more abundant than kaolinite in most of the sandstones, vermicular and book habit kaolinite pore fillings are fairly common in the associated sandstone lithoclasts where they appear to be the products of feldspar alteration. This may indicate a differing diagenetic path for the matrix and lithoclasts in which the feldspar alteration was more alkaline in the former than the latter. The lithoclasts sometimes show better feldspar survival than the matrix (because the clasts are tighter and less permeable?) but the opposite also occurs; comparisons are complicated by the inherently differing proportions of feldspar. Some of the tight sandstone clasts occasionally show a preferential replacement of the feldspar by carbonate cements, perhaps assisted by earlier solution or weakening of the feldspar by kaolinitisation. Rare thin sections exhibit unaltered feldspar grains which have syntaxial overgrowths.

Minor syntaxial quartz overgrowths occur in the sandstones of units 1 and 2. Here they are limited to the more enclosed pore spaces but they become more common and extensive in unit 3a, below which they are usually common to abundant and lead to lowered porosities. Quartz overgrowths are absent or much less conspicuous in the sandstone lithoclasts of facies C (except perhaps in the more penecontemporaneous clasts?) when compared with the matrix sandstones. Original quartz grain boundaries are usually shown by 'dusty' traces but very occasionally they are picked out by illite (?) grain coatings and sometimes concentrations of opaque (pyrite) crystals. Two to three percent of pyrite is common in most of the sandstones where it sometimes occurs preferentially at grain or bioclast boundaries or forms small patches of 'cement'.

Thin sections of shales exhibited a strong bedding-parallel orientation of elongate silt and very fine sand grains, small amounts of phosphatic debris, kerogenous laminae, trace amounts of glauconite and a complete absence of bioclastic debris. Common ovate 'pellets' (about 50μ in long dimension), apparently composed of clay minerals but of uncertain origins, were also noted in some of the shales.

The diagenetic history of the sandstones in well 16/17-4A was clearly complex and the relative chronology is difficult to establish. Complicating factors include: the control of carbonate cementation at least partly by the original bioclast distribution, the possibility of at least a partly separate diagenetic evolution of the matrix and lithoclasts, the relative importance of depth versus sedimentologic factors, and the possibility of there having been several phases of carbonate and silica cementation, dissolution and/or clay mineral growth. However, a possible diagenetic history is presented below.

(a) Separate evolution of subarkosic lithoclasts prior to their final sedimentation (e.g. weathering and kaolinisation of K-feldspar?)?

(b) Compaction, partial pressure solution of bioclasts

(c) Solution and reprecipitation of bioclasts as replacive calcite cements (feldspars or altered feldspars attacked more easily? cementation of pores formed by earlier solution of feldspars?)

±(d) Rare early formation of illite grain coatings and opaque coatings

(e) Formation of authigenic quartz overgrowths

(f) Feldspar alteration, formation of illite pore fills or feldspar solution leaving open pores

(g) Illite grain coatings

(h) Illite and/or kaolinite pore fillings

(i) Compaction; increase in packing density/concavo-convex grain contacts.

These individual stages probably occurred with a considerable degree of overlap and possible as several discrete generations. The early carbonate cements in step (c) could have undergone subsequent solution to create pore spaces which were subsequently infilled or remained as porosity. The lower permeability of the lithoclasts (except where increased by feldspar dissolution) apparently partly exempted them from various diagenetic processes occurring in the matrix sandstones. A separate phase of late minor calcite and/or dolomite cementation may have taken place.

All the petrographic observations presented above are similar to those briefly reported by Stow et al. (1982) for the nearby Brae Oilfield.

Discussion of 16/17-4A lithofacies

The cores clearly consist of two similar fining upward sequences. The fining upwards trend is reflected in the distribution of sandstone grain sizes (especially by sorting), the overall lithological ratios (Appendix 7.1), the distribution of clasts and pebbles, the mean thickness of the graded units (Table 7.1) and the mean maximum values of clast and pebble size (Table 7.1). Both of the fining upward sequences exhibit similar ratios of the facies A to C, where the lithologically intermediate facies B occurs as a relatively thin transition between thicker sandstone-shale and clast-bearing sandstone units. The distribution of coarse-tail graded (C.T.G.) units in facies B and C indicates a symmetrical pattern illustrating a facies transition not only from B to C, but also C to A (see Appendix 7.1).

All of the facies lack evidence for deposition in a shallow water regime. The range of sedimentary features in facies A is very similar to that described from fine-grained turbidites by Stow & Shanmugam (1980) which suggests that facies A was deposited in a setting compatible with the levee or channel-mouth to interchannel or basinward (e.g. inter-fan) regions of a submarine fan. The rippling observed in the form of the thin sandstone lenticules could 'theoretically' be the result of wave influence, but given other criteria (see later) this does not appear likely. The coarse-tail graded units which characterise facies B probably represent primarily grain flow - liquified flow (and partly turbidity current?) deposits like those observed elsewhere in this study (see Chapters 4 and 6), and hence relatively proximal resedimentation. The sandstones of facies C appear to be representative of the variety of deposits produced by the spectrum of redepositional processes (from rock-fall avalanches - debris flows - grain flows - liquified flows to turbidity currents) which typify highly proximal submarine fan sedimentation in a base of fault scarp setting (see Surlyk, 1978; Chapter 6). The low diameter of the cores and the lack of back-up data preclude more detailed analysis of facies C.

The facies interpretations given above are 'identical' to those made for the coeval sediments of the Brae Field by Stow et al. (1982) whose mudstone, sandstone and conglomerate facies groups are essentially equivalent to facies A, B and C respectively. This is reassuring since their work was based on far more extensive and varied data than was available for this study! By comparison with Stow et al. (1982) the sand-stone clasts in facies C may be primarily of Devonian age, but macroscopic and petrographic observations indicate that at least some of the clasts are relatively penecontemporaneous reworked Jurassic sediments.

PALYNOFACIES ANALYSIS OF WELL 16/17-4A

Kerogen trends

Percentage particle abundance determinations of kerogen composition were carried out on 118 slides from well 16/17-4A, including 77 normal shale samples, 30 'anomalous' argillaceous samples (e.g. shale or clay clasts, argillaceous matrix, claystones, etc.) and 11 sandstone samples (see Appendix 7.2). Robertson Research carried out 51 organic carbon determinations and 33 Rock Eval. pyrolysis assays on selected samples (Appendix 7.3). Only the hydrogen indices from the Rock Eval. data are presented here; the oxygen indices supplied to me were all very low values (mainly between 1 and 3 and all less than 11) which would imply that the organic matter was composed predominantly of algal material and devoid of vitrinitic phytoclasts. Since by visual determination this is very clearly not the case, instrumental or experimental error has been assumed and the oxygen indices discarded. The data for each group of sample lithologies was treated separately and the results were all considered in the context of the previously defined lithofacies and informal lithostratigraphic units.

(a) Shale samples

The means and standard deviations of the main kerogen parameters were calculated for each unit and/or subunit (see Appendix 7.4) and the results are summarised in Table 7.2. All the shales are very organic-rich with organic carbon values ranging from 3-30% by weight. Except for subunit 1a the ratio of A.O.M. to woody material is fairly stable (around 20:60) and the palynomorph content is consistently low. Foraminiferal linings are rare (always less than 1%) but the foraminifer: palynomorph ratio is in fact comparatively high (averaging $\sim$0.3) because of the very low palynomorph content. Many of the lithological divisions in the sequence are reflected in differences in the kerogen characteristics, but the latter also reveals trends which have no lithological counterparts. This is especially true in unit one where a sharp change in the A.O.M. to wood ratio (also reflected in the hydrogen indices) was used to define two subunits. The lithological symmetry and similarity between units 1-3 and 4-6 is not reflected in any comparable way in the kerogen data except perhaps by a possible upward increase in the Wu/Wd ratio.

Table 7.2 Summary of means of main kerogen parameters in shale samples from Units 1 to 6, Well 16/17-4A (see also Appendices 7.1 and 7.4)

Core	Unit	Facies	Org.C.	H.I.	A.O.M.	Tp	I	Tw	Wu/Wd
1-3	1a	A	10%	606	50%	1.5%	4%	37%	0.14
3-5	1b	A	10	330	28	1.5	8	55	0.04
5-6	2	B	10	270	21	1.9	7	60	0.09
6-11	3a	C	13	380*	17	1.2	17	58	0.09
12	3b	C	17	239	19	1.2	14	58	0.29
13-14	4	A	6	208	18	1.1	16	58	0.34[†]
14-15	5	B	11	253	20	1.7	14	58	0.2
16-20	6a-c	C	ND	ND	20	2.8	17	54	0.2

N.B. Horizontal lines indicate major changes in values

Abbreviations as used elsewhere

* One determination only N.D. Not determined

† In Unit 4 Wu/Wd ratios are higher in the upper part than the lower - see Appendix 7.4

Within the phytoclast components it is noted that the distribution of the black wood (inertinitic) fraction is only partly correlated with the other phytoclast materials.

(b) Anomalous argillaceous and sandstone samples

The main kerogen parameters of the anomalous argillaceous and sandstone samples were each separately compared with the mean values of the shale samples in the lithological unit or subunits in which the samples occur. The general results of these comparisons are recorded in Table 7.3. Compared with the unit or subunit shale mean values the anomalous argillaceous samples are significantly different; they are generally less organic-rich, have proportionately less phytoclasts and proportionately more palynomorphs. The sandstone samples are comparatively more variable but also tend to have lower proportions of phytoclasts, although to very varying extents. The latter trend is very atypical when contrasted with the other lithological kerogen trends observed during this study (compare with Chapter 4 and 6). Compared with the background shales the anomalous argillaceous samples have values of a more distal aspect (higher palynomorphs, lower phytoclasts); the lower organic richness may be due to higher mineralic dilution and/or reflect deposition under less reducing conditions in up slope environments (this possible contradiction presumably reflects the variety of the anomalous samples).

Palynomorph trends

An extensive study of the palynomorph trends in well 16/17-4A was prevented by the very low palynomorph densities which were immediately apparent from the kerogen observations. Only a few of the total kerogen slides which were prepared for this study contained sufficient palynomorphs to make counting worthwhile and most of these represented atypical lithologies. In order to help overcome these problems Robertson Research allowed me to examine a suite of 10 previously prepared palynostratigraphic (oxidised) slides, although unfortunately they could not provide any

Table 7.3 Comparison of anomalous argillaceous and sandstone sample kerogen characteristics with those of the corresponding unit (shale) sample means

A Anomalous argillaceous samples (P/4A, 14, 22, 38, 43, 54, 65, 71, 73, 74, 76, 85, 87, 100, 101, 102, 103, 108, 109, 121, 129, 132, 143, 147, 149, 151, 153, 159, 161)

Parameter	Mean Factor	Trend	Occurrence
Org.C	x2-4	down	100% of samples
Tp	x3	up	87%
Tw	x2	down	83%
I	x3	down	90%
Wu/Wd	x2-3	down	90%

B Sandstone samples (P/4A 11B, 42, 45, 50, 63, 95, 96, 97x, 97z, 148)

Parameter	Mean Factor	Trend	Occurrence
Tp	0-0.6	down	54% of samples
Tw	0-40	down	82%
I	0-26	down	82%
Wu/Wd	-	down	100%

details of the exact sample lithology (the latter being especially significant in the pebbly, clast-bearing sandstone intervals). As the A.O.M. content of the kerogens in this well generally averages about 20% it is unlikely that the low Tp (total palynomorph) values result from 'masking' and this was confirmed by the comparatively poor palynomorph densities even in the oxidised preparations. The results of the palynomorph counts are presented in Appendix 7.5.

It is obviously necessary to exercise caution when trying to interpret trends based on as few samples as were available here, but it is encouraging that the data from the palynomorph counts appears to be non-random from sample to sample and that such trends as apparently exist are not entirely correlated with the lithofacies variations in the cores. In general terms the palynomorph assemblages are dominated by sporomorphs and plankton is virtually absent except in the upper parts of facies B. Within the sporomorph category the pteridophyte spores 'dominate' in units 5 and 6 but decline through unit 4 in which the bisaccates become the most abundant sporomorphs. The dominance of bisaccates continues up into unit 3a, where they apparently undergo a rapid decline and pteridophyte spores increase again peaking in unit 2 and the lower part of unit 1a. Note that although the bisaccates reach up to 70% of the total palynomorphs, pteridophyte spores never exceed 50% and are usually subordinate to the sphaeromorphs.

Palynofacies interpretation and discussion

The combination of high proportions of phytoclasts, low proportions of palynomorphs and high organic carbon values in the shales from well 16/17-4A argue for highly proximal sedimentation. This is in good agreement with the sedimentological interpretation of the sequence as a fault scarp-fan association. In such a setting, however, the complex interactions of fan progradation, sea-level changes, synsedimentary tectonics and the interaction of adjacent fans make more sophisticated interpretations of palynofacies untenable on such a limited data base. As was noted earlier very few simple and easily explainable lithofacies-palynofacies correlations are apparent from this study but some of the empirical trends observed from unit to unit might be of some assistance in well correlation.

One of the most conspicuous features of the palynomorph data was the apparent restriction of the plankton to the upper portions of facies A (a similar distribution to that noted for bioclasts). Traditionally this might be thought of entirely in terms of 'degrees of marine influence' but I would suggest that this was not the main cause in this case. The nature of the shales in well 16/17-4A (their consistently laminated character, the total absence of bioturbation and benthic macrofossils, and the abundance of very strongly fluorescent alginitic/bituminitic debris within the A.O.M. when observed under U.V. fluorescence) strongly suggests a consistently anaerobic to anoxic bottom water environment during the sedimentation of all six lithological units. In such a setting one would only expect dinocysts if they were redeposited from up-slope (the limited data base available, however, suggests that redeposition is not associated with higher plankton contents as,for example,it was at Brora).

In the sedimentological sense facies A occurs at the top of fining upwards sequences and corresponds to a time of least redeposition; the sediments therefore have a <u>distal</u> aspect but it is entirely possible that the facies actually corresponds to shallower water (due to fan build up or progradation) and more <u>proximal</u> conditions in the palaeogeographic sense. Under such conditions more redeposition of dinocysts would be likely to occur although resedimentation of sand (related to tectonic activity?) might well be very slight. Had the ripples in the sandstone layers of facies A been wave induced (indicating a lower mixed layer environment] many more dinocysts would have been expected. The general paucity of plankton is here interpreted to be a consequence of fault controlled sedimentation which results in a close juxta position of fluvial source area and a permanently stratified marine basin separated by only a thin coastal zone (due to a steep bottom slope). Under such circumstances there would only be a very poorly developed seasonally stratifying and de-stratifying watermass in which cyst-forming dinoflagellates might have lived. Although the high proportion of phytoclast material in the shales argues for a significant fluvial influence which might have inhibited or diluted the marine plankton, there does not appear to be a direct relationship between the phytoclast and dinocyst contents (contrast units 1a and 4). The predominant plankton may have been blue-green algae which were apparently common to abundant in the stratified basinal waters.

Changes in the proportions of black wood, pteridophyte spores, bisaccate pollen and the Wu/Wd ratio in well 16/17-4A are not interpretable on the basis of the data available here. The changes are not simply correlated with lithofacies and it is possible that they reflect independently varying factors such as sea level and/or climate. The pronounced kerogen trend in unit 1 used to define the two subunits may be an expression of the mid-Volgian event observed elsewhere (e.g. Rawson & Riley, 1982) and reflect increasing anoxia in the Viking Graben bottom waters and/or a change in sea level.

A somewhat comparable study of the palynofacies of submarine fan sediments was made by Habib (1979a) on unit 4 (Late Barremian to Mid-Cenomanian) of DSDP Site 398 (see also Arthur, 1979; de Graciansky & Chenet, 1979). At this site units 4b and 4c each consist of a fining upward sequence of mudstones, claystones or shales with interbedded sandy or calcareous debris flows (mud-supported conglomerates) and turbidites. The lower part of these sequences (with the highest proportion of re-sediments) are characterised by an 'exinitic palynofacies' (see Chapter 2) and the upper part a 'micrinitic palynofacies'. The exinitic palynofacies has a high phytoclast content and a rich sporomorph assemblage dominated by pteridophyte spores of proximal aspect and the micrinitic palynofacies consists of fine carbonaceous phytoclast debris, a distal gymnosperm pollen dominated sporomorph assemblage and higher proportions of plankton. The richest organic carbon values ($\leq$3.0%) are associated with the exinitic palynofacies, within which the plankton has been heavily diluted and the other marine organic matter oxidised. It is apparent that redeposition of clastic materials is accompanied by redeposition of terrestial organic matter and subsequently higher organic carbon values. Bottom conditions remained predominantly oxidising at Site 398 except within the Cenomanian interval of unit 4a where A.O.M. is well preserved and organic carbon values reach up to 9.0%.

Unit 4 at Site 398 is quite similar to the sequence in Well 16/17-4a except that it is relatively more distal and exhibits more obvious relationships between lithofacies and palynofacies. The analogy would be far greater if the background conditions in Unit 4 had been like those which existed in the Cenomanian. Habib (1982) shows a 'tracheal palynofacies' as occurring seaward of the exinitic palynofacies; this relationship is supported by the distal sporomorph assemblage of the 'tracheal palynofacies' and its higher plankton content. The high phytoclast, low palynomorph assemblage characteristic of Well 16/17-4A has no equivalent in Habib's present scheme as it is evidently too proximal to occur in deep sea sediments, but the secondary phytoclast maxima recorded by Habib in the 'tracheal palynofacies' may partly reflect tapping and more direct redeposition of such a proximal palynofacies (by passing), since one would expect continuing offshore fractionation of the <u>exinitic</u> palynofacies to result in a <u>decreased</u> phytoclast: palynomorph ratio.

SEDIMENTOLOGICAL AND PALYNOLOGICAL OBSERVATIONS ON CORES 1-6, WELL 16/17-1

<u>Lithological description</u>

Cores 1 to 6 of Well 16/17-1 are composed of approximately 112.5' of "Late Jurassic" sediments of similar facies to those described from

Well 16/17-4. The first three cores consist of interbedded sandstone-shales and the remaining three cores pebbly clast-bearing sandstones, the whole representing an apparent fining upward sequence like those observed in Well 16/17-4A (note, however, the large gaps in core coverage). Unlike the coeval sediments in Well 16/17-4A,those in Well 16/17-1 do not fall easily in to three facies,and no facies B equivalent(with clearly defined coarse tail graded pebbly sandstone units)is present.

The sandstones and shales in cores 1 to 3 are most similar to facies A in Well 16/17-4A,but the sandstones tend to be somewhat more argillaceous, calcite cementation and bioclastic debris are much more common, and up to 20% of the cores consist of pebbly and or clast bearing, mostly structureless (?) sandstones (see Fig. 7.4). In other respects, however, (e.g. the range of sedimentary structures, etc.) the sandstones and shales are identical to those of facies A in Well 16/17-4A, with individual sandstone beds ranging up to 6 cm in thickness, and uninterrupted shale units up to 4 cm (but generally ∿1 cm). Where present, clasts in core 1 are composed almost entirely of platey shale or argillaceous sandstone fragments and are associated with horizons enriched in quartizitic granules and/or pebbles and bioclastic debris (principally bivalve fragments). In cores, 2, 4, 5 and 6 the predominant clast type consists of grey-green micaceous laminated sandstone but other sandstones occur as clasts including some reworked, white calcite-cemented quartzites, and all are accompanied by more rounded quartzitic pebbles. From core 4 to 6 there is an apparent increase in the mean maximum clast size from 4 cm to 9 cm and there is also an increase in clast density with core 6 being clast-supported. The matrix of the pebbly clast bearing sandstones in cores 4 to 6 is argillaceous and dark grey-green in colour; common pale green claystone clasts and chips are also characteristic in these cores. A dark green colour was also noted in bedded shales in core 1 between 11689-75' and 11690.5'. A belemnite fragment was recorded in a pebbly sandstone unit at the base of core 2. The sediments are petrographically very similar to those in Well 16/17-4A.

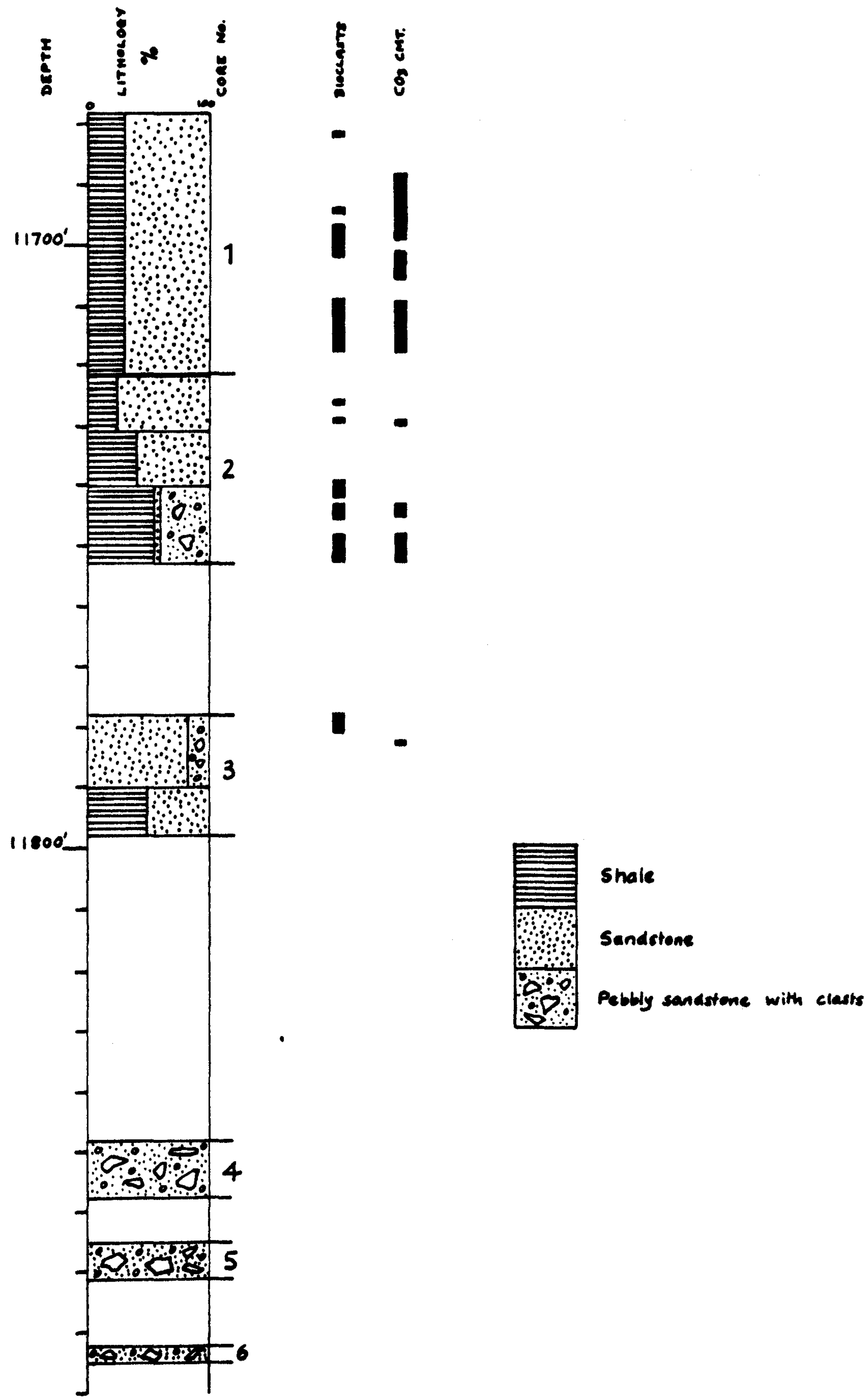


Fig. 7.4 Lithological composition of cored interval from Well 16/17-1.

The sedimentological interpretation of cores 1-6 in Well 16/17-1 is essentially the same as that given for the coeval sediments in 16/17-4A. The much greater abundance of bioclastic material in cores 1-3 compared with facies A of Well 16/17-4A suggests, however, that the upslope source area of redeposited sediment was possibly less fluvially influenced or at least experienced a lower clastic dilution of bioclastic materials. This could possibly be explained entirely in terms of the respective positions of Wells 16/17-4A and 16/17-1 relative to the apex of the fans through which they were drilled. In fans with essentially single point sources of sediment, relatively small lateral displacements at right angles to the direction of sediment transport may produce marked 'distal' shifts in lithofacies characteristics.

Palynofacies observations on Well 16/17-1

Percentage particle abundance determinations of kerogen composition were carried out on 21 slides from Well 16/17-1 and Robertson Research made six organic carbon determinations and three Rock Eval. pyrolysis assays on selected samples (see Appendices 7.6, 7.7 and 7.8). The results are summarised in Table 7.4. The general kerogen character of the samples is similar to that of the samples from 16/17-4A although the proportion of A.O.M. tends to be generally higher and the proportions of 'black wood' and palynomorphs somewhat lower. As the correlation of even lithological mega-sequences is extremely difficult in fault-scarp fan associations (see Stow et al. 1982) it would be totally unwise to attempt to correlate the fining upward sequence in 16/17-1 with one of those in 16/17-4A on the basis of the kerogen data. It is possible, however, that a correlation based on organic geochemical data might be more meaningful and on this basis it is noted that the samples from cores 1 to 3 in 16/17-1 are most similar to those from Unit 4 in 16/17-4A. No suitable samples were available to examine the palynomorph characteristics of the sediments in the cores from 16/17-1 (note the very low palynomorph values) but if the correlation between bioclasts and dinocysts observed

Table 7.4 Mean kerogen characteristics of shale samples from cores 1-3, Well 16/17-1, and core 13 Well 16/29-2X.

<u>16/17-1</u>

	Org.C	H.I.	A.O.M.	Tp.	I	Tw	Wu/Wd
Core 1	5.3(3)	460(1)	39	1.1	2	53	0.02
2			51	0.6	3	42	0.02
3	7.7(2)	340(2)	30	0.7	9	57	0.07
1-3			31	0.9	4	50	0.03

<u>16/29-2X</u>

	Org.C	H.I.	A.O.M.	Tp	I	Tw	Wu/Wd
Core 13	9.6(13)	306(4)	68	2.7	5	20	0.15

N.B. Numbers in parentheses are the number of determinations in facies A in Well 16/17-4A is a general phenomenon in this area (note that it does not apply at Brora) relatively high proportions of plankton might be expected. The kerogens from cores 1 to 3 all show A.O.M. rich in alginitic-bituminitic debris when examined under ultra violet fluorescence.

SEDIMENTOLOGICAL AND PALYNOLOGICAL OBSERVATIONS ON CORE 13, WELL 16/29-2X

Lithological description

Core 13 from Well 16/29-2X consists of 30 feet (10265-95') of Kimmeridgian shales from the Kimmeridge Clay Formation in the northern North Sea (age supplied by Robertson Research). The core is made up of a monotonous sequence of dark grey, micaceous, sometimes pyritic, organic-rich shale which is either fissile or shows a slightly subconchoidal fracture. The shale is featureless except for occasional concentrations of pyrite and the presence of carbonaceous debris and/or fish scales on some bedding surfaces. No macroscopically visible sand laminae were noted and no macrofossils were recorded. The shales are laminated in thin section.

Palynofacies observations

Percentage particle abundance determinations of kerogen composition were carried out on 18 samples from Well 16/29-2X and Robertson Research made 18 organic carbon determinations and four Rock Eval. pyrolysis assays on selected samples (see Table 7.4 and Appendices 7.6 to 7.8). The shales are all organic-rich (averaging about 10.0% Org.C) and have A.O.M. dominated kerogens with only about 25% phytoclasts. Like the other Phillip's well samples, under ultra violet fluorescence the A.O.M. in the shales is seen to contain an abundance of strongly fluorescent alginitic-bituminitic debris. The proportion of palynomorphs is too low for any meaningful quantitative observations.

When constrasted with the kerogen data for Wells 16/17-4A and 16/17-1 it is clear that kerogen assemblages in the shales from 16/29-2X are of a much more distal aspect. This is in agreement with the total absence of sandstone interbeds in 30 feet of core which could reflect sedimentation in an area remote from coarse clastic sediment sources or on an intra-basinal high out of the reach of turbidity currents. The laminated nature of the shales, the absence of benthic macrofossils, the high organic carbon and A.O.M. values and the strong fluorescence of the organic matter all point to deposition under anaerobic or anoxic bottom waters.

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Appendix 7.1 Lithofacies composition of core 16/17-4A

Core	Thickness (m)	Facies	Unit	% Shale	% Ordinary sst.*	% C.T.G. ssts.*	% massive, pebbly ± clasts
1	0.9	A	1	≤33?	≤77?	0	0
2	16.3	A	1	"	"	0	0
3	14.6	A	1	25-33	67-75	0	0
4	18.4	A	1	10	90	0	0
5	10.7	A	1	} 6-10	90	0	0
5	7.6	B	2		40	50	0
6	12.9	B	2	<5	50	30	15
6	5.4	C	3a	0	0	0	100
7	18.3	C	3a	0.1	0	0	99.9
8	14.4	C	3a	<5	22	3	70
9	1.5	C	3a	0	0	0	100
10	11.4	C	3a/3b*	0.1	11	33	56
11	13.1	C	3b	2.3	42	25	31
12	18.6	C	3b	2.6	8	33	56
13	18.3	A	4	50	50	0	0
14	16.0	A	4	40	60	0	0
14	2.8	B	5	10	60	31	0
15	10.8	B	5	40	51	9	0
16	18.2	C	6a	<5	7	2	68
17	9.5	C	6b	0.3	2	0	98
18	4.6	C	6b	0	28	0	72
19	18.3	C	6c	0	2	50	48
20	11.0	C	6c	0	5	20	75

N.B. C.T.G. units = coarse-tail graded sandstone units
'Ordinary sandstones' do not include the non-pebbly upper parts of C.T.G. units
C.T.G. units in core 10 occur only in the lower third
Figures are calculated from detailed notes and measurements

Appendix 7.2 Percentage particle abundance kerogen spectra, Well 16/17-4A
(No. counts per sample = 500)

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/4A 2	12057-10"	Shl	34.6	7.6	5.0	0	2.0	0	0	0.4	46.2	4.2
4	~12060-10"	"	27.8	6.2	2.8	0	0.2	0	0	0	57.4	5.6
6	12068-6"	"	32.6	8.8	4.2	0	0.6	0	0	0.4	44.4	9.0
7	12075-6"	"	35.2	4.2	5.0	0.2	1.0	0.4	0	0.2	47.2	6.6
8	12081'	"	35.6	4.8	3.6	0	0.6	0	0	0	48.4	7.0
9	12088'	"	29.6	4.0	2.2	0	1.2	0	0.4	0.4	54.8	7.4
10	12091'	"	31.0	2.2	4.0	0	1.0	0	0.4	0.2	54.4	6.8
11B	~12067'-12092	Mic.V.F.sst.	34.6	1.4	4.8	0	1.8	0	0.6	0	50.8	6.0
12	12102-3"	Shl	25.0	3.4	4.6	0	0.6	0	0	0.2	60.6	5.6
13	12108-3"	"	26.6	7.4	2.2	0	1.2	0	0.2	0.4	58.4	3.6
14X	~12113-4"	Arg.sst + clasts	71.2	0.2	0.2	0.2	4.2	0	1.8	0	11.2	11.0
15	12117-5"	Shl	33.6	2.8	3.4	0	0.4	0	0	0.6	52.2	7.0
16	12126-8"	"	44.4	3.4	6.4	0	1.6	0	1.0	0.4	32.4	10.4
17X	12133-10"	" (Folded)	15.6	0.6	3.8	0.2	2.8	0	0.4	0.4	69.0	7.2
18	12135-6"	"	43.0	5.2	5.8	0	1.2	0	1.2	0.2	37.4	6.0
19	12140'	"	39.0	3.4	5.0	0.2	1.8	0	0.2	0.2	44.2	6.0
20	12143-2"	"	35.4	3.0	8.2	0	1.0	0	1.0	0.2	44.4	6.8
21	12149-9"	" (silty)	57.2	2.4	6.2	0	0.8	0	0.4	0.1	25.8	7.0
22	12155-5"	Thin shls.in sst.	63.0	2.0	8.8	0.2	1.4	0	0.6	0	8.0	6.0
24	12170-4"	Shl	45.2	0.8	8.0	0	1.0	0	0.4	0.2	26.0	18.4
25	12180-4"	"	59.4	1.6	8.8	0	1.6	0	0.2	0.2	18.0	9.2
26	12183-5"	"	52.4	7.4	14.8	0	1.2	0	0.2	0	14.4	9.6
27	12192'	"	53.8	1.0	7.4	0	1.2	0	0.2	0	29.2	7.2
28	12201-4"	"	54.6	1.4	10.0	0	0.4	0	0.2	0.2	27.2	6.0
29	12206-10"	"	50.8	2.2	8.4	0	1.6	0	0.2	0.4	29.2	7.2
30	12212-2"	"	53.0	2.0	5.6	0	2.0	0	0.6	0.2	29.4	7.2
31	12220'	"	50.2	1.4	5.2	0	1.2	0	0	0	35.8	6.2
33	12224'	"	40.0	1.8	5.6	0	0.4	0	0	0.4	41.0	10.8
34	~12231-10"	"	61.0	1.0	6.6	0	1.0	0	0	0	25.6	4.8
35	~12241-2"	"	43.8	3.0	6.2	0	1.2	0	0.4	0.2	37.6	7.6
36	12247-2"	"	62.2	2.4	7.8	0	0.8	0	0	0	22.4	4.4
37	12255'	"	46.6	2.6	4.4	0	2.0	0	1.0	0.2	38.0	5.2

APPENDIX 7.2 (cont)

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/4A 38	∿12257-8"	Sst + cly clasts	14.0	2.8	0.8	0	4.4	0	11.8	0	57.8	8.4
40	∿12274-1"	Shly. sst.	68.8	2.4	5.4	0	1.4	0	0.2	0	18.0	3.8
42	∿12277-3"	V.F. sst.	1.6	0	0	0	1.8	0	0	0	68.4	28.2
43	∿12278-9"	Clay clast	24.7	0.7	3.0	0.35	3.3	0	1.7	0.35	55.3	11.0
45	∿12280-5"	F-M sst.	44.8	1.2	6.0	0	0.8	0	0.4	0	40.4	6.4
46	12281-4"	Sst + thin shls.	57.4	2.0	11.6	0	1.4	0	0.4	0	21.0	6.2
48	∿12285-3"	Cly clast	34.2	3.4	6.2	0.2	18.8	0	3.6	0	11.6	22.0
50	∿12289'	Sst. clast	8.8	0	4.4	0	1.2	0	0.2	0	81.4	4.0
52	12294-6"	Shl. + sst.	39.4	2.6	5.2	0	1.3	0	0.4	0.4	44.6	6.0
53	12298'	Shl	55.2	9.8	6.2	0	0.8	0	0.6	0	12.2	15.2
54	∿12305'	Clay clast	35.0	2.0	5.0	0	1.0	0	0.6	0.2	44.6	11.6
55	12306-6"	Shl.	56.0	7.0	9.6	0	2.6	0	0.2	0.4	9.6	14.6
57	12314-7"	"	54.6	4.2	6.2	0.2	1.4	0	0.4	0.8	18.4	13.8
60	12370'	"	58.0	4.0	8.0	0	1.6	0	0.2	0	17.0	11.2
63	∿12372'	Sst. clast	57.1	5.7	10.9	0	3.4	0	1.1	0	13.7	8.0
65	12364'	Shl. clast	17.5	0.5	4.5	0	6.0	0	0.5	0	57.0	14.0
67	12385'-6"	Sst. clast	3.0	0	2.5	0	7.5	0	0	0	81.0	6.0
71	12412-6"	Arg. sst.	41.8	2.0	5.8	0	4.0	0	0.4	0	34.4	11.6
72	12413'	Shl.	53.8	6.8	10.2	0	0.8	0	0.2	0.4	16.2	11.6
73	12423-6"	Arg. sst.	45.6	4.4	6.8	0	6.6	0.2	1.6	0	15.8	19.0
74	12426-3"	" "	37.2	2.8	8.4	0	4.6	0	2.0	0	34.8	10.2
75	12426-3"	Cly. clast	37.2	3.2	8.8	0	10.0	0.8	1.6	0	21.2	17.2
76	12248-6"	Arg. sst.	47.4	4.6	7.0	0.4	6.4	0.2	2.8	0	15.4	15.8
83	12480-6"	Shl. + sst.	43.0	0	43.5	0	0	0	0	0	9	4.5
85	∿12497'	Carb. clast	9.6	0	2.2	0	1.4	0	0	0.2	80.8	5.8
87	12502-2"	Shl. (clast?)	15.2	0	5.4	0	2.0	0	0.2	0	54.2	23.0
92	12529-2"	Shl.	53.2	5.4	11.6	0	0.8	0	0.2	0	21.4	7.4
93	∿12537'	"	60.2	4.6	11.2	0	2.0	0	0	0.4	15.8	5.8
94	12546-8"	"	53.2	4.4	14.8	0	0.6	0	0.6	0.4	21.0	5.0
95	∿12550-5"	F-M sst + arg.clast	51.2	6.4	11.0	0	0.8	0	0.2	0	25.4	5.0
96	∿12552'	Sst. + arg. clasts	2.8	0.4	1.6	0	6.8	0	0.8	0	85.6	2.0
97X	∿12576-1"	VF-F sst.	1.4	0	0.6	0	0.8	0	0.2	0	45.4	0.8

APPENDIX 7.2 (cont)

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/4A 92Z	∿12576-1"	Sst. clast	14.3	0	54.0	0	7.9	0	0	0	13.5	10.3
98	∿12585-8"	Shl.	54.2	7.4	14.4	0	1.2	0	0	0	15.2	7.6
99	12590-10"	"	50.4	10.8	14.0	0	1.2	0	0.2	0.2	14.0	9.2
100	∿12593'	Shl. clast	31.3	0.7	8.7	0	1.0	0	0	0	50.3	8.3
101	∿12598-9"	Clay	31.0	0	24.0	0	11.0	0	0	0	16.0	18.0
102	∿12598-9"	Clayst.	46.2	5.4	11.4	0	1.6	0	0.2	0.2	24.0	11.0
103	12599-4"	Shl/cly (clast?)	35.4	1.4	4.8	0	9.4	0	1.2	0	21.2	26.6
104	∿12604-5"	Shl.	40.0	15.0	20.4	0	0.6	0	0.2	0.2	18.4	5.2
105	12611-4"	"	40.0	12.2	12.0	0	1.4	0	0	0	26.8	7.2
106	12613-3"	"	42.4	16.8	6.8	0	1.0	0	0.2	0	19.2	13.6
108	∿12634'	Arg. sst.	65.6	8.4	13.6	0	0.8	0	0	0	7.2	4.4
109	∿12634'	Mdst. (pyrite)	77.8	1.2	7.6	0	5.0	0	0	0	2.6	5.8
110	12637-10"	Shl.	37.6	17.6	17.0	0	1.2	0	0	0	13.0	13.6
111	12642'	"	42.4	9.2	17.6	0	1.6	0	0.6	0.2	22.8	5.6
112	12651'	"	28.0	3.6	27.2	0	0.8	0	0	0	23.0	7.4
113	12661-10"	"	47.6	15.8	16.4	0	0.6	0	0.2	0	12.6	6.8
114	12670'	"	41.0	17.0	16.6	0	1.0	0	0	0	11.8	12.6
115	12672-10"	"	50.0	24.6	11.4	0	0.6	0.2	0.2	0.2	6.6	6.2
116	12681'	"	41.0	16.6	14.6	0	2.2	0.2	0	0	19.6	5.8
117	12689-6"	"	45.6	20.4	13.8	0	1.4	0	0.2	0.4	11.6	6.6
118	12694'	"	36.8	30.8	16.2	0.2	0.6	0.2	0	0.2	7.6	7.4
120	12702'	"	39.6	14.0	13.6	0	0.8	0	0	0	28.2	3.8
121	∿12705-4"	Arg. sst.	46.8	2.0	21.2	0	1.2	0	0.2	0	24.8	3.8
122	12706-10"	Shl.	42.4	18.0	18.2	0	0.6	0	0	0.2	18.0	2.6
123	12716-6"	"	48.6	8.6	12.0	0	0.4	0	0.2	0	24.6	5.6
125	12724-3"	"	38.6	10.8	14.4	0	1.0	0	0	0	29.0	6.2
126	12731-2"	"	39.4	10.4	15.6	0	1.8	0	0.2	0.2	28.0	4.2
127	12740-6"	"	53.4	5.2	15.2	0	0.2	0	0.2	0	20.0	5.8
128	12743-7"	"	61.2	5.2	11.6	0	0.8	0	0	0	14.2	7.0
129	12743-8"	Arg. C. sst.	9.8	0.2	1.2	0.4	45.4	0	6.0	0	7.0	30.0
132	∿12749-7"	Arg. sst.	32.6	0.4	1.6	0.8	16.2	0	7.4	0	15.0	26.0
133	12754-10"	Shl.	56.2	10.6	14.8	0	0.8	0	0.2	0	9.6	7.6

APPENDIX 7.2 (cont)

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/4A 134	12756'	Shl.	50.8	4.8	16.2	0	0.8	0	0.4	0.2	23.2	3.6
135	12758-8"	"	52.2	7.0	12.4	0	0.6	0	0	0.2	20.8	6.8
136	12767'	"	43.0	11.2	15.4	0	1.4	0	0.6	0	20.6	7.8
137	12775-5"	"	51.0	8.6	15.8	0	1.4	0.2	0.4	0	13.6	9.0
138	12782-10"	"	44.0	10.6	12.2	0	1.2	0.4	0.6	0.2	23.2	7.6
140	12785-9"	"	45.6	14.2	17.8	0	2.6	0.2	1.0	0.2	13.2	7.2
142	12789-3"	"	43.0	9.0	9.0	0	0.6	0	0	0	34.2	4.2
143	12846'	Clayst.	39.0	8.3	16.3	0	2.0	0	1.0	0.3	26.0	7.0
144	12847-6"	Sst. + shl.	43.0	9.4	18.2	0	1.2	0	0.2	0	23.6	4.4
145	12852-10"	Shl.	41.2	9.8	15.0	0	0.6	0	0	0	29.8	3.6
146	12856-5"	"	40.8	11.8	22.0	0	2.6	0	0.4	0	18.0	4.4
147	12858-6"	Shl. (C.G.U.)	48.6	10.8	12.6	0	3.6	0	1.2	0.2	16.4	16.6
148	12860-4"	M.sst.clast	22.4	1.6	6.2	0	2.4	0	0.2	0	64.2	3.0
149	12866'	Shl. (C.G.U.)	49.0	3.4	10.2	0	4.4	0	0.8	0.2	24.0	8.0
151	12877-4"	Shl. clast?	12.0	0.2	1.2	0	21.2	0	6.4	0	36.2	22.8
152	12907-5"	Shl.	56.2	8.0	21.2	0.2	1.8	0	0	0	9.8	2.8
153	12924'	Arg. clast	29.8	2.4	4.2	0	10.6	0	4.2	0	26.2	22.6
157	12977-5"	" "	42.0	0.2	0.6	0	9.6	0	3.8	0	30.4	13.4
159	12966'	" "	21.6	0.2	2.6	0	9.2	0	5.6	0	31.2	29.6
161	13004'	Siltst. clast	33.5	2.3	5.8	0	0.8	0	0	0	54.6	3.5

APPENDIX 7.3 Organic Carbon and Hydrogen Indices, Well 16/17-4A

(for sample depths and lithologies see Appendix 7.2)

Sample	Org.C	H.I.	Sample	Org.C	H.I.
6	9.8	863	94	12.5	-
7	10.3	-	95	8.0	-
8	10.5	425	98	8.6	225
9	11.2	-	104	12.7	253
10	9.0	917	106	29.8	-
12	9.4	-	110	4.6	-
13	9.0	345	111	7.2	-
14X	2.8	-	112	3.4	-
15	12.9	-	113	6.0	124
16	7.2	776	115	7.4	-
18	14.8	-	116	3.1	109
19	11.0	-	117	6.1	-
20	9.4	307	120	4.3	135
21	8.2	-	121	3.6	-
22	3.9	187	122	2.9	-
25	12.1	262	123	6.9	-
27	8.4	-	125	11.1	271
29	11.8	596	127	2.7	-
30	13.3	-	128	7.6	402
33	12.5	364	132	4.3	-
35	9.7	240	133	12.4	-
40	12.4	245	135	10.2	299
52	10.0	365	136	12.4	-
53	5.1	-	140	8.6	207
55	11.8	199			
87	6.2	338			
93	13.9	380			

APPENDIX 7.4 Statistical summary of main kerogen parameters in shale samples from Well 16/17-4A

Unit One (Facies A), 12055' to 12255', Samples P/4A 2-37

		A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
Subunit 1a (Samples 2-20; 12055'-12143.2')	No.	15	15	15	15	15	12	6
	$\bar{x}$	50.1	1.5	4.4	37	0.14	10.4	606
	S.D.	9.1	0.9	1.6	7.7	0.07	1.9	252

		A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
Subunit 1b (Samples 21-37; 12149.75'-12255')	No.	15	15	15	15	15	8	5
	$\bar{x}$	27.8	1.5	7.5	55.1	0.04	10	330
	S.D.	7.6	0.7	2.5	6.8	0.03	2.9	145

Unit Two (Facies B), 12255' to 12322.5', Samples P/4A 38-57

	A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
No.	6	6	6	6	6	4	3
$\bar{x}$	20.6	1.9	7.4	59.9	0.09	9.8	270
S.D.	11.4	0.4	2.4	9.0	0.05	2.9	70

Unit Three (Facies C), 12322.5'-12636', Samples P/4A 60-106

		A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
Subunit 3a (Samples 60-94; 12370'-12546.7')	No.	6	6	6	6	6	2	1
	$\bar{x}$	16.7	1.2	16.6	57.8	0.09	13.2	380
	S.D.	4.1	0.6	12.2	7	0.02	0.7	

		A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
Subunit 3b (Samples 98-106; 12585.75'-12613.25')	No.	5	5	5	5	5	3	2
	$\bar{x}$	18.7	1.2	13.5	57.8	0.29	17	239
	S.D.	4.5	0.2	4.4	3.7	0.1	9.2	

Unit Four (Facies A), 12636' to 12748.5', Samples P/4A 110-128

	A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
No.	16	16	16	16	11* 5+	13	5
$\bar{x}$	18.2	1.1	15.7	57.6	0.41 0.18	5.7	208
S.D.	7.1	0.6	3.6	9.6		2.3	113

* Samples 110-112, + Samples 123-128

Unit Five (Facies B), 12748-5' to 12790', Samples P/4A 133-142

	A	Tp	I	Tw	Wu/Wd	Org.C	H.I.
No.	8	8	8	8	8	4	2
$\bar{x}$	19.8	1.7	14.2	57.8	0.2	10.9	253
S.D.	7.2	1.0	2.6	4.4	0.07	1.6	46

Unit Six (Facies C), 12846' to 13053', Samples P/4A 144-152

	A	Tp	I	Tw	Wu/Wd	Org. C	H.I.
No.	6	6	6	6	6	0	0
$\bar{x}$	20.3	2.8	16.5	54.L	0.2		
S.D.	6.4	1.7	4.3	3.0	0.07		

APPENDIX 7.5 Percentage palynomorph composition of samples from Well 16/17-4A

(key to codes as elsewhere, see page

Depth	Approx. lith.*	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	Counts
12058'	F-M sst.	1.9	1.3	2.1	8.3	0.2	0.3	0.3	0.6	3.4	4.6	0.6	10.3	0	7.1	59.3	678
12095'	Mic. sst.	1.0	0.5	0.5	5.0	0	0	0	0	2.5	4.4	0.5	15.3	0	7.9	62.5	203
12113'	P/4A-14	0.3	0.3	0	0	0.6	0	0	0	1.3	5.2	0	24.2	0	19.6	48.1	306
12212'	P/4A-30	0	0	0	0	0	0	0	0	0.6	7.9	0.6	15.8	0.6	17.0	57.0	165
12285'	P/4A-48	0	0	0	0	0	0	0	0	0	4.0	0	18.4	0	17.3	60.3	375
12295'	sst/shl	0.6	0.9	0	0.9	0	0.9	0.3	0	1.6	13.9	1.3	16.7	0.3	15.8	46.1	317
12440'	Arg.pebbly sst.	0	0	0	0	0	0.6	0	0.6	0	5.9	0	5.0	0.3	5.6	81.9	338
12502'	Pebbly sst.	0	0	0	0	0	0	0	0	0	66.9	0	0	0	0.3	32.8	299
12577'	Peblly sst.	0	0.2	0	0.2	0	0	0	0.2	0.7	63.7	0.2	2.0	0	3.7	28.9	410
12650'	sst/shl	0.6	5.7	0	0.9	0	0	0	1.0	0	68.7	0	1.9	0	6.3	14.3	315
12702'	sst/shl	1.2	1.8	0	23.4	0	0	0	0.6	1.8	30.3	0.4	9.0	0.4	12.6	18.7	509
12743'	P/4A-129	0	0	0	0	0	0	0	0.3	0	4.3	0	32.3	0	9.5	53.8	368
12786'	P/4A-140	0	0	0	0	0	0.4	0	0	1.4	12.6	0	23.7	0.4	16.2	45.4	278
12877'	Pebbly sst	0	0.5	0	0.2	0	0	0	0	0	13.6	0.2	29.5	0	20.5	35.4	404

* Apart from samples 14, 30, 48, 129 and 140 the remainder are Robertson Research samples whose exact lithology is unknown. Samples from pebbly sandstone units may be, or include, shale or argillaceous clasts

N.B. A bisaccate dominated assemblage (like the samples at 12502', 12577' and 12650') was also noted in a sample from 12858.5' (a shale lying immediately above a C.T.G. sandstone unit and corresponding to sample P/4A-147).

Appendix 7.6 Percentage particle abundance kerogen spectra for samples from Wells 16/17-1 and 16/29-2X (based on 500 counts per sample)

16/17-1

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/1-2	11679'	Shl.	54.0	0.4	2.4	0	0.6	0	0	0.2	37.2	5.2
3	11681'	"	55.4	0	1.2	0	0.4	0	0	0	39.8	3.2
6	11688-7"	Arg. sst.	47.2	0.6	1.4	0.2	0.4	0	0	0.2	45.6	4.4
7	∿11690'	Grn. shl.	78.6	1.8	2.0	0	1.8	0.2	0	0.4	8.6	6.6
8	11692-10"	Shl.	54.2	1.8	1.4	0	2.8	0	0	0.4	32.2	7.2
12	11698-2"	Shl. clast.	32.2	0.4	1.4	0	1.4	0.4	0	0.8	60.4	3.0
14	11704'	Shl.	37.6	0.4	3.8	0	1.2	0	0	0	51.8	5.2
17	11708-5"	"	45.2	0.6	2.0	0	0.8	0	0	0	46.2	5.0
18	11715'	"	47.0	0.2	1.8	0	0.4	0	0	0.4	47.0	3.2
19	11719-11"	"	43.6	1.6	1.2	0	0.4	0	0	0.2	50.2	2.8
20	11722-4"	"	47.8	0.6	1.8	0.2	0.6	0	0.2	0.4	47.2	1.2
21	11725-4"	"	46.4	0.6	4.2	0	0.8	0	0	0.4	44.4	3.2
23	1173-6"	"	31.4	0.6	1.6	0.2	0.2	0	0	0	64.2	1.8
24	11737-7"	"	40.6	1.4	2.6	0	0.4	0	0	0	52.2	2.8
25	∿11739-7"	Arg. sst.	66.0	0.3	1.0	0	0.3	0	0	0	31.0	1.3
26	∿11741-6"	" "	34.6	0.6	0.9	0	0.3	0	0	0	60.6	3.1
27	11747-8"	Shl.	39.5	1.3	5.7	0.2	0.5	0.2	0	0	49.0	3.7
29	11778'	"	56.8	2.4	9.6	0	0	0	0	0	29.2	2.0
31	11790-4"	"	51.5	3.3	6.0	0	1.0	0	0	0.5	36.3	1.3
32	11798'	"	52.3	5.8	11.3	0.2	0.8	0.2	0.2	0	24.8	2.7
33	11851-6"	Arg. sst.	64.8	6.8	9.0	1.0	2.5	0	0	0.5	10.0	5.5

16/29-2X

P/2-1	10265-67.5'(T)	Shl.	15.0	6.0	6.2	0	2.8	0	1.0	0.6	65.4	2.8
2	10265-67.5'(B)	"	14.5	3.0	3.5	0.2	2.2	0	0.3	0.3	69.8	6.2
3	10267.5-70'(T)	"	17.2	4.8	9.0	0	4.0	0	1.4	1.6	57.8	4.2
4	10267.5-70'(B)	"	16.0	0.6	4.6	0.4	2.8	0.2	0.4	0	71.6	3.4
5	10270-73'(T)	"	17.0	2.2	3.2	0	2.0	0.2	0.6	0.2	69.8	4.8

Appendix 7.6 (cont)

16/29-2X

Sample	Depth	Lithology	Wd	Wu	I	C	UP	P	M	F	A	Pd
P/2-6	10270-73'(M)	Shl.	16.2	2.4	6.6	0.2	1.4	0	0	0.2	68.4	4.6
7	∿10274'	"	19.0	2.0	5.4	0.4	2.4	0	0.4	1.2	64.8	4.4
8	∿10275'	"	17.6	1.6	3.2	0	3.0	0.2	0	0.6	67.0	6.8
9	∿10278'	"	18.2	0.6	2.8	0	1.4	0	0.4	0.2	70.4	6.0
10	10279-82'(M)	"	14.2	1.6	4.6	0	1.8	0	0.2	0.6	72.8	4.2
11	10279-82'(B)	"	17.0	1.6	5.6	0.2	1.4	0.2	0.2	0.2	69.6	4.0
12	10282-85'(M)	"	22.4	1.6	3.2	0	2.0	0	0.2	0	64.2	6.4
13	10285-87-5'(M)	"	21.6	0.6	7.2	0	0.4	0	0	1.0	65.8	3.4
14	10285-87.5'(B)	"	24.2	2.3	4.3	0	1.7	0	0.2	0.7	64.0	2.7
15	10287.5-90.5'(M)	"	17.6	2.2	6.2	0	3.6	0	0.2	0.6	66.4	3.2
16	10287.5-90.5'	"	13.2	2.2	4.6	0.4	4.4	0.4	0.2	0.2	73.2	1.2
17	10290.5-93'(M)	"	18.2	1.2	4.6	0	1.6	0	0	1.0	70.4	3.0
18	10293-95'	"	15.8	1.0	3.6	0	2.0	0	0	0.2	75.2	2.2

N.B. B, M and T indicate the base, middle and top of the sample interval; exact positions unknown due to the broken, rubbly nature of the core

Appendix 7.7 Statistical summary of main kerogen parameters of shale samples from Wells 16/17-1 and 16/29-2X

16/17-1

	A.O.M.		Tp		I		Tw		Wu/Wd		
	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	$\bar{x}$	S.D.	No Samples
Core 1	39.1	13.1	1.1	0.8	2.0	0.8	52.8	11.9	0.02	0.01	8
2	51.4	6.9	0.6	0.2	3.2	1.6	42.0	5.8	0.02	0.01	5
3	30.1	4.7	0.7	0.5	9.0	2.2	57.4	1.9	0.07	0.03	3
1-3	41.3	12.8	0.9	0.7	3.7	3.0	50.3	10.8	0.03	0.03	16

16/29-2X

Core 13	68.1	4.1	2.7	1.2	4.9	1.6	19.6	2.8	0.15	0.11	18

Appendix 7.8 Organic carbon and Hydrogen Indices for samples from Wells 16/17-1 and 16/29-2X (for sample depths and lithologies see Appendix 7.6)

16/17-1

Sample	Org.C	H.I.
P/1-12	7.1%	-
17	6.3	460
18	5.0	-
19	4.8	-
29	8.4	367
31	6.9	313

16/29-2X

P/2-2	8.5%	236
3	8.0	-
4	9.8	-
7	8.5	-
8	9.6	337
9	14.7	-
11	8.7	-
12	10.0	332
13	10.4	-
14	9.2	-
15	8.3	-
17	10.5	-
18	8.7	320

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CHAPTER EIGHT

Conclusions

In this chapter I would like to concentrate mainly on the comparison of the different sediment sequences in terms of their palynofacies characteristics. Summaries of the general findings on each of the sequences are given at the end of the relevant chapters and are not repeated here.

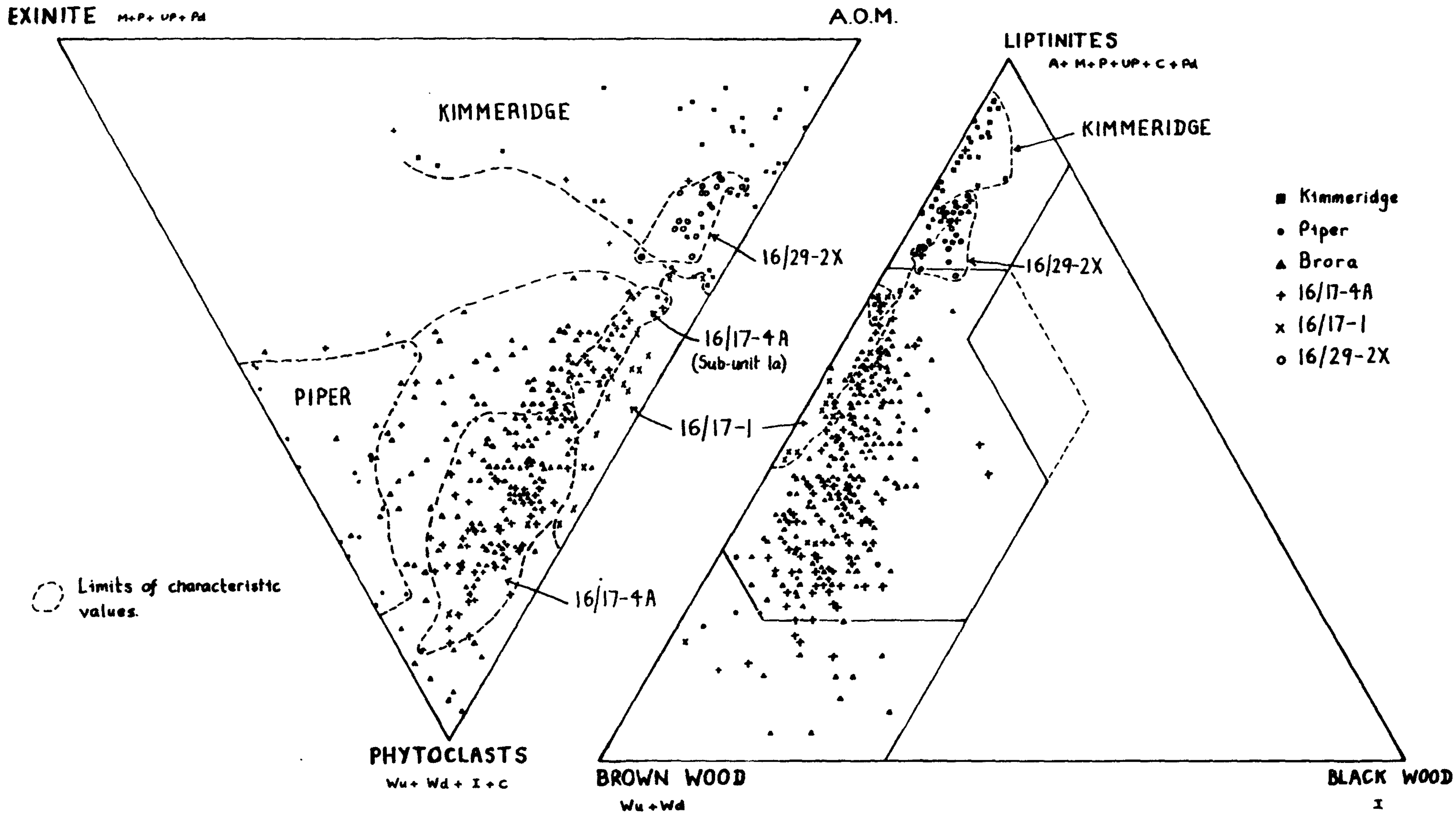
Palynomorph trends

Because of the low number of oxidised palynological preparations made available for this study (and the generally high A.O.M. contents), only limited comments can be made here on the palaeoenvironmental utilisation of palynomorph trends. I believe, however, that this study has demonstrated the morphotype sorting of plankton and sporomorphs both within shales and between shales, silts and sandstones. I have also shown clearly that, except where they are redeposited, dinoflagellate cysts are notably less abundant in permanent basinal settings and during meromictic episodes in shelf settings. The value of analysing palynomorph trends has also been more conclusively demonstrated elsewhere (Tyson, 1984b and unpublished observations on Leg 77 D.S.D.P.).

Kerogen trends

To compare all the shale data from the different sections I have utilised two kinds of triangular diagram to plot the data (Fig. 8.1). The liptinite-brown wood-black wood plot (Fig. 8.1A) used for determining the kerogen prefix (see Fig.2.3), indicates most of the sediments are 'sapro-humic' apart from the Type Kimmeridge Clay and Phillips 16/29-2x shales. There is considerable overlap on this diagram but the Kimmeridge, 16/29-2x and 16/17-1 data do plot in relatively distinct fields. The plot also demonstrates the general correlation between the black wood and brown wood components. The A.O.M.-phytoclast-exinite (A.P.E.) plot (Fig. 8.1B) shows far greater discriminating power between the different data sets. With the exception of the sediment-ologically very similar Brora and Phillips 16/17 data, all the different sections plot rather distinctly. The diagram also shows subunit la in well 16/17-4A as being markedly different from the other subunits.

Fig. 8.1 Triangular plots of all shale kerogen data (percentage particle abundances)



The logic behind the descriminating power of the A.P.E. plot is given in Fig. 8.2, which also shows the palaeoenvironmental fields suggested by this study. This diagram is based on fundamental logic and should be expected to have general applicability for epeiric sediments, but the fields will not, of course, be identical for all data sets. It should be noted that the detailed examination of the palynomorph component would allow further resolution and refinement of the palaeo-environmental fields and hence discriminate between palaeoenvironmentally non-specific kerogen data.

Stratigraphic and/or regional plots of kerogen data on an A.P.E. diagram should result in the ability to delineate

(1) The palaeogeographic polarity (onshore-offshore trends)

(2) The relative proximity to sources of fluvial input (and/or sediment redeposition)

(3) Transgressive-regressive facies trends (in perhaps lithologically 'uniform' sequences)

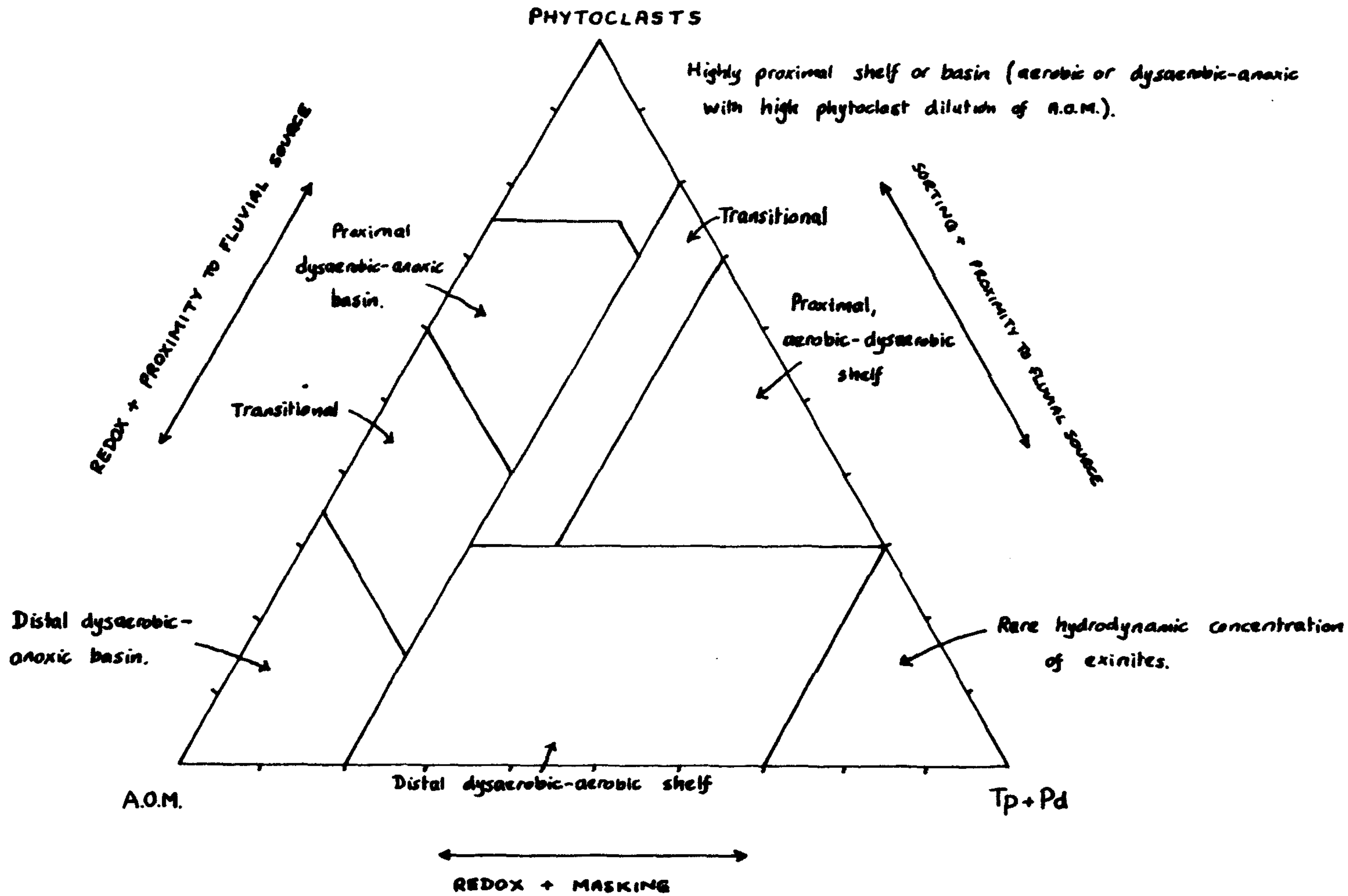
(4) The hydrocarbon source-rock potential (with u.v. analysis of (A.O.M.)

(5) Redox changes within the basin

and (6) Help to define the basin-shelf (bottom water-mixed layer) transition.

It will be noted that the highest organic carbon values tend to plot in the A.O.M. corner of the A.P.E. diagram. High T.O.C. values may, of course, also occur along the A-P axis, where high preservation potential gives high marine organic contents, and the supply of phytoclast material boosts this value even higher. To improve the comparison between the various sections some allowance has to be made for the varying organic richness of the sediment. I have attempted this by making the rather crude assumption that the amount of marine organic carbon (M.O.C.) present in these shales is represented by the T.O.C. multiplied by the percentage particle abundance of non-collinitic A.O.M.

Fig. 8.2 Distribution of palaeoenvironmental fields on the A.O.M.-Phytoclast-Exinite (A.P.E.) plot.



This should yield a minimum estimate of M.O.C. and is sufficient for comparative purposes. I have also assumed that the amount of non-marine organic carbon (N.M.O.C.) is equivalent to the T.O.C. multiplied by the proportion of non-A.O.M. material.

The resulting ranking of the various sections is shown in Fig. 8.3. Allowing for the general overlap of values for Brora and the 16/17 wells, one can see that the M.O.C. trend corresponds with the proximal-distal trend indicated by the A.P.E. plot alone. It will be noted that high (>3.0%) values of M.O.C. are limited to the most basinal and/or distal facies.

Ranking of the N.M.O.C. values for the different sections do not produce simply the reverse of the M.O.C. trend. Also, although one would expect the N.M.O.C. trend to correlate largely with proximity to fluvial sources, it does not ideally correspond with the general palaeoenvironmental proximal-distal ranking of the sections. Unlike what one might expect, subunit 1a of 16/17-4 has more N.M.O.C. than 16/17-1, and with 16/29-2 has more N.M.O.C. than Piper. The major N.M.O.C. contents (= supply?) are all clearly associated with the fault scarp-fan associations bordering the west and east Shetland Platform areas. By comparison the Piper (Lower Shale) appears to be an open shelf facies further removed from fluvial inputs of N.M.O.C., or supplied by a smaller (less vegetated?) hinterland.

An interesting question is posed by the M.O.C. data and ranking. Apart from Piper all the other data come from shales which are all sedimentologically very similar, although the Brora shales are probably a dysaerobic-anaerobic facies while the Phillips shales and Kimmeridge oil shale are all at least dysaerobic and probably anaerobic-anoxic facies. Elsewhere (Tyson, 1984a) I have reviewed information which suggests that the preservation of laminations (i.e. the principle sedimentological criterion for determining palaeo-oxygenation) appears to be more sensitive to redox changes than are geochemical characteristics.

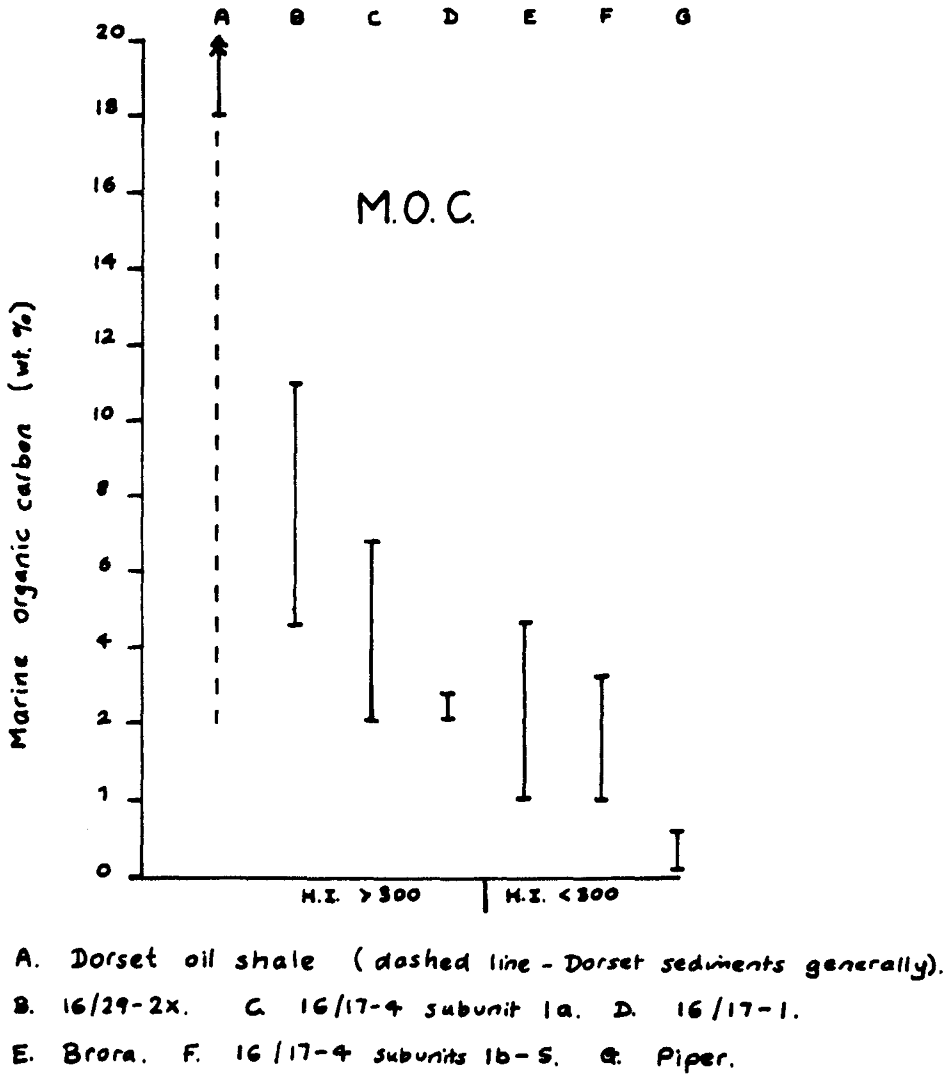
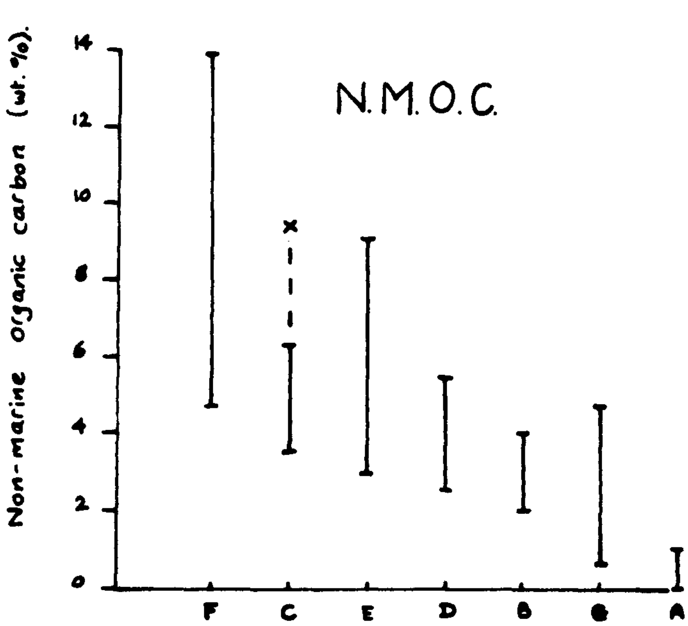


Fig. 8.3 Comparison of Upper Jurassic sections according to estimated contents of marine (MOC) and non-marine (NMOC) organic carbon.



Accepting also that the main site of M.O.C. preservation is at the sediment-water interface (especially in 'shelf' settings), and that A.O.M. is essentially autochthonous, should there be any significant changes in preservation potential between these sediments? An increasing degree of preservation (perhaps related to temporal variability in redox conditions?) could explain the M.O.C. differences between the Phillips shales and Kimmeridge sediments, but the differences are considerable.

A distinct possibility exists that there were original differences in M.O.C. input (i.e. productivity), such that beyond a base level of about 2-3% M.O.C., (associated with increased preservation resulting from the attainment of at least dysaerobic conditions), further variation is productivity related. For the Dorset sediments it is possible that a supply of semi-amorphous marine macrophyte debris might be responsible for boosting the T.O.C. values, although in calculating the M.O.C. range I excluded the problematic fibrous material (F.M.) component.

As noted before, the high M.O.C. values occur in the more distal or basinal facies. This is in fact something of a generalisation for subunit 1a of 16/17-4A. In this well Unit 1 is clearly indicative of a distal facies shift, but neither the lithologically very similar subunit 1b or Unit 4 show M.O.C. values like those of subunit 1a. This indicates secular, stratigraphic differences are influencing the values. However, returning to the general point, what does this correlation of distal facies and high M.O.C. values signify? The possibilities are:-

(a) Increased M.O.C. preservation is associated with distal, deeper and more stable oxygen deficient environments

(b) The more distal facies were deposited below waters of higher productivity. This is more difficult to explain than if the marginal sediments showed higher M.O.C. values. Perhaps it suggests basin centre eutrophication with nutrient-rich bottom waters being at least periodically accessible to the plankton (see also Chapters 2 and 5).

(c) The M.O.C. trend reflects a gradient of decreasing clastic dilution in a proximal-distal direction. Weight percent T.O.C. values provide only a relative measure of concentration; what is required is the absolute amount of T.O.C./M.O.C. independent of clastic dilution effects, i.e. the T.O.C. accumulation rate. Unfortunately I do not at present have the necessary data to calculate this parameter.

In general, it is probable that the M.O.C. trend between the sections is mainly a product of gradients in preservation, clastic dilution and possibly also productivity as well. There is not an especially good correlation between the trends of M.O.C. and hydrogen indices, although M.O.C. values of over about 2-3% seem to be associated with H.I. values over 300. The restricted nature of this correlation reflects the fact that:-

(a) exinitic material was not included in the M.O.C. calculation (as most of the exinitic material is not marine derived, although some allowance was made for Piper).

(b) differing degradation states and original composition of the A.O.M. were not taken into account (except collinite was excluded)

(c) variations in the geochemical character of the phytoclast material were also not taken into account

(d) the kerogen counts were not formally size-classed.

Comments

I would now like to return to a point raised in Chapter One. Near the beginning of this project I wished to see if palynofacies could discriminate between the Piper and/or Allt na Cùile and Lothbeg sandstones which some geologists have considered similar facies. A problem with this is that there is very little argillaceous material to sample in the Allt na Cùile sandstone body (as defined here) and its lateral relationships with both more argillaceous and more diagnostic facies are obscure. The Allt na Cùile sandstone could be a shallow marine (almost deltaic) body in fault contact with the deep water shales and sandstones adjacent to it. Clays from the Allt na Cùile body yield only poor information and cannot even conclusively prove a marine origin. However, the contrast between the Piper and Lothbeg sandstones (as deduced from the shales within which they are intercalated) is marked. The palynofacies show that the Piper Formation is part of a dominantly aerobic, fairly proximal shelf sequence, while the Lothbeg sandstone occurs within dysaerobic-anaerobic, basinal (but proximal) shales. On the basis of palynofacies alone it would have been possible to have speculated on the likely geometry and orientation of these sandstones if one were called upon to do so.

However, this thesis is about the multi-disciplinary approach, and palynofacies should be used to enhance sedimentological interpretations and not as a technique which can be applied without sedimentological control. Palynofacies, however, does provide a method of assessing source rock characteristics which can be used with a degree of independence and in the absence of geochemical data. It is ideally suited to integrate sedimentological, palaeoecological and geochemical data into a cohesive story and it is this which should be regarded as its main strength and hope for the future. Clearly the more quantitative the treatment the more meaningful the results that are to be had. Counting kerogens in palynofacies preparations may be tedious but I hope I have demonstrated its considerable value.

Regional setting of the Kimmeridge Clay Formation and its correlatives

Much of my appreciation of this topic is derived from regional stratigraphic and palaeogeographic syntheses of the Boreal Late Jurassic-earliest Cretaceous which will be presented elsewhere. As a conclusion to this thesis I would, however, like to give a brief summary of the Late Jurassic palaeoenvironmental changes.

The Kimmeridge Clay Formation was deposited within a broad epeiric sea at a time characterised by considerable tectonic activity in the

North Sea and its environs. This tectonic activity undoubtedly resulted in a palaeobathymetrically diverse basin characterised by stable shallow water platforms, broad, downwarped deep shelf areas (e.g. southern England and the western southern North Sea basins), and rapidly subsiding deep water grabens (especially the northern North Sea). This differing tectonic regime combined with the Late Jurassic-earliest Cretaceous eustatic trends to produce regional variations in sedimentary facies. These variations are rather subdued to the extent that the dominant preserved facies is shelf or basinal mudstone-shales

Review of regional and stratigraphic data indicate that some of the boreal deep water graben areas became prone to oxygen deficiency and the deposition of organic-rich mudrocks as early as the late Oxfordian, when most shelf basins were still shallow and aerobic. The British shelf basins did not begin to 'deepen' until after the _cymodoce_ zone. There appears to have been a widespread deterioration in the level of oxygenation in the shelf and basin waters (resulting from increased or more stable stratification) during the _eudoxus_ zone. In this study this trend can be detected at both Kimmeridge and Brora. Further intensification of oxygen depletion appears to have taken place during the mid-Volgian, as seen for example in the Type Kimmeridge Clay in Dorset. It is clear from the cyclic facies development in the shelf basins that the eustatic and tectonic conditions made these basins prone to deoxygenation but that the course of this process was controlled by climatic variables.

In the Dorset basin, and in many other shelf areas as well, the mid-Volgian event was prematurely terminated by the advent of the latest Jurassic-earliest Cretaceous Portlandian-Purbeckian regression. This at least partly corresponds with diminished subsidence of the shelf areas, although graben subsidence may have increased. The latter is suggested by the way in which the Volgian-Ryazanian apparently oversteps some of the previous Oxfordian-Kimmeridgian structural highs in the Norwegian North Sea. Accepting the latter, the increasing restriction of the overall marine area at the Jurassic-Cretaceous boundary probably lead to even more severe oxygen depletion culminating in deposition of the 'hot shale' facies in many graben areas. The general regressive setting at this time, however, will have resulted in progradation of shelf wedges into the graben basins and in some areas the consequent associated shallowing and/or reoxygenation and/or clastic dilution will have prevented formation of typical 'hot shales'.

At some point, approximately in the mid to late Ryazanian,the deep, stagnant graben areas were flushed out and hot shales were replaced by frequently condensed, calcareous, aerobic sediments. This termination of the organic-rich mudrock deposition is synchronous throughout the North Sea, Greenland and West Siberian basin. It is an event of major magnitude and significance. Changing climate may have had something to do with it (in general terms from the arid Purbeckian phase to the more humid Wealden one). However, the scale of the event and its sharpness lead me to suspect a partly tectonic answer. Perhaps a crucial deepwater connection was established which allowed a surge of cooler waters to flow down through the graben system. Perhaps such an event combined with general climatic cooling which passed through a critical point where the pre-requisites for meromixis in the grabens was irrevocably destroyed. To solve this problem, if it can be solved at all, will require more data from the North Sea - including coring of the appropriate intervals. Unfortunately, if understandably, oil companies rarely invest in coring shale sequences.

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