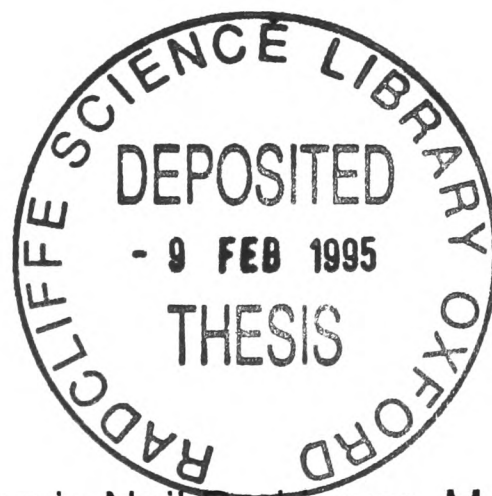


A Thesis Presented for the Degree of Doctor of Philosophy

The Sequence Stratigraphy of the Lower Jurassic of Western Europe

Volume 1: Text



Dennis Neil Parkinson, M.A.
Wolfson College & Department of Earth Sciences,
University of Oxford

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ABSTRACT

Dennis Neil Parkinson
Wolfson College

Doctor of Philosophy
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The Sequence Stratigraphy of the Lower Jurassic of Western Europe

The Lower Jurassic stratigraphy of seven contrasting areas in western Europe is compared in order to seek evidence for a pan-Western European stratigraphic forcing mechanism. Sequence-stratigraphic models are discussed and emphasis is placed upon the differing response of sedimentary systems in "accommodation space-dominated" and "supply-dominated" settings.

Spectral gamma-ray data from clastic successions in the Wessex and Cleveland basins (England) are used to elucidate vertical trends. A proximal-distal model for control of Th/K ratios is advanced. Stage-frequency regressive-transgressive cycles in the two areas are shown to be closely correlative.

Sedimentological logs and spectral gamma-ray data are presented for the carbonate ramp into turbidite sequence of Peniche (Portugal) and for a new exposure of the Pliensbachian-Toarcian in southern Germany. Systematic variation in clay mineralogy across Europe is suggested.

Cycles in the Lower Jurassic of the North Viking Graben (Norwegian North Sea) are examined using wireline log correlation and the stratigraphic evolution of the Tethyan Rift in the Western and Southern Alps is reviewed. X-ray diffraction studies of the Pliensbachian-Toarcian interval in the Southern Alps are presented in order to elucidate sediment supply to the pelagic realm.

The cycles observed in the English sections appear to be manifest widely across western Europe in a variety of tectonic and sedimentary settings. Sharp basinward facies shifts (candidate sequence boundaries) do not appear to be synchronous between basins.

$^{87}\text{Sr}/^{86}\text{Sr}$ analysis of belemnites from the Portuguese and German sections confirms the regional applicability of the results of Jones (1992) and the utility of this technique in long range correlation. Carbon and oxygen analysis of the same material supplements the data of other workers and a direct relationship is suggested between relative sea level and organic carbon burial in the Early Jurassic.

Long Abstract

The Lower Jurassic stratigraphy of seven contrasting areas in Western Europe is compared in order to seek evidence for a pan-Western European stratigraphic forcing mechanism. The Lower Jurassic was selected for this work because of the robust European ammonite biostratigraphic scheme, enabling inter-basinal correlation with a resolution of some 0.5Ma, and because of the objective of testing long-range correlation in a world where it seems likely that glacioeustasy can be eliminated. The seven study areas discussed are representative of the Wessex and Cleveland Basins (England), of the Lusitanian Basin (Portugal), of Tethyan rifts preserved in the western (French) and southern (Italian) Alps, of Southern Germany and of the North Viking Graben of the (Norwegian) North Sea.

Portable gamma-ray spectrometry data are presented covering the accessible parts of the Lower Jurassic sections of Dorset and Yorkshire and selected intervals of the Portuguese and south German sections. In Yorkshire there is a strong correspondence between sedimentological indications of proximity and elevated Th/K ratios which provides support for the use of Th/K ratio as a proximal-distal indicator. Spectral gamma-ray data and, in particular, Th/K ratios have then been used to elucidate proximal-distal relationships in fine-grained parts of the Yorkshire, Dorset, Portuguese and German sections which are difficult to interpret from sedimentological data alone. A model is proposed which reconciles the stratigraphies of Dorset and Yorkshire and suggests that they could be the product of a common stratigraphic forcing mechanism characterised by Hettangian-Early Sinemurian transgression, Late Sinemurian regression, earliest Pliensbachian transgression (with an Early Pliensbachian regressive pulse), Late Pliensbachian regression, potentially with several higher-frequency events, and Early Toarcian flooding.

The spectral gamma-ray data set also provides interesting insights into the occurrence of anoxia in the various basins (via the measurement of uranium concentration), useful examples of the interplay between elevated uranium concentrations and depressed Th/K ratios at "maximum

flooding surfaces", such as the Jet Rock of Yorkshire, and a suggestion of systematic variation in Th/K ratios across Europe from values around 3 in Portugal, through 4-5 in Germany and Dorset to 6-7 in Yorkshire. This may be interpreted to reflect latitudinal climatic effects or suspended sediment transport effects on a European scale.

The Lower Jurassic section at Peniche, Portugal is logged and described in detail for the first time and a log through a new roadcut exposing the Pliensbachian-Toarcian of Southern Germany is also presented. $^{87}\text{Sr}/^{86}\text{Sr}$ measurements on ammonite-controlled belemnite skeletal material from these sections confirm the regional applicability of the $^{87}\text{Sr}/^{86}\text{Sr}$ curve for Early Jurassic sea water developed using English material by Jones (1992) and the validity of the long-range ammonite correlations. It has then been possible to use $^{87}\text{Sr}/^{86}\text{Sr}$ measurements to determine the age of critical material from Portugal where no ammonites are available. Detailed logging and cliff photography of the Late Toarcian turbidite sequence at Peniche demonstrates significant differences from the work of Wright & Wilson (1986) and a model is suggested to explain the observations in terms of temporal alternations of clastic and carbonate supply to the fan.

The stratigraphic evolution of the Tethyan Rift in the Western and Southern Alps is briefly described from a combination of literature studies and reconnaissance fieldwork. Whole-rock X-ray diffraction studies of the Pliensbachian-Toarcian interval in the Breggia Gorge are presented and provide a useful method of monitoring siliciclastic sediment supply to the pelagic realm.

The final study area is the North Viking Graben of the North Sea. The three-dimensional distribution of Lower Jurassic sediments is interpreted using data from 68 petroleum exploration boreholes. New sequence, sand isopach and paleogeographic maps are presented based upon this interpretation.

Comparison of the stratigraphic evolution of the study areas suggests that many elements of the stratigraphic signature identified in the Dorset and Yorkshire sections are also manifest in other sections throughout Europe. The character of the European Early Jurassic stratigraphic signature

is discussed and compared with the sea-level curves of Hallam (1988) and Haq et al. (1988). Most notably, it appears that events with a "second order" (in the sense of Vail et al., 1990) periodicity are the most pervasive, whereas higher-frequency "third order" events are rather difficult to correlate. This runs counter to the idea that third order events are "global" and second order events are the product of "local tectonics". It is also notable that sharp basinwards facies shift, candidate sequence boundaries in the interpretative scheme of, for example, Van Wagoner et al. (1990) do not appear to correlate between basins.

Additional quantitative insights into Early Jurassic stratigraphic forcing mechanisms appear to be provided by $\delta^{13}\text{C}$ measurements. These are presented for skeletal carbonate from Portugal and Germany and from whole-rock measurements in Italy. When combined with the data of other workers they strongly suggest, despite significant data scatter, real temporal variation in the carbon-isotopic composition of European Early Jurassic sea water of some 3 per mil, with a periodicity approximating to the duration of a stage. It is proposed that this variation reflects variation in the rate of organic carbon burial resulting primarily from variations in sea-level and basin compartmentalisation consequent upon rifting. One might therefore predict links between the carbon-isotope signal and the stratigraphic signature of the Early Jurassic elucidated by the inter-basinal comparison of stratigraphies which forms the basis of this thesis. Such links appear to be present.

The principal conclusions of this thesis are that stratigraphic events of approximately stage-frequency can be correlated between the study areas and that there is a link between the observed stratigraphic cyclicity and the carbon isotope record. This work should provide a useful platform for inter-continental correlation in the future using similar techniques, together with the sea-water strontium isotope curve, which provides a vital independent geochronometer for long-range correlation studies.

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Contents

(see Volume 2 for figures)

Page No.

Short Abstract	
Long Abstract	
Acknowledgements	

Part I Background

Chapter 1 Introduction.....	1
Structure of this Thesis	2
Chapter 2 The Theory of Sequence Stratigraphy	4
Regressive-Transgressive and Reciprocal Sedimentation Models	5
Tectonic and Sediment Supply Complications	8
Stacking Patterns and Surfaces Associated with Reciprocal Sedimentation	10
Long Range and Global Correlation	14
Chapter 3 Overview of the Western European Lower Jurassic.....	17
Global Tectonics and Paleogeography	17
Western European Tectonics and Paleogeography	18
Previous Regional Correlations	19

Part II. Stratigraphy of the Detailed Study Areas

Chapter 4 The Wessex Basin.....	25
Blue Lias	26
Shales-with-Beef	30
Black Ven Marls	31
Belemnite Marls	33
Green Ammonite Beds	34
Three Tiers to Junction Bed	34
Downcliff Clay and Bridport Sands	37
Synthesis & Conclusions	37
Chapter 5 The Cleveland Basin.....	40
Calcareous Shales to Siliceous Shales	41
Pyritous Shales and Ironstone Shales	43
Staithes Sandstone and Cleveland Ironstone	45
Grey Shales to Peak Mudstone	47
Peak Mudstone to Dogger	48
Synthesis & Conclusions	49
Chapter 6 The Lusitanian Basin	51
Coimbra Formation	52
Brenha Formation, Unit A	54
Brenha Formation, Unit B	55
Brenha Formation, Unit C	57
Brenha Formation, Unit D	58
Brenha Formation, Unit E	61
Brenha Formation, Unit F	62
Brenha Formation, Unit G	63
Post- <i>bifrons</i> Turbidites	64
Strontium Isotope Stratigraphy	67
Conclusions	68

Chapter 7	South Germany	70
	The Aselfingen Section	71
	The B27 Roadcut	73
	Conclusions	76
Chapter 8	The Western Alps	77
	The Bourg d'Oisans-La Mure Area	78
	The Digne Area	83
	Conclusions	84
Chapter 9	The Southern Alps	87
	Conclusions	90
Chapter 10	The North Viking Graben	91
	Data Interpretation Proceedure	92
	The Well Correlations	95
	Stratigraphic Development	100
	Discussion and Conclusions	104
Part III . Synthesis		
Chapter 11	Comparison of Study Areas	106
	Hettangian-Early Sinemurian	106
	Late Sinemurian-Early Pliensbachian	107
	Late Pliensbachian	108
	Toarcian	112
	Discussion & Conclusions	113
Chapter 12	Carbon Isotopes: An Independent Sea Level Curve?.....	115
	Validity and New Data	116
	Theoretical Basis	118
	Comparison with Qualitative "Sea-Level" Curves	121
Chapter 13	Conclusions	124
Appendices		
	1. Gamma-Ray Spectrometry	126
	2. Strontium Isotope Mass Spectrometry	133
	3. Other Analytical Techniques	136
	Oxygen and Carbon Isotope Mass Spectrometry	136
	X-ray Diffraction	137
	Heavy Mineral Analysis	137
	Stains & Peels	138
	4. Timescales	139
References		141

Part I Background

Chapter 1. Introduction

By its narrowest definition, sequence stratigraphy is a method of stratigraphic analysis proposed by P. R. Vail, and co-workers at the Exxon Corporation's research laboratories in the 1970's and '80's (Payton (ed.), 1977; Wilgus et al. (eds.), 1988). The method has two principal components: a sedimentation model driven by relative sea-level change, whereby periods of high relative sea-level and proximal accumulation alternate with periods of proximal sediment by-pass and distal accumulation; and an assertion that the high-frequency relative sea-level changes which drive the model are global in extent. This narrow definition does not, however, encompass much of the work taking place in the renaissance in stratigraphy which has been stimulated by the ideas of the Exxon school. Many workers would, for example, reject sea level as the principal forcing mechanism to explain their observations (e.g. Galloway, 1989a) or reject global correlation (e.g. Hubbard, 1988). Some of these workers would still call their work "sequence stratigraphy", others would avoid the term because of the connotations of the narrow definition.

The title of this thesis is intended to reflect both the narrow and broad definitions of sequence stratigraphy. In the narrow sense, this thesis is a specific investigation of the proposition that synchronous pan-European relative sea level change was a key control on the development of the Lower Jurassic stratigraphy of Western Europe. Few well-documented examples of the regional correlation of depositional sequences exist and yet they are an essential pre-requisite of attempts at global correlation, since it is only by careful regional work that we have a clear idea of the key events which we would hope to correlate globally.

In the broad sense, the approach to the studied sections has very much been that of the new school of stratigraphy: an emphasis upon the identification of key surfaces bounding stratigraphic packages and upon the identification of systematic trends, such as progradation and retrogradation, within those packages.

The Lower Jurassic was selected for study for two principal reasons. Firstly, the rocks were deposited within a period of earth history during which the effects of glacio-eustatic sea level change would be expected to be least evident (Hallam, 1975): if global correlation can work at this time it can work at any time. Secondly, excellent long-range ammonite control is available for correlation, reflecting both the relative non-provinciality of ammonites during the Early Jurassic and over 150 years of careful biostratigraphical effort.

It was clearly impossible to examine the whole of the Lower Jurassic over the whole of western Europe within a project of this duration, yet it was considered vital to maintain significant spatial and temporal coverage if the results were to be meaningful in terms of the hypothesis to be tested: long-range correlation. The strategy chosen was to select seven study areas from around Europe, commonly single sections, which offered contrasts of proximal and distal settings, sediment accumulation rates and facies (fig. 1.1). Wherever possible the sections were to have good biostratigraphic control and relatively continuous outcrop. Each section studied is but a small sample drawn from a complex basin history deserving of at least one thesis in itself (and in some cases already the subject of several). Lack of contextual information may make interpretation more difficult in some cases, but it does not in any way invalidate the inter-section comparisons.

Structure of this Thesis

Chapters 2 and 3 (Part 1) provide essential background material. Chapter 2 reviews the principal ideas in sequence stratigraphy, ideas which underpin much of the discussion in subsequent chapters. Chapter 3 is an overview of the western European Lower Jurassic, in order to place the studied sections in context.

Chapters 4 to 10 (Part 2) describe the observations upon which this thesis is based and include interpretations of a rather local nature. The approach differs in each section depending on the

nature of the stratigraphy and the data available. The new data presented in this thesis are summarised in figure 1.2.

Chapters 11 to 13 (Part 3) attempt to draw the arguments together. Chapter 11 compares the sections and demonstrates the need for a pan-western-European stratigraphic forcing mechanism, going on to develop a qualitative sea level curve from the stratigraphic observations. Chapter 12 discusses systematic variation in the Early Jurassic carbon isotope curve and how this might relate to relative sea-level change. The principal conclusions of the study are summarised in Chapter 13.

Details of analytical procedures, essential background information on timescales and tables of raw data, will be found in the appendices. The reader unfamiliar with the zonal and subzonal subdivision of the Lower Jurassic may find it helpful to turn out page 140 of this volume.

Chapter 2. The Theory of Sequence Stratigraphy

This chapter reviews the principal ideas in sequence stratigraphy, ideas which underpin much of the discussion in subsequent chapters. It also serves as an introduction to the vocabulary of sequence stratigraphy and to how that vocabulary will be used in this thesis.

A large number of factors may influence the stratigraphic record and many of them are interdependent (fig. 2.1a). They may, however, be simplified into two groups: those factors which affect the amount of space available in which to put sediment ("accommodation space") and those factors which affect sediment supply. The first section of this review considers the case of constant sediment supply and spatially uniform tectonic subsidence. It looks at the stratal patterns produced simply by raising or lowering relative sea-level, thus establishing the principles of what will be described as the "regressive-transgressive" and "reciprocal sedimentation" models. Complications introduced by more complex tectonic environments and changing sediment supply are then briefly addressed in the second section. The third section, on "stacking patterns and surfaces", looks at the field and subsurface observations which are relevant to sequence-stratigraphic interpretation. The chapter then concludes with a discussion of the theoretical issues relating to global correlation using sequence stratigraphy.

Many of the sediments examined in Part II of this thesis accumulated below storm wave base, in what will be described as "supply-dominated" settings: areas where accumulation rate was governed more by the availability of sediment than by space to put the sediment in. It is important to recognise the differing response of sedimentary systems in "erosion-", "accommodation space-" and "supply-dominated" settings (fig. 2.1b) to the factors described below. It should also be noted that Schlager (1993) has used "supply-dominance" in a rather different sense from that used here. Schlager discusses the relative importance of sediment supply and sea-level change in building stratigraphic sequences, and uses "supply-dominance" for sequences where

variations in sediment supply is considered to be the most important factor in controlling regression and transgression. Schlager's "supply-dominated" sequences may thus still accumulate in the region of "space-domination" in the sense used here.

Regressive-Transgressive and Reciprocal Sedimentation Models

For the purposes of the following discussion, **eustatic sea-level** may be defined as sea-level relative to a single, fixed global datum, for example the centre of the earth (fig. 2.2a). Eustatic sea-level varies spatially in response to inhomogeneities in the earth's internal structure (fig. 2.2b), and also varies temporally, with the changing volume of the ocean basins (e.g. due to varying mid-ocean ridge volume) and the changing volume of ocean waters (e.g. due to continental glaciation). **Relative sea-level** may be defined as sea-level relative to a local datum, such as basement. **Water-depth** is the distance between the sea surface and sea bottom (fig. 2.2a).

Conditions of tectonic subsidence and static or rising eustatic sea-level will give rise to rising relative sea-level, thus creating space for the accumulation of sediment: **accommodation space**. The way in which space is filled is a function of sediment calibre and energy levels in the receiving basin. Essentially, however, if there is an excess of sediment supply over space creation then sedimentary systems will **prograde** and the sequence will be described as **regressive**. If sediment supply and space creation are in balance then sedimentary systems will **aggrade** and facies belts will remain stationary. If sediment supply is inadequate to keep pace with the creation of accommodation space then sedimentary systems will **retrograde** and the sequence will be described as **transgressive**. Note that throughout this thesis, regression and transgression refer to the basinwards and landwards migration of facies belts and not to the extent of marine incursion over land (see "coastal onlap", below).

The system can clearly be forced to oscillate between the states of regression and transgression by varying the rate of eustatic sea-level change, the rate of basement subsidence or the rate of sediment supply. This produces regressive-transgressive cycles, which have been recognised as

a principal building-block of stratigraphy since the early 19th Century. It is generally impossible to distinguish the relative influence of the three principal controlling factors when interpreting a single stratigraphic section (Kendall & Lerche, 1988), and it is for this reason that most of the discussion in chapters 4-10 of this thesis is couched in terms of regression and transgression rather than sea-level change.

In the regressive-transgressive model accommodation space is created continuously, though at varying rates. In contrast, tectonic uplift, or falling eustatic sea-level, may cause accommodation space to be destroyed. This leads to more complex sediment stacking patterns whereby intervals of proximal deposition and distal starvation alternate with intervals of distal deposition and proximal by-pass (Barrell, 1917): a pattern sometimes known as "reciprocal sedimentation" (e.g. Kidwell, 1988). Under these circumstances we replace a two-fold division of the stratigraphy (into regressive or transgressive components) with a three-fold division (fig. 2.3b):

1. Intervals when relative sea-level is rising quickly enough both to create accommodation space and to promote transgression. This is equivalent to the transgressive part of the regressive-transgressive model and generates the "Transgressive Systems Tract" (TST) of Posamentier & Vail (1988)

2. Intervals when relative sea-level is rising, but not rising fast enough to overcome sediment supply and promote transgression. This is equivalent to the regressive part of the regressive-transgressive model and would be described as a "Highstand Systems Tract" (HST) by Posamentier & Vail (1988)

3. Intervals when relative sea-level is falling. Accommodation space is destroyed and sediment is forced to by-pass the former shelf. This generates an offlapping package (the "Shelf Margin Wedge" (SMST) of Posamentier & Vail (1988) which is not present in the regressive-transgressive model. Successive offlapping packages have been referred to as the products of "forced regression" (Posamentier et al., 1992).

Posamentier & Vail (1988) also offer a variant on the above reciprocal sedimentation model. It relates to the observation, common on seismic reflection profiles from high-subsidence, ramp-type, settings and illustrated in figure 2.3, that sediment bodies tend to prograde basinwards with sigmoidal "foresets". The top of the "sigmoid" equates with the "depositional shoreline break" of Van Wagoner et al. (1988), which they define as "a position on the shelf, landward of which the depositional surface is at or near base level, usually sea level, and seaward of which the depositional surface is below base level". In the sense used in this work, the point marks the transition from space-dominated to supply-dominated sedimentation. Figure 2.3b illustrates a case where relative sea-level fell, but did not fall below the depositional shoreline break of the underlying sequence. The by-pass surface in this situation is described as a "type 2" sequence boundary by Van Wagoner et al. (1988). Posamentier & Vail (1988) suggest that a very different situation will arise if relative sea-level does fall below the depositional shoreline break of the underlying depositional sequence: there will be no shelfal accommodation space and sediment will be delivered directly into deep water to develop a "Lowstand Fan" ("LF" in fig. 2.3c). The by-pass surface in this case is known as a "type 1" sequence boundary. Posamentier & Vail's model suggests that as relative sea-level begins to rise again, a new shelfal system may build out below the old depositional shoreline break, a system which they refer to as the "Lowstand Prograding Wedge" ("LPW", fig. 2.3c).

We may develop this discussion further in two directions. Firstly, how important is relative sea-level fall in the development of the stratigraphic record, i.e: how often must we appeal to a reciprocal rather than a regressive-transgressive model to explain our observations? Secondly, do these models adequately describe, even to a first approximation, the way real sedimentary systems respond to relative sea-level change? Reflection seismic data provide numerous real examples of reciprocal sedimentation (e.g. Bally, 1989), however these largely relate either to Neogene or younger sediments, where one would expect to see a dominant glacioeustatic signature (Bartek et al., 1991) and/or to tectonic environments where tectonic hinges could be anticipated (see below). It is by no means given that relative sea-level fall was a common feature of stratigraphic

development in the epeiric seas of the European Jurassic and each case must be taken on its own merits.

The response of sedimentary systems to sea-level change continues to generate much controversy. The two areas in which the models of Posamentier & Vail (1988) have received most criticism relate to the way in which base-of-slope features are generated and the way in which carbonates respond to base level change. Galloway (1989b) suggests that mass wasting during relative sea-level highstands plays a significant role in the production of apparent "Lowstand Fans". Schlager (1989) points out that, up to certain critical levels, relative sea-level rise in carbonates can promote productivity, thus encouraging progradation: one example of where sea-level and sediment supply are not independent (see below). Conversely, relative sea-level fall can kill a carbonate platform and extinguish supply.

Tectonic and Sediment Supply Complications

The above discussion assumes that subsidence is spatially uniform. Hence a eustatic rise is indistinguishable from basin subsidence. It is most unusual for this to be the case in reality, and most stratigraphic models (e.g. Pitman, 1978; Posamentier & Vail, 1988) assume that subsidence-rate decreases proximally, eventually becoming negative (i.e. uplift). The cross-over point between subsidence and uplift is known as the **hinge line**. Models incorporating a hinge line are realistic in the sense that subsidence and loading of elastic lithosphere results in a peripheral bulge (Watts, 1982) hence uplift and subsidence are intimately linked. Similar results, from a stratigraphic viewpoint, may be produced by footwall uplift associated with active rifting (Barr, 1987). One important effect of a hinge basinward of the onlap point is to force maximum rate of sea-level rise to coincide with maximum onlap (fig. 2.4). In the absence of a hinge, onlap will continue so long as relative sea-level is rising, i.e. so long as basin subsidence rate exceeds the rate of eustatic sea-level fall (fig. 2.4a). With a hinge, the situation is more complex. We can consider a suite of relative sea-level curves (1 to 9, fig. 2.4b), which represent a sinusoidal eustatic curve (with the form of curve 5) combined with various rates of subsidence (curves 1-4, relative

sea-level generally increasing with time) and uplift (curves 6-9, relative sea-level generally decreasing with time). Curves 1-9 may be thought of as different points on the tectonically-tilting slope, "9" most proximal, "5" at the hinge and "1" most distal. Deposition proximal of the hinge will occur whilst relative sea-level is rising, i.e. within the envelope A-A'-A" (fig 2.4b). Maximum onlap will occur at the time represented by A', which is the time of maximum rate of eustatic sea-level rise. Curve 9 represents a point on the slope where uplift rates are sufficiently high that no relative sea-level rise is recorded, simply a stationary value of the rate of relative sea-level fall. The stratal patterns produced by a hinged system are represented in figure 2.4c.

Hinges basinward of the onlap point enable offlap to take place without a eustatic sea-level fall: an increase in tilt rate will suffice. This is the basis of various tectonically-driven reciprocal sedimentation models (Watts, 1982; Cloetingh, 1985). Tectonically-driven models also have the potential to introduce important feedbacks and dependencies into the system (Gaffin, 1992): for example offlap leading to distal loading, flexure and further offlap or the link between tilting and increased sediment supply.

In most cases it is impossible to alter relative sea-level without also affecting sediment supply. Relative sea-level change will affect supply directly by altering the long-profile of rivers (Schumm, 1993) and indirectly by changing climate, for example by changing the earth's albedo (Donn & Shaw, 1977). Most basin-forming processes will simultaneously affect both accommodation space and sediment supply: for example, footwall uplift, flexure or an encroaching thrust front. The work of Schlager on the complex relationships between carbonate systems and relative sea-level has already been noted. The response of sedimentary systems to changing base level may be neither immediate nor linear, introducing the possibility of chaotic behaviour.

The above complexities have led Smith (1994) to describe the stratigraphic record as the chaotic output of a "stratigraphy machine" which convolves the various input variables: sea-level change, climatic variability, tectonics etc., in an entirely unpredictable manner. In subsequent chapters of this thesis it is hoped to demonstrate some coherence in the output of Smith's stratigraphy

machine between different basins across Europe. This will be used to infer the influence of a strong pan-European forcing mechanism. The extent to which it is legitimate to use this to extract one of the inputs to the "machine", a sea-level curve, must still be in doubt.

Stacking Patterns and Surfaces Associated With Reciprocal Sedimentation

The products of reciprocal sedimentation can clearly be recognised on seismic data, but how might they be recognised in the field? This question has recently been reviewed by Van Wagoner et al (1990) for siliciclastics and Schlager (1992) for carbonates. It is essentially a matter of recognising stacking patterns and key surfaces.

Shallow marine sediments are commonly organised into small-scale shallowing-upwards cycles, variously known as "Punctuated Aggradational Cycles" (Goodwin & Anderson, 1985) or parasequences (Van Wagoner et al., 1990). The way in which these cycles relate to each other vertically and horizontally has been described as the "stacking pattern" (Van Wagoner et al., 1990). In accommodation space dominated settings, if sediment supply and accommodation space creation are in balance then aggradational stacks will be created, recognisable by the even spacing of small-scale cycles. Where the rate of accommodation space creation outpaces sediment supply, retrogradational stacks will be created, characterised by upward-thickening of small-scale cycles and the inclusion of progressively more distal facies within each cycle. Where sediment supply exceeds the rate of accommodation space creation then progradational stacks will result, characterised by upward-thinning of small-scale cycles and the inclusion of progressively more proximal facies within each cycle. The decreasing rates of accommodation space creation during deposition of the Highstand Systems Tract might be expected to cause aggradational patterns to give way upwards to progradational patterns, whereas the reverse might be expected for the Shelf Margin Systems Tract (fig. 2.5). Retrogradational stacks would be a feature of the Transgressive Systems Tract. If small-scale cycles are not present, then aggradation may still be interpreted from the stability of facies in space and time in contrast to sections exhibiting rapid facies belt migration.

As noted above, and as is common in the literature, the discussion above concentrates on accommodation space-dominated settings. Such settings, whilst often allowing unambiguous sedimentological and sequence-stratigraphic interpretation, rarely allow for good long-range correlation as pelagic fossils are uncommon. In supply-dominated settings, such as many of those discussed in this thesis, the above rules must be reversed: decreasing rates of shelfal accommodation space creation result in increasing supply to the basin, hence the toes of "sigmoids" may show thickening-upwards patterns in the basins equivalent to thinning-upwards patterns on the shelf. The converse is true of increasing rates of shelfal space creation (fig. 2.5).

The three key surfaces associated with the reciprocal sedimentation model (Van Wagoner et al., 1988) are: 1) the **sequence boundary**, associated with relative sea-level fall and commonly illustrated as separating the Highstand Systems Tract from the Shelf Margin Systems Tract; 2) the **transgressive surface** or **first flooding surface**, which represents the point at which relative sea level rise is sufficient to overcome progradation and thus divides the Shelf Margin Systems Tract from the Transgressive Systems Tract; and 3) the **maximum flooding surface**, which represents the point at which rates of base level rise are no longer sufficient to hold back progradation and thus marks the boundary between the Transgressive and Highstand Systems Tracts.

Most proximally, the sequence boundary is unambiguous: it is an unconformity resulting from offlap driven by relative sea-level fall. The time represented by the unconformity will decrease distally: the time when erosion begins gets later and the time when redeposition begins gets earlier. The unconformity will converge distally on what has been described as its "correlative conformity" (Mitchum Jr. et al., 1977). The timing of this correlative conformity may be considered in a manner exactly analogous to the discussion of the timing of maximum onlap (above). If subsidence is uniform, relative sea level will be falling so long as the rate of eustatic fall exceeds the rate of basin subsidence. The age of the correlative conformity will therefore be sometime on the falling limb of the relative sea-level curve, after the maximum rate of relative sea level fall (fig.

2.6a). Sequence-stratigraphic models (e.g. Jervy, 1988) commonly assume that a tectonic hinge is present and that subsidence-rate increases basinwards. As with the previous discussion of onlap, one can consider a suite of curves representing the combination of a single sinusoidally-varying eustatic sea-level curve and a range of linear tectonic subsidence rates (curves 1-9, fig. 2.6b). Erosion will occur distal of the hinge point so long as relative sea-level is falling, i.e. within the envelope B-B'-B" (fig 2.6b). The unconformity will pass into its correlative conformity at the time represented by B', which is the time of maximum rate of eustatic sea-level fall. Curve 1 represents a point on the slope where subsidence rates are sufficiently high that no relative sea-level fall is recorded, simply a stationary value of the rate of relative sea-level rise.

There are two critical points. Firstly, the existence of a sequence boundary, and its timing, is entirely dictated by the relative sea-level curve. Unlike the maximum or first flooding surfaces, it is independent of sediment supply variations. Secondly, given the model of linear subsidence and a sinusoidal sea-level curve, the correlative conformity dates the maximum rate of fall of eustatic sea-level, independent of tilt-rate. It is for these reasons that the sequence boundary is favoured by some authors (Vail et al., 1977; Haq et al., 1988) for use in global correlation. Clearly, convolving non-linear subsidence, which is likely to be the norm over longer geological time-periods, with a sinusoidal eustatic curve will displace the stationary value of the relative sea-level curve in time and hence alter the date of the correlative conformity.

There are also practical problems with the identification of sequence boundaries where they become conformable. Van Wagoner et al. (1990) suggest that they may be located by a sharp upwards change to more proximal facies. However, it seems likely that in many cases the first erosion products from a relative sea-level fall will be seen in the basin before the maximum rate of relative sea-level fall is attained (fig. 2.7a). This is implicit in the diagrams of Posamentier et al. (1988) and explicit in the work of Cant (1990). It also seems likely that sharp facies shifts may result from the attainment of critical energy levels in the basin, for example the impingement of storm wave-base upon the sea bottom, and that sharp-based sand bodies may be generated by autocyclic processes. One of the conclusions of the present study is that, for the Lower Jurassic

of western Europe, sharp upwards changes to more proximal facies can rarely be correlated at the ammonite-subzonal level between basins (Chapter 11).

The transgressive surface would be predicted to be marked, in areas of space-domination, by distal facies sharply overlying more proximal facies. Such surfaces might be expected even with a simple regressive-transgressive model, in that progradation would be expected to generate a broad, low gradient shelf and coastal plain which would then be flooded rapidly by even a small relative base level rise. The principal widely-recognised complication associated with the transgressive surface is the "transgressive surface of erosion", more opaquely known as a "ravinement surface" (Swift, 1968). This concept recognises the importance of an erosive shoreface in the transgressive process, which can erode the previous deposits and leave shelfal facies resting erosively on more proximal facies as the evidence of transgression. In supply-dominated settings the transgressive surface would generally be marked by a sharp upwards decrease in sediment supply although the transgressive surface of erosion might generate its own supply peak.

The maximum flooding surface is widely acknowledged as the most useful surface for practical local correlation of outcrop and subsurface data whether the regressive-transgressive or reciprocal sedimentation models are being applied (Frazier, 1974; Galloway, 1989a; Vail, 1990). The term "maximum flooding" is somewhat confusing, as the surface actually marks the point at which the rate of relative sea-level rise is unable to hold back progradation for a given sediment supply. Onlap may continue after this point as long as relative sea-level is rising. Given a curvilinear relative sea-level history, the maximum flooding surface will occur at some time after the maximum rate of sea-level rise. The utility of the maximum flooding surface lies in its ease of recognition. It is marked by the most distal facies in space-dominated settings and by a supply minimum in supply-dominated settings (which reach their maximum extent at this time). The stratigraphic products of low sediment accumulation rates are reviewed by Jenkyns (1971), Loutit et al. (1988) and Kidwell (1991). They include the development of fossil concentrations, firm- and hard-grounds and the

promotion of authigenic processes such as concretion formation. In conditions of low sediment supply, biogenic reworking may cause net loss of sediment (Hesselbo & Palmer, 1992).

The maximum flooding surface is also known as the "downlap surface", a term which refers to its classical expression on seismic profiles. It is appropriate here to speculate upon some of the properties of the downlap surface, since these bear upon the interpretation of intra-basinal topography in supply-dominated settings, which is relevant to a number of the study areas in this thesis. The first point is that intra-basinal highs are likely to suffer the most extreme condensation in periods of low sediment supply (rapidly rising base level), this phenomenon is potentially confusable with winnowing due to the impingement of wavebase on the high (rapidly falling base level). Increasing sediment supply may cause inundation of intra-basinal highs (fig. 2.7b).

The second point is that onlap patterns in starved half-graben relate not just to fault growth rate (Prosser, 1993), but to the balance between fault growth and sediment supply. Onlap patterns occur when supply exceeds growth rate, offlap patterns when growth rate exceeds supply. Hence seismic unconformities (genetically, "downlap surfaces") may be created either by increasing the rotation rate with constant supply or by decreasing the supply rate with constant rotation rate (fig. 2.7c).

Long Range and Global Correlation

The above discussion is concerned with the interpretation of individual sedimentary piles, which is an essential prelude to any correlation exercise and is the preoccupation of Part II of this thesis. However, the principal objective of this thesis is to use these local interpretations to examine the evidence for long-range synchronicity of stratigraphic events. It is with this in mind that we now turn to a brief review of the theory of long-range correlation using sequence stratigraphy.

The quest for a "natural" stratigraphy with which to bring order to our understanding of earth history is as old as stratigraphy itself. The modern distinction between lithostratigraphy and

chronostratigraphy was not a feature of the work of early stratigraphers, such as William Smith, who saw their stratigraphic units as encompassing both spheres (Rupke, 1983). As we came to know more of the stratigraphic record, the similarities between areas were gradually subsumed by the differences, although some authors continued to develop the ideas of long range tectonic cycles (Stille, 1924; Sloss, 1963).

Serious attempts to create a "natural" global stratigraphy were rekindled by the work of Vail et al. (1977). They identified sequence boundaries on seismic lines, dated them using well penetrations as close as possible to their "correlative conformities", and estimated the relative magnitudes of the inferred relative sea-level changes from onlap-offlap relationships (fig. 2.8a). They felt that similarities between widely separated datasets justified the erection of a "Global Cycle Chart". The work gave rise to two linked criticisms: that the observations were inadequate to support the hypothesis, and that no mechanism for synchronous global sea level change could be found for certain periods of geological time. Numerous alternative hypotheses were generated to explain the observations in terms of local tectonic effects (Pitman, 1978; Watts, 1982; Parkinson & Summerhayes, 1983; Cloetingh et al., 1985). Despite the objections, however, Vail et al.'s ideas clearly found some resonance with stratigraphers who had compared rocks on an inter-continental scale, e.g. Hancock & Kaufmann (1979), and numerous papers by Hallam discussed in Chapter 3 of this thesis.

Vail et al.'s (1977) published data were certainly weak and suggested that the curve was a composite of various sections from around the world rather than extracting the "lowest common denominator" (fig. 2.8b). The "coastal onlap chart" of Haq et al. (1988) has continued to be elaborated at a rate which many would feel is not commensurate with global biostratigraphic resolution or sequence-stratigraphic understanding. If the observations and their interpretation in terms of sea-level are correct then they appear to demand global changes of sea level of period 0-5Ma and 10's of metres magnitude: quite feasible in an "ice-house", but much less feasible in a "greenhouse" world (Donovan et al., 1979).

The position adopted in this thesis is that the case for a "natural" global stratigraphy is still far from proven, but that the prize is worth a great deal of labour.

Chapter 3. Overview of the Western European Lower Jurassic

The objectives of this chapter are to place the sections studied in context and to review previous work on the regional correlation of the Lower Jurassic.

Global Tectonics and Paleogeography

In global terms, Early Jurassic western Europe was located at around 30°N in the north-western corner of "Paleotethys" (fig. 3.1). Pangea was largely intact at this time, being surrounded by relatively old "Pacific" type, and therefore deep, oceans (Wyatt, 1988) which may explain the relatively high continental freeboards associated with the Permian and Triassic. The Triassic-Jurassic tectonic *leitmotiv* is the break-up of Pangea, and it is to rift-related foundering which one would immediately turn as the primary cause of widespread Jurassic transgression.

Basaltic volcanism, associated with rifting, is found in the Newark Supergroup (onshore eastern United States: Manspeizer, 1988). More spectacular volumes of Lower Jurassic basaltic material, associated with the separation of Africa and Antarctica, were extruded to form the Karoo complex of south-east Africa and the equivalent rocks of Dronning Maud Land, Antarctica (Cox, 1989). This volcanism may have played a key rôle in the $^{87}\text{Sr}/^{86}\text{Sr}$ history of Early Jurassic seas (Appendix 2).

The absence of glacial facies (Hallam, 1985) suggests the absence of permanent low latitude ice on the Early Jurassic globe, though Frakes and Francis (1988) have advanced an alternative view based on putative dropstone evidence. Climatic models for the Lower Jurassic (Chandler et al., 1992) suggest that, whilst Lower Jurassic climate may not have been particularly "equable", as it is often described (e.g. Hallam, 1975), it was probably warmer than that of today: by perhaps 5-10°C on average, and much more at high latitudes. Vegetation categories at the latitudes of this study probably ranged between desert in the south and mixed deciduous-evergreen forest in the north

(Chandler et al., 1992). The "cul-de-sac" location at the head of Paleotethys may have induced a propensity for anoxia due to removal from sources of intermediate water in a manner analogous to the modern Indian Ocean (Fleet et al., 1987).

Western European Tectonics and Paleogeography

North-west European rifting is dominated by Triassic and Late Jurassic events (fig. 3.2). The exceptions to this are the basins of the Tethyan Rift: Morocco, the Scotian Margin and the Ligurian Tethys (chapters 8 & 9). This supports the interpretation of a "zig-zag" Early Jurassic Tethyan rift stretching from the Caribbean Tethys (largely transform), through the Atlantic Tethys and the Gibraltar-Maghreb-Sicilia transform to the Ligurian Tethys (fig. 3.3). Rifting described in the basins of Northern Europe (chapters 4, 5, 7 & 10) should be seen as peripheral to this activity. The oldest sediments so far found on oceanic crust along the Tethyan Rift are Callovian-Oxfordian in age (Lemoine, 1983).

North-West European early Jurassic paleogeography is conventionally cast as a series of "highs" associated with the ancient crystalline massifs, and separated by fault-controlled depocentres (fig. 3.4). Most of these highs can be substantiated as islands by facies evidence, at least in the earliest Early Jurassic: the Pennine-Grampian High (Chapter 5), the Iberian High (Chapter 6), the Amorican and Central Massif Highs (Cariou et al., 1991), and possibly the Welsh High (Chapter 4). The London-Brabant High may have been submerged throughout the Early Jurassic (Chapter 4).

None of these highs can have been major siliciclastic source areas as the largest was similar in size to present-day Britain. For significant siliciclastic supply we must look to the Fennoscandian and possibly to the Laurentian Shields. Lower Jurassic deltaic sediments are extensively developed on Haltenbanken, offshore Norway (fig. 3.4, see also Karlsson, 1984) and other siliciclastic entry points may have been located further south along the Norwegian coast (Chapter 11) and in the Polish-Danish Trough. The seas of Northern Europe can thus be envisaged as supplied largely by fine-grained suspended sediment from the north and east. To the south the climate must have

been warmer and the seas may also have been clearer (less suspended sediment load) and better supplied with oxygen and nutrients from Paleotethys. Carbonates flourished in the shoal areas here (e.g. Bernoulli & Jenkyns, 1974).

Facies interpretations (e.g. Part II of this thesis) suggest that water depths over much of the area, except for the floors of major half-graben, must have been intermittently within reach of at least storm wavebase. This implies water depths, analogous to modern shelves, of up to about 200m (Johnson & Baldwin, 1986). Sellwood (1972b) recognised tidal flat deposits from the Sinemurian of Bornholm, Denmark and Livbjerg and Mjøs (1989) suggested a tidal sand ridge origin for the Pliensbachian sands of Oseberg area in the North Viking Graben (Chapter 11). Most sandstones, however, are massive and bioturbated or, more distally, preserve starved oscillation ripples and scours (Chapters 4, 5 & 11). Analogous facies occur on modern shelves such as the Oregon Shelf (Kulm et al., 1975) or the South Texas Shelf (Berryhill, 1986). Sand is principally deposited from suspension during storms and then reworked by waves and organisms. The configuration of the European Shelf in Jurassic times was probably not favourable for the development of strong tides. Meteorological currents are therefore the favoured mechanism for sediment transport. Tidal processes may still be manifest in estuaries where other sources of energy are not available (Johnson & Baldwin, 1986) but clear examples of these in the Lower Jurassic are certainly not common.

Previous Regional Correlations

The recognition of cyclicity in the British Jurassic dates back to the 19th Century (Phillips, 1871) and was extended to the base of the Lower Jurassic by Arkell (1933). These were clay-sandstone-limestone cycles at what we might call "2nd Order" scale (following Vail et al., 1991). In Dorset (Chapter 4), the two cycles relevant to the Lower Jurassic were that from Sinemurian marls through Upper Pliensbachian sands to the Marlstone and that from Upper Toarcian marls through uppermost Toarcian sands to the Middle Jurassic Inferior Oolite. Cycles at a somewhat finer scale were recognised in the Lower and Middle Jurassic of Lorraine by Klüpfel (1916) and, for the

Sinemurian-Lower Pliensbachian interval, traced from north-west, through southern Germany and into France by Frebold (1927). The accepted explanation for these observations was widespread synchronous crustal movement (Stille, 1924).

The outstanding modern worker in the field has been Hallam, whose ideas on Jurassic sedimentary cycles have been developed in a number of papers since 1961 (fig. 3.5). Hallam's 1961 paper uses the key horizons of the Dorset Lias (Chapter 4) to divide the Lower Jurassic into eleven "Klūpfelian cyclothems" and demonstrates that these are also critical horizons in other British outcrops, in the eastern Paris Basin and in Germany. By 1978 this scheme had been developed into a eustatic sea level curve for the Jurassic and evidence from around the world began to be incorporated. This comprises mostly a series of transgressions: the Hettangian in Siberia and Chile; the Early Sinemurian in Alberta and Mexico; the Late Sinemurian in Chile-Argentina and other parts of the Andes; the Early Pliensbachian in East Greenland and North Africa; the Late Pliensbachian in Siberia, the Caucasus, northern Iran and Japan and the Toarcian in Saudi-Arabia, Pakistan and Madagascar. These observations appear superficially compatible with the idea of progressive rifting outlined previously in this chapter, but Hallam clearly felt that there was more synchronicity between the transgressions than a purely tectonic model would admit. The general transgressive nature of the Lower Jurassic is clearly shown by Hallam's (1969) measurement of the extent of marine facies on world maps (fig. 3.6).

The most detailed account of Hallam's views on Early Jurassic eustasy is contained in his 1981 paper. The curve is essentially that published in 1978, but more substantiation is given at the zonal level. The cusps on Hallam's 1981 curve may be summarised as follows:

1. Base *liasicus* Zone. Deepening is interpreted at the base of the "Liasicus Shale" within the Blue Lias of Dorset (and elsewhere in the UK), as discussed in Chapter 4. Hallam (1961) suggests that a similar facies change also exists in South West Germany.

2. Basal Sinemurian. A similar event to (1), represented by the "Bucklandi Shale" of Dorset (Chapter 4). Hallam suggests that this event is minor and much subordinate to:
3. *Semicostatum* Zone. A sea level rise: the sharp facies shift at the top of the "Blue Lias" in southern Britain is equated with widespread global evidence for transgression at this time.
4. Mid-*oxynotum* Zone. The "Coinstone" horizon of Dorset and its equivalent hiatus in South Germany is interpreted as recording a sea-level fall. No evidence is presented for the existence of this event outside Europe.
5. End Sinemurian. The "Hummocky" horizon of Dorset and its German equivalents are interpreted as a sea level rise which Hallam also feels may be identifiable in South America.
6. *Ibex-davoei* Zonal boundary. A possible eustatic fall inferred from the sparsity of records of the *ibex* and *davoei* Zones outside Europe. Hallam does not appear to refer to the distinct facies transitions at this time in the English sections (chapters 4 & 5).
7. Basal *margaritatus* Zone. A slight eustatic fall, followed by a rise. The fall is not discussed, but the rise is interpreted from the shift to argillaceous facies in Europe - southern Germany (Chapter 8) and apparently in the Cantabrian Mountains of Spain - and records of transgression in Japan, Siberia, Oregon (U.S.A.) and the North American Arctic. Hallam suggests that the regressive facies of north-west Europe are exceptional and may be the result of local tectonics.
8. Base Toarcian. Deepening interpreted from facies changes in western Europe, the American Arctic and eastern Asia and from marine incursions in the Middle East, East Africa and Alberta. The accelerated sea-level rise shown by Hallam at the base of the *falciferum* Zone presumably reflects the particularly widespread recognition of organic-rich facies at this level.

9. Top *bifrons* Zone. A eustatic fall interpreted from reworked horizons in north-east France and south Germany, from the Raasay Ironstone in Scotland and from facies interpretation in the Chilean-Argentinean Andes.

In his most recent (1988) paper on the subject, Hallam moved to a sinusoidal rather than cusped sea-level curve, based on an acknowledgement that lowstands may be as long as highstands but represented in the record by less rock due to condensation. The 1988 curve no longer recognises significant events at the base of the *liasicus* Zone ("1", above), or within the *ibex/davoei* Zones ("6"). The events near the base of the Sinemurian ("2" and "3") have been smoothed into one, apparently basal Sinemurian, sea-level low. A further effect of smoothing out the cusps appears to be the introduction of a sea-level fall at the base of the Pliensbachian ("5"). This is not discussed in the text, but would bring the interpretation of two similar horizons - "4" and "5" - into line. A similar smoothing effect has promoted a low-point on the relative sea level curve at the end of the Pliensbachian.

The general approach used by the "Exxon School" of sequence stratigraphers has been outlined in Chapter 2. The early Global Cycle Chart of Vail et al (1977) does not attempt to subdivide the Lower Jurassic, save for placing a major basinward shift in coastal onlap at the top of the Hettangian (the base of their "Jurassic Supercycle"), and a less prominent basinward shift at the top of the Toarcian (fig. 3.7). The published seismic-stratigraphic study which includes this interval (Todd & Mitchum, 1977) provides little support for even this simple scheme. It looks at two areas: the Gulf of Mexico and the continental margin of North-West Africa. In the former area Triassic (and possibly lowermost Jurassic, based on plant macrofossils) continental red beds are overlain by anhydrite and salt which is apparently assigned to the Middle Jurassic on the basis that "worldwide experience indicates that Lower Jurassic sequences are areally restricted...[whilst]...worldwide experience with global cycles indicates Middle Jurassic rocks to be comparatively widespread...". In West Africa the stratigraphy is little better controlled: brachiopod data apparently suggest that red beds extend into the lowermost Jurassic. Of the overlying four marine sequences, the first is dated as Sinemurian and the last as Callovian. The intervening three sequence boundaries

appear to be unconstrained by biostratigraphy. The evidence seems to offer support for little more than the Lower Jurassic, locally specifically Sinemurian, transgression reported by Hallam (see above).

Vail and Todd (1981) offer the only other published seismic-stratigraphic study of the Lower Jurassic. The two areas studied here were the North Viking Graben (see Chapter 10) and the Inner Moray Firth (offshore North-East Scotland). Two seismic-sequence boundaries were identified in the North Viking Graben (fig. 3.8). One was tied to the non-marine - marine transition, then thought to be of earliest Sinemurian age, and the other was tied to the top of the Dunlin Group. The latter shows distinct truncation and is equivalent to the "mid-Cimmerian" unconformity (chapters 5 & 10). An intermediate sequence boundary, based only on well data, was placed at the base of the late Pliensbachian Cook Formation. The Lower Jurassic interval in the Inner Moray Firth was acknowledged as being too thin for seismic stratigraphy, but a major Late Pliensbachian hiatus over the Beatrice field was equated with the "basal Cook" sequence boundary.

These examples serve to emphasise the limitations of the seismic-stratigraphic method where sequences have not been expanded by high subsidence rates. More recent studies of the Lower Jurassic by Exxon School stratigraphers have concentrated on outcrop data, specifically (as listed in Haq et al, 1988) the Dorset and Yorkshire sections examined for this thesis, the type section for the Pliensbachian in South Germany (somewhat expanded, but essentially the same sequence as that discussed in Chapter 8) and the type section of the Toarcian at Thouars (France). The details shown on the chart of Haq et al. (1988) are therefore based on essentially the same European outcrops discussed by Hallam and described in this thesis. Differences are largely matters of facies interpretation, to which Haq et al. bring the predictions (and prejudices) of the reciprocal sedimentation model discussed in Chapter 2.

In summary, the case for a global Early Jurassic sea-level curve independent of local rifting does not, from the published record, appear convincing. The starting point for a global curve must be a thorough understanding of the well-dated sequences in western Europe. Both Hallam (1988) and

Haq et al (1988) have interpreted this data set, yet their interpretations differ substantially (fig. 3.7). Neither authors provide sufficient detail to enable their curves to be evaluated against the data without an intimate knowledge of European Lower Jurassic stratigraphy on the part of the reader. In this context, the objectives of this thesis are:

- To document a number of key sections and discuss the commonly equivocal interpretation of the facies in terms of sequence stratigraphy.
- To add quantitative data to the debate wherever possible.
- To assess the extent to which the data support the erection of a pan-western-European sea-level curve.

Part II Stratigraphy of the Detailed Study Areas

Chapter 4. The Wessex Basin

The Lower Jurassic rocks of southern England were deposited in an east-west oriented extensional basin known as the Wessex Basin (fig. 4.1). In Early Jurassic times the basin was bounded to the north by the Welsh Landmass and the London Platform. Neither of these "highs" have been shown unequivocally to have been emergent. The "proximal" facies of the Lower Jurassic in south Wales (Trueman, 1920, 1922, 1930; Hallam, 1960) have been re-interpreted as a mass-flow deposit by Ager (1986). Onlap of the London Platform has been demonstrated from borehole evidence by Donovan et al. (1979), but this may be submarine onlap (fig. 2.7b). Deposition in the saddle between the Welsh Landmass and the London platform was, to a first approximation, continuous from Hettangian to Toarcian times and Brandon et al. (1990) have erected a borehole correlation for the Hettangian-Pliensbachian interval across this saddle, connecting the Wessex and Yorkshire Basins. A compilation of data by the British Geological Survey (Whittaker (ed.), 1985) shows a series of east-west extensional faults defining a series of half-grabens which step down towards the Wessex Basin axis in the English Channel and control Lower Jurassic sediment thickness.

The exposures on the Dorset coast, resulting from Tertiary inversion, are classic sections for Jurassic stratigraphy. Not only is ammonite control excellent (Cope et al., 1980), but the sections have been discussed so widely by previous workers (Chapter 3) that they may be regarded as "reference sections" against which other areas are compared. Hesselbo & Jenkyns (in press) comprehensively document the previous lithostratigraphic and biostratigraphic literature.

This chapter is concerned with the interpretation of Dorset Lower Jurassic stratigraphy in terms of varying sediment supply and relative sea-level change. The interpretations are principally based on stratigraphic logs by Hesselbo & Jenkyns (in press), together with new spectral gamma-ray measurements and sedimentological observations. The portable spectral gamma-ray tool has proved to be a valuable way of elucidating vertical change in otherwise cryptic mudrock

successions (see Appendix 1 for a discussion of the technique and the generalities of interpretation).

The sections discussed may be located in figure 4.2. Details of access are to be found in House (1989). The major lithostratigraphic units (fig. 4.26) are first described in stratigraphic order. A synthesis is then attempted.

Blue Lias

The lowermost unit of the Dorset Lower Jurassic is known as the Blue Lias (Lang, 1924). Crudely, the unit comprises decimetre interbeds of limestone and marl (fig. 4.3) which may be traced laterally for 100's of metres. In detail, five rock types may be distinguished (Hallam, 1960; Weedon, 1986): limestone, light and dark marl, laminated shale and laminated limestone; the limestones being formed by early diagenetic cementation of light marls and, occasionally, of laminated shales. These rock types are organised at a number of scales: Weedon's power-spectral analysis of the limestone-marl rhythms showed that they are grouped into cycles of 51 and 85 cm in thickness, which he suggested might correspond to the precession and obliquity astronomical cycles. Plots using Weedon's data (fig. 4.5a) show the most open bed spacing in the *planorbis* and higher *rotiforme* ("*bucklandi*") Subzones and marked bed tightening in the *angulata* Subzone. Further smoothing of the data (fig. 4.5b) emphasises a bed-spacing cyclicity at the scale of ammonite subzones. Such interpretations assume, of course, that the limestone-marl alternations are fundamentally primary. This is undoubtedly true in many cases, as demonstrated most convincingly by the piping of light into dark marl by sediment burrowers. Hallam (1986) expresses reservations, however, citing an apparent correlation between sequence thickness and number of limestone beds as evidence for primary diagenetic control. Weedon (1987) offers a number of alternative explanations for this, perhaps the most convincing of which, for the case of the Blue Lias, is that significant but biostratigraphically unresolved hiatuses occur within the succession.

Interbed frequency in the Blue Lias is greater than spectral gamma-ray tool resolution (Appendix 1). The Blue Lias is not, therefore, ideally suited to this method of analysis. For completeness, however, half-metre-spaced measurements were made through the section (except for the *liasicus* Zone which was inaccessible at the time of logging) and these yielded useful results. All log traces are highly erratic (fig. 4.6), reflecting the interbedded nature of the sequence. Th and K co-vary rather strongly (fig. 4.7a) and the regression line passes close to the origin. As will be seen, this is a feature of many mudrock units and suggests that the Th and K reside in a clay mineral assemblage which, to a first approximation, is of constant composition, diluted by varying amounts of other non-gamma-ray emitting components. The gradient of the regression line characterises the clay mineral assemblage. Detailed inspection of figure 4.6 reveals the reason for the scatter in the correlation: there is a subtle upward increase in the Th/K ratio from *angulata* Zone times into the overlying Shales-with-Beef. The significance of systematic variation in Th/K ratio will be developed at greater length subsequently in this thesis.

As an aside, the dangers of interpreting only a total gamma-ray curve should be noted (this type of interpretation will be necessary in Chapter 10). The upwards-increase in Th/K ratio in the post-*liasicus* Zone sediments reflects an upwards-increase in Th, with K roughly constant. The increase in Th produces an upwards-increase in total counts which, without the sedimentological log and gamma-ray spectrum, might be interpreted to reflect an increase in shaliness or organic-matter content rather than a change in clay-mineralogical composition.

The Blue Lias spectral gamma-ray data set is somewhat unusual in that there is weak covariance not only between Th and K, but also between Th and U (fig. 4.7b). Small quantities of uranium may be associated with the detrital clay assemblage: Myers (1987) suggested a Th/U ratio for "normal", i.e. low organic-carbon, marine shales in the range 3-5 based on his own measurements (154 samples from British Carboniferous and Jurassic sections, mean $\text{Th/U} = 3.9 \pm 0.7$) and those of Adams and Weaver (1958; 67 samples of various ages, mean $\text{Th/U} = 3.8 \pm 1.1$). The much higher levels of uranium in the Blue Lias ($\text{Th/U} \approx 1$) are likely to be associated with high organic-matter contents and indeed Weedon (1986) reports average organic carbon contents of 2.8% for

the dark marls and 5.7% for the laminated shales. The covariance of U and Th in the Blue Lias may therefore be interpreted as illustrating a link between periods of high terrigenous mud input and high organic matter preservation, as discussed below.

The sequence-stratigraphic interpretation of the Blue Lias is equivocal, turning upon whether the section was "supply-dominated" or "space-dominated" (see Chapter 2 for a general discussion of this issue).

The fully marine fauna with abundant pelagic elements, such as ammonites and ichthyosaurs, locates the section in an offshore marine setting. The carbonate-rich rocks are fine-grained and contain no obvious platform-derived elements such as oolites or reworked skeletal grains. As has been noted, they are organised at a variety of scales (rather than randomly as might be expected with resedimented carbonates) and there is generally no sedimentological evidence for resedimentation (though distinctive sharp-based beds such as "Intruder", fig. 4.6, may have a turbidite origin: see Hesselbo & Jenkyns, in press). Weedon (1986) presents microscopical evidence to suggest that the carbonate-rich rocks originally comprised coccoliths, sedimented as zooplankton faecal pellets. The balance of current evidence appears, therefore, to support a dominantly pelagic, rather than platform-derived, origin for the carbonate.

Lateral persistence of beds at outcrop and lack of evidence for high-energy processes suggest a supply-dominated setting. Weedon (1987) suggested that increased terrigenous supply would dilute the carbonate with siliciclastics and might both introduce organic matter and promote organic productivity through the influx of nutrients, thus leading to dysaerobic bottom conditions (as evinced by the limited bioturbation) and high organic-matter contents. This model is consistent with the suggestion that carbonate-rich intervals were deposited at the lowest rates and also with the correlation between clay content and organic-matter enrichment which is implied by the spectral gamma-ray data.

The outcrop data suggest, therefore, that the Blue Lias was essentially a quiet-water deposit with pelagic carbonate being intermittently diluted by influxes of terrigenous clastics. That is, a supply-controlled model. Under this model thin-bedded, carbonate-rich intervals, notably that within the *angulata* Zone would be interpreted to result from record failure due to starvation. The model is capable of logical extension to the overlying Shales-with-Beef, where a major increase in terrigenous clastic input appears to have occurred. This supply-dominated model must, however, be reconciled with the following observations. Firstly, Donovan et al (1979), based on borehole correlation, demonstrated that the periods of most extensive onlap of the London Platform correspond with the deposition of the more argillaceous units. Donovan et al. suggest that the onlap was controlled by the water depth in which sediment could accumulate, hence the onlapping argillaceous units would be associated with relatively high rates of relative sea level rise, a situation likely to suppress, rather than enhance, argillaceous supply. An alternative is to suggest that onlap was controlled by sediment supply rather than accommodation space (fig. 2.7b). This equates periods of onlap with periods of high argillaceous supply, which in turn might be related to low rates of accommodation space creation in the sediment source area. This interpretation is within the definition of a "downlap surface" as used by Exxon-school stratigraphers (Chapter 2). The Blue Lias - Shales-with-Beef boundary would be interpreted as such a surface by this school (P. R. Vail, pers. comm.).

A second, and unresolved, conflict with the supply-controlled model is that Copestake & Johnson (1989), reviewing Lower Jurassic foraminiferal assemblages, suggested that new taxa and numerical increases in longer-ranging species might be associated with sea-level highstands and extinctions with lowstands. They used these criteria in support of *liasicus* and *semicostatum* Zone flooding and an *angulata* Zone lowstand: the converse of the supply-controlled interpretation advanced above.

Shales-with-Beef

This is the least-studied interval on the Dorset coast, partly because of poor exposure and partly because it is an apparently monotonous mudrock sequence, alleviated only by subtle variations in the carbonate content of the marls, by very occasional nodules and by diagenetic "beef" calcite layers.

The Shales-with-Beef (Lang, et al., 1923) mark a sharp increase in sediment accumulation rates (as measured by the thickness of ammonite zones and subzones) compared with the underlying Blue Lias. As discussed above, this increased accumulation rate is apparently achieved by the dilution of (?pelagic) carbonate by an influx of terrigenous mud. This may be seen on the spectral gamma-ray log (figs. 4.8 & 4.26) by an increase in both Th and K concentrations, though it is masked on the total gamma-ray curve by a sharp decline in uranium counts, reflecting a decline in preserved organic matter content (see above). Speculatively, lower sea-level related to the influx of clastics, may have increased sea-bottom energy levels resulting in increased agitation and more complete oxygenation. Uranium values appear to recover somewhat above "Little Ledge" (fig. 4.8).

No distinct trends in thorium and potassium concentration are apparent over the Shales-with-Beef interval and Th/K correlate closely (fig. 4.9). However, when presented as part of the run of data through the Lower Lias (fig. 4.26), it is apparent that the Shales-with-Beef form part of a trend of increasing Th/K ratios which reflect increasing K but more rapidly increasing Th. This suggests that, in addition to any changes in clay content, there are also systematic changes in the bulk mineralogy of the clay fraction. The generalities of the interpretation of Th/K ratios are discussed in Appendix 1. Burial diagenesis is an unlikely cause of the variation seen in the Dorset Lower Lias as the vertical interval under consideration is only some 150m and the Th/K trend is reversed in the overlying Belemnite Marls. Climatic influence is a possibility: as discussed in Appendix 1, increased Th/K may be associated with intense hydrolysis, a feature of hot and wet climates. Oxygen isotope data (fig. 4.10) tend to argue against this interpretation, suggesting a cooling

climatic trend through the Lower Lias with no reversal during the deposition of the Belemnite Marls. However, these data are not unequivocal: there is much scatter and uncertainty over the extent of a diagenetic imprint and there is no reason to expect a simple relationship between sea temperature and hinterland climate given the complexities of the earth's climate. Myers (1987) suggested changes in clay provenance to explain similar data from the Upper Jurassic. Certainly any satisfactory explanation must take into account the relationship between Th/K ratio and facies. Low Th/K ratios might, in some way, be related to low accumulation rates, but there is no relationship to bed spacing in the Blue Lias, and Th/K ratios tend to increase towards the top of the Belemnite Marls, where bed spacing becomes closer (see below). Perhaps the most attractive explanation is a relationship between Th/K and proximity to source, a model discussed theoretically in Appendix 1 and developed more extensively for the Yorkshire Lower Jurassic in Chapter 5. If proximity is the cause of the Lower Lias Th/K trend then the supra-Blue Lias and supra-Belemnite Marls supply increases would also mark a shift to more proximal facies. This would be a strong argument for supply control on sedimentation: in a space-dominated setting, increases in the rate of space creation would be expected to cause facies belts to retreat rather than to advance.

Black Ven Marls

The relatively high accumulation rates of the Shales-with-Beef are generally maintained into the Black Ven Marls (Lang & Spath, 1926) and there is no marked facies distinction between the two units. There are, however, two important omission surfaces within the Black Ven Marls where key ammonites have not been found: the "Coinstone" horizon, where the *denotatus*, *simpsoni* and *oxynotum* Subzones are not represented, and the "Hummocky" horizon, at the top of the unit, where the *macdonelli* and *aplanatum* Subzones are not represented.

Elevated Th and K contents demonstrated by the spectral gamma-ray data (figs. 4.11 & 4.26) suggest that the Black Ven Marls are somewhat richer in clay relative to carbonate than underlying units. Th/K plot (fig. 4.12a) shows strong covariance with a further incremental increase in the

Th/K ratio over equivalent plots for the Shales-with-Beef and Blue Lias, which according to the model for the proximal-distal significance of the Th/K ratio, advanced above, may be indicative of continued progradation.

Background levels of uranium enrichment are lower than in the Shales-with-Beef (fig. 4.26). However, there is a significant uranium excursion in the *obtusum* Zone, which correlates well with organic carbon measurements over the same interval (fig. 4.13), corroborating the idea that uranium and organic matter are closely associated (Appendix 1). Both Th and K levels are somewhat lowered around the upper *birchi* Subzone and the *obtusum* Zone, suggesting a negative correlation between clay and organic matter accumulation. This is the converse of the relationship found in the Blue Lias and perhaps suggests a different control on organic carbon enrichment.

Smoothing of the data (fig. 4.14) suggests subzone-frequency sediment supply variations similar to those revealed by bed-spacing studies of the Blue Lias. Numerous further examples of such variation will be seen in this work.

The origin of the Coinstone and Hummocky omission surfaces has been a cause of much controversy, Sellwood (1972b) and Hallam (e.g., 1988) favouring winnowing by the interaction of base level with the sediment and Haq et al. (1988) favouring starvation at a time of low sediment supply. Omission surfaces of precisely the same age are known in southern Germany (Chapter 8). The micro-stratigraphy of the Coinstone surface has recently been reviewed by Hesselbo and Palmer (1992). They demonstrate convincing evidence of biogenic erosion and exhumation of the concretions but admit that this does not resolve the issue of supply- versus space-control, which can only be addressed by looking at the surface in context. If the Shales-with-Beef and Black Ven Marls represent progradation, then space-controlled omission would be a logical culmination for the progradational cycle. Interpretation of the Coinstone as the result of lack of accommodation space also enables stratigraphic events in Dorset and Yorkshire to be reconciled. These issues will be further explored below.

Belemnite Marls

Above the Hummocky horizon, the Belemnite Marls (Lang et al., 1928) mark a sharp return to decimetre-scale limestone-marl interbeds and relatively low sedimentation rates, similar to those of the Blue Lias, the principal differences being greater differentiation of the carbonate-rich beds in the Blue Lias and the presence of scours filled with bioclastic debris in the Belemnite Marls, interpreted as due to storm action by Sellwood (1970). As with the Blue Lias, vertical organisation at a number of scales may be demonstrated: 37.5 cm couplet cycles grouped into variable length (191-705cm) bundles have been described by Weedon & Jenkyns (1990), and the limestone marl couplets clearly tighten towards the top of the unit (fig. 4.15).

Again, as with the Blue Lias, the key problem from a sequence-stratigraphic viewpoint is supply-versus space-limitation of sediment accumulation. Couplet tightening suggests space control as one would expect declining supply to affect the relative proportions of the terrigenous and carbonate components. Space-control is further supported by the evidence for winnowing. The couplet tightening culminates in the Belemnite Bed (fig. 4.16a), which would, in a space-controlled setting, be predicted to be the product of winnowing. It must be noted, however, that the features of this bed rather tend to suggest a low-energy starvation surface: no strong preferred belemnite orientation (fig. 4.16b), phragmacones often intact and common pyrite.

Considering the spectral gamma-ray data (fig. 4.17), there is a subtle upwards increase in Th which is not matched by an upward increase in K and results in an upward increase in Th/K ratios. This supports Walker's (1991) observation of an upwards increase in the kaolinite/illite ratio in the Belemnite Marls which he interprets as reflecting increasing proximity. This association between increased proximity and condensation further supports a space-controlled hypothesis. Despite upward-increasing thorium, uranium counts tend to decline upwards, (with significant positive excursions in the *polymorphus-brevispina* Subzones). Hence Th/U ratios also decline upwards, interpreted as a reduction in preserved organic carbon (see discussions above and Appendix 1).

Green Ammonite Beds

The Belemnite Marls are overlain by a stratigraphically expanded and monotonous mudrock sequence known as the Green Ammonite Beds (Lang, 1936); the relationship between these mudrocks and the underlying limestone-marl unit being analogous to that between the Shales-with-Beef and Blue Lias.

Total gamma-ray counts increase upwards through the Green Ammonite Beds (fig. 4.19), interpreted to reflect an increasing terrigenous supply (at the expense of carbonate) through the unit: the Th/K ratio remains constant. Interestingly, there is a suggestion in the data that this upward increase in terrigenous content occurs in two cycles, the top of the first cycle being near the "Red Band", the most distinctive marker bed in a section otherwise devoid of key horizons.

Th/K ratios are somewhat higher than those in the Black Ven Marls, but the Th/K crossplot (fig. 4.20) shows that this is due to a positive Th-axis intercept, rather than a steeper slope: the slope of the crossplot is actually very similar to that for the Shales-with-Beef (fig. 4.9). This observation suggests the presence of an independent Th-rich component, an observation which holds for both the carbonate-poor post-ibex Zone sediments of Dorset and the carbonate-poor sediments of Yorkshire (Chapter 5). The site of the thorium is unknown, though a significant heavy mineral component appears unlikely in rocks of this fine grain size. The change may mark the tapping of an alternative source of sediment, presaging the arrival of coarse clastics in the Middle and Upper Lias.

Three Tiers to Junction Bed

This interval (fig. 4.26) represents a greater thickness, some 125m, than those discussed above. However, ammonite subzones suggest that it represents a similar time span. The section is difficult of access, cropping out mostly high in the cliffs of Golden Cap and Thorncombe Beacon,

so only limited spectral gamma-ray data have been obtained. Fortunately, these data are less critical than in the Lower Lias as the introduction of coarse siliciclastics makes lithological distinctions less subtle and the section appears relatively straightforward to interpret in sequence-stratigraphic terms.

The Three Tiers (fig. 4.21a) mark the first coarse siliciclastics deposited in the basin during the Jurassic. They are three grey, metre-scale beds of carbonate-cemented argillaceous siltstone (fig. 4.21b). They show no vertical grainsize trend and contain common *Thalassinoides*. The Three Tiers are overlain by the Eype Clay. Unfortunately no spectral gamma-ray data were obtained more than 15m below the Starfish Bed (fig. 4.26), so it is not possible to draw firm conclusions about lithological trends in the lower part of the Eype Clay, between the Three Tiers and the Eype Nodule Bed (the prominent nodular horizon shown some 40m above the Three Tiers). However, it is clear that by the upper part of the Eype Clay, for which gamma-ray data are available, total counts have declined significantly, reflecting:

1. Reductions in Th, K and U, interpreted to result from increasing dilution by coarse detrital minerals, especially quartz.
2. Slight reduction in Th/K, which may reflect the clay-mineralogical changes discussed above, but may also be complicated by the introduction of K-rich detrital phases such as feldspar.
3. Constant but high (≈ 5) Th/U ratios suggesting little organic enrichment and small amounts of uranium associated with the clay fraction.

The Starfish Bed is the lowest bed of the Down Cliff Sands. It is a fine-grained, micaceous siltstone, similar to one of the Three Tiers and distinctive due to the preservation of ophiuroids on its base (Ensom, 1984; Goldring & Stephenson, 1972). Approximately 1m beneath is a shell concentration known as Day's Shell Bed. Hesselbo (pers. comm.) has suggested that this may indicate winnowing caused by base-level lowering which preceded the influx of coarse-grained

sediment which now forms the Down Cliff Sands. Initial influx must have been sudden, perhaps storm driven, so as to overwhelm the ophiuroid colony.

Above the Starfish Bed, the Down Cliff Sands are heterogeneous: grey with wispy cross-lamination defined by alternations of silt and very fine sand: evidence for the higher-energy processes associated with "space-dominance". They are capped by the Margaritatus Stone, a bioclastic grainstone dominated by bivalve fragments and calcareous algae with some quartz silt. Similar lithologies are found on top of the other principal sand packages in the sequence: the Thorncombe Sands (Thorncombiensis Bed) and the Bridport Sands (Scissum Bed). These high-energy carbonate lithologies are interpreted as representing condensation due to lack of accommodation space. Both the Margaritatus Stone and the Thorncombiensis Bed are overlain by mudrocks, interpreted to represent renewed deepening.

Renewed accommodation space creation following deposition of the Margaritatus Stone is indicated by some 2m of mudrock known as the Blue Band. The overlying Thorncombe Sands are yellow very fine-grained sandstones, locally with low-angle cross lamination. Hummocky cross-stratification has been described by Sellwood et al. (1970). A number of concretionary horizons are present, which may represent pauses in accumulation, most notably the large "doggers" near the top of the unit which may have developed whilst the sediment-water interface remained static during Thorncombiensis Bed times.

The Marlstone/Junction Bed is a complex unit of condensed, but biostratigraphically complete limestone (Jackson, 1922; Jackson, 1926). The lower part, the Marlstone, comprises the *spinatum* and *tenuicostatum* Zones and includes high energy elements, especially oolites. The upper part, the Junction Bed in the strict sense (*falciferum* Zone to *dispansum* Subzone), comprises nodular limestones. A distinct mudrock interval is present between the Thorncombiensis Bed and the Marlstone/Junction Bed which, unless significant amounts of stratigraphy have been lost, places the Marlstone/Junction Bed firmly in the context of an interval of condensation due to starvation. The elapsed time and evidence of high energy processes suggests, however, greater

complexity. It is possible that *margaritatus* Zone flooding left this part of the basin starved of clastics such that a *spinatum* Zone regression was marked only by reworked oolites (the Marlstone). The Junction Bed may represent maximum flooding in the same, clastic-starved setting.

Down Cliff Clay and Bridport Sands

The Down Cliff Clay demonstrates the resumption of clastic deposition in *levesquei* Subzone times. The Down Cliff Clay passes transitionally upwards into the Bridport Sands (fig. 4.22). These are very fine-grained argillaceous sandstones with prominent carbonate-cemented bands which are thought to reflect slight primary differences in sediment texture (Bryant et al., 1988). Grainsize measurements (Hesselbo & Jenkyns, in press) demonstrate that the most marked increase in grainsize occurs at bed 4. The beds, as defined by the cemented bands, tighten towards the top of the unit, suggesting progradation.

The Bridport Sands are, predictably, marked by lower total gamma counts than underlying finer-grained rocks. Once again, we see that Th falls disproportionately to other elements, so that not only do Th, K and U counts all decline, but Th/K and Th/U ratios also decline (fig. 4.22). Although strictly beyond the scope of this thesis, the Inferior Oolite is interesting in that a fall in total counts (resulting from a fall in Th and K) masks a marked rise in U counts. The uranium may be associated with phosphate enrichment in this unit.

Synthesis & Conclusions

Cyclicity is evident in the Dorset stratigraphy at a wide variety of scales. This synthesis will focus on cycles and events which involve more than a single subzone and hence have some potential for regional correlation. It should be noted, however, that these cycles unfold against a background of Milankovich-scale variability and that the spectral gamma-ray studies presented above show that

the ammonite subzones may themselves represent sedimentary cycles of approximately 0.5 million years duration. The key events under discussion are summarised in figure 4.24.

The Hettangian, Sinemurian and Early Pliensbachian are developed in mudrocks with moderate to high contents of fine-grained carbonate. Periods of high carbonate content and low deposition rate (Blue Lias, Belemnite Marls) alternate sharply with periods of low carbonate content and high deposition rate (Shales-with-Beef and Black Ven Marls, Green Ammonite Beds). Within the Blue Lias, the *angulata* Zone is marked by very low accumulation rates and may represent a hiatus: higher accumulation rates (as measured by bed spacing rather than zonal thickness) are found in the *liasicus* and late *bucklandi* Zones. Within the Black Ven Marls, significant biostratigraphic hiatuses are present at the top of the *obtusum* Zone and at the top of the *raricostatum* Zone. Accumulation rates decline towards the top of the Belemnite Marls, where there is a belemnite concentration, the Belemnite Bed, of *valdani* Subzone age. The key issue, from a sequence-stratigraphic viewpoint, is whether these changes in accumulation rate represent changes in the rate of accommodation space creation or changes in the rate of sediment supply. If the former, then intervals such as the Shales-with-Beef represent periods of rapidly rising sea-level. If the latter, then intervals such as the Shales-with-Beef might represent falling sea-level or changes in some other factor critical to sediment supply, such as climate.

Sedimentological criteria do not readily resolve this dilemma. There is little evidence of high energy processes operating, but this may be due to the unavailability of coarse grainsizes rather than an absence of accommodation-space control. Conversely, winnowing in muds may be an inevitable consequence of low supply rates and does not necessarily imply the impingement of wave-base processes. The balance of the petrographic evidence suggests a pelagic origin for the carbonate which, if true, leads to the interpretation of the correlation between carbonate-rich and relatively condensed sediments in terms of supply dominance, with pelagic carbonate periodically diluted by terrigenous input. The use of Th/K as a proximal-distal indicator has been suggested in this chapter and will be developed subsequently. If this is a valid approach to the interpretation of these data then further general support is provided for the supply-dominated model: the

Belemnite Marls and Blue Lias have lower Th/K ratios than the mud-rich intervals and would therefore be interpreted as more distal. However, the correlation between bed-tightening and progradation in the Belemnite Marls has been discussed above and tends to favour a space-controlled interpretation for that interval.

The model illustrated in figure 4.25 (and used to construct figure 4.24) is an attempt to reconcile these observations. The Blue Lias is interpreted as unequivocally supply controlled. Relative sea-level fall is then suggested as a mechanism to increase terrigenous supply for deposition of the Shales-with-Beef and the lower part of the Black Ven Marls, which form a progradational mudrock sequence infilling the available space and leading to the impingement of wavebase at the Coinstone horizon. As relative sea level rises once more, mudrock deposition resumes in *raricostatum* Zone times, but flooding at the end of the Sinemurian leads to record failure due to starvation (the Hummocky). The Belemnite Marls are interpreted to result from renewed progradation, but into less space and, perhaps linked to the end Sinemurian flooding episode, less terrigenous supply.

Renewed flooding is interpreted to have taken place at the top of the Belemnite Marls, leading to the creation of accommodation space into which the coarsening-upwards cycles of the Middle and Upper Lias were able to prograde.

Chapter 5. The Cleveland Basin

The Cleveland Basin (Dingle, 1971) is a Mesozoic depocentre bounded to the west and north by the Paleozoic rocks of the Pennines and to the south by the attenuated Jurassic sequences of the Market Weighton High (fig. 5.1a). Offshore and subsurface data now enable this depocentre to be placed in context and it can be seen to form the north-west corner of the Southern North Sea Basin, with perhaps 4km of Jurassic section offshore in the basin centre (fig.5.1b). The extent to which the Pennine - Mid-North Sea High was actually emergent during the Jurassic are unclear. However, paleocurrent and facies studies in both the Middle Jurassic (Alexander,1986) and Lower Jurassic (Howard,1985) suggest sediment derivation from the north and west. Proximal-distal relationships suggested by spectral gamma-ray data from the present study support this supply direction for the Lower Jurassic (see below).

As with the Dorset sections (Chapter 4), continuous exposure is available from coastal sections exhumed by late Cretaceous and Tertiary inversion. Ammonite biostratigraphic control is excellent (Cope et al., 1980), enabling very detailed comparison of events with the Wessex Basin, some 450km to the south. Again, the interpretations below are principally based on recent stratigraphic logs by Hesselbo & Jenkyns (in press), who also summarise the long history of interpretation of these strata, dating back to Tate and Blake (1869) and Simpson (1884). Hesselbo & Jenkyns' logs have here been supplemented by new spectral gamma ray measurements and sedimentological observations. The sections discussed may be located using figure 5.2a. Rawson & Wright (1992) provide practical details of access.

Significant spectral gamma-ray work has already been undertaken on the Yorkshire Lower Jurassic: Myers (1987, 1989) logged the Cleveland Ironstone, Grey Shales and Jet Rock at Staithes (fig. 5.1a), Owen (1990) logged the Staithes Sandstone to Alum Shales, using the same section as Myers, together with the exposures between Saltwick Bay and Whitby. Both used

dominantly bedding-plane measurements. Owen's data are unpublished and I am grateful for her permission to reproduce them here. Van Buchem (1990) logged the Calcareous Shales to "Banded Shales" at the localities of this thesis. Despite these previous measurements, it was considered to be worthwhile producing a unified data set covering as much as possible of the Lower Jurassic using a consistent approach and tied to new, detailed sedimentological logging.

The format of this chapter is the same as for the Wessex Basin: a description of the principal lithostratigraphic units (fig. 5.26) followed by a synthesis. Lithostratigraphic nomenclature follows Powell (1984) except for the adoption of the recommendations of Rawson & Wright (1992) on the nomenclature of the Mulgrave Shale.

Calcareous Shales to Siliceous Shales

Robin Hood's Bay (fig. 5.2a) forms the core of an anticline. The oldest rocks in the area are therefore exposed in the foreshore in the centre of the bay. They are dark grey calcareous mudstones with occasional decimetre-scale limestones, assigned to the *semicostatum* Zone of the Sinemurian. Subsurface data (Ivimey-Cook & Powell, 1991) suggest that similar facies extend to, and beyond, the base of the Jurassic. These "Calcareous Shales" (Buckman, 1915) pass transitionally upwards into more silty sediments, the "Siliceous Shales". The Siliceous Shales contain numerous scours (fig. 5.2b), especially in the *oxynotum-raricostatoides* Subzone interval. Scour bases may preserve a crinoid lag. Scour fills commonly comprise discordant bundles of climbing-ripple laminated sand. More persistent decimetre-scale carbonate-cemented thin sands also occur. Sellwood (1970) believed that these defined the tops of 1.4-3.6m coarsening- and shallowing-upwards cycles ("parasequences"). Certainly the facies suggest intermittent impingement of storm wavebase (van Buchem et al, 1992). Beds 70 and 72, strictly within the overlying Pyritous Shales sequence, are ferruginous and resemble a weak development of the Cleveland Ironstone facies (see below).

Spectral gamma-ray data for the Calcareous Shales (fig. 5.3) are somewhat unreliable, as they were obtained from ledges in the foreshore. The upward increase in total counts into the Siliceous Shales is, however, corroborated by the subsurface data and by Van Buchem (1990), who extended his log stratigraphically downwards using the Calcareous Shales exposure at Redcar. Th/U ratios are high (≈ 4 -5, fig. 5.3 & 5.4b), suggesting the absence of organic matter enrichment (Appendix 1). Th and K co-vary (fig. 5.4a) and Th/K ratios show no systematic trend, suggesting that the source of the Th and K signals is of constant composition and simply varies in amount. This observation tends to exclude both changes in clay mineralogy and the introduction of a radioelement-bearing heavy mineral phase as the origin of the upwards-increase in total counts. The trend is best explained by an increase in the proportion of terrigenous mud to carbonate from the Calcareous to the Siliceous Shales.

The Th/K crossplot (fig. 5.4a) shows a positive intercept on the Th axis, a feature of most of the Yorkshire dataset (except for the Grey Shales/Jet Rock, which may be too distal, see below). Myers (1987), in his study of the Upper Pliensbachian and Lower Toarcian parts of the section, suggested that this might be due to high levels of kaolinite (recorded in the Middle Lias mudstones by Catt et al., 1971), although high levels of kaolinite in the Pyritous Shales do not appear to be associated with high Th/K (Van Buchem et al., 1992).

The Siliceous Shales, then, represent a period of increased terrigenous input and, at least intermittently, higher energy conditions when compared with the underlying Calcareous Shales. Variations in deposition rate in the Yorkshire Lower Jurassic are not so marked as in the Dorset section. However, it appears from the relative thickness of the ammonite subzones that the coarser-grained intervals (including the Siliceous Shales, fig. 5.26) accumulated more slowly than the finer-grained intervals. This supports an interpretation of relative sea-level control whereby sea-level falls, and not only stimulates terrigenous supply but also restricts the space available for accumulation. A simple increase in terrigenous input would be expected to increase accumulation rates.

Pyritous Shales and Ironstone Shales

The Pyritous Shales (Buckman, 1915) are dark grey mudrocks with ferruginous concretions (fig. 5.5a) which Van Buchem (1990) suggests might represent cemented silt lenses. Total gamma-ray counts increase into the Pyritous Shales, largely as a result of U enrichment (figs. 5.6 & 5.26). There is a subtle increase in K counts and a decrease in Th counts which results in a sharp decrease in the Th/K ratio. Th/U ratios are not low enough to indicate organic matter enrichment unequivocally (Appendix 1), but they are significantly higher than in either the Siliceous Shales beneath or the upper Ironstone Shales to Grey Shales above. Th/K ratios increase upwards through the Pyritous Shales, a systematic increase which continues into the overlying Ironstone Shales.

The Ironstone Shales may be divided into two distinct units. The lower unit (the "Banded Shales" of Van Buchem et al., 1992) comprises light and dark mudstone interbeds, strongly resembling the Belemnite Marls of Dorset, though with occasional ferruginous concretion horizons (fig. 5.5b). The U-enrichment levels of the Pyritous Shales continue through the Lower Ironstone Shales. The light and dark mudstone interbeds of the lower Ironstone Shales tighten towards a silty bed rich in belemnites (the Belemnite Bed, fig. 5.8). Th/K ratios increase towards the Belemnite Bed. Given that the bed is rather silty it was considered possible that the elevated Th/K ratios could be due to the introduction of a Th-bearing heavy mineral phase rather than a change in clay mineral chemistry. Only small quantities of such minerals might be needed: Fertl (1979) reports concentrations of Th in zircon in the range 100-2500ppm, so that 3ppm Th in the rock might be produced by 0.1% zircon. This possibility was investigated by heavy mineral analysis (Appendix 3). The rock proved to contain approximately 10% (wt/wt) of grains >53 μ m, mostly quartz and feldspar. Approximately 1% of the coarse fraction (i.e. 0.1% of the total rock) was "heavy" relative to tetrabromoethane (s.g. 2.9), but thin-section inspection showed this to comprise almost

entirely pyrite and authigenic barite. A few tourmalines but no zircons were found, thus excluding the possibility of significant influence by heavy minerals upon the spectral gamma-ray results.

The upper part of the Ironstone Shales comprises relatively monotonous mudstones with occasional ferruginous concretions, though Hesselbo & Jenkyns (in press) detect silt-free/silt-rich cycles at a parasequence scale. Total gamma counts decline through the Ironstone Shales towards the Staithes Sandstone (fig. 5.26), but this cannot simply be explained as a result of reduced clay content. Uranium declines sharply at the top of the lower Ironstone shales and K declines more rapidly than Th, resulting in an upwards increase Th/K ratios. This fits with the lithostratigraphic observations: whilst the passage to the Staithes Sandstone is transitional, it is not transitional over the entire thickness of the Ironstone Shales, as might be imagined from inspection of the total gamma-ray curve. Rather, there is a clear maximum in Th/K at the Belemnite Bed, followed by a decline to a point near the top of the Ironstone Shales and then a rise into the Staithes Sandstone.

If the Siliceous Shales represent higher-energy deposition during a relative sea level lowstand, then the above observations enable the construction of a robust sequence-stratigraphic model for the Pyritous Shales - Ironstone Shales sequence. The facies of the Pyritous Shales are compatible with accelerated rates of accommodation space creation which push facies belts shoreward (Sellwood, 1972) and allow more rapid accumulation rates: the Pyritous Shales may have accumulated entirely within the *taylori* subzone. The Th/K ratio is lower in the lowermost Pyritous Shales than in the Siliceous Shales, compatible with more distal deposition. Th/K rises through the Pyritous Shales and Ironstone Shales suggesting progradation. Beds tighten and ammonite subzones reduce in thickness as the rate of accommodation space creation is reduced, culminating in the condensed and sandy Belemnite Bed where, it is suggested, an omission surface below biostratigraphic resolution may be present. Th/K ratios suggest that the upper Ironstone Shales represent a renewed phase of increased space creation prior to the development of the Staithes Sandstone.

It should be noted that the above model is critically dependent upon the Th/K data. Without these, the Belemnite Bed could be explained as a result of reduced sediment supply rates and winnowing due to starvation. That is, one could interpret one simple cycle of base level change from low rates of space creation during deposition of the Siliceous Shales, through high rates of space creation reaching a maximum at the Belemnite Bed, to low rates of space creation for the Staithes Sandstone. The correspondence between the lower Ironstone Shales of Yorkshire and the Belemnite Marls of Dorset (Chapter 4) is remarkable. Both comprise a tightening-up interval of more- and less-marly interbeds culminating in a belemnite concentration of *valdani* Subzone age. Both exhibit an upwards-increase in Th/K. It is difficult to justify radically different sequence-stratigraphic interpretations for the two units and this is a powerful argument for accommodation rather than supply control on Belemnite Marl accumulation rates (at least as high in the stratigraphy as the Dorset Belemnite Bed).

Staithes Sandstone and Cleveland Ironstone

Immediately north of Robin Hood's Bay, where it was logged for this study, the Staithes Sandstone (fig. 5.12a) comprises dominantly bioturbated silt with beds and lenses of fine-grained sandstone (Howard, 1985). Concretionary horizons, locally ferruginous, are also present. The sandstones fill scours and form decimetre-scale beds. Sharp erosive bases are common, in places with shell hash. The sandstones are commonly internally laminated, the laminations being arranged into discordant, climbing bundles with many of the features of oscillation ripples (de Raaf et al., 1977). Sandstone beds occasionally fine upwards. The facies strongly suggests the intermittent impingement of storm wavebase. Sandstones may be grouped into metre-scale bedsets (most notably beds 59/60 & 6, figs. 5.12 & 5.13) which may define intervals of more frequent storms. Spectral gamma-ray data (figs. 5.13 & 5.14) show, as expected, low total counts due to the low clay content. Th/K ratios are high, consistent with the interpretation of these ratios

as a proximal-distal index. The highest Th/K ratios are attained around "bed" 6, near the base of the *stokesi* Subzone.

The Cleveland Ironstone may be divided into three units. The base of the stratigraphically lowest unit is marked by a sharp reduction in grain size from the underlying Staithes Sandstone. The unit then coarsens up and culminates in the Avicula Seam: a band of nodular, ferruginous limestone. The second unit of the Cleveland Ironstone is similar to the first, coarsening upwards to the Raisdale Seam. The top of this cycle is, however, somewhat coarser and includes rippled sand lenses, some minor scouring and an horizon of gutter casts. Coarsening-upwards in the uppermost unit of the Cleveland Ironstone is less distinct in the field, but an upward decrease in clay content is evident from the spectral gamma-ray log (fig. 5.15). *Chondrites* are locally abundant in this unit and belemnites are common.

The presence of fine-grained sediments within the Cleveland Ironstone (and, except for the ironstones, reduced Th/K ratios), suggests that this unit is transgressive relative to the underlying Staithes Sandstone. The association of the ironstones with the coarser parts of the cycles and the proximal passage to thicker and richer ironstone development (see below) suggests that the ironstones were associated with regressive pulses within this overall transgressive trend. In this connection it should be noted that Chowns (reported in Cope et al., 1980, see also Howard, 1985) has proposed an unconformity of basin-wide significance, with its correlative conformity at the *gibbosus-apyrenum* Subzone boundary, i.e. near the base of the Pecten Seam. This is synchronous with the base of the Marlstone in Dorset.

Comparison of the spectral gamma-ray log for the Staithes Sandstone/Cleveland Ironstone with that obtained by Owen (1990) from the same interval at Staithes, some 20km to the north west, is instructive (fig. 5.17). Owen used the same instrument with the same calibration settings as used in this study, so the results should be directly comparable. The Staithes Sandstone at Staithes has somewhat higher Th/K ratios than at Hawsker Bottoms, where it was studied for this thesis.

This suggests the same proximal-distal relationships as the sand/mud ratio (Howard, 1985). The ironstones are more thickly developed in the Staithes area and all register the very high (>20) Th/K ratios noted by Myers (1987) and interpreted by him as a result of Th-enrichment associated with lateritic soils in the source area of the iron-enriched material. At Hawsker Bottoms the ironstones are much less well-developed and only the Pecten Seam registers these very high values (Th/K=13), other "ironstones", notably the Avicula Seam give the low total counts and background Th/K ratios of normal limestones. Again, high Th/K values correspond with facies which would be interpreted as more proximal using other lines of evidence.

Grey Shales to Peak Mudstone

The subtleties of the lithological distinctions in the Toarcian mudrocks have contributed to an unstable and bewildering lithostratigraphic nomenclature for this interval. The terms used here are as shown on the logs of Hesselbo & Jenkyns (in press), who follow Knox (1984), Powell (1984) and Rawson & Wright (1992). A further complication arises in that a complete Lower Jurassic section can only be assembled by crossing the Peak Fault (fig. 5.1a). This throws down to the east and Jurassic sediments are likely to thicken on the downthrown side (Milsom and Rawson, 1989), a factor of great relevance when considering sediment accumulation rates. The cross-fault correlation used here is based on Howarth (1962).

The Grey Shales to lowermost Bituminous Shales were logged on the west side of the Peak Fault. Total gamma-ray counts increase into the Grey Shales due to the increase in clay content compared with the Cleveland Ironstone (figs. 5.18 & 5.26). The Jet Rock is marked by significant uranium enrichment, as would be expected from its high organic matter content (Myers, 1987, measured up to 20% organic carbon). The Jet Rock is also marked by very low Th/K ratios, suggesting a distal setting. Owen (1990) noted that the fall in Th counts tends to compensate for the rise in U counts so that this important event is not evident from the total gamma-ray log: commonly the only measurement available from a subsurface data set (compare Chapter 10). The

lowest Th/K ratio is found near the top of the Jet Rock (the top of the *exaratum* Subzone). The peak of U-enrichment is somewhat more diffuse but is essentially coincident with the Th/K minimum.

High levels of U enrichment and low Th/K ratios continue into the Bituminous Shales where the log was continued on the east side of the Peak Fault (fig. 5.20). In the overlying Alum Shale, U counts and Th/K ratios return to the levels found in the Grey Shales (fig. 5.26) . This signature continues through the Peak Shales (fig. 5.22). Hesselbo & Jenkyns (in press) note the presence of more marly interbeds in the lower part of the Peak Shales. The carbonate cement may reflect greater permeability consequent upon the somewhat coarser grainsizes reported for this interval by Knox (1984) This may be a subtle reflection of a wider *variabilis* Zone regressive event (Chapter 11).

Th/K ratios in the Toarcian section logged here remain somewhat higher than those in the more proximal section at Staithes (fig. 5.16), again supporting a NW-SE trend from proximal to distal settings.

Peak Mudstone to Dogger

No spectral gamma-ray data were obtained over the lower part of this section. Quantitative grainsize and clay-mineralogical data are, however, provided by Knox (1984, fig. 5.24). These data clearly show the coarsening-upwards trend from the upper part of the Peak Mudstone to the Yellow Sandstone. Knox's clay-mineralogical data are interesting the light of the limited Th/K data which have been recorded, and which show an upwards increase in Th/K ratio through the interval (fig. 5.26). The Th/K trend is consistent with the empirical model (for mudrocks) that Th/K ratios increase with more proximal settings, but in this case the result is clearly not due to an association between high Th/K ratios and kaolinite: kaolinite content decreases from values of around 50% in the Peak Mudstone to around 10% in the upper part of the Blea Wyke Sandstone. One possibility

is the upwards-increasing proportion of chlorite, thought by Knox to represent an original input of bertherine mud. Hassan et al. (1976) report Th/K values for chlorite similar to those for kaolinite. With sandstones, of course, the complicating effects of feldspar and heavy mineral content must also be considered.

The upper part of the Blea Wyke Sandstone (the Yellow Sandstone Member of Knox, 1984) is strongly bioturbated, contains abundant belemnites and is compatible with a lower shoreface environment of deposition. There is a sharp contact, marked by a shelly lag, with the overlying Middle Jurassic Dogger Formation sandstones. On the western side of the Peak Fault the Dogger rests directly upon the Alum Shale (Rawson & Wright, 1992). There is, therefore, a significant structural unconformity at the base of the Dogger: the mid-Cimmerian event of Underhill & Partington (1993). The unrecorded time at Blea Wyke, the section relevant to this thesis, is bracketed by the discovery of *moorei* Subzone fossils in the Yellow Sandstone Member (Cope, Getty et al., 1980a) and *opalinum* Zone fossils in the Dogger (Cope, Duff et al., 1980).

Synthesis & Conclusions

The accumulation of most of the Yorkshire Lower Jurassic appears to have been limited by the availability of accommodation space, rather than the availability of sediment supply. Periods of relatively slow space creation, when facies belts were allowed to prograde, were marked by relatively slow accumulation rates (as measured by ammonite subzones), coarser-grained sediments, high energy processes and high Th/K ratios. These alternated with periods of relatively rapid accumulation, marked by mudrocks with low Th/K ratios. The boundaries of these cycles tend to be transitional, with few grounds for the precise placement of sequence boundaries in the sense discussed in Chapter 2.

The lowermost cycle, Calcareous Shales-Siliceous Shales-Pyritous Shales (fig. 5.25) is clearly defined by both the facies and Th/K evidence. The next cycle, Pyritous Shales-lower Ironstone

Shales-upper Ironstone Shales, is more contentious and rests strongly on the Th/K evidence, which argues against the Belemnite Bed being a starvation surface despite its basinal, "condensed" aspect.

The upper Ironstone Shales-Staithes Sandstone-Grey Shales is another unequivocal cycle, although the Th/K data are again important to the interpretation of the Cleveland Ironstone as part of the transgressive limb, with regressive pulses giving rise to the individual ironstone seams.

The most distal facies appear to be associated with the *falciferum* Zone, and specifically with the top of the *exeratum* Subzone.

Spectral gamma-ray data from the Yorkshire Jurassic illustrate how complex can be the origins of the total gamma-ray count, and in particular demonstrate the importance of changing Th/K. There is a strong correlation between Th/K and distal/proximal relationships as interpreted from facies and this relationship may be extrapolated to provide useful insights into the sequence stratigraphic evolution of this and other sections. Comparison between the detailed interpretation outlined above and the total-count gamma-ray curve provides an excellent case-study of the problems of interpreting subsurface gamma-ray data: in particular the subtleties of the Lower Pliensbachian interval are largely obscured on the total-count curve and the signal from important organic-rich Jet Rock interval is suppressed by the coincidence of high U-concentration and low Th/K. Where clay mineralogical data have been obtained, for the Lower Pliensbachian and Upper Toarcian, they show that a simple model of Th/K as a proxy for kaolinite/illite is not adequate to explain the spectral gamma-ray data and much further work on both the sites of radionuclides in fine-grained sediment and the actual distribution of radionuclides in modern systems will be required before we can make unequivocal interpretations of spectral gamma-ray data sets.

Chapter 6. The Lusitanian Basin

The Lusitanian Basin is a roughly north-south trending rift which extends both onshore and offshore western Portugal (fig. 6.1). Lower Jurassic strata are usually buried to depths of several kilometres, but a small number of fragmentary exposures are available, due largely to the effects of diapirism in Triassic salt. The most stratigraphically complete of these exposures is at Peniche, some 70km north of Lisbon (fig. 6.2). Virtually continuous coastal exposure is available here and it is this section which will be considered below. It should be noted that the Triassic-Jurassic Lusitanian Basin was only some 100km wide in the area of Peniche and that the outcrop lay perhaps 10km from the western margin of the basin, currently demarcated by the basement outcrop of the Berlenga Islands (fig. 6.2). Hallam (1971) suggested a westerly provenance for at least the Toarcian terrigenous clastic sediments and this has been supported by subsequent workers (Wright and Wilson, 1984).

The Peniche section is summarised in figure 6.42. The sedimentology of the lowest beds seen, the Coimbra Formation, has been studied in detail by Watkinson (1989). They are capped by a prominent hardground which separates them from the overlying Brenha Formation. The lowermost 230m of the Brenha Fm. includes a variety of distinctive lithostratigraphic units informally labelled A-G in this study, and ranging in age from the *raricostatum* Zone (latest Sinemurian) to the *bifrons* Zone (mid-Toarcian). The sedimentology of this interval has not previously been studied in detail, save for a description by Hallam (1971). The biostratigraphy is, however, well known due largely to the work of Mouterde (1955). The mid- to Late Pliensbachian interval contains an especially rich ammonite fauna described most recently by Phelps (1985). The post-*bifrons* -Zone section comprises some 250m of carbonate turbidites which have been studied sedimentologically by Wright & Wilson (1984). Their age is poorly known.

In contrast with the British sections, detailed stratigraphic logs were not available. Basic description of the section therefore assumes much greater priority in this than in other study areas. Work up to and including the *bifrons* Zone is entirely the responsibility of the author. Work above the *bifrons* zone was undertaken jointly with S. P. Hesselbo and extends beyond the scope of this thesis in providing excellent three-dimensional exposure of a mixed carbonate-siliciclastic submarine fan. Outcrops may be located using figure 6.3.

Belemnites were collected against the new logs wherever they were available. The initial objective of this was to test the long-range applicability of the strontium-isotopic curve for Lower Jurassic seawater developed by Jones (1992) from English belemnite material, and to use Jones' curve to provide an alternative method of correlation where ammonites are scarce, particularly near the base of the Peniche stratigraphy. A short section below summarises the results. Details of the strontium-isotopic analyses will be found in Appendix 2. During the course of the present study the potential importance of belemnite-derived carbon isotope data to the overall interpretation became apparent, and carbon and oxygen isotope analyses were performed on the Portuguese belemnites. This aspect of the study is developed in Chapter 12.

Spectral gamma-ray measurements were made in the mudrocks in order to elucidate fine-scale cyclicity.

Coimbra Formation

The top of the Coimbra Formation is marked by a prominent hardground which is best seen on the west side of the Papoa headland (fig. 6.3 & 6.4a). The hardground may be traced eastwards, across the headland, via a series of small quarries and is well exposed again in the small cove to the north-west of the ruined Forte da Luz (fig. 6.3 & 6.4b). The Coimbra Formation also forms the Ilhéu da Papoa (fig. 6.3), which is separated from the "mainland" by an agglomerate vent which utilises a fault. The base of the Coimbra Formation is unseen.

Lithologically, the Coimbra Formation is dominated by medium to thickly bedded, locally tabular cross-laminated, peloidal grainstone (fig. 6.5a). The peloids are fine-grained and tend to be "over-packed", with fine prismatic coatings and a drusy spar matrix. Some fine-grained angular quartz fragments are present, locally forming the cores of peloids (fig. 6.5b). Bioclasts, principally bivalves, are common in some beds, especially towards the tops. Some beds grade up into coarse bioclastic grainstone. Occasional beds of soft, yellow, very fine grained sandstone are present with corroded and angular quartz grains and some fragmented bioclasts (fig. 6.5c).

The bivalves suggest open marine conditions, although, as they are fragmented, washover fan facies cannot be excluded. The peloids with quartz cores, and the cross-lamination, suggest a high energy shoal. Clearly, there was intermittent access to terrigenous material. These conclusions are broadly in line with those of Watkinson (1989), who interpreted a complex of sub- and inter-tidal environments on the basis of a more comprehensive petrographic and sedimentological study of the Formation.

In the lower part of the Coimbra Formation, thick beds of peloidal, bioclastic or micritic limestone appear to alternate randomly, with variably developed silty partings. Near the top, however, distinctive cycles spanning several metres can be recognised (figs. 6.6 & 6.7). Beds thicken and clean up through bioclastic to peloidal limestones. The last, and best-developed, of these cycles has a sandstone at its base and culminates in the top-Coimbra hardground. Speculatively, these cycles may reflect changing rates of accommodation space creation. The lowest rates of space creation, at the base of the cycles, may have exposed the carbonate shoals and promoted the influx of terrigenous material. Subsequent accelerated space creation rates would have lead to increased rates of accumulation, suppression of siliciclastic supply and a return to carbonate shoal environments. The enhanced development of these cycles, with clearer facies differentiation, towards the top of the Coimbra may represent overall accelerated rates of space creation at this time: a logical precursor to flooding at the top Coimbra hardground surface.

The hardground at the top of the Coimbra Formation (fig. 6.8) is marked by oyster and serpulid encrustation, bivalve borings (locally with bivalves in place) and large spiral scours (demonstrably primary on the basis of internal encrustations) which may be the resting traces of fish. There is no evidence for subaerial exposure or significant erosion at the hardground, which is here interpreted to mark a deepening event, between the shoal facies of the Coimbra Formation and the more distal facies of the Brenha Formation (see below). Hallam (1971) assigns the top of the Coimbra Formation to the *obtusum-oxynotum* Zones, though the basis of this is unclear. Mouterde (1955) recovered echioceratids from the overlying Unit A sediments, which he assigned unequivocally to the *raricostatum* Zone. Sr-isotopic results and their comparison with biostratigraphically well-constrained data from Britain are discussed at greater length at the end of this chapter. However, the highest $^{87}\text{Sr}/^{86}\text{Sr}$ ratio recorded from a belemnite within post-Coimbra sediments was 0.707427, compatible with a *jamesoni* or late *raricostatum* Zone date. Oysters encrusting the hardground yielded 0.707521 and 0.707476, suggesting that the youngest encrusters were extant in *oxynotum* Zone times. The flooding event at the top of the Coimbra Formation appears therefore, to be of *oxynotum* Zone, or perhaps slightly older, date: approximately equivalent in age to the Coinstone of Dorset (Chapter 4).

Brenha Formation, Unit A

Brenha Formation Units A to C were logged on the east side of the Portinho da Areia do Norte from the basal hardground, described above (fig. 6.3). Unit A (figs. 6.7 & 6.9a) is some 8m thick and comprises medium interbeds of three facies:

- 1) Soft, yellow, discontinuously laminated, argillaceous bioclastic packstone/wackestone with abundant, oriented, fine-medium grained bivalve and echinoderm fragments, common fine sand-silt grade angular quartz and minor phosphate and glauconite (fig. 6.9b).

2) Hard, grey, bioclastic lime mudstone (fig. 6.10a).

3) Hard, grey bioclastic packstone with burrow mottling, abundant bivalve and gastropod fragments and subordinate crinoid fragments (fig. 6.10c).

Rare thin beds of organic-rich micrite are also present. Small belemnites and an ammonite were found near the base of the section. Vertical and U-shaped burrows are common.

The fauna suggests an open marine setting. Bioclastic packstone, mud, thin beds and vertical burrows are all compatible with a setting between storm and fairweather wavebase. With its siliciclastic-carbonate alternations the facies is entirely consistent with an interpretation as the distal equivalent to the Coimbra Formation. Sharp transgression, marked by the basal Brenha hardground is suggested.

Brenha Formation, Unit B

Unit B is approximately 45m thick. It was logged largely in the low cliffs south of Unit A, although the middle part of the Unit had to be logged in the foreshore with only approximate ties to the cliff across a boulder-strewn beach. The contact between Unit A and Unit B is unseen in the Portinho da Areia do Norte but most of the 2m gap can be filled with reference to the section on the opposite side of the headland at the Forte da Luz, where the obscured strata may be seen to be developed in Unit B facies. There is no evidence of discontinuity although the facies boundary is sharp.

Unit B is dominated by monotonous medium- bedded blocky grey micritic limestone (figs 6.7 & 6.11a). Beds are separated by wavy and sometimes clearly stylolysed partings which pinch and occasionally anastomose. Chert beds are common in the interval 15 to 20m above the base of Unit B, occurring in association with the preservation of cross-laminated bundles, the cross

laminae being defined by very fine peloids (fig 6.11b). The chert beds tend to have sharp, scoured bases and transitional tops, commonly with evidence of bioturbation (fig 6.12). This leads to the suggestion that the unit was pervasively bioturbated and that only the occasional bed escaped this, presumably due to rapid deposition. Redistributed silica was preferentially deposited in the non-bioturbated beds, possibly due to higher permeabilities there.

Apart from microspar replacements of sponge spicules, presumed to be the source of the silica in the cherts, fragmentary gastropods and very occasional whole rhynchonellids, Unit B is faunally barren. Mouterde (1955) places the Sinemurian-Pliensbachian boundary within the Unit by lithological analogy with the coastal section at Sao Pedro de Muel, some 40km to the north (fig. 6.2)

The scarcity of marine fauna (unless it has been lost by dissolution, and there is no reason to suppose that this unit should have suffered particularly badly relative to its neighbours), combined with the micritic texture, suggests a restricted setting. This interpretation is supported by some of the details: the micrograded peloidal grainstone described above is similar to that described by Wilson (1975, Plate XI B) from restricted marine shoals and, whereas sponges are not environmentally diagnostic, similar fabrics to those seen in Unit B are known from the Portland Beds of Southern England, where they occur in an unambiguously peri-littoral facies association (Coe, 1992). It is possible, therefore, that Unit B represents minor regression within the overall transgressive trend from the Coimbra Formation to Unit D.

Some systematic organisation within Unit B is evident from the bed spacing (fig. 6.13), particularly thin-bedded intervals occurring immediately above the chert rich interval (around 25m above the base) and near the top of the Unit.

Brenha Formation, Unit C

Unit C is transitional from Unit B. The base of Unit C has been placed at the first occurrence of soft grey argillaceous beds, accompanied by an abundant and diverse trace fossil assemblage. Unit C is approximately 30m thick and was logged in the cliff section immediately south of Unit B, with some foreshore exposure being required near the base.

The unit principally comprises medium interbeds of two lithologies (figs. 6.7 & 6.14a): impersistently laminated, soft yellow/grey argillaceous micrites (locally associated with diffuse lenses of grey marl near the top of the unit) and hard grey micrites. Some fragmentary punctate brachiopods, echinoderm fragments, phosphatic fragments and rare small belemnites occur. A third facies: a hard grey bioclastic wackestone (including bivalve and gastropod fragments), commonly with *Thalassinoides*, occurs towards the base of the Unit. Bedding is tabular but impersistent at the scale of outcrop, with local slumping (fig. 6.14a). The domal lamination (fig. 6.14b) which is prominent in Unit C, but also occurs in many of the other units requires some comment here. Superficially, the domes have many of the features of stromatolites, including upward increase in amplitude. However, shell fragments and burrows cross-cutting the fabric (fig. 6.15a), together with domes growing both downwards and upwards from fissures (fig 6.15b) unambiguously indicate a diagenetic origin.

Mouterde's (1955) Unit 6 lies in the middle of Unit C, as defined here. Mouterde places Unit 6 in the *jamesoni* Zone (basal Pliensbachian) and records *Phricodoceras* aff. *taylori*, which suggests the *taylori* Subzone. *Uptonia* sp., indicative of the *jamesoni* Zone, has been recorded at the base of the overlying Unit D (Mouterde, 1955).

Unit C is interpreted as intermediate and transitional, between the faunally barren micrites of Unit B and the faunally rich marls with thin limestones of Unit D (see below). The increasingly abundant pelagic fauna within the unit supports a gradual transition from restricted to fully marine

environments. The transitional nature of lithological changes and the absence of coarse-grained facies suggests deposition on a low energy carbonate ramp. Minor slumping is consistent with deposition on a gentle slope.

Brenha Formation, Unit D

Unit D, approximately 55m thick, is transitional with Unit C. Its base is picked where grey marls, rather than limestones begin to dominate the stratigraphy. The lower part was logged in continuity with Unit C. The upper part was logged in the cliff to the south-west of Portinho da Areia do Norte (fig. 6.3), the two sections being linked by foreshore exposure, available at low tide. The base of this exposure, Phelps' (1985) bed 44, is easily matched to the cliff. The top of the foreshore exposure is less easy to match but matching appears possible utilising two prominent limestones underlain by a pyritised-ammonite bed (Phelps' beds 80 & 84). Both the foreshore and south-western cliff section are cut by a number of minor normal faults which hamper bed-by-bed correlation. Bed numbers (fig. 6.7) are from Phelps (1985), supplemented below bed 12 by Rocha & Soares (1987). As with other sections of the log, Mouterde's (1955) units have also been marked. It should be noted that Mouterde provides only a text description and therefore the placing of these units is an interpretation. The ammonite zones and subzones of figure 6.7 have been taken from Phelps (1985) for the *ibex* and *davoei* Zones, from Rocha & Soares (1987), for the base of the *jamesoni* Subzone, and from Mouterde (1955) in all other cases.

Unit D comprises dominantly medium and dark grey marls, locally with "paper shales", containing abundant ammonites and belemnites. Carbonate interbeds of two types are found. The first, most common towards the top of the Unit, is represented by hard, grey micrites; tabular, nodular or concretionary, locally associated with *Chondrites* and pyrite. The second type of carbonate interbed, most common towards the base of the Unit (fig. 6.16a), is represented by medium-hard, brown-grey carbonate mudstones, weathering yellow, locally with a granular texture which may extend into the surrounding marl (fig. 6.16b). Under the microscope, a range of matrix textures is

present in this latter carbonate type, from micrite with patches of dolomitic microspar (commonly centred upon clasts and fractures) to complete microspar replacement. Grains include peloids, bivalves, brachiopods and the granules, obvious at outcrop, which, in thin section, comprise micrite with irregular tubes suggestive of calcareous algae. There is potential for confusion here as this fabric is described as "grumeleuse", i.e. "lumpy" or "gritty", in the French literature (Elmi et al., 1988; Mouterde, 1955). It is not, however, the carbonate microfabric "structure grumeleuse" described by Bathurst (1975, pp 511-513) as clots of fine-grained crystalline calcite in a matrix of granular calcite. The granules at Peniche are cm-scale features and closely resemble the "grapestones" illustrated from the modern Bahama Bank by Tucker & Wright (1990), see figure 16.17. These modern grapestones are aggregates of grains initially bound together by cyanobacteria, algae or encrusting foraminifera. They form in areas where waves and currents are sufficient to remove silt but not sand: generally areas under 3m in water depth.

A spectral gamma-ray log was measured through Unit D and down into Unit C (fig. 6.18). Th and K co-vary strongly (fig. 6.19a) and there is little evidence for secular variation in Th/K, apart, perhaps, from somewhat lower values around the *luridum* Subzone (fig. 6.19). As in previous chapters, this covariance is interpreted as indicative of time-invariant clay mineral composition. There is also little systematic variation in Th/U ratio, although the variance of these data is greater (fig 6.19b). There is some evidence for lower Th/U above the *luridum* Subzone, which may have two causes: lower concentrations of Th in the *davoei* and late *ibex* Zone sediments and higher concentrations of U in the *margaritatus* Zone sediments.

Some caution must be exercised in the interpretation of the absolute concentration, rather than the ratio data, as the *davoei* and late *ibex* Zone data were obtained from bedding planes on the foreshore: there is therefore a potential bias towards more carbonate-rich sediments which are resistant to weathering (see Appendix 1). However, there is a convincing upwards increase in total counts which extends through the foreshore section and continues through the cliff exposure of the *margaritatus* Zone. This is the uppermost of two cycles of upwards-increase in total counts,

the lower extending from the base of the gamma-logged section up to the mid-*ibex* Zone. Given the constant Th/K ratios discussed above, the primary explanation of the total-count cycles is likely to lie in changing clay content. Clay/carbonate ratio increases through the *jamesoni* Zone and peaks in the *masseanum* Subzone, it reaches a minimum in the early *luridum* Subzone and then increases again towards the top of Unit D, where there may be a contribution to total counts from increased uranium concentrations.

Carbonate ramp systems commonly pass distally into terrigenous mud (Burchette and Wright, 1992), though the source of that mud is unclear in many cases. The most commonly cited modern analogue is the Persian Gulf, where the transition to terrigenous mud, largely supplied axially by the Tigris and Euphrates, occurs in approximately 40m of water (Purser, 1973). In the Brenha Formation at Peniche, the transition to terrigenous mud observed from Unit B through to Unit D, accompanied as it is by an increase in pelagic fauna, is suggested to represent a gradual transition from proximal to more distal environments of deposition.

The spectral gamma-ray data suggest that the overall scheme of deepening is interrupted in the *ibex* Zone, where the more calcareous sediments could reasonably be interpreted as more proximal. Changing Th/K ratios may be indicating subtle changes in the nature of the mud supply between the two cycles. There may be a connection between this postulated regressive phase and the abundance of the "granule" facies (described above) and particularly a feature of the *valdani* Subzone. The grapestone belt of the Great Bahama Bank covers an area transitional between the agitated oolite shoals and more protected muddy environments (Tucker and Wright, 1990, p11). If the "granules" do represent grapestones then they may be a further indication of minor regression.

Brenha Formation, Unit E

Unit E rests with a sharp contact on Unit D (fig 6.20a). It is exposed in the roughly strike-parallel cliffs on the south side of the Portinho da Areia do Norte (fig. 6.3). Individual beds may be traced along the cliff (fig 6.20b) and a log assembled by descending fishermen's steps at various points.

The unit comprises tabular, medium interbeds of lime mudstone (light grey, hard, with occasional micritised bioclasts, silt-grade quartz and mica, fig. 6.21a) and argillaceous wackestone (medium grey, soft, yellow-mottled, abundant silt-grade mica and quartz, fig. 6.21b). Two more marly intervals occur, at approximately 7 and 19m above the base of the Unit. These provide useful marker horizons but bed spacing data (fig. 2.22) provide little support for the suggestion that the marly intervals are an expression of systematic organisation or "cyclicity" within the unit. Belemnites are abundant throughout. Near the top of the unit they are concentrated into scours and are commonly found to be encrusted by oysters and serpulids, suggesting long exposure on the sea bottom and commensurate with sediment winnowing (Doyle and MacDonald, 1993).

Unit E is similar in aspect to Units A and C. Whilst the ammonite and belemnite fauna indicate that it is still fully marine, the influx of quartz silt, the return to carbonate rather than clastic mud dominance and the presence of scours suggest a basinwards shift of facies relative to Unit E.

Beds around the Unit E-Unit F boundary yield an abundant and biostratigraphically diagnostic ammonite fauna. The facies transition takes place over a few thin beds (fig. 6.7) The boundary between Unit E and Unit F in this study has been placed on the top of Mouterde's (1955) bed 15c, coincident with the Pliensbachian-Toarcian boundary as described in Rocha et al. (1987). This slightly revises the original work of Mouterde (1955), who placed the boundary above his bed 15e. Mouterde (1955) suggests that the lower limit of what is here described as Unit E is coincident with the base of the *spinatum* Zone. He does not, however, record age-diagnostic forms until the lower of the two marly horizons, which contains both *Amaltheus margaritatus* and

Paltoleuroceras . On this basis the sharp base of Unit E has here been placed in the latest *margaritatus* Zone rather than at the *margaritatus-spinatum* Zonal boundary.

Brenha Formation, Unit F

Unit F (fig. 6.7) is equivalent to Wright & Wilson's (1984) Unit 1. It was logged in the small cove to the west of the Portinho da Areia do Norte (fig. 6.23a) and comprises dominantly medium and dark grey silty marls (fig 6.23b) with comminuted shell debris, quartz and mica. Belemnites, common at the base, are succeeded upwards by abundant ammonites. The upper part of the unit has only rare macrofauna.

The background of silty marls is punctuated not only by occasional nodular and concretionary limestone beds, which might be expected in this facies, but also by sandstone beds, varying between centimetres and up to a metre in thickness. The sandstones are medium to very coarse grained with angular to subangular quartz grains (composite and strained), minor feldspar and a carbonate cement (fig. 6.24a). The thinner sandstones have erosive bases with gutter casts and other sole marks suggesting NW-SE transport directions. The thicker sandstones are convex-up (figs. 6.24b & 6.25a), with convex-up internal bedding, suggesting that the sandbodies had original relief rather than filling channels. In one case (fig. 6.25a & b) a convex-up sand body is overlain by a very coarse-grained deposit with clasts up to cobble size "frozen" in matrix, clearly demonstrating that the gravity flow which deposited the bed had some shear strength at the time of deposition (Pickering, et al., 1988, p. 27).

Spectral gamma-ray measurements in Unit F (fig. 6.26) again show strong Th/K covariance (fig. 6.27a), with a rather low Th/K ratio (though higher than Unit D). Despite the dark colour of the rocks and the abundant disseminated pyrite reported by Wright & Wilson (1985), there is little uranium enrichment, Th/U ratios being between 3 and 4 in the *tenuicostatum* Zone and 4 and 6 in the *falciferum* Zone. This suggests that critical values of anoxia for uranium fixation were not

exceeded (Appendix 1), possibly an important factor in the failure to find significant petroleum accumulations in the Lusitanian Basin since, as will be seen in Chapter 11, the early Toarcian is a potentially attractive horizon for petroleum source rock development. There is a distinctive peak in total gamma-ray counts in the early part of the *falciferum* Zone, which reflects a maximum of clay content at this time.

The return of marly sediments at the top of Unit E is interpreted to mark a return to the distal settings of Unit D times. This interpretation is supported by the pelagic fauna, the evidence of dysaerobic, if not anaerobic, bottom conditions, nodular horizons suggesting periodic starvation and the absence of shallow water sedimentary structures. Intermittent small sediment gravity flows continued to supply coarse siliciclastics to the basin. Wright & Wilson (1984) drew attention to the anomalous nature of the sand bodies: coarse sand lenses, in some ways suggestive of proximal, perhaps crevasse splay, turbidites but embedded in a muddy sequence and in a distal setting relative to the overlying progradational fan sequence discussed below. The interpretation preferred here, based upon the massive, convex-up character of the sand bodies and the local evidence of shear strength, is one of largely debris-flows, similar to those described by Prior et al (1984). Such debris flows may be associated with the collapse of the basin margin during the early phases of Toarcian rifting: the rifting event which is familiar from other circum-Tethyan areas (e.g. Chapter 8) and may be required in order to create space for the thick overlying submarine fan succession.

Brenha Formation, Unit G

This is the lower part of Wilson & Wright's (1984) Unit II. It was logged to the west of the cove described above (Unit F), in the cliffs below Station of the Cross VI (stone tablets indicating "Stations of the Cross" have been erected along the northern coast of the Peniche peninsula and provide useful landmarks).

Unit G comprises thin to medium interbeds of nodular limestone and medium grey marl (fig. 6.28a), laterally persistent over 10's of metres. Some limestones contain abundant thin-shelled bivalves (fig. 6.28b), typical of Mediterranean pelagic facies (H. C. Jenkyns, pers. comm.). Ammonites are locally abundant. Deep water conditions are inferred on the basis of the low supply rates associated with nodularity and the presence of a pelagic bivalve fauna. Coarse siliciclastic supply appears to have terminated at this time.

Unit G marks a profound change in depositional style. Below Unit G we are able to monitor the shifting facies belts of a carbonate ramp involving the creation of perhaps 250m of accommodation space in around 15 million years. Above Unit F there is space for the development of a submarine fan system of several km radius and some 250m in thickness (described below), suggesting that subsidence significantly outpaced sediment supply in mid-Toarcian times. The combination of deepening and change of sedimentation style is felt to mark significant rifting (c.f. fig. 6.1).

Post-bifrons Zone Turbidites

The rocks exposed westwards from Unit G, around the western and southern cliffs of the Peniche peninsula, were noted as "allodapic" by Hallam (1971) and interpreted as a thick prograding carbonate submarine fan sequence by Wright & Wilson (1984; 1982). The sequence is poorly constrained biostratigraphically: limestones near the top of yield the foraminifer *Nautiloculina* which suggests that the sequence may extend up into the Middle Jurassic (or younger), though the range of the genus is poorly understood (M. Simmons, pers. comm.). A solitary belemnite from near the top of Unit 3 yielded a strontium isotope ratio of 0.707254, consistent with a late Toarcian age (see below). Given the lack of biostratigraphic constraint, it was important for this thesis only to confirm the generalities of Wright & Wilson's model and the progradational character of the sequence. However, the following impressions stimulated more detailed work:

1. The importance of pulses of coarse-grained siliciclastic sedimentation.
2. The possibility that some units which were interpreted as "stacked" by Wright & Wilson were actually lateral equivalents.
3. The lack of fining up in individual beds and lack of systematic thickening or thinning of bedsets.
4. The lack of distinct channelling.
4. The presence of well developed tabular cross-beds suggestive of sustained flows.

Figure 6.29 is a log of the turbidite section from the base of Wilson & Wright's Unit 3 (which occurs immediately above Unit G, described above) to the lighthouse at Cabo Carvoeiro (fig. 6.3). The logged section can be followed on the annotated photomontages (figs. 6.31-6.36). Wright & Wilson's (1982) Unit 3 can be clearly recognised and was logged at 29m compared with their 30m. The unit dominantly comprises 3-30cm beds of fine-grained peloidal and oolitic grainstone (fig. 6.30a) interbedded with bioturbated grey silty marls. The grainstone beds tend to be grouped into packages of 2-3m in thickness, but there is little evidence for systematic thickening or thinning upwards. As is evident from the photomontages, the grainstone beds are laterally persistent for at least 100's of metres. They tend to have sharp bases, commonly with tool, flute and groove marks. Graded bedding is uncommon, perhaps due to a limited distribution of available grain sizes: although most beds have transitional tops due to bioturbation, the immediate superposition of a further grainstone bed can prevent this. Bioturbation tends to penetrate the bases of beds less than about 5cm thick and to progressively eliminate beds less than about 4cm thick. Thicker beds tend to be coarser and include more siliciclastic material. Rarely, cross-lamination is present at the base of beds and one exceptionally thick bed (78cm) grades upwards from clast to matrix support.

These observations support Wright & Wilson's (1984) interpretation of their Unit 3: "turbidity currents carrying sediments originating from carbonate shoals which interrupted a background sedimentation of hemipelagic muds. The turbidites resemble those deposited on outer-fan and basin-plain settings".

Wright & Wilson's (1982) Unit 4 extends upwards from the top of Unit 3 to the fault immediately east of Cerro do Cão. Facies are broadly similar to Unit 3, but the turbidite beds are thicker and coarser-grained, often with loaded bases. As with Unit 3, the turbidite beds tend to be organised into 2-3m packages, and this interval includes the only convincing example of a thickening-upwards package (between horizons 3 and 4, see log and fig. 6.30b). The turbidite/background ratio is considerably higher in Unit 4 than in Unit 3, due to the reduced thickness of marly sediment between the turbidite packages. Minor erosional down-cutting also occurs.

Wright & Wilson's Units 4 and 5 are separated by a fault. However, a satisfactory cross-fault correlation may be possible (fig. 6.30b), enabling a continuous log to be made. Further upwards progress then requires beds to be traced around the headland of Cerro do Cão (fig. 6.33), where the log can be extended vertically using various accessible stacks in the region of Station of the Cross X (fig. 6.33, section G-G'). The section continues to comprise medium and thick-bedded turbidites, as in Unit 4, but marly background sediments are almost completely excluded and the turbidites are amalgamated. It is critical to note that this part of the log (above surface β , fig. 6.29) is approximately 1km west of the lower part, as well as stratigraphically younger. Furthermore, figures 6.34 and 6.35 demonstrate that the stratigraphic interval between surfaces β and C can be traced a further 1km to the lighthouse at Cabo Carvoeiro. It appears, then, that Units 5, 6 and the lower part of 7 of Wright & Wilson (1984) are stratigraphically equivalent rather than superposed.

At the lighthouse (Wright & Wilson's Unit 7) beds are of 2-5m in thickness and much coarser than observed at the same stratigraphic horizon further to the north east. There is a clear alternation of siliciclastic and carbonate-dominated units and channelling is evident, especially at the base of the

siliciclastic-dominated units (fig. 6.37). Large scale cross-bedding, usually tabular but occasionally sigmoidal or trough-shaped, is common. This alternation of facies persists along the southern coast of the peninsula (fig. 6.36), with approximately another 100m of stratigraphic thickness being exposed before outcrop ends at the "Citadela" in Peniche town. Carreiro da Joana (fig. 6.38a & b) provides good exposure of an erosive clastic-carbonate contact, with carbonate clasts entrained in the base of a clastic-filled channel.

In summary, the observations presented here confirm Wright & Wilson's model of a prograding carbonate fan. However, the total recorded thickness of that fan must be reduced from 300 to some 200m due to the lateral equivalence of part of the section. This lateral equivalence provides an interesting example of horizontal variation in a small coarse-grained fan, with lateral passage over a few km from well-defined channels cut by powerful currents to rather ill-defined cut-and-fills, similar to those described by Pickering (1982) from the late Precambrian of north Norway. A possible model is that carbonate sediments were shed during relative sea level highstands from a productive carbonate platform to yield a fan built from "compensation cycles" (Pickering, et al., 1988, fig 7.4) without well defined channels (fig. 6.39a). Coarse siliciclastics, periodically introduced during sea level lowstands, cut their own discrete channels into the top of this fan, concentrated along the fan axis (fig. 6.39b). No systematic study was made of the transport directions on the fan: the trend to more thinly-bedded sediments from south-west to north-east could be proximal-distal or axial-peripheral. The large clastic channel observed at the lighthouse (fig. 6.37) has a generally SW-NE orientation, but the sinuosity of the channels is unknown. No evidence was found to support different transport directions for the siliciclastics and carbonates.

Strontium Isotope Stratigraphy

Jones (1992) derived a $^{87}\text{Sr}/^{86}\text{Sr}$ curve for Jurassic and Cretaceous seawater based on the analysis of belemnite and oyster material from Britain. The excellent subzonal ammonite control in the Pliensbachian at Peniche, combined with abundant well-preserved belemnites, represented

an excellent opportunity to attempt to replicate part of his curve at a rather distant location. Given that the curve could be replicated (modern oceans have homogenous $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, see Appendix 2), there were opportunities to use it to refine the dating of the older parts of the Peniche section, where ammonite control is much poorer.

Belemnites were analysed according to the procedure in Appendix 2 and dated by interpolation between the ammonite control shown in figure 6.7. Figure 6.40 compares the results with those of Jones (1992). The Pliensbachian part of Jones' curve has been replicated in some detail, including a suggestion of the mid-Pliensbachian (~191Ma) levelling and possible reversal of the curve to which Jones draws attention. Given the precision with which samples of biostratigraphically known age fit Jones' data (better than $\pm 1\text{Ma}$, i.e. certainly with zonal precision) it is tempting to interpret the spread of data at and near the top-Coimbra hardground (fig. 6.41) as representing genuine condensation or stratigraphic omission. This is largely supported by the stratigraphic order of the samples: the two highest Sr-isotope values (oldest dates) are from oysters which encrust the hardground surface and belemnite samples from the lower part of the Brenha Formation yield significantly younger dates. Only sample P1, a solitary belemnite from the upper part of the Coimbra Formation, is anomalous in that it yields a *raricostatum* Zone date, when an *oxynotum* Zone or older date would be predicted. More detailed analyses across this surface would clearly be instructive. The single sample from the post-*bifrons* Zone turbidites falls on a rather "flat" part of the curve, however, its Sr-isotopic composition is consistent with its presumed stratigraphic age.

Conclusions

- The $^{87}\text{Sr}/^{86}\text{Sr}$ curve of Jones (1992) can be replicated in detail in the Pliensbachian of Portugal, where good subzonal ammonite control is available. This confirms at least the regional significance of this curve. Given suitable material the curve can be used to date material of unknown age to at least zonal precision.

- At Peniche, a well-exposed carbonate ramp sequence enables regressive-transgressive cycles to be tracked from late Sinemurian to *falciferum* zone times (fig. 6.42).
- Spectral gamma-ray studies suggest lower Th/K ratios than in the UK area, even allowing for proximal-distal relationships. This may be attributable to climate, and will be discussed at greater length when additional German data can be included (Chapter 7). Absence of U-enrichment in the *falciferum* zone suggests that despite relatively distal facies, anoxic conditions did not prevail in this area at this time.
- The *bifrons* zone marks a change in sedimentary style which is most likely to reflect basin tectonics. Above this zone there is excellent three-dimensional exposure of a prograding turbidite system. The mixed carbonate-siliciclastic nature of this system, reflecting intermittent bypassing of the carbonate source area, is emphasised (in contrast to the description of Wright & Wilson, 1984).
- Re-measuring of the turbidite system, and cliff photography demonstrates that some of the sections measured by Wright & Wilson (1984) are lateral equivalents rather than vertical successors.

Chapter 7. Southern Germany

The Jurassic rocks of Southern Germany outcrop along the line of the Swabian and Franconian Alps (fig. 7.1a). These hills are formed by a series of south-eastward dipping scarps which, as with so many of the European Mesozoic outcrops, result from Tertiary events: here a combination of the uplift of the shoulders of the Rhine Graben and the flexural forebulge of the Alpine Chain.

The attractions of the area for the purposes of this thesis are twofold. Firstly, the facies and thickness of the Lower Jurassic here place it in sharp contrast with the more directly rift-related settings discussed in other chapters. The Lower Jurassic varies in thickness between 50m and 150m along the length of the outcrop (Urlichs, 1977), and is largely developed in marls with thin limestones and some thin sands in the Hettangian. The thin succession, without marked lateral variation and with no marginal facies suggest a lack of tectonic activity and a remoteness from coarse-clastic input, as brought out in typical paleogeographic maps (fig. 7.1b). The second attraction of the area is that the history of biostratigraphic research rivals that of southern England, so that confident correlation at the ammonite subzonal level is possible.

The disadvantage of the area is that outcrop is very poor. The current comprehensive understanding of the stratigraphy is the result of painstaking work over many years using fragmentary and often temporary exposures. The best continuous section is a stream section above the village of Aselfingen (fig. 7.2). This is at the southern end of the outcrop and is thin. It does, however, show the essential features of the stratigraphy and is reviewed in some detail below.

The outcrop thickens northwards, reaching a maximum around Stuttgart. The second section examined for this thesis is a new exposure due to the re-routing of the B27 road between Balingen and Hechingen (fig. 7.2). Only the interval from the *jamesoni* Zone at the base of the

Pliensbachian to the *falciferum* Zone of the Toarcian is exposed, but this is also the interval best exposed and dated in the other study areas of this thesis and hence is of considerable importance. New spectral gamma-ray measurements were obtained through this section and a suite of belemnites collected with a view to both $^{87}\text{Sr}/^{86}\text{Sr}$ dating (reported in this chapter) and $\delta^{13}\text{C}$ measurements (reported in Chapter 12).

The south German Lower Jurassic is traditionally divided using the system of Greek letters (α – ζ) introduced by Quenstedt (1843).

The Aselfingen Section

This section is exposed in the valley of the Aubach, the stream which enters the small village of Aselfingen from the north (fig. 7.3). The following account relies heavily upon the review by Ulrichs (1977), which is the source of all of the biostratigraphic information.

The Hettangian (α_1 & α_2) is not well exposed at Aselfingen. Bloos (1976) has examined this unit in detail along the length of the outcrop shown in figure 7.2. It ranges between 2 and 30m and comprises thin lenses of sand (up to 200km long x 70km wide x 4m thick), commonly capped by iron oolites and reworked calcareous concretions. The sand lenses are separated by marls and mudrocks (fig. 7.4a). Einsele & Bayer (1991) have interpreted this sequence in terms of relative sea level-induced submarine erosion on a shallow shelf. It is noteworthy that these cycles of relative sea level change occur at subzonal frequency or somewhat greater: cycles of similar frequency have been noted in the Dorset Hettangian (Chapter 4). The most laterally continuous of the reworked horizons (apart from that at the base of the whole sequence) is the basal-*angulata* Zone "Oolithenbank" (basal α_2). The most laterally extensive sandstone (the Hauptsandstein) is of mid-*angulata* Zone age.

At Aselfingen, as elsewhere in South Germany, there is virtually no sand development in Lower Jurassic rocks younger than the Hettangian. The Sinemurian - Toarcian interval essentially comprises alternating units of relatively condensed medium-bedded limestone and relatively expanded marl (fig. 7.5). As argued with the transition from the Blue Lias to the Shales-with-Beef in Dorset (Chapter 4), expansion of the clastic-rich intervals relative to the carbonate-rich intervals suggests the dominance of supply control rather than accommodation space control upon accumulation rates.

The α_3 unit is the first of the limestone-dominated, relatively condensed, intervals. At Aselfingen the marly interbeds are very thin but they thicken northwards and bituminous shales occur towards the top of the unit (Urlichs, 1977). The sharp transition to dark grey marls (β) occurs within the *stellare* Subzone, which is much expanded compared with the units both above and below (fig. 7.5). The "Obliquabank" limestone at the top of the Lias β is of *raricostatum* Zone age: the *denotatus* Subzone and *oxynotum* Zone appear to be absent. These are the same missing ammonite subzones as those missing at the "Coinstone" horizon in Dorset (Chapter 4). The *aplanatum* and *macdonneli* Subzones, absent at the "Hummocky" horizon in Dorset, may also be missing above the Obliquabank.

The lower Pliensbachian γ unit comprises medium interbeds of limestones and marls, the marls being somewhat thicker than the limestones. The transition to the overlying marl unit (δ) is apparently time-transgressive along the length of the outcrop, occurring in various places between the base of the *stokesi* Subzone and within the *subnodosus* Subzone (Urlichs, 1977). Nodule horizons occur throughout δ . Persistent limestone beds recur in the *hawskerense* Subzone at Aselfingen (fig. 7.5), marginally earlier (*apyrenum* Subzone) elsewhere.

Relatively good outcrop of the Lias β and the lower Lias γ at Aselfingen enabled some spectral gamma-ray measurements to be made. Total counts decline slowly through the marl and then sharply into the limestones. The primary reason for this is that both Th and K concentrations are

declining, presumably due to a decrease in the proportion of clay (fig. 7.5). However, there are also systematic changes in clay mineralogy: Th/K ratios are somewhat lower towards the top of β . They are also lower in the γ limestones. The "Obliquabank" shows a markedly higher Th/K ratio. This further emphasises the anomalous nature of the "Obliquabank", which is associated with the biostratigraphic hiatuses discussed above. In Appendix 1 and Chapters 4 & 5, the idea of Th/K ratio as a proximity indicator was developed. If that interpretation is adopted here then it suggests increasingly distal conditions associated with the marling-upwards of the Lias β , but a marked proximal shift associated with the "Obliquabank". This supports the attribution of the biostratigraphic hiatuses to base level winnowing rather than starvation.

Uranium levels remain approximately constant and rather high (around 4ppm) throughout the section. The effect of this is to produce low Th/U ratios in the Lias γ , where Th levels are low. Th/U ratios at this level suggest anoxia (Appendix 1) and tend to cast the Lias γ as a less "extreme" example of the Posidonienschiefer (see below).

The base of unit ϵ (the "Posidonienschiefer") is marked by a sharp transition to laminated organic-rich shales with tabular organic-rich limestones, traceable basinwide. The limestones are best developed in the *falciferum* Zone. The facies is extinguished at the top of the *bifrons* Zone, with a return to more "normal" marly sediments with concretionary horizons.

The B27 Roadcut

The roadcut for the B27 (fig. 7.7a & b) enabled a new section through the Lias γ to Lias ϵ to be logged (fig. 7.8). I am grateful to Dr. M. Franz of the Geological Survey of Baden-Württemberg for providing me with his unpublished ammonite data on this section and to Richard Heberle of the Staatliche Bauleitung for granting access and providing me with road survey information.

Since no biostratigraphic data on the outcrop have yet been published, a somewhat extended justification for the dates assigned on figure 7.8 is necessary. According to the notes of Dr. Franz, the prominent ledge between gamma-ray readings eleven and twelve yielded *Prodactylioceras davoei* and the overlying marls, *Amaltheus stokesi*, accurately locating the base of the *margaritatus* Zone. Marls immediately below the ledge yielded *Androgynoceras* "c.f. *capricornus*", extending the *davoei* Zone downwards by approximately 1m. Below this, marls yield *Acanthopleuroceras valdani* and *Acanthopleuroceras maugenesti* (*ibex* Zone), together with *Uptonia* "ex. gr. *angusta/jamesoni*" and *Coeloceras pettos* which suggest that at least the upper part of the *jamesoni* Zone is present. Dr. Franz notes *Echioceras* sp. from the roadcut floor, suggesting that the entire *jamesoni* Zone is present. This interpretation accords well with the Sr-isotopic data obtained from belemnites within the section (fig. 7.11): restricting the extent of the *jamesoni* Zone would reduce the age of the youngest samples obtained by up to 2 million years and cause significant divergence from the Sr-isotope curve for UK material developed by Jones (1992)

The top of the *margaritatus* Zone may be drawn within 1m by the last recorded *Amaltheus* and the first *Pleuroceras*. The *stokesi* Subzone may be distinguished near the base where, between gamma-ray readings fourteen and sixteen, marls with *Amaltheus stokesi* give way to marls with *Amaltheus margaritatus*. The top of the *spinatum* Zone can be drawn within some 1.5m by the last occurrence of *Pleuroceras* and the first *Dactylioceras*.

The Sr isotope data of figure 7.11 are plotted using linear interpolation between the zonal and subzonal boundaries discussed above. As with the Portuguese data (Chapter 6), the results tend to confirm the long-range (at least pan-European) validity of Jones' (1992) seawater Sr isotope curve and to suggest that, given suitable material, interbasinal correlation to ± 1 Ma using Sr isotopes should be possible. It is notable that the same positive excursion between 189 and 190.5Ma, away from a steadily decreasing trend, is observed in the German data set as well as in the UK and Portugal data sets. Overlap of late *davoei* (~190Ma) and *ibex* Zone (~192Ma) Sr

isotope values means that the excursion cannot be removed by stretching the timescale and duplication of the result in different sections provides further support for the reality of this event. Its possible significance is discussed by Jones (1992). Essentially, it requires a significant input of radiogenic strontium to the world oceans in mid-*davoei* Zone times. This could be achieved by increased sea-floor spreading rates (with an associated increase in the hydrothermal flux of Sr) or an increase in either the flux or ^{87}Sr composition of river waters flowing into the oceans. Jones notes the possible influence of flood basalts and associated doming, which could provide both elevated Sr-isotope ratios and increased riverine flux: the first phases of Karoo volcanism occur at approximately this time (Baksi, 1990).

Th/K ratios remain relatively constant through the section (fig. 7.8), with a maximum around the mid-*margaritatus* Zone, which also marks the maximum in both Th and K and the maximum Th/U ratio: all indicating a maximum of clay content. This is consistent with a maximum of clay supply in the most proximal setting (see above). Notably, there is a second peak of clay supply around the *spinatum-tenuicostatum* Zone boundary. The Posidonienschiefer shows typically "enriched" Th/U ratios, as one might expect with the high observed organic-matter content (Appendix 1).

Addition of the German data to the total Lower Jurassic spectral gamma-ray data set demonstrates an interesting latitudinal trend in Th/K ratios (fig. 7.10). As has already been noted, there is considerable variation within sections, which has been attributed largely to proximal-distal relationships. Overall, however, the more northerly sections appear to have higher Th/K ratios. Two possible explanations are:

1. Latitudinal (climatically controlled) variation in clay mineral provenance: each clay mineral system being locally sourced.

2. Large-scale transport effects associated with the origin of the majority of the Lower Jurassic suspended sediment load from clastic source areas in northern and eastern Europe, as discussed in Chapter 3.

These observations offer exciting possibilities for understanding the distribution of Lower Jurassic clays if the data set could be expanded and, perhaps, the distribution of radionuclides in modern analogues better understood.

Conclusions

- The South German Lower Jurassic accumulated in a supply-dominated setting with base-level impinging probably only in Hettangian and in late Sinemurian times.
- Periods of high supply are marked by high clay input and rather faster accumulation rates. Periods of low supply are marked by medium-bedded limestones and rather low accumulation rates. The key events are summarised in figure 7.12.
- Spectral Gamma data provides useful information on clay content and mineralogy. The addition of German data to the total spectral gamma-ray dataset shows gross trends across Europe which may represent climatic or large-scale sediment transport effects.

Chapter 8. The Western Alps

Alpine tectonics have exhumed large volumes of sedimentary rock originally associated with the Jurassic-Cretaceous passive margins of Tethys (fig. 8.1a & b). A large body of work (e.g. Bernoulli, 1964; Lemoine & de Graciansky, 1988) now documents:

1. Lower and Middle Jurassic tilted fault blocks of the rift phase, with their marked lateral facies variations.
2. Late-Middle or early-Late Jurassic ophiolites of the spreading phase, with associated foundering of carbonate platforms at the continental margins.

The geology of the rift phase on the European margin of Tethys is especially well exposed in the external zones of the Western Alps, in south-east France. Reduced deformation in this area appears to be a result not only of a location external to the Alpine orogeny, but also of the small circles of closure in the Western Alps corresponding with the small circles of opening so that there is no climbing of lateral ramps as seen further east, for example in Switzerland (de Graciansky, pers. comm.). Biostratigraphic control is not as good as in other study areas and interpretation is complicated by the effects of Alpine compression. However, the Western Alps provide a unique opportunity to see deep into Early Jurassic half-graben, with well in excess of 1km of Lower Jurassic sediment. It is useful to see whether the stratigraphy of these deep and contemporaneously-extending half-grabens may be related to the more gently subsident and more obviously base-level-controlled stratigraphies examined in other chapters of this thesis. Study of the stratigraphy of the Western Alps also provides an opportunity for the critical examination of French work on long-range correlation in the Lower Jurassic which was published close to the completion date of this thesis (de Graciansky et al., 1993).

Fieldwork in the Western Alps has been largely confined to understanding the work of earlier authors, although some key sections have been re-logged and new petrographic data obtained. Two areas will be examined in detail (fig. 8.2). The Bourg d'Oisans-La Mure area provides the best-preserved examples of Lower Jurassic tilted fault blocks and, despite poor biostratigraphic control, serves as a useful introduction to the key events of the Lower Jurassic in the Western Alps. The Digne area is structurally more complex, so that the extensional structural elements cannot be observed directly but must be inferred by comparing sections now separated by thrusts. Biostratigraphic control is better in this area, but it should be constantly remembered that the rarity of fossils in the Alps results in a temptation to correlate stages using lithostratigraphy which few authors appear able to resist. This is emphasised below by the free use of quotation marks.

The Bourg d'Oisans-La Mure Area

The two north-south trending bands of Lower Jurassic outcrop in this area have been related to two inverted Lower Jurassic half-graben by Bas (1988). The localities key to this interpretation are shown on figure 8.3a and Bas' interpretation in terms of original half-graben is shown in figure 8.3b.

Bourg d'Oisans lies in the bottom of a valley which slices through the northern end of the eastern Lower Jurassic outcrop. The valley walls provide a seismic-scale cross-section of a Lower Jurassic half-graben, now significantly shortened (Gillcrist, 1988, see also fig. 8.4a & b). The northern wall of the valley enables one see that "basement" on the hangingwall block is dissected by a number of small normal faults which were sealed early in the history of the graben fill. These normal faults may be examined in more detail on the plateau above the ski resort of Alp d'Huez (fig. 8.5a). Crystalline basement is here capped by thin "Triassic" dolomites which exhibit common sabkha-style features such as pseudomorphs after evaporite minerals, flat-pebble breccias and stromatolitic lamination (fig. 8.5b). Further sections through the Triassic - lowermost Jurassic

interval are available in the mountains to the south of Bourg d'Oisans. Pinto-Bull (1988) has logged a number of these and records between under 2m (la Chave) and nearly 30m (les Clottous) of Lower Jurassic carbonates overlying both Triassic sediments and metabasalts, the thinner Lower Jurassic sediments tending to overlie the metabasalts and the thicker early Jurassic sediments to overlie Triassic sediments. Although minor faulting may have been taking place at this time, the thickness variations are consistent with carbonates simply infilling a slowly drowning topography created on the Triassic sabkha by the extrusion of the basalts. The sections at la Chave and le Paletas were visited for this thesis. The carbonates comprise wavy, medium to thickly bedded tabular grainstones. Grains include pellets, oolites and bioclasts (fig 8.6a). Pinto-Bull's detailed petrography demonstrates an upwards-trend to a dominance of bioclasts and sponge spicules over pellets and oolites, which suggests a transition from somewhat higher energy conditions with a restricted fauna to lower energy conditions with a diverse fauna, i.e: a deepening trend. The top of the carbonates is marked by a sharp transition to marls with thin beds and nodules of fine-grained limestone. Pinto-Bull (1988) reports *Caloceras* from the carbonates of the la Chave section, which supports an early Hettangian date. His discovery of *Schlotheimia angulata* in post-flooding marls of the Lac du Vallon section demonstrates that, at least at that locality, the marked deepening which appears to mark the onset of major rifting took place in the *angulata* Zone (i.e. in latest Hettangian times). Pinto-Bull also notes the earlier discovery (by J. Schade) of a *Schlotheimia* in the top bed of the la Chave section. An ammonite was found on the top surface of the top limestone bed at la Chave during logging for this thesis (fig 8.6b). Unfortunately it has not proved possible to identify this specimen unequivocally: it is most likely to be an *Arnioceras*, which would suggest later flooding, in the *semicostatum* or early *obtusum* Zones, however it may be a *Caloceras*, more compatible with the Hettangian flooding suggested by Pinto-Bull (ammonite identification courtesy of Prof. Desmond Donovan, University College, London).

Two subdivisions of the Lower Jurassic graben fill are recognised in the literature (e.g. Barf  y & Gidon, 1984): the "Calcareous Lias" (Hettangian to Carixian) and the "Schistose Lias" (Domerian

to lower Aalenian). Together, these units comprise well in excess of 1km of limestone-marl alternations and marls ("Ls" and "Lc", fig. 8.4b). The eastern half-graben has largely been excavated by erosion down to the Calcareous Lias: however, some "Schistose Lias" has been preserved, especially adjacent to the Ornon fault. Near la Chalp, olistoliths of Triassic metabasalt and dolomite and Lower Jurassic limestone have been found within the "Schistose Lias" (fig. 8.7a & b), demonstrating the presence of sea-floor topography on the fault scarp at this time.

The more westerly of the two inferred Jurassic half-graben in the area has been reconstructed by Bas (1985). His model (fig. 8.8a) emphasises three good sections: Laffrey, Pont du Prêtre and St. Michael en Beaumont. The section exposed beneath Napoleon's statue at Laffrey comprises somewhat less than 15m of cross-laminated crinoidal grainstones penetrated by neptunian dykes and resting on Triassic rocks. Bas reports ammonites from various sections through these condensed "Calcaires de Laffrey" which range in age from the *liasicus* to *bifrons* Zones.

The Pont du Prêtre section is very poorly dated and the stratigraphy is crude. However, it offers the only insight into Sinemurian stratigraphy in the deep graben. The grey fine-grained limestone-marl alternations near the base of the section (fig. 8.8b)) are assigned to the "Sinemurian-Lotharigian" (i.e. early to late Sinemurian in the English sense) by Barféty et al. (1988). These are overlain by perhaps 50m of "Lotharigian" (i.e. Late Sinemurian) thick-bedded, rusty red, blocky limestones (fig. 8.9a) and over 100m of "Carixian" grey limestone-marl alternations (fig. 8.9b). Near the top of these Bas (1985) reports an *Oistoceras*, suggesting the *figulinum* Subzone. The "Domerian" comprises thicker-bedded limestones without marls (fig. 8.9c). Bas reports bivalve and echinoderm fragments from these rocks.

The St. Michael en Beaumont section is well exposed in a roadcut, rather better dated than the Pont du Prêtre section and less cryptic sedimentologically. Bas (1985) divides the section into Units D to H. Unit D comprises medium to thick interbeds of dark grey (weathering orange-brown) limestone and dark grey marl (fig. 8.10a). Beds are tabular and laterally persistent at the limited

scale of the outcrop: less than 10m. Occasional distinctive beds of crinoidal grainstone are found (fig. 8.10b), which Bas interprets as turbidites shed from Laffrey-like carbonate factories on the crests of the fault blocks (see above). Slumped units are also present. Bas interprets a *Lytoceras* found near the base of overlying Unit E (determined by R. Mouterde) to suggest a Carixian age for Unit D, but D could clearly extend down into the Sinemurian.

Unit E is yellower in overall appearance, a result of greater argillaceous content. Beds also show greater lateral impersistence with some scours. Crinoidal grainstones still occur, especially near the base. Slumps are absent. Ammonite evidence suggests that the top of the unit lies within the *margaritatus* Zone in the north-west European sense.

Units F and G comprise thick bedded and massive grainstones (fig. 8.11a). Crinoid debris is abundant, but shell fragments and, locally, wood also occur. There is some burrow mottling and cross lamination. Unit G is distinguished by the presence of metre-scale matrix-supported cobble beds, including cobbles of crystalline basement (fig. 8.11b). Unit H marks a sharp transition to marls. Bas records a *Cotteswoldia* (Late Toarcian) from near the top of Unit H as exposed and *Pleuroceras spinatum* from nearer the base. This suggests that the transition to marls takes place in Late Pliensbachian, probably *spinatum* Zone times.

In summary, two Lower Jurassic half-graben may be seen in the Bourg d'Oisans area. The thin sabkha facies of the Triassic attest to the peneplanation of crystalline basement prior to rifting. The very earliest stages of rifting were marked by basaltic volcanism and numerous small faults. Variable, but thin, shallow-marine carbonates of Hettangian age infill this small-scale topography. In *angulata* Zone times, extension focused on a small number of much larger faults which define the major half-graben. Most of the small faults were abandoned and shallow-marine carbonates were sharply overlain by marls. The graben fill is characterised by limestone-marl alternations with intercalated debris flows, slumps and crinoidal grainstone turbidites, presumably sourced from the crests of the tilted fault blocks. On close examination, however, the graben fill is not monotonous.

Both half-graben show a switch to more marly sedimentation sometime in the Domerian: apparently in the *spinatum* Zone at St. Michael en Beaumont. This is the transition from the "Calcareous" to the "Schistose" Lias. Olistoliths in the Bourg d'Oisans half-graben suggest topography on the fault scarp during this period. In the western half-graben it is also possible to demonstrate that this switch to marly sedimentation was preceded in the Late Pliensbachian by the reverse trend: an upwards change from limestone-marl interbeds to massive limestones. The Pont du Prêtre section demonstrates that a similar upwards succession from limestone-marl interbeds to massive limestones may occur in the Sinemurian.

The alternation of marl-rich and marl-poor sediments in the half-graben may be interpreted in terms of supply variations from the fault block crests. Low supply, perhaps associated with relative sea level highstand on the crests, would result in the dominance of limestone-marl rhythms- "background sedimentation" - in the graben. Carbonate factories would be active on the crests, providing thin crinoidal grainstone turbidites. Starved, but, perhaps still rotating dip-slopes could generate slumps. Fault growth under starved conditions might leave topography on the fault scarps suitable for the generation of olistoliths. In contrast, high supply from the crests, perhaps associated with relative sea level lowstands, would yield thick debris flows which would include both reworked crinoidal grainstones and, in extreme cases, reworked lithified material, including crystalline basement. As has been pointed out by Schlager (1989) one must always, with carbonates, consider the alternative explanation of highstands generating more material for the basin due to increased carbonate productivity. This alternative appears to be excluded in this case, due to the association of basement clasts with periods of high supply. The same observation tends to exclude climatic control: deep erosion without base level change on the small shoals and islands between fault blocks is somewhat difficult to envisage. Supply control to the half-graben by relative sea level change on the crests of the fault blocks could, of course, be achieved by either eustatic or tectonic means: periods of fault block rotation and consequent footwall uplift would be equivalent in their supply effects to eustatic lowstands.

The Digne Area

A number of well exposed Lower Jurassic sections are available in the area to the north of Digne (fig. 8.12a). Again, significant lateral facies changes are present and have been interpreted in terms of a shoal and basin topography (fig. 8.12b), although it is necessary to restore the intervening thrust. The condensed section at Barles and the expanded section at la Robine have been re-examined for this thesis.

The Barles section is some 250m thick. The "Rhaetian-Hettangian" is characterised by thick beds of blocky, medium grey fine-grained limestone interbedded with units comprising cm-scale, often anastomosing, interbeds of fine-grained limestone and marl (fig. 8.13a). The "Sinemurian" (i.e. early Sinemurian in the English sense) comprises rubbly, nodular limestones with pectinids and *Gryphaea*. More blocky limestones return in the "Lotharingian", commonly with siliceous nodules (fig. 8.13b). These pass up into the massive, siliceous and bioclastic limestones of the Carixian. The lowermost Domerian (*stokesi* Subzone?) is reported as missing at the top of this interval (Debelmas, 1983, see also fig. 8.13c). The base of the preserved Domerian is marly with occasional limestones, the most prominent of which reaches 7m in thickness. The Domerian culminates in a limestone of over 17m. There is a shell concentration (pectinids, belemnites and ammonites) at the base. A major hiatus is reported at the top, with only late Toarcian fossils present. This is followed by a thin late Aalenian bed and then mid-Bajocian sediments (Debelmas, 1983).

Although much expanded, the la Robine stratigraphy is remarkably similar: "Hettangian" thick bedded limestones pass up into "Sinemurian" thin bedded limestone-marl alternations and nodular limestones. Within the limestone-marl alternations there is a bed of *semicostatum* Zone age with a remarkable fossil concentration (fig. 8.14a), almost exclusively of the ammonite *Coroniceras multicosatum* (Corna, et al., 1990). The "Lotharingian-Carixian" comprises massive limestone (which is a very prominent landscape feature in the Digne area, fig. 8.14b). The

Domerian is developed mostly in marls with a prominent limestone at the top (fig. 8.14c). Lower Toarcian marls with a *bifrons* Zone (Mouterde et al., 1966) limestone are locally developed but these lap out onto an omission surface (fig. 8.14d), locally famous for its ichthyosaur remains.

Many aspects of stratigraphic development in the Digne area are shared with the Bourg d'Oisans area, 100km to the north. The "Hettangian" is relatively thin and carbonate-rich, passing into much more marly sediments in the "Sinemurian". The "Lotharingian" is more carbonate-rich in the Digne area, and possibly also in the Bourg d'Oisans area. Critically, however, there is no evidence for redeposition in the Digne area and the thick, tabular bedded, laterally persistent bioclastic limestones are more consistent with the deposits of carbonate platforms (c.f. Jaquin et al., 1991). Hence, if relative sea level lowstands were responsible for the "Lotharingian" and late Domerian limestone developments in the two areas, they probably acted in different ways: the first by reworking sediment, the second by forcing the regression of a carbonate platform. The two systems might be expected to respond to relative sea level rise in different ways. Rising relative sea level over a half-graben topography might immediately reduce reworking, but could actually promote carbonate platform development (Schlager, 1989) until the system "gave up" due to excessive rates of rise (fig. 8.16). This might explain the extensive development of a Carixian carbonate platform in the Digne area synchronously with a period of starvation in the Bourg d'Oisans area. It might also explain the apparent delay between *spinatum* Zone marls around Bourg d'Oisans and Toarcian marls around Digne.

Conclusions

- The remnants of the northern passive margin to Jurassic-Cretaceous Tethys are well preserved and well documented in the external zones of the Western Alps. The Bourg d'Oisans area shows the development of simple half-graben, the stratigraphy of the Digne area suggests more gradual topography with no redeposition and opportunities for the progradation of carbonate platforms.

- The onset of rifting in late Triassic-Hettangian times was marked by basaltic volcanism and displacement on numerous small faults. The topography was infilled by a thin transgressive carbonate sequence of Hettangian age. Major rifting began in the late Hettangian or early Sinemurian, perhaps in *angulata* Zone times. It utilised only a few of the earlier Triassic-Hettangian faults, others being abandoned.
- The rift fill stratigraphy in the Bourg d'Oisans area shows evidence for supply variations from the fault block crests. The Late Sinemurian and Late Pliensbachian appear to have been the times of high supply. These variations may be explained by pulses of block rotation and footwall uplift, or by eustatic changes.
- The stratigraphy of the Digne area reflects periods of carbonate platform progradation and flooding. Flooding is a feature of the Early Sinemurian, early Late Pliensbachian and Toarcian (fig. 8.17).

It is appropriate at this stage to compare these conclusions with those of de Graciansky et al. (1993), who have the benefit of immensely deeper knowledge and experience of French stratigraphy and who included the Barles and la Robine outcrops in their study of depositional cycles in the southern Subalpine Jurassic. The regressive-transgressive cycles described in this thesis are essentially the "second order" cycles of de Graciansky et al. (fig. 8.18). They believe that the "Sinemurian" deeper-water facies extend upwards into the *obtusum* Zone and place peak transgression in the *stellare* Subzone, though the basis of this is unclear. They recognise the Late Sinemurian - Early Pliensbachian regression, as shown in figure 8.17, but do not discuss the possibility (which has been developed in this thesis by comparing graben fill deposits from further north with those of the Digne area), that the progradation of Carixian platform limestones could be the result of relative sea-level rise rather than relative sea-level fall. The Late Pliensbachian and Toarcian events of de Graciansky et al. are identical with those described here. In addition to the unequivocal "second order" cycles, de Graciansky et al. document a number of more subtle

events within the transgressive part of the Hettangian-Sinemurian cycle and within the latest part of the Late Pliensbachian regression (fig. 8 .18).

Chapter 9. The Southern Alps

Chapters 4-8 present numerous examples of changes in accommodation space and sediment supply in the Lower Jurassic of western Europe. The Late-Pliensbachian - Toarcian sediments of the Southern Alps were briefly examined in order to see if any of these same supply changes could be detected in the pelagic realm.

Lower Jurassic half-graben, analogous to those described from the Western Alps in Chapter 8, were recognised in the Southern Alps of Switzerland by Bernoulli (1964). These half-graben are separated from their Western-Alpine counterparts by a band of ophiolites (Laubscher and Bernoulli, 1977). The Southern-Alpine half-graben are therefore interpreted to represent what was to become part of the southern margin of Tethys (fig. 8.1a). The ophiolites appear to be of Mid- or Late Jurassic age, so that in Early Jurassic times the fault block systems of the Western and Southern Alps were for the most part probably still linked (Lemoine, 1983).

Two half-graben are recognised in the area between lakes Maggiore and Como, separated by the "Lugano Swell" (fig. 9.1a & b). The section examined in this chapter is that exposed in the Breggia Gorge, between Lugano and Como (fig. 9.2). Here we see sediments deposited deep in the Monte Generoso trough, the eastern of the two half-graben. Excellent biostratigraphic control is available due to the work of Wiedenmayer (1980). The bulk of the graben fill, the "Lombardische Kieselkalk" comprises monotonous limestone-marl interbeds, commonly with bands and nodules of chert and locally with slumps and turbidites (fig. 9.2a, see also Bernoulli, 1964). The facies is similar to that described for the same interval in the St Michael en Beaumont area (Chapter 8). Weedon (1987) has analysed the bed thickness data and suggests the presence of 100Ka and 21Ka orbital cycles.

A striking change occurs in late Pliensbachian times. Sedimentation rates are much reduced and, around the base of the *subnodosus* Zone, the sediments become markedly reddened (fig. 9.3b). The minimum accumulation rate is reached in the *tenuicostatum* Zone, which is under 0.5m thick (Wiedenmayer, 1980). Above this, the sediments comprise red marls and nodular limestones (fig. 9.3c). The Domerian and Toarcian may, therefore, be interpreted as periods of truly pelagic accumulation, isolated from both periplatform carbonate and hemipelagic mud. The termination of carbonate supply may have been due to tectonic compartmentalisation or the drowning or exposure of nearby carbonate platforms (Bernoulli & Jenkyns, 1974).

Two aspects of the Breggia Gorge section - bed spacing and mineralogy - were examined in order to seek evidence for supply changes which might relate to those observed in previous chapters. In addition, the opportunity was taken to obtain a further suite of stratigraphically controlled $\delta^{13}\text{C}$ measurements.

Bed spacing data (fig. 9.4) serve to reinforce the upwards-decrease in accumulation rate through the Domerian. There is a possible increase in bed spacing during the *gibbosus* and *apyrenum* Subzones.

X-ray diffraction analysis (appendix 3) was performed on bulk powders collected throughout the section as a rapid quantitative method of assessing bulk mineralogy. No attempt was made to examine the clay fraction independently: this has been done rigorously but with a relatively coarse sample set by Deconinck & Bernoulli (1991). Carbonate levels are high throughout - around 80% - and much of the remainder of the rock is quartz (fig. 9.5). Quartz content declines through the lower part of the section, with the lowest quartz content around the lower part of the *spinatum* Zone. Surprisingly, quartz content recovers to rather high levels in the Toarcian.

The base of the section (*?ibex* Zone), the *stokesi* Subzone and the *falciferum* Zone appear to show elevated kaolinite and/or feldspar and/or dolomite, in contrast with intervening zones which

are more calcite-rich. It is possible that clays may also provide an important source of magnesium for dolomitisation. Hence this too, may be an indicator of elevated clay levels. The kaolinitic nature of the *falciferum* Zone sediments was also noted by Deconinck & Bernoulli (1991). These variations are subtle and should be confirmed by more detailed analysis including decalcification and clay separation, complex techniques which were judged to be beyond the scope of this thesis. However, there is an interesting suggestion of increased terrigenous supply to the pelagic realm in the *ibex* Zone, *stokesi* Subzone and *falciferum* Zone.

The bulk powders used for XRD were also subjected to stable isotope analysis using the methods described in appendix 3. The general trend from $\delta^{13}\text{C}$ values around +2 per mil in the late Pliensbachian to +2.5 in the Toarcian described by Jenkyns & Clayton (1986) was confirmed. However, it proved impossible to duplicate their negative excursion in the *spinatum* and *tenuicostatum* Zones (fig. 9.6a). Analytical error could immediately be excluded: a number of Jenkyns & Clayton's samples were re-run according to the procedure in appendix 3 and the earlier results were reproducible to better than ± 0.3 per mil (fig. 9.6b). Another possibility was collection error: Jenkyns and Clayton collected whilst Wiedenmayer's bed numbers were still evident on the rock. This was not true for this thesis and therefore some slippage could have occurred. It is not, however, possible simply to "slide" the two data sets into correspondence. The two data sets interpenetrate on a plot of $\delta^{13}\text{C}$ versus $\delta^{18}\text{O}$ (fig. 9.7) and there is little evidence for covariance, hence there is no obvious evidence for mixing with low $\delta^{18}\text{O}/\delta^{13}\text{C}$ diagenetic phases, nor is there any evidence that one data set is more "altered" than the other, though of course the low values of $\delta^{13}\text{C}$ observed by Jenkyns and Clayton tend towards the values of organic-rich shales, whose low $\delta^{13}\text{C}$ values may be related to the bacterial oxidation of organic matter.

The reconciliation of the two data sets remains unresolved. Jenkyns & Clayton noted over 1 per mil variation across a single 4cm sample from the *apyrenum* Subzone at Breggia which they suggested might be due to the oxidation of varying (but small) amounts of organic matter. One possibility in these finely interbedded lithologies is that sample bias might result in one data set

recording oceanographic changes whilst the other recorded changes in organic matter content. Only further sampling at bed-by-bed level could resolve this. Pending further work, it should be noted that Jenkyns and Clayton also recorded their negative excursion in the Val Ceppelline section, some 20km from Breggia and that the negative excursion has also been documented in the Valdorbia section some 400km to the south (Emmanuel, reproduced in Baudin, 1989).

Conclusions

- In general, the Late Pliensbachian and Toarcian of the Monte Generoso half-graben was a period of pelagic sedimentation with low supply of both periplatform carbonate and terrigenous clay.
- Sediment accumulation rates had reached a minimum by *tenuicostatum* Zone times due to a reduction in carbonate supply. The post-*tenuicostatum* increase in supply included a significant terrigenous clay component. That clay component had a significant proportion of kaolinite.
- Further periods of somewhat higher clay supply may be tentatively recognised in the *ibex* Zone and *stokesi* Subzone. This result would need to be confirmed using more sophisticated analytical techniques beyond the scope of this thesis.
- The rise in $\delta^{13}\text{C}$ values from Late Pliensbachian to Toarcian times reported by Jenkyns & Clayton (1986) was confirmed. It was not possible to reproduce their negative $\delta^{13}\text{C}$ excursion in the *spinatum* and *tenuicostatum* Zones. This remains unexplained but may be due to systematic sampling bias. It underlines the problems of establishing a true oceanic $\delta^{13}\text{C}$ signature from whole rock data and the need for very large sample sets (see Chapter 12).

Chapter 10. The North Viking Graben

The Viking Graben is the northern-most arm of the North Sea Jurassic rift (fig. 10.1). The emphasis in this study is on the northern end of the Graben, between 60-62°N. The principal data base comprises 68 petroleum boreholes from Norwegian waters, which penetrate the Lower Jurassic and for which wireline log data are publicly available.

Between 60 and 61°N the Graben runs approximately north-south with major normal faults throwing down towards the Graben axis on both the eastern and western shoulders (fig. 10.2). Tilted fault blocks on the western shoulders provide the structures for such oil fields as Brent, Ninian and Alwyn, whilst blocks on the eastern shoulder provide the structures for Oseberg and Brage. As with Dorset (Jenkyns & Senior, 1991), regional seismic lines demonstrate that the major growth on the North Viking Graben faults is of Late Jurassic age (Stewart et al., 1992) though, as will be seen, the same faults are also important in controlling Early Jurassic sedimentation. North of 61°N the Graben axis turns onto a SSW-NNE trend, thought to reflect underlying Caledonide structures (Stewart et al., 1992). Here the Graben is markedly asymmetrical, with major faults to the north-west (which define the structural high known as the "Tampen Spur" and create the structures for the Statfjord and Gullfaks oil fields) and a more gentle "roll-over" geometry to the south-east, rising onto the Horda Platform (fig. 10.2). At the northern limit of the study area, structures turn again north-south and extension is focused on a narrow trough known as the Sogn Graben.

The study area is characterised by some 200-300m of Lower Jurassic strata, a great deal more in the deepest parts of the Graben. The lithostratigraphy has been summarised by Vollset & Doré (1984), see figure 10.3. The lowermost unit, the Statfjord Formation, comprises largely coarse siliciclastic sediments which mark the end of Triassic-earliest Lower Jurassic continental deposition. The marine Lower Jurassic, the "Lias", is known as the Dunlin Group. Over much of

the basin it is mudrock-dominated with an important Late Pliensbachian shallow-marine sandstone known as the Cook Formation. The mudrocks above the Cook Fm. are known as the Drake Fm. In the westernmost (United Kingdom) part of the study area it has proved possible to subdivide the mudrocks below the Cook Fm. into a lower, siltier, Amundsen Formation and a higher, finer-grained, Burton Formation (Deegan & Scull, 1977; Richards et al., 1993). However, as will be seen, this distinction is difficult to apply basin-wide. In the east of the study area, sandstones are developed over a much greater stratigraphic range, extending throughout the Pliensbachian. The sandstone package beneath the Cook Formation in this area is known as Johansen Formation.

The North Viking Graben differs from the other study areas of this thesis in that the data are derived entirely from the subsurface. The advantage of such a data set is that it provides greater insights into the three-dimensional architecture of sediment bodies and the nature of shifting facies belts than is possible in most outcrop work. The disadvantages are poorer biostratigraphical resolution and the larger interpretative leaps required in what is essentially a "remote sensing" study. Because of these disadvantages it is necessary to precede a review of Early Jurassic stratigraphic development of the North Viking Graben with a detailed look at the data available and the way in which they have been interpreted.

Most published studies of the North Viking Graben Lower Jurassic concentrate on individual field areas (Dreyer & Wiik, in press; Johnson & Stewart, 1985; Livbjerg & Mjøs, 1989). Steel (1993) has developed a regional subdivision with many similarities to that described here, but based on a smaller well database which does not permit the three-dimensional mapping of depositional systems.

Data Interpretation Procedure

Natural Gamma-Ray, Sonic, Density and Neutron logs for all studied wells were plotted out and organised into a series of interlocking correlation panels (figs. 10.8 to 10.25). The locations of

these panels are shown on figure 10.7. "Quadrangles" are shown in large numbers on this and subsequent maps. Within a Quadrangle, Blocks are numbered from 1 (NW corner) to 12 (SE corner). The small numbers on the map are well numbers. Hence, by convention, the well at the NW end of correlation line "P" in figure 10.7 is 30/2-1: Quadrangle 30, Block 2, Well 1. The interpretation of wireline logs in terms of lithology is discussed in detail by Rider (1991) and summarised in figure 10.4.

The alternation, at a scale of 10's of metres, of silty and sandy intervals with more muddy, possibly occasionally organic-rich, intervals is evident on most of the logs and is interpreted as representing an alternation of more proximal and more distal facies respectively. It would clearly be possible to define log-based lithostratigraphic units and to correlate these. The problems of such a correlation are, however:

1. Unit boundaries tend to be transitional and to change character, offering numerous alternatives for the pick.

2. Unit boundaries are time-transgressive almost by definition, since the same facies transition is unlikely to take place synchronously across the whole study area (see Ager, 1983, Chapter 6 for an elegant discussion).

These problems are largely avoided by basing the correlation on the interpretation of "maximum flooding surfaces" (Chapter 2) and indeed this is the practice which has been found effective by most exponents of wireline log correlation in stratigraphy (see Galloway, 1989b; Loutit et al., 1988). Maximum flooding surfaces are generally suggested by gamma-ray maxima as a proxy for maximum clay content and/or uranium enrichment associated with poorly oxygenated bottom conditions (Wignall and Myers, 1988). The pitfalls of this approach will be appreciated from discussions elsewhere in this thesis (Chapters 4 & 5 and Appendix 1): for example, uranium enrichment is also a feature of phosphatic lags associated with winnowing, gamma trends may be

influenced by changing clay mineralogy rather than total clay content and uranium peaks resulting from anoxia may be masked by reduced terrigenous clay content associated with a distal location.

Once maximum flooding surfaces have been identified and tied around the basin it is possible to characterise the interval between the maximum flooding surfaces in order to look at lateral variations. This has been attempted for this study by two methods. Firstly, an estimate of the proportion of sand between two flooding surfaces; secondly, the description of log-based facies.

The simplest method of estimating sand content would be to use a single gamma-ray value, for example proposing that lithologies with a gamma-ray response of >80 API Units are sandy. However, comparison of intervals which are described as sandy from core material with their gamma responses (fig. 10.5a) shows that this yields extremely inconsistent results, probably due to the variable mica and/or heavy mineral content of the sands. A more reliable method is to compare the density and neutron logs as described in Rider (1991): given appropriate choice of scales the two logs will overlie when all the hydrogen nuclei (measured by the neutron log) are accounted for by water in the pore space (measured by the density log). An increase in the clay content will increase the neutron log reading due to the presence of bound water, without a concomitant increase in the porosity, hence the two curves will separate. The practice used in this study was to establish "clean sand" where the curves overlie and "mudstone" at maximum curve separation and then to take a sand cut-off half way between these two extremes.

The log-based facies scheme looks at the log signature of each cycle in terms of "progradational" (P), "aggradational" (A), "retrogradational" (R) and "complex" (C) character as interpreted from cleaning-upwards and other log motifs (fig. 10.5b). As an example, a unit comprising two stacked progradational cycles overlain by a complex interval would be described as "P2C".

A three-fold subdivision of the Dunlin Group, based on the correlation of maximum flooding surfaces, is possible throughout the study area. The three informal units have been labelled

J12/14, J16 and J18 following BP internal company practice. Correlations were tested against biostratigraphy wherever possible, but in many wells no adequate biostratigraphic data were available.

Picks for maximum flooding surfaces, estimates of sand thickness and facies codes for each well are summarised in figure 10.6. These data have been used to plot the maps described below. Before this description, however, individual correlation panels will be discussed in order to demonstrate the foundations of the subsequent interpretation.

The Well Correlations

A. (fig. 10.8) This panel runs along the crest of the Gullfaks fault-block, down the back of the block and up onto the Statfjord fault-block. The lithostratigraphy of Richards et al. (1993) is well expressed in this area and well 33/9-1 is a "reference well" in the standard Norwegian stratigraphic reference (Vollset & Doré, 1984). Identification and correlation of the maximum flooding surfaces which mark the base of J12/14, J16 and J18 is unambiguous. There is no clear expression of a maximum flooding surface at the top of J18, probably due to erosional truncation at the "mid Cimmerian unconformity" (Underhill & Partington, 1993). In the absence of a clear "Top J18" pick, the lithostratigraphic pick for the base of the Brent Group (Middle Jurassic coarse siliciclastics) has been recorded.

J12/14 in this area is a fine-grained unit. The lower part includes numerous thin limestones and from cuttings descriptions is siltier than the upper part, which comprises monotonous dark grey to reddish-grey non-calcareous mudstones (Vollset & Doré, 1984). The transition commonly takes place in two steps (labelled "A" and "B" on fig 10.8), which are expressed with differing relative prominence in different wells. This leads to ambiguity in the pick for the top of the Amundsen Fm. The transition at "A" somewhat resembles that between the Blue Lias and the overlying Shales-

with-Beef in Dorset (compare fig. 4.26) and may have a similar interpretation in terms of increasing clastic supply resulting from the progradation of clastic systems in a distal setting.

J16 comprises siltstones and sandstones, commonly arranged into one, two or three coarsening/cleaning upward cycles. The uppermost cycles commonly contain numerous thin calcareous horizons. In 34/10-1 the unit is thin and calcareous throughout, perhaps reflecting low depositional rates on the crest of the Gullfaks block. 29/3-1 shows multiple coarsening/cleaning up cycles of 10-15m in thickness.

B. (fig. 10.9) This panel illustrates further wells on the back of the Gullfaks block, mostly to the south-west of those seen in panel A. 33/9-1 is on the foot wall of the major block which forms the south-eastern boundary of the Statfjord oil field, 33/12-6 is on the hanging wall. The Lower Jurassic section can be seen to thicken gradually up the back of the Gullfaks block towards 34/10-2 at its crest: clear indication of fault control upon deposition. A threefold subdivision of the stratigraphy can still be maintained with confidence. J12/14 appears to be less calcareous and more silty than in wells on panel A, emphasising the top of the Amundsen Formation (pick "B", fig. 10.8, rather than pick "A"). As with well 29/3-1 on panel A, J16 on panel B comprises a larger number of thinner cycles in the area of this panel than in wells to the north-east.

C. (fig. 10.10). Panel C runs northwards from Gullfaks and then out along the Tampen Spur to the north-east. All three units are thin here, though easily distinguishable. J12/14 is indivisible until the section thickens slightly in 34/2-2.

D (fig. 10.11) This panel links wells on the crest of the Statfjord and Statfjord North structures, i.e. wells somewhat to the north and west of those seen previously. The correlation into 33/9-6 & 10 provides some insights into the loss of J12/14 section: it appears that the upper, finer-grained part is lost into the condensed "gamma spike" at the top of the unit, whilst the calcareous base,

with its characteristic "bee hive"-shaped log profile is preserved onto the high. 33/9-6 also shows some evidence of an intra-J16 event, evident in many wells in subsequent panels.

E (fig. 10.12) Panel E shows two further wells at the flanks of the Viking Graben with condensed sections similar to those in the previous panel.

F (fig. 10.13) This panel provides the first part of a link across the Viking Graben which is continued in panel G. At the easternmost end of this panel, well 34/8-1 shows that J16 sands are no longer organised into simple coarsening/cleaning up cycles. Sand packaging is much more complex, although 10-15m coarsening/cleaning cycles capped by calcareous beds can be discerned. Note the erosive removal of J18 from the area of 34/10-7 on the crest of the Gullfaks structure.

G (fig.10.14) Panel G crosses the Graben and shows the continuation of the complex J16 log signatures encountered in 34/8-1. Near the top of J16 there is commonly a distinctive coarsening/cleaning upwards cycle of 20-40m thickness. The pick for the top of J16 has consistently been taken above this cycle, resulting in the interpretation of J18 missing by erosion at 35/11-1. However, the fine-grained interval at the base of the coarsening/cleaning upwards cycle provides a reasonable alternative (see figure). Consistent subdivision of J12/14 is impossible in this area but in well 35/8-2, and to the south and east of this well, the unit becomes very sandy (note the convergence of the density-neutron logs).

H (fig. 10.15) Panel H, for completeness, shows the thin Lower Jurassic sequences preserved in wells in the extreme north-east of the study area, where control is available from wells drilled on the Agat accumulation. Triassic and Jurassic sequences lap out onto basement in this area so that by wells 36/1-1 & 2 the Lower Jurassic is absent.

I (fig. 10.16) This panel traces the correlation from the eastern end of panel G, southwards across the Troll Field area. J12/14 remains sandy and by the 31/2-1 well these sands have become organised into a distinctive "bow"-shaped log signature. The sands are known as the Johansen Formation and 31/2-1 is the type well for this formation (Vollset & Doré, 1984). Intra-Johansen events can be recognised (e.g. "A" and "B" in 31/2-2) which may relate to the lithostratigraphic subdivisions on the opposite side of the Graben discussed above (panel A).

The J16 interval thins somewhat from north to south and by 31/2-2 we see the return of a distinctive coarsening/cleaning upwards profile. Wells 31/6-1 and 31/6-3 are critical to the interpretation which follows as they demonstrate the pinch-out of J16, with J18 mudrocks resting directly upon the J12/14 "bow".

This panel also demonstrates the presence of thin sandstones with calcareous beds within J18, an occurrence unique to the south-eastern part of the study area which, together with a clear top-J18 gamma-ray maximum in this area, lends support to the definition of J18 as a genuine and complete regressive-transgressive cycle (see discussion of panel A, above).

J (fig. 10.17) Panel J shows further wells from the south-east Troll Field area, illustrating the absence of J16. Thin J16 is interpreted as being present in 31/6-6.

K (fig. 10.18) Panel K provides further data for the north-west Troll Field, area where all three major regressive-transgressive cycles are well developed. 31/5-2 again (see panel G) illustrates the problem of the top J16 pick, which may be taken high (2127m) or low (2172m). The former is considered to be most consistent with other wells.

L (fig. 10.19) Panel L takes the correlation westwards again from Troll, across the Brage and Oseberg blocks to Hild Field on the western margin of the Graben. It allows discussion of a few generalities before each of the areas which it crosses is examined in detail.

J12/14 decreases in sandiness westward towards the thick basinal sequence of 30/9-4. The stepwise facies changes noted in the fine-grained J12/14 sequences of the Tampen Spur area return in this well. On the opposite side of the Graben, in the Hild Field area, the "bow" signature of J12/14 reappears, suggesting that this feature is not unique to a single sediment entry point: an important observation if vertical facies changes are to be interpreted in terms of basin-wide relative sea level change rather than autocyclic processes.

A shallow pick for the base of J18 is maintained. This has the effect of including increasingly fine-grained rocks into the top of J16. Biostratigraphic data may subsequently show this correlation to be incorrect, but this would not materially affect the conclusions of this study.

M & N (figs. 10.20 & 10.21) Panels M and N provide greater detail in the Brage Field and northern Oseberg Field areas respectively.

O (fig. 10.22) Panel O runs southwards from the Brage block to the Oseberg block. Wells such as 30/3-4 and 30/6-19 lend support to the high top J16 pick as they place two clear coarsening/cleaning upwards cycles into J16, leaving a "bow"-shaped J18 similar to that in the Troll area. Continuing this scheme into the Oseberg Field area, however, does mean that only the bottom part of J16 is sandy. The sand becomes very thin by 30/9-5, at the southern end of this panel. The Cook Sand in the Oseberg area has been mapped out in detail by Livbjerg & Mjøs (1989) and their work is discussed in the next section of this chapter.

P (fig. 10.23) Panel P runs northwards from the Brage Field area and further illustrates the strong coarsening/cleaning upwards style of the J16 sandstones along this structural high. Note the relatively thick (over 15m) development of sand within the J18 unit of the 30/3 area.

Q (fig. 10.24) The two wells in this panel show the deepest part of the North Viking Graben yet penetrated. The Dunlin Group in 30/11-4 is over 600m thick. The three cycles can still be clearly recognised in this well, although J16 is expressed as a well log "bow" in mudrocks. The mid-Dunlin group sandstones present in 30/7-7 are presumed also to be of J16 age. These blocky, locally thinning-up packages of sandstone are quite different in character from the stacked coarsening-up cycles seen elsewhere within J16. Given their context, in the Graben axis, they may represent sediment gravity flows.

R (fig. 10.25) Panel R provides further detail in the Hild Field area.

Stratigraphic Development

This section describes the main features of the stratigraphic development of the study area, as determined from the log correlation exercise described above. For completeness, a brief description of the lowermost Jurassic units, Tr80 and J00 (based largely upon the work of Parkinson & Hines, in press) is also included.

Tr80-J00

Triassic sediments in the area comprise thick sequences of alluvial sandstones and shales with occasional brackish incursions, characterised by grey-brown mudstones containing green algae and rare acritarchs. The last of these incursions, lithostratigraphically the "Upper Lunde A" of Nystuen et al. (1989), has been dated as of late Rhaetian age in the Tampen Spur area (Snorre Field) on the basis of its palynomorph assemblage (Eide, 1989). It may be interpreted as the marginal marine expression of a maximum flooding surface and, as such, forms the base of the Tr80/J00 regressive-transgressive cycle. The top of the Tr80/J00 cycle (base of J12/14) is picked in marine mudstones, at the maximum flooding surface near the base of the Dunlin Group (see correlation panels, discussed above). The pick corresponds with a gamma-ray maximum, locally

associated with the development of thin limestones. It is characterised biostratigraphically by the Late Sinemurian dinocyst *Liasidinium variabile* (D. Ewen, pers. comm.)

The stratigraphy of the Statfjord Formation is summarised in figure 10.26. Its base is a major regional basinwards facies shift, with the alluvial/brackish mudstones of the Upper Lunde A overlain by the stacked braided stream sandstones of the Raude Member. Sand body density and interconnectedness locally increase upwards within the Raude Member (Johnson and Stewart, 1985), which may be interpreted as evidence of progradation. There is a relatively minor transgression near the top of the Raude Member (taken as the top of Tr80), however the overall regressive trend continues with a sharp upwards change to stacked, amalgamated fluvial channel sandstones which are assigned to the lower part of the Eriksson Member (fig. 10.26). These sandstones mark maximum regression in the Tr80/J00 cycle. The upper part of the Eriksson Member comprises more argillaceous alluvial plain sediments which Johnson and Stewart (1985) interpreted as indicative of floodplain alluviation consequent upon relative sea-level rise. These sediments are locally separated from the marine mudrocks of the Amundsen Formation by shoreface sandstones assigned to the Nansen Member. The Nansen Member is known, at least locally, to be of Sinemurian (*semicostatum* to *raricostatum* Zone) age (Copestake & Johnson, 1989). It is commonly interpreted as time transgressive (e.g. Vollset & Doré, 1984) and may be underlain by a transgressive surface of erosion.

J12/14

The J12/14 regressive transgressive cycle is picked between the inferred maximum flooding surface immediately above the top of the Statfjord Formation and the inferred maximum flooding surface immediately below the base of the coarse-grained "Cook Sandstone" interval. The latter is characterised by several biostratigraphic events understood to be of Late Pliensbachian (*margaritatus* Zone) age, including the inception of the *Nannoceratopsis gracilis/senex* dinocyst association, the extinction of the agglutinated foraminifer *Haplophragmoides lincolnensis* and an

abundance of the benthic foraminifer *Dentalina matuitina* (D. Ewen, pers. comm.). Figures 10.27-10.30 show the details of this unit, as derived from the log correlation exercise.

Figure 10.27, the total sequence isochore, shows that these fully marine rocks extended across the whole of the study area, being thickest in the axis of the Viking Graben. On the Tampen Spur, where the unit is entirely fine-grained, thicknesses tend to be consistent within an individual fault block and to increase stepwise between fault blocks, an observation which is consistent with the seismic data and also with the style of Lias mudrock deposition seen, for example, in Dorset (Chapter 4).

Thick J12/14 on the eastern flank of the Viking Graben reflects thick "Johansen" sand development (figs. 10.28 & 10.29). The sand thick in the vicinity of the Troll field is thought to locate the principal sediment entry point into the study area. Sand thicknesses in excess of 50m are maintained to the north and south of this area, though log signatures differ, tending to be more aggradational than "bow-shaped". The thick sands to the north and south of the Troll area are inferred to define a shore-parallel trend. Towards the Viking Graben, log facies become more complex and sand/shale ratios decline, suggesting a passage from inner- to outer-shelf environments (fig. 10.30). Well-developed log "bows" and somewhat higher sand thicknesses in the 30/3 & 30/6 area may reflect shoaling controlled by the underlying fault blocks.

J16

The J16 cycle is picked to embrace the coarse-grained "Cook Sand" interval. Its top is commonly a strong gamma-ray spike, associated with an abundance of algal sphaeromorph clusters, interpreted to be of mid-Early Toarcian (*falciferum* Zone) age (D. Ewen, pers. comm.)

As with the J12/14 cycle, J16 thickens towards the axis of the Viking Graben. In contrast with J12/14, however, it is *possible* to contour the isopach without invoking the known principal faults

(fig. 10.31). At the regressive-maximum of the sequence, sand is ubiquitous except for the extreme north-west of the study area (fig. 10.32).

A combination of log facies and sand thickness patterns enable some conclusions to be drawn about the paleogeography of the period (figs. 10.33 & 10.34). The stacked coarsening/cleaning cycles in the south-east of the area are interpreted to indicate that the sediment entry point identified in J12/14 times persisted into J16. In southern 31/6, however, the cycle is completely absent, suggesting sediment by-pass (see discussion of correlation panels "I" and "J", above). The sand/shale ratio thickens to the north west, accompanied by a transition to complex log signatures. The inference is that there was little proximal accommodation space and that most sand was driven offshore. Data for the deeper parts of the Graben are limited, but suggest that much of this offshore sand transport was directed northwards, perhaps deflected by shoals in the 30/3, 30/6 area. Sand may have been deposited as a "deceleration sheet" where offshore flows expanded (McCave, 1972). Little sand seems to have found its way into the southern part of the Graben axis, save for the sands found in well 30/7-7, which, it is suggested, are redeposited. The paleogeography shown in figure 10.35 seems a plausible, though clearly not unique, reconciliation of the limited data in this south-western area. A muddy, by-passed, inner shelf is suggested. Such coasts, and their sediment dynamics, have been described by Rine & Ginsburg (1985).

J18

The J18 cycle embraces the dominantly fine-grained interval at the top of the Lower Jurassic sequence. The base of the cycle is well defined (see above). The top is less well defined and frequently truncated by the base of the Middle Jurassic Brent Group (see discussion of correlation panel "A"). This truncation limits the significance of the unit isochore (fig. 10.35) for the interpretation of paleogeography.

Widespread marine shale deposition, extending even into areas previously by-passed, suggests significantly higher relative sea level than in previous cycles. The regressive maximum does introduce some thin sand, however (fig. 10.36). As with older cycles, sand is concentrated in the Troll area. The isopachs suggest the presence of isolated offshore sand bodies, though it is difficult to imagine how such sand bodies would be sourced, given that a transgressive mud blanket isolates them from the underlying Cook Sand (transgressive reworking being the common origin for offshore isolated sands: Johnson & Baldwin, 1986). It is possible that the sands represent a further period of shelfal by-pass, analogous, on a smaller scale, with the Cook Formation sandstones. It is not, however, possible to demonstrate this on the correlations presented here.

Discussion and Conclusions

Four major regressive-transgressive cycles (fig. 10.37) can be defined in the Lower Jurassic of the North Viking Graben. These are similar to Megacycles 3-6 of Steel (1993).

The Hettangian and much of the Sinemurian Stages are developed in non-marine facies and are largely progradational in aspect. A brackish incursion in late Rhaetian times defines the base of the Tr80/J00 cycle and marine transgression in Late Sinemurian times defines the top.

The J12/14 cycle is developed in marine facies. Coarse siliciclastics prograded into the basin in the Troll Field area but are assumed to have been trapped there by high proximal rates of accommodation space creation: marine mudrocks were deposited over most of the basin at this time (fig. 10.38). Progradation was gradually replaced by retrogradation. The maximum flooding surface which terminates the J12/14 cycle is of *margaritatus* Zone age.

The J16 cycle is associated with the spread of coarse siliciclastics across the whole of the Graben. Simultaneously, the Troll Field area appears to have been by-passed (fig. 10.38). These two

observations suggest a forced regression in the sense of Plint (1988). That is, a regression as a result of falling relative sea level rather than merely autocyclic progradation. This interpretation has also been advanced for the upper part of the Cook Formation by Steel (1993) on the basis, again, of sandstone extent, together with other criteria which include the relatively coarse grainsize of the sand and its sharp base (Steel suggests that it is locally erosive).

Sand deposition is terminated across the basin by a major transgression which appears to peak in the *falciferum* Zone. The gamma-ray spike with which this maximum flooding surface is associated is usually located near to the top of the Cook Sandstones and there is no evidence from the data available to include significant *tenuicostatum* Zone mudrocks above the Cook Sandstones, within the J16 cycle: the inference being that sand deposition extends into the Toarcian. As will be seen in Chapter 11, this is anomalous when compared with other European sections. It may be an artefact of poor biostratigraphic control.

Part III Synthesis

Chapter 11. Comparison of Study Areas

Synthesis of the observations in Part II of this thesis begins with an analysis of the similarities and differences between the sections and poses the question "are we right to seek a common forcing mechanism?". In order to facilitate this discussion, the simplified stratigraphic summaries for each study area have been brought together in figure 11.1.

Hettangian - Early Sinemurian

The major events of the Hettangian - Early Sinemurian are:

1. Collapse and drowning of shallow water carbonates on the Tethyan margins, associated with major rifting, around the end of the Hettangian (Bernoulli and Jenkyns, 1974). Specifically, in the case of the Bourg d'Oisans Area (Chapter 7) this appears to have taken place within the *angulata* Zone.
2. End-Hettangian cessation of space-dominated siliciclastic deposition in South Germany in favour of condensed, probably supply-dominated carbonate deposition.
3. Sinemurian marine incursion into the North Viking Graben (Chapter 10)
4. An increase in accumulation rate, associated with an increase in the proportion of mud to carbonate, in the *semicostatum* Zone of Dorset.

The common thread of these observations is an increase in the rate of accommodation space creation. This is commonly manifest by a sharp upwards change from more proximal to more distal facies, but these facies changes are not synchronous across the study area. The observations are

entirely consistent with a period of rifting, with individual rift systems operating at slightly different times. The Dorset stratigraphy, as interpreted here, is anomalous in that a sediment supply increase is suggested: an observation which one would associate with a decrease in accommodation space creation. The alternative, accommodation-space controlled interpretation of Dorset stratigraphy is difficult to sustain (Chapter 4). A possible explanation is that a local effect of rifting can be to stimulate sediment supply by footwall uplift (Chapter 2).

As well as the gross facies changes, both of the well constrained Hettangian-Lower Sinemurian sequences, South Germany and Dorset, show evidence of more subtle variations at zonal and subzonal frequencies. The *angulata* Zone appears to be the most regressive part of the Lias α_{1-2} in south Germany. It is also the most condensed part of the Blue Lias in Dorset, as suggested by bed thickness data. Again, there are similarities over continental distances but a common accommodation-space model for the two stratigraphies would require a relatively unattractive space-dominated interpretation of the Blue Lias.

Late Sinemurian - Early Pliensbachian

The Late Sinemurian is marked by the introduction of relatively coarse siliciclastics in Yorkshire (the Siliceous Shales). Redeposited and platform carbonates overlie marls in the Western Alps. The Dorset and South German sections are notable for their two remarkably synchronous omission surfaces, with deposition taking place only in the early *obtusum* and early *raricostatum* Zones. If the omission surfaces are interpreted as due to lack of accommodation space, rather than starvation, then all of these stratigraphies can be reconciled, at least qualitatively, by postulating falling sea-level. This brings down more proximal facies in the more highly subsident areas and leads to winnowing in the less subsident areas.

The stratigraphy of the Early Pliensbachian is less straightforward to reconcile, but can largely be interpreted in terms of sea-level rise (or less rapid fall). This would explain the Pyritous Shales of

Yorkshire, the post "Unit B" transgression in Portugal and the condensed limestones of the Lias γ in South Germany. It would also explain the Belemnite Marls of Dorset if these are taken to represent a facies more distal than the underlying Black Ven Marls (fig. 4.25). In order to sustain the reconciliation, one has to postulate that the Early Pliensbachian platform carbonates of the Alps were able to "keep up" with the postulated sea level rise and hence do not record transgression.

In Chapter 5 it was suggested that the lower Ironstone Shales of Yorkshire represent a relative base-level fall. Dorset stratigraphy is so strikingly similar that one would wish to postulate similar events there. One might infer that this fall was relatively subtle within the context of the general rise discussed above, as it does not introduce significant coarse siliciclastics in either area, in contrast to the events of the Late Pliensbachian (see below). As discussed in Chapter 6, there is some spectral gamma-ray evidence for a contemporaneous regression within Unit D in Portugal and this event may also be marked by the "Johansen" (J12/14) regression of the North Viking Graben (Chapter 10).

Late Pliensbachian

The regressive style of the Late Pliensbachian is in strong contrast to overlying and underlying sediments in many areas and provides amongst the most persuasive evidence for a pan-European forcing mechanism. In general terms, this is the time of the first sand in the Dorset section (The Three Tiers) and the Yorkshire Section (Staithes Sand). Major regressive sandbodies are also developed in the North Sea (Cook Sandstone). The supply-dominated sections of South Germany see an increase in mud supply (Lias δ). Relatively shallow water carbonates are developed in Portugal ("Unit E") and in the Western Alps (with shedding into the deep half-graben).

In detail, the interval is rather complex (fig. 11.2). The most unequivocal and best-constrained section, at least with regard to the *margaritatus* and *davoei* Zones, is Dorset, where spectral gamma-ray data from the Green Ammonite Beds (*davoei* Zone) show a clear upwards increase in terrigenous input, possibly in two steps. This culminates in three unequivocal cycles of siliciclastic transgression-progradation in the *stokesi* and *subnodosus* Subzones, each apparently more regressive than the previous one. The final zone of the Pliensbachian is marked in Dorset by the condensed Marlstone facies. The facies is difficult to interpret but may be a manifestation of relative sea-level fall where terrigenous supply is unavailable: perhaps on the crest of an isolated fault block. Hence the facies may be a response to the same relative base level fall which stimulated the Unit E limestones of Portugal, the end Pliensbachian limestones of the Western Alps and the Cook Sands of the North Sea, which also seem to be of *margaritatus* Zone age or younger.

The *davoei* Zone is also unambiguously regressive in Yorkshire, although the first sand is older: around the *capricornus-maculatum* Subzonal boundary. The late *stokesi* and *subnodosus* Subzones in Yorkshire see the replacement of the Staithes Sandstone with the Cleveland Ironstone facies. The strong polarity of the Cleveland Ironstone cycles: transgressive marls and regressive iron-rich limestones was emphasised in Chapter 5. They may equate with the subzone-wavelength cycles of Dorset, but represent much less net accommodation space creation in each cycle, presumably due to lower total subsidence rates.

In Portugal, the *davoei-margaritatus* Zone interval, equivalent to the Green Ammonite Beds to Thorncombe Sands of Dorset, is developed entirely in the marls of "Unit D". Spectral gamma-ray studies suggest that the interval is regressive, as in Dorset. The sharp upwards change to the more proximal facies of Unit E occurs within or at the top of the *margaritatus* Zone: certainly younger than the *stokesi* Subzone, and hence younger than the first sand in Dorset or Yorkshire. It is possible that the two more marly intervals within Unit E (Chapter 5) reflect something of the subzone-frequency variation noted in Yorkshire.

In South Germany, the sudden influx of terrigenous mud, which may, as discussed above, be associated with relative sea-level fall, occurs at the base of the *margaritatus* Zone. Maximum regression within the Lias ∂ marls, from spectral gamma-ray evidence, appears to have been reached in the mid or late *margaritatus* Zone.

In the Western Alps section at Barles, where age control permits some detailed comparisons, massive carbonate development was finally terminated at the end of the *davoei* Zone and there is a hiatus. The *margaritatus* and *spinatum* Zones demonstrate renewed flooding, with at least one minor cycle before the major end-Domerian carbonate.

There are two approaches to the synthesis of the above observations, which might be described as the "lowest common denominator" and "highest common multiple" approaches. The former, and most conservative approach, certainly supports Late Pliensbachian regression. The most marked basinwards facies shifts associated with that regression span a range of dates between the base of the *davoei* Zone (the base of the Green Ammonite Beds in Dorset) and the base of the *spinatum* Zone (the approximate base of the Unit E limestones in Portugal). Most sections show evidence for more than one cycle and, in particular, there is a tendency for facies shifts at the base of the *margaritatus* (Dorset and South Germany) and the base of the *spinatum* Zones (Dorset, Portugal and probably the North Viking Graben). The "highest common multiple" approach would seek evidence to equate multiple cycles within the regression. Following this approach we have:

1. Maximum transgression in the late *ibex* Zone of Dorset, Yorkshire and Portugal.
2. A sharp basinward-shift at the base of the *stokesi* Subzone. The evidence from Dorset and South Germany is unequivocal, but we might also suggest an increase in terrigenous input in Portugal (fig. 6.18) and note the most proximal facies of the Staithes sandstone (fig. 5.13).

3. Transgression followed by renewed regression by the end of the *stokesi* Subzone: demonstrated by the Eype Clay-Down Cliff Sand cycle of Dorset, but also perhaps manifest in the first Cleveland Ironstone Cycle of Yorkshire. The suggested decrease in clastic input on the gamma-ray log of Portugal and the *margaritatus* Zone limestones of the Barles section may represent this cycle or "3", below.

4. Transgression followed by renewed regression within the *subnodosus* Subzone: demonstrated by the Blue Band-Thorncombe Sand cycle of Dorset, but perhaps also manifest as the *subnodosus* cycle of the Cleveland Ironstone, and in Portugal and France as noted above.

5. *gibbosus* Subzone transgression and regression. This is really only manifest in the Cleveland Ironstone. The other sections are best interpreted as:

6. *gibbosus* Subzone transgression followed by *spinatum* Zone regression, as suggested by the Marlstone Rock succession of Dorset and, as discussed above, our best estimate of the timing of the sharp basinwards facies shifts in Portugal and the North Sea.

This high-frequency cyclicity is verging upon one cycle per subzone. Cyclicity at this frequency is notable in many of the sections in this thesis and it is tempting to speculate that the very faunal turnover which provides our measure of time is linked in some way to the sedimentary packaging: a blurring of the distinction between lithostratigraphy and chronostratigraphy which would have been familiar to 19th century geologists (Chapter 2). There are problems, however, in correlating cyclicity at this frequency. The proximal facies are commonly non-fossiliferous and one is faced with successive transgressive events each containing a different fauna. Such observations cannot distinguish between cycles which are in phase or 180 degrees out of phase. This is analogous to the problem of aliasing (e.g. Pias & Mix, 1988). That is, in order to correlate cycles, temporal resolution of at least half the cycle wavelength is required. None the less, there is a

strong suggestion that, as proposed by de Graciansky et al. (1993), there is more to the late Pliensbachian story than simply inter-basinal evidence of regression. There is also a common thread of higher-frequency cyclicity which would warrant further investigation.

Toarcian

A sharp transgressive facies shift at or near the base of the Toarcian is one of the most striking features of Lower Jurassic stratigraphy in Europe (see also Hallam, 1981). In Yorkshire and South Germany the most distal facies may be represented by the organic-rich rocks of the *falciferum* Zone, a Zone which is marked globally by exceptionally high levels of organic carbon preservation (Jenkyns, 1988 and Chapter 12). Limited grainsize data for Yorkshire (Gad et al., 1969) might suggested a somewhat later transgressive maximum, in the *bifrons* Zone. In the North Viking Graben, the prominent Toarcian gamma-ray maximum appears to be of *falciferum* Zone age (Chapter 10). In the Western Alps, whilst the expanded sections clearly show the contrast between late Pliensbachian limestones and Toarcian dark marls, the timing of the most distal facies is uncertain. In Portugal, whilst the basal Toarcian facies shift is distinctive, the deepest water was probably attained in *bifrons* (Unit G) times, though, as noted in Chapter 6, this probably represents a period of significant rifting in the area.

The Yorkshire Toarcian shows gradual coarsening-upwards from mid-*variabilis* Zone times onwards, culminating in sandstones in the *levesquei* Zone. The *variabilis* Zone is the base of the prograding carbonate turbidites of Portugal and sees the return of more "normal" marly sediments following deposition of the Posidonienschiefer in Germany. In Dorset there is a sharp increase in sediment accumulation rates at the base of the *levesquei* Subzone, culminating in Bridport Sand deposition by the middle of the *levesquei* Zone. The late Toarcian is thus generally marked by regression.

Locally, there are hints of more subtle events within the late Toarcian regression: the distinctive lower *variabilis* Zone sediments of Yorkshire, the suspected mid- to Late Toarcian regressive peak in the North Viking Graben and the *bifrons* Zone limestone in the Western Alps have been mentioned above. A top *bifrons* eustatic fall has been suggested by Hallam (see Chapter 3) and three late-Toarcian sequence boundaries have been interpreted by Haq et al (1988). The *variabilis* Zone is also a time of high siliciclastic supply to South Germany (Ulrichs, 1977). In the light of this knowledge we may return to the Peak Shales of Yorkshire (Chapter 5) perhaps slightly more confident of a regressive interpretation.

Discussion & Conclusions

An attempt at a qualitative synthesis of these observations discussed above is presented in figure 11.3, which also shows the curves of Haq. et al (1988) and Hallam (1988) for comparison. The evidence is compelling for a common Late Pliensbachian regression, Early Toarcian transgression and Late Toarcian regression. Common motifs in the earlier Lower Jurassic demand more contrived interpretations of the stratigraphy, but suggest Hettangian and Early Sinemurian transgression, Late Sinemurian regression, and transgression in the earliest Pliensbachian with at least one well defined Early Pliensbachian regressive-transgressive pulse within the overall Pliensbachian regressive trend. These common elements of western European Lower Jurassic stratigraphy present a strong argument for a regional stratigraphic forcing mechanism operating at around stage-frequency ("second order" in the sense of Vail et al., 1991). The arguments become less compelling with higher frequency events, though there are parts of the stratigraphy where concerted events of a frequency less than 3Ma ("third order" in the sense of Vail et al., 1991) can be demonstrated: the two Late Sinemurian omission surfaces and the complexities of the Late Pliensbachian being the best examples. The observations serve to emphasise two important aspects of the Lower Jurassic stratigraphic record (compare Vail et al., 1991):

1. There is no clear hierarchy of cycles in the sense that low amplitude, high frequency cycles are superimposed on the back of high amplitude, low frequency cycles. Rather, the curve is ragged, with particular higher-frequency events having considerable impact upon the stratigraphy: for example in the Late Sinemurian and Late Pliensbachian.

2. There is no suggestion that "second order" cycles are a result of local basin-forming processes and that "third order" cycles are of more regional importance. Evidence suggests that the converse is true.

A further relevant generalisation is that, whilst sharp basinwards facies shifts are common on the regressive limbs of the inferred regressive-transgressive cycles, these shifts (candidate sequence boundaries in the scheme of Van Waggoner, 1990) are commonly not synchronous between basins.

Chapter 12. Carbon Isotopes: An Independent Sea Level Curve?

Chapter 11 reached the conclusion that the different study areas had sufficient similarities to justify seeking a pan-western European (at least) stratigraphic forcing mechanism. The dangers of working back from the output of the "stratigraphic machine" to its inputs were discussed in Chapter 2. The individual regressive-transgressive curves of figure 11.1 represent an insight into relative sea-level change and sediment supply in individual basins. The abstraction of figure 11.3 cannot be taken as a sea-level curve, but merely as an insight into the nature of the stratigraphic forcing function, as yet poorly known. In this chapter a further, independent and quantitative insight into that function is offered, in the form of a carbon isotope curve for the Early Jurassic.

The measurement of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ was one of the methods used by Jones (1992) to test for diagenetic alteration of the skeletal carbonate which he was using to derive a $^{87}\text{Sr}/^{86}\text{Sr}$ curve for Jurassic-Cretaceous seawater (Appendix 2). A by-product of Jones' work was therefore a suite of measurements of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ on Jurassic and Cretaceous belemnite and oyster material, mostly from the Dorset and Yorkshire sections described in this thesis. Inspection of this dataset for the Lower Jurassic (fig. 12.1) shows that, despite considerable scatter, there is clear cyclicity with the stage-length periodicity of the postulated stratigraphic forcing function (Chapter 11). There are also indications of, for example, Early Toarcian and mid-Sinemurian peaks. This raises the possibility that carbon isotope data might offer an independent approach to the Early Jurassic stratigraphic forcing function and indeed, as will be discussed below, there is reason to suppose that the carbon isotope curve may reflect sea-level stand.

In this chapter we first look at the validity of Jones' UK data as a measure of the carbon-isotopic composition of Early Jurassic seawater. Comparisons are made with data from elsewhere in western Europe to seek further support for the trends in the UK data set. Secondly, we look at the

theoretical basis for the curve, addressing the issue of what is being measured and how it might relate to sea-level. Thirdly we compare the curve with qualitative sea-level curves for the Early Jurassic proposed by other authors and with the results of Chapter 11. Analytical details will be found in appendices 2 and 3

Validity and New Data

Initial encouragement may be derived from the fact that the bulk of Jones' measurements (fig. 12.2a) cluster around the slightly negative $\delta^{18}\text{O}$, slightly positive $\delta^{13}\text{C}$ values determined for Tethyan pelagic carbonates by Jenkyns & Clayton (1986), themselves compatible with general values for marine carbonate (Arthur et al., 1983). Points straying from this tight cluster could be taken to indicate diagenetic alteration. This would be especially true of negative outliers which might reflect calcite precipitation associated with higher temperatures and/or isotopically light diagenetic fluids (Jenkyns & Clayton, 1986). Such fluids may be associated with some meteoric waters (Arthur et al., 1983) and the degradation of organic matter during burial (Irwin et al., 1977). Only six data points are clearly anomalous on this basis, and these have been excluded from subsequent analysis.

Jones (1992) examined the possibility of using his determinations of $^{87}\text{Sr}/^{86}\text{Sr}$ and iron and manganese concentrations as tracers of diagenetic alteration where stable isotope concentrations had not been reset to an extent detectable by the $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ crossplot. He suggested that elevated Fe and Mn concentrations might reflect the more advanced stages of organic matter diagenesis; Fe and Mn being common in diagenetic fluids derived from siliciclastic rocks and being precipitated in carbonate phases during methanogenesis and decarboxylation (due to the absence of reduced sulphur). If the data are plotted showing iron content (fig. 12.2b) it is not clear that samples with elevated iron levels depart in any significant way from the trend of the rest of the data, except that three of the four samples which clearly do not fit with the bulk of the data set have iron concentrations greater than 300ppm and the fourth has iron greater than

150ppm. Again, on purely empirical grounds, these samples have been excluded from further analysis, taking the total number of samples excluded to 10 (<6% of the total).

In the end, the most powerful evidence for the data representing a primary signal is the existence of coherent trends across several hundred metres of stratigraphy and independent of the primary lithological subdivisions. Of course, the scatter is considerable: approximately 2 per mil. This may be a result of partial diagenetic alteration (Spaeth et al, 1971) or original biological fractionation, perhaps related to taxonomy (Sælen & Karstang, 1989). The Dorset and Yorkshire data are separated in figure 12.3a. The two datasets complement each other: the early part of the section is not available in Yorkshire (Chapter 5) and the late Pliensbachian-early Toarcian is highly condensed in Dorset (Chapter 4). Where the curves overlap, there is good correspondence, with a mid-Sinemurian high, end Sinemurian low and mid-Pliensbachian high. The Yorkshire data tend to plot slightly higher than the Dorset data which Jones (1992) suggested might be due to the incorporation of more methanogenic carbonate in Yorkshire.

Portugal provides good coverage of the Pliensbachian, confirming the low values at its base, rising to a Toarcian peak (fig 12.3b). There is a suggestion that this is not a smooth rise, but that there is an intermediate peak around the *davoei* Zone (191Ma). Unfortunately the detail in this area of the curve is not well resolved due to poor foreshore exposure (Chapter 6). The German data (fig. 12.4a) provide further evidence for the rising trend in the Pliensbachian and for a *davoei* Zone excursion.

Finally, Jenkyns & Clayton's (1986) whole rock data can be seen to be part of the same picture (fig. 12.4b). Their *falciferum* Zone excursion is rather more strongly expressed in the belemnite data than in their whole-rock studies: this is in accord with the slight diagenetic alteration which they detect and the inference that their samples represent minimum values for Toarcian seawater. The controversial negative excursion which Jenkyns and Clayton (1986) suggest for the *spinatum*

and *tenuicostatum* Zones (Chapter 9) is also supported when all the available data are posted together.

An attempt has been made to draw out trends in the data by fitting a smoothing spline (fig. 12.5). The computer package employed for this uses a smoothness parameter, λ . As the value of λ decreases, the error term of the fit has more weight and the fit becomes more curved (SAS Institute, 1989). The method provides an objective approach to minimising the deviation of data from the fitted curve, but the choice of λ (in this case is $\lambda=1$) is a subjective judgement about what data trends are real.

To summarise, the tight clustering of the data without obvious "exotic" isotopic compositions and the systematic variation with age strongly support the inference that the data presented represent the primary isotopic composition of Early Jurassic seawater. This is further supported by the repeatability of the observations across the various study areas in western Europe. This repeatability simultaneously suggests that the carbon-isotope curve is of at least plate-wide significance. A further important perspective is that marked positive excursions, such as that in the *falciferum* Zone of the Toarcian, which have been labelled "Oceanic Anoxic Events", and which are known to be associated with regional organic-carbon burial (Schlanger and Jenkyns, 1976) may be seen as part of a continuum of change in oceanic carbon isotope composition rather than as anomalies set against a stable background.

Theoretical Basis

The carbon cycle has been the subject of much recent study due to concern over the environmental effects of rising atmospheric carbon dioxide levels. A version of the cycle is shown in figure 12.6. An increase in the $\delta^{13}\text{C}$ value of the oceans would require more isotopically light carbon to be sequestered in one of the light carbon reservoirs. The principal ways of achieving this would be by:

1. Increased organic carbon burial
2. Increased biomass
3. Increased atmospheric CO₂

Alternatively, small increases in the present day $\delta^{13}\text{C}$ of the oceans could be achieved by adding large masses of the slightly heavier carbon currently sequestered in carbonate rocks. This is not, however, a feasible mechanism for effecting the 3-4 per mil excursions which we seek to explain here.

If the mass of dissolved carbon in the oceans is taken as $35 \times 10^{18}\text{g}$ (Arthur et al., 1983), then the size of the necessary mass transfers to change the isotopic composition of the ocean can be calculated using the following equation, which simply states that the sum of the volumes of each reservoir multiplied by their isotopic compositions must remain constant before and after the transfer.

$$35\delta^{13}\text{C}_{\text{initial}} = T\delta^{13}\text{C}_{\text{sink}} + (35 - T)\delta^{13}\text{C}_{\text{final}} \quad (1)$$

or

$$T = \frac{35(\delta^{13}\text{C}_{\text{initial}} - \delta^{13}\text{C}_{\text{final}})}{(\delta^{13}\text{C}_{\text{sink}} - \delta^{13}\text{C}_{\text{final}})} \quad (2)$$

Where $\delta^{13}\text{C}_{\text{initial}}$ is the initial isotopic composition of the ocean
 $\delta^{13}\text{C}_{\text{sink}}$ is the isotopic composition of the sink from/to which material is
being transferred
 $\delta^{13}\text{C}_{\text{final}}$ is the final isotopic composition of the ocean
T is the mass transferred in $\text{g} \times 10^{18}$

Solution of this equation for the likely sinks of isotopically light carbon (fig 12.6) shows that, in order to effect each 1per mil increase in the $\delta^{13}\text{C}$ of the oceans, some $1.2 \times 10^{18}\text{g}$ must be transferred to -23/-25 per mil carbon sinks (organic carbon and the biosphere). To effect the same change using sinks of only -7per mil (the atmosphere) requires transfers of some $4 \times 10^{18}\text{g}$. The latter figure represents a more than seven-fold increase in atmospheric carbon dioxide. If the entire effect were to be produced by expanding the biosphere, it would require a three-fold increase in biomass for each 1per mil change in oceanic $\delta^{13}\text{C}$.

If the effect is to be achieved using organic carbon burial then the numbers, whilst still large, are not quite so dramatic. Assuming 2% Total Organic Carbon and a density of 2.7gcm^{-3} , $1.2 \times 10^{18}\text{g}$ of organic carbon represents approximately $22 \times 10^{12}\text{m}^3$ of rock or deposition of some 4m of rock over an area the size of Europe. Such changes appear to have been effected over periods of perhaps 1Ma. Given the global extent of organic rich rocks in the early Toarcian (Jenkyns, 1988), together with the fact that many of these rocks have organic carbon contents well in excess of the 2% suggested above, and that the oceanic record from the period has been lost due to subduction, these numbers for organic carbon burial seem entirely realistic.

Given that the most realistic mechanism for effecting 3-4 per mil shifts in oceanic $\delta^{13}\text{C}$ is probably by increased organic carbon burial, what are the controls that burial? In particular, which of those controls could give rise to parallel effects in the $\delta^{13}\text{C}$ record and in the stratigraphic signature of the Early Jurassic (as discussed in Chapter 11 and portrayed in figure 12.7)?

The controls on organic carbon preservation are complex. In the marine realm the basic requirement is a flux of organic matter to the sea bottom which is sufficiently great that demand for oxygen for its oxygenation exceeds the available supply. To this requirement we can probably add the need to take the organic matter out of the zone of oxidation relatively rapidly by burial (Demaison & Moore, 1980), though in some circumstances high levels of organic matter preservation may be associated with low sedimentation rates, perhaps because highstands of sea

level, which potentially increase the flux of organic matter and decrease bottom-water oxygen levels, also decrease sediment supply (Loutit et al.,1988). Examples of organic carbon enrichment which may have occurred at relatively low (Black Ven Marls, fig. 4.26) and relatively high (Blue Lias, fig. 4.26; Mulgrave Shale, fig. 5.26) stands of sea-level will be found in this thesis.

The flux of organic matter to the sea bed is a function of organic productivity and the consumption of organic matter by oxygenation in the water column. Productivity may be enhanced by increasing the nutrient supply, which is potentially sensitive to such factors as upwelling, hinterland climate and runoff. Critically for this study, productivity may also be enhanced by the development of broad shelves (Jenkyns,1988), which might be expected during periods of relatively high sea-level stand. Consumption of organic matter in the water column leads to an oxygen minimum zone, for example between 150-1000m in the modern Atlantic (Kennet, 1982). This zone may have been much expanded in an ice-free world without sources of cold, well-oxygenated polar bottom waters, and may have overlapped onto the shelf during periods of high sea-level stand (Schlanger and Jenkyns,1976).

Various models have been proposed to create bottom-water anoxia. A prime requirement is clearly to be beneath wave and storm mixing. Additionally, basin restriction is probably important. Present-day oceanic circulation patterns mean that the Indian Ocean is a "cul-de-sac", with far-travelled, oxygen-depleted waters. The Early Jurassic Tethys may have had a similar pre-disposition to anoxia (fig. 3.1). Bottom-waters in restricted basins may become stagnant and oxygen-depleted, either due to a buoyant freshwater cap (from fluvial input), or due to evaporation, feeding the bottom with dense, saline waters (Fleet et al, 1987).

Comparison with Qualitative "Sea-Level" Curves

This section compares the $\delta^{13}\text{C}$ record of the Early Jurassic with three studies of the Early Jurassic stratigraphic signature: those of Hallam (1988, 1981) and Haq et al (1988), and that

developed in Chapter 11 of this thesis (fig. 12.7). Whilst, as previously discussed, the interpretations of Hallam and Haq et al. are not always the interpretations preferred here, it is hoped that comparison with the work of other authors' lends some objectivity to the debate.

Correspondence with Hallams's curve is most immediately striking. The end Hettangian and end Sinemurian lows are well reproduced, as is the small cycle between lows at the top of the *davoei* and *spinatum* Zones, though there is much scatter in the $\delta^{13}\text{C}$ record here. The *falciferum* Zone peak and subsequent decline can also be clearly seen on both curves. The curves look rather different in the late Sinemurian, but this may easily be explained using rate effects: Hallam has interpreted sea level lowstands at the horizons of the "Hummocky" and "Coinstone" in Dorset (Chapter 4), but these horizons could well reflect times at which the rate of sea level fall is at a maximum (as suggested by slope of the $\delta^{13}\text{C}$ curve), times when base level is likely to interact with the sediment-water interface in areas of relatively low subsidence rate.

Comparison with the Haq et al. curve is less easy. The best starting point is to look at the major sequence boundaries, which are those identified in the early seismic-based studies (c.f. Vail, 1981). "202" represents the end Hettangian lowstand, "195" the end Sinemurian lowstand, "188.5" the top *davoei* Zone lowstand and "177" the end Toarcian lowstand. The end Pliensbachian lowstand on the $\delta^{13}\text{C}$ curve is recorded only as a minor sequence boundary ("186.5") on the Haq et al curve. It is possible to interpret each of these sequence boundaries to be reflected in declining values of $\delta^{13}\text{C}$.

The Sinemurian comparison is confused by Haq et al's interpretation of the top of the Blue Lias ("199.5") and the Coinstone ("196.5") as downlap surfaces (see Chapters 4 and 11). This presumably tempts them to insert a sequence boundary between the two, which they do in the *turneri* Zone. There is little stratigraphic evidence for this sequence boundary in the sections examined for this thesis. The *turneri* Zone marks a peak on the $\delta^{13}\text{C}$ curve.

The "191" sequence boundary does not feature on Hallam's curve. Haq et al. derive it by interpreting the Belemnite Beds of Dorset and Yorkshire as representing a time of sea level fall. This is supported in this thesis, see Chapters 4, 5 & 11, but is not a feature of Hallam's work. The $\delta^{13}\text{C}$ curve is rising at this time.

The general correspondence between the $\delta^{13}\text{C}$ curve and the key events of the Early Jurassic stratigraphic record is striking. It is tempting to suggest some form of common control, or controls, the two most obvious candidates, not necessarily independent, being rifting and sea-level. Rifting was undoubtedly a feature of the Hettangian-Sinemurian and probably also of the Toarcian (Jenkyns & Senior, 1991). It is likely to have resulted in both basin compartmentalisation, with consequently reduced circulation, and flooding, with increased shelfal area. These effects could be reflected both in the sequence-stratigraphic history and the record of organic-carbon burial, as discussed above. Regressive intervals in the stratigraphic record appear to be marked by declining volumes of organic-carbon preservation, potentially due to reduced shelfal areas and basin oxygenation.

Chapter 13. Conclusions

The principal aim of this thesis has been to document and compare the sequence-stratigraphic interpretations of seven contrasting Lower Jurassic sections in western Europe.

A corollary of the excellent biostratigraphic control provided by the pelagic ammonite fauna of the Lower Jurassic is that sequence-stratigraphic interpretations of the distal facies are often ambiguous. The differences between previous authors are due, not to differences over which surfaces are important, or to differences over correlation, but to differences in the interpretation of the sequence-stratigraphic significance of the key surfaces. Some progress has been made in this area by the addition of new data, particularly using spectral gamma-ray studies, which appear, via Th/K ratios, to provide an empirical index of proximal-distal relationships. Further work on the significance of Th/K ratios in modern and near-modern sediments where sediment transport pathways are better understood would greatly enhance the use of this technique. Where ambiguities of interpretation remain, it is hoped that they have been fully exposed for debate. Previous discussion has also placed heavy reliance on the interpretation of the Dorset and Yorkshire stratigraphies. Whilst these are important, it is hoped that this thesis has illustrated the diversity of stratigraphy which must be accommodated within a pan-European stratigraphic model.

The principal result of the long-range comparisons presented here is that the stratigraphies of the different study areas are not independent products of local basin history, but rather that significant aspects of stratigraphic development are "orchestrated" on at least a plate-wide basis. This "signature" of the European Early Jurassic has been discussed in Chapter 11 and summarised in figure 11.3. The signature of the early Lower Jurassic is rift-associated transgression. This is reversed during the Late Sinemurian. The Early Pliensbachian tends to be transgressive relative to the Late Sinemurian and Late Pliensbachian. The Late Pliensbachian is everywhere a time of significant regression, with some evidence that the regression was "forced" by an actual relative

sea-level fall (see Chapter 10) and some suggestion that the long-range correlation of subzone-frequency events may be possible. The well-known mid-Toarcian flooding event (Jenkyns, 1988) finds wide expression in the studied sections.

New carbon isotope data have been derived from the measurement of belemnite material. These data flesh-out the data of Jones (1992) and suggest that the carbon-isotopic composition of Early Jurassic seawater is closely linked to the stratigraphic signature of the Early Jurassic as suggested not only in the present work but in the work of Haq et al. (1988) and Hallam (1988). The reasons for this may be that sea-level highstands, reflected in the stratigraphy, also promote sequestering of organic carbon by creating broad and highly productive shelves (Jenkyns, 1988). A further factor may be basin compartmentalisation associated with rifting.

It must be remembered that, in looking at the Early Jurassic stratigraphic signature and at the carbon isotope curve, we are looking at two products of the "stratigraphy machine" (Smith, 1994). They are likely to be the complex output of a number of inputs, one of which may be eustatic sea-level change, others of which may be regional tectonics or climatic change. What can be demonstrated is that, whatever the principal forcing mechanism(s) is(are), they operate on at least a plate-wide basis. The observed output curves have, in general, a rather long period (5-10Ma) but are rather "ragged", with no correlation between period and amplitude. This is in contrast with the common sequence-stratigraphers simplification of high-amplitude/long-period, "second order" curves, held to be a product of local tectonics, convolved with low-amplitude/short-period, "third order" events which are suggested to be eustatic (Vail et al., 1991). The observations for the Early Jurassic, presented here, suggest that it is the "second order" effects which are pervasive, though further work will be required to see if they are merely pan-European or truly global. The Sr-isotope results described in this thesis demonstrate the potential of this technique to provide an independent chronometer which would be potentially invaluable in such long-range correlation studies.

Appendices

Appendix 1. Gamma-Ray Spectrometry

Portable gamma-ray spectrometry was found to be a useful supplement to conventional sedimentological logging where subtle vertical variations were sought. The method is particularly useful in mudrocks, where it provides objective and quantitative insights into clay content, clay composition and (via uranium content) organic matter concentration. An additional advantage is that it facilitates comparison with borehole data sets where natural radioactivity is routinely recorded.

The portable gamma-ray spectrometer is widely used in mineral prospecting. Its rarity in sedimentological studies to date justifies some theoretical discussion. Further details will be found in Myers (1987).

Principles of the Technique

The principal naturally-occurring gamma-emitters are Thorium (Th^{232}), Potassium (K^{40}) and Uranium (U^{238}). K^{40} emits gamma rays centred on only one energy level (1.46MeV) during its decay to Ar^{40} . Both Th^{232} and U^{238} have more complex decay chains, which result in the emission of gamma rays of a number of different energies. In practice, the spectrum of gamma ray energies presented to a detector is further broadened by interactions between emitted gamma rays and the surrounding rock and atmosphere, for example by Compton Scattering, resulting in the energy spectra illustrated in figure A1.1a). The gamma-ray spectrometer operates by measuring the magnitudes of scintillations produced as gamma rays are absorbed by a suitable crystal connected to a photomultiplier. Scintillation magnitude is proportional to gamma ray energy.

Measurements in this study were made using an Exploranium GR-256 spectrometer (serial no. 1523, property of BP Research). This uses a sodium iodide crystal detector and sorts the detected scintillations into one of 256 equally-sized energy classes ("channels 1-256"). The detector incorporates a Cs^{137} reference source which emits gamma rays centred on 662KeV. This peak in the spectrum is automatically assigned to channel 55 which means that the channels then measure energy levels of 0 to 3.11MeV in steps of approximately 12KeV. The Cs source thus acts as an automatic gain control, protecting against variations due, for example, to temperature changes or component wear. Note that for long counting times the Cs source can saturate the peak channel so that the Cs peak breaks into two (fig. A1.2). This can affect the accurate assignation of the Cs peak to Channel 55 and be a major source of error on some Exploranium instruments (D. Talbot, pers. comm.). Tests at the British Geological Survey calibration facility demonstrated that this was not a problem with the instrument used here.

There are essentially three steps in the conversion of the recorded spectrum to element concentrations:

1. A suitable window is selected which isolates, so far as is possible, gamma ray energies from a single source (U, Th or K). The approximate positions of suitable windows can be predicted theoretically from known decay series as discussed above. In practice, interaction between the gamma-ray spectra of the different elements, and instrument effects, mean that the windows must be fine-tuned empirically. The windows used in this study are shown in figure A1.2 on spectra recorded at the British Geological Survey calibration facility at Keyworth in Nottingham.

2. Background count values for each window are subtracted. Background radiation arises from cosmic sources and impurities within the detector assembly. Values for background radiation are best established by counting over water at least 2m deep and 5m from shore in a fibreglass boat: wooden boats yield a significant potassium count (P. Roberts, BGS, pers. comm.).

Background count values used in this study were established by suspending the instrument on a boom from a fishing boat offshore N. E. England (K. J. Myers, pers. comm.)

3. Count values in each window are corrected for contributions from the decay series of other elements (Løvborg & Mose, 1987). For example, low-energy gamma-rays from thorium and uranium decay will appear in the potassium window and contribute to the counts there. Calibration is achieved by counting on concrete pads, such as those at the British Geological Survey facility. These enable optimum and repeatable counting geometries to be attained (see below) and are "spiked" with Th-, K- and U-bearing phases so that one radionuclide is in vast excess over the others in a given pad.

Background counts and calibration constants used in this study are tabulated in figure A1.3a. Further details of the calibration procedure will be found in Løvborg (1984).

Sources of Error

1. Counting statistics. Radioactive decay is a statistical phenomenon and hence there are inherent uncertainties associated with its measurement. Løvborg & Mose (1987) derive the following equation for the calculation of gamma-ray spectrometer precision:

$$t \approx (r_i + 2w_i + b_i)/(pr_i)^2$$

Where t is the counting time in minutes to achieve a precision of $\pm 100p\%$ (1σ), given a net count rate r_i from element i to be assayed, an interfering count rate of w_i from other elements in the element i window and a background count rate of b_i in the element i window. Applying this equation to the typical rocks of this study (U=3ppm, K=3%, Th=15ppm) suggests that precisions of better than $\pm 10\%$ could be expected for count times of 1 minute or less for Th and K. Poorer

precision is predicted for U, with count times of around 6 minutes required for $\pm 10\%$ and 2.5 minutes for $\pm 15\%$.

An alternative, empirical, approach to measurement precision is to make a number of repeat measurements at the same station. This was done in the Black Ven Marls of Dorset (fig. A1.3b). Four-minute counts, most commonly used in this work, yielded 2σ errors at a given station (i.e. with constant counting geometry) of approximately $\pm 7\%$ for K, $\pm 10\text{-}15\%$ for Th and $\pm 30\%$ for U.

2. Counting geometry. This is by far the largest source of error in practice. The spectrometer is calibrated by standing the detector on a flat surface where gamma rays reach the detector from the geometry shown in figure A1.4a. This arrangement is rarely achievable on natural outcrops and any departure from this geometry will lead to errors by adding or subtracting rock from the volume investigated by the instrument (fig. A1.4b). Major rock overhangs and measurements less than 50cm from the base of cliffs were avoided in this study but some ledges and hollows are unavoidable in practice. Such poor geometries are indicated on the logs presented in this thesis by open triangles. An estimate of the errors involved from "average" poor geometries can be made by repeat measurements at different localities along a bed (fig. A1.3b). These suggest 2σ errors of the order $\pm 15\text{-}30\%$, with again, U being the poorest quality assay.

Clearly, these errors are considerably in excess of instrument precision and it is these errors which are thought to represent more closely the actual repeatability of individual measurements in this thesis. It is because of these significant errors that emphasis is placed only upon the large-scale trends in the data and not upon individual or small groups of count values.

Resolution

Given the counting geometry discussed above, the resolution of the technique is governed by the relationship between the detector and bedding (fig. A1.4c). Bedding plane measurements,

for example on wave-cut platforms, give maximum resolution (15-30cm). Minimum resolution is obtained parallel to bedding when the results represent an average of 1 metre of rock.

For most of the sections studied for this thesis suitable bedding-plane exposures were not available, particularly when one considers that all beds must be measured, not just the harder beds which tend to form good platforms. Bedding-parallel measurements at half or one metre vertical spacing provide overlapping "moving average" radioelement concentrations adequate to define the vertical trends sought in this study. The approximate angle between the detector and bedding is shown on the logs.

Generalities of Interpretation

Specific interpretations of the spectral gamma-ray data obtained for this thesis are advanced in Chapters 4-7. It is appropriate here to discuss the general background against which these interpretations have been made.

Current understanding of the distribution of U, Th and K in mudrocks may be summarised as follows:

- High uranium concentrations are commonly associated with high organic-matter concentrations (e.g. Schmoker & Hester 1983, and fig. 4.13). The association is complex: uranium can be directly adsorbed onto organic matter and is also precipitated from seawater under anoxic conditions, a reaction promoted by hydrogen sulphide, itself resulting from organic matter decay (Swanson, 1960).
- Potassium is a major component of rocks, and hence of soils. However, intense hydrolysis in hot and wet climates tends to remove K in solution and leave kaolinitic (K-free) clays. Griffin et al. (1968) have demonstrated that the distribution of kaolinitic clays in the world oceans is strongly

controlled by the climate of the sediment source area . It has also been established that differential transport of clay species plays an important role in their distribution in marine environments, with onshore-offshore trends from kaolinite to smectite domination being demonstrated for the Niger (Porrenga, 1966) and Amazon (Gibbs, 1977) shelves.

- Thorium is largely insoluble in natural waters; it is therefore transported out of the soil-forming environment either in the heavy mineral component or adsorbed onto clays (Langmuir, 1980). Michel (1984), by investigating the distribution of Th in granite and its overlying soil, has shown that a significant proportion is likely to follow the latter route. The importance of Th in the clay mineral fraction is supported by high concentrations of Th in mudrocks, such as those in this study, where heavy minerals are not present. The data of Hassan and Hossin (1975) suggest that there is partitioning of Th between clay mineral species, with kaolinite having higher Th concentrations than mixed layer and 2:1 clays.

This understanding leads to a number of ideas for the interpretation of spectral gamma-ray data. Firstly, it is common for uranium to be treated as an independent component which is a measure of anoxia and/or organic matter concentration. Secondly, where Th and K co-vary (as they commonly do, see Myers (1987) and most of the data sets in this thesis) they can each be used to indicate the total clay volume in, for example, mixed clay-carbonate systems. Thirdly, systematic variations in Th/K may be interpreted to reflect changing clay-mineral assemblages. In particular, following the arguments outlined above, increased Th/K may be interpreted to indicate more kaolinitic assemblages, which in turn may be interpreted to reflect more proximal settings (Myers, 1987) or increasingly hot and wet climates (Myers, 1989). The data presented in this thesis may be similarly interpreted.

Critically, these interpretations have two weaknesses: they rely upon direct links between clay mineralogy and radionuclide distribution, and they do not account for possible diagenetic effects. Whilst the spectral gamma-ray studies undertaken for this thesis have provided invaluable

quantitative data on trends within the mudrock units, it is apparent that the interpretation of the data would be much more concrete given greater knowledge of the actual distribution of Th, K and U in modern marine environments and of the sites of Th, K and U within the complex fabric of a mudrock, including the diagenetic phases.

Results

Spectral Gamma Ray results were obtained from the Dorset, Yorkshire, Portugal and South German study areas and are discussed in detail in Chapters 4-7. Figures A1.5 to A1.8 tabulate the results numerically.

Appendix 2. Strontium Isotope Mass Spectrometry

Strontium-isotopic analyses were carried out on belemnite and oyster skeletal material from Portugal and Germany (see Chapters 6 and 7). This work was stimulated by that of Jones (1992), who measured the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of belemnite and oyster skeletal carbonate collected from classic British localities where the stratigraphic age of the sampled horizons is tightly constrained by ammonite biostratigraphy. Most samples for his Lower Jurassic curve were collected from the Dorset and Yorkshire sections described in Chapters 4 & 5 of this thesis. The objectives of the strontium isotope work reported here were:

- To attempt to replicate Jones's curve on other European sections well-constrained by ammonites, thereby confirming that Jones' curve reveals a primary oceanographic signal and demonstrating the Europe-wide validity of both the curve and the ammonite correlations.
- To use strontium isotope data to supplement existing ammonite-based correlations between the studied sections.

Theoretical Basis

^{87}Sr is a stable isotope formed by the β decay of ^{87}Rb with a half-life of 4.99×10^{10} years. Its abundance is conventionally expressed relative to non-radiogenic ^{86}Sr . The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of a system is therefore a function of its Rb/Sr ratio and of time. Different rocks have widely differing Rb/Sr ratios as a result of melt fractionation: in particular, continental crust has a significantly higher Rb/Sr ratio than the upper mantle. This leads to significant differences in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of crustal ($^{87}\text{Sr}/^{86}\text{Sr} \approx 0.716$) and mantle ($^{87}\text{Sr}/^{86}\text{Sr} \approx 0.703$) rocks, differences which are becoming more marked with time (Faure, 1986; Elderfield, 1986).

The oceans have a very low Rb/Sr ratio which, combined with the long half-life of ^{87}Rb , means that very little ^{87}Sr is generated there. Hence, the oceanic $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is essentially a function of the inputs: rivers dissolving continental silicate rocks and recycled limestones, and hydrothermal circulation systems dissolving oceanic crust (fig. A2.1). Modern oceans are found to be homogeneous for $^{87}\text{Sr}/^{86}\text{Sr}$, a result of the long residence time of Sr (2-4Ma; Hodell et al., 1989) in the oceans compared with the oceanic mixing time of approximately 1-2Ma (Broecker & Peng, 1982). Modern marine carbonate rocks and carbonate skeletal material yield similar $^{87}\text{Sr}/^{86}\text{Sr}$ ratios to modern seawater, which encourages the view that the marine $^{87}\text{Sr}/^{86}\text{Sr}$ signal is accurately recorded by Ca-bearing precipitates.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of ancient oceans has been investigated by measuring the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of marine carbonate rocks and carbonate skeletal material, most notably by Burke et al. (1982). Their curve (fig. A2.1b) demonstrates considerable variation in marine $^{87}\text{Sr}/^{86}\text{Sr}$ ratio through Phanerozoic time, presumably due to the differing fluxes and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the various inputs discussed above. There is considerable scatter in this dataset, interpreted by Burke et al. to result largely from diagenetic alteration of the samples. The major contribution of Jones (1992) was to demonstrate that, given a carefully selected suite of stratigraphically-controlled belemnite and oyster material and meticulous sample preparation, the curve was capable of great refinement (fig. A2.2a). Jones (in press) suggests that the stratigraphic resolution of the Early Jurassic Sr-isotope curve may be as good as $\pm$ one or two ammonite subzones ($\pm 1\text{Ma}$).

Analytical Method

Belemnites and oysters analysed for this thesis were prepared using the method of Jones (1992), summarised in figure A2.3. Samples were run on the VG Isomass 54E mass spectrometer in the Department of Earth Sciences, Oxford. Filaments were loaded by first etching with phosphoric acid and then rinsing water into the dry microbeakers containing the precipitated strontium,

placing a drop of this solution onto the filament and then heating until glowing red and evaporated.

Sources of Error

The E & A and NBS 987 standards were run with each turret. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 20 runs of NBS 987 during the study period are shown in figure A2.2b. Error bars on individual samples represent two standard errors about the mean of the individual $\text{Sr}^{87/86}$ ratios analysed. This is a measure of "intra-run" precision and gives a good indication of the quality of an individual data point. Values of less than $\pm 15 \times 10^{-6}$ (2σ) were regularly obtained for all samples unless there was a problem with the sample, for example too little Sr on the filament. The "inter-run" precision is a more reasonable measure of the overall reproducibility of the results. The values in figure A2.2b yield an overall reproducibility of $\pm 29 \times 10^{-6}$ (2σ), which is comparable to the results of Jones (in press) for 26 replicate analyses using the same machine ($\pm 23 \times 10^{-6}$, 2σ).

Jones (in press) presents powerful arguments for normalising the data from each turret according to the value of the standard obtained in that turret run. However, the more conventional and simple procedure of normalising to the average value of NBS measured over the period of investigation ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710256 \pm 5.10^{-6}$, 2σ) has been used here.

Fe and Mn were analysed for all samples as a test of possible diagenetic alteration. These elements were selected because of their low concentrations in seawater and in modern marine CaCO_3 shell materials and relatively high concentrations in many diagenetic fluids (Jones, in press). All samples here passed Jones' (in press) critical hurdle of <150ppm Fe. However, in most cases the most convincing test of the primary nature of the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is that they define smooth curves within the analytical error of the data and that they replicate the curve of Jones with such precision (Chapters 6 and 7). The new data are tabulated in figure A2.3.

Appendix 3. Other Analytical Techniques

Oxygen and Carbon Isotope Mass Spectrometry

Oxygen and carbon isotope analyses were performed on belemnite (and some oyster) skeletal carbonate from Portugal and Germany (Chapters 6 & 7). Whole rock stable isotope analyses were performed on material from the Southern Alps (Chapter 9).

Clean belemnite chips were extracted from the Strontium isotope processing chain as shown on figure A2.3. 2-3 chips were crushed to flour in an agate pestle and mortar and a small quantity of this flour transferred to microbeakers for analysis. In the case of whole rock samples, the small quantity of material required for isotopic analysis was extracted from samples of approximately 50g which had been reduced to flour using a toughened steel pestle and mortar and an agate ball mill. All samples were treated for 30 minutes with H_2O_2 and then washed with acetone prior to analysis in order to denature any organic matter present. Analysis was performed using the VG PRISM mass spectrometer in the Department of Earth Sciences, Oxford. Samples were reacted on-line using purified H_3PO_4 . Samples were calibrated to PDB using an internal laboratory standard of known isotopic composition and NBS19. Precision, based on the reproducibility of replicate standards is ± 0.1 per mil (e.g. Corfield et al., 1991). Repeat analysis of whole rock samples in this study suggests a precision of ± 0.2 for $\delta^{13}\text{C}$ and ± 0.32 for $\delta^{18}\text{O}$ (2σ).

Analytical results are tabulated in figures A2.4, A 2.5 and A3.4. They are discussed in Chapter 12.

X-ray Diffraction

X-ray diffraction analysis was undertaken on whole-rock samples from the Breggia Gorge in the Southern Alps (Chapter 9). The objective was a rapid reconnaissance study to determine gross vertical trends in terrigenous input and clay mineralogy. Detailed quantitative clay mineralogy involving dissolution of carbonate and separation of clays was considered to be beyond the scope of this thesis.

Analysis was performed on a split of the same whole-rock flours used for carbon and oxygen isotope analysis (see above). Samples were packed into aluminium cavity mounts and analysed on the PW1710 diffractometer at the Postgraduate Research Institute for Sedimentology at the University of Reading using Copper K-alpha radiation. A typical diffractogram shows peaks indicating calcite, dolomite and quartz (fig. A3.1). Small quantities of feldspar, mica and kaolinite were identified in some specimens.

Diffractograms were interpreted by measuring peak heights above background, multiplying by calibration "H" factors (fig. A3.2), totalling the factored heights and calculating the percentage of each mineral relative to the total (Hooton & Giorgetta, 1977). Calculated percentages are tabulated in figure A4.3.

Heavy Mineral Analysis

Heavy minerals were isolated from the Belemnite Bed in Yorkshire (Chapter 5) in order to seek potential non-clay carriers of thorium.

Approximately 100g of rock were crushed to pass through a "100-mesh" (212 μ m) sieve. 10% acetic acid was used to dissolve the carbonate component (apatites being soluble in HCl), Teepol and NaCO₃ were then added to disperse the clays and samples were agitated in an ultrasonic

bath before decanting off the clear liquid and drying. The sample was then split to make a separate investigation of the >63 μ m fraction and the 53-63 μ m fraction. The latter was of particular interest since zircons, prime carriers of Th, are prone to radiation damage which then gives rise to very fine fractured grains (M. Mange, pers. comm.). Each split was passed through appropriate sieves and then added to tetrabromoethane (s.g. 2.9), stirred and centrifuged. It was then possible to freeze the bottom of the test tube in liquid nitrogen, decant off and filter the light fraction and then unfreeze the heavy fraction and filter this. Samples were washed through the filters with methanol, dried and mounted for inspection. Samples were weighed at each stage to provide semi-quantitative abundances. The results are described in Chapter 5.

Stains & Peels

Carbonate petrography was an important part of the investigations in Chapter 6. Staining was undertaken on some specimens to investigate carbonate mineralogy and enhance the appearance of grains. Thin sections were etched using 2-5% HCl for approximately 1 minute and then stained for 2-3 minutes using a 3:2 mixture of Alizarin Red-S and Potassium Ferricyanide diluted 50% with water (Dickson, 1966).

Peels were taken by etching polished slabs in 10% HCl for 2-3 minutes. The peeling medium was acetate sheet and acetone.

Appendix 4. Timescales

Ammonites provide the primary method of time correlation in the Lower Jurassic and all the observations in this thesis have been related to the subzonal scheme of Dean et al. (1961), incorporating the modifications suggested by Cope, Getty et al. (1980) and Ivimey-Cook & Donovan (1983). Empirical observation has long convinced ammonite workers of the validity of their biozones for time correlation, the most persuasive observation being the consistency of the sequence of ammonite faunas in widely separated localities (Hallam, 1975 p.9). Strontium isotope results reported in this thesis provide some measure of corroboration for this assumption (see Chapters 6, 7 and Appendix 2).

Numerical ages have been assigned to the quantitative data in this thesis in order to facilitate data processing and comparison between sections. This has been done by linear interpolation between the biostratigraphic boundaries marked on the logs. Thus the conclusion to be drawn from a sample labelled 188.30Ma is that, at a best estimate, it was collected from, or measured in, the middle of the *apyrenum* Subzone and is younger than a measurement or sample labelled 188.31Ma. The chronostratigraphic accuracy of the assignation is clearly no better than definition of the *apyrenum* Subzone which may be ascertained from the log of the section under consideration.

Little time-stratigraphic accuracy is claimed. Ages have been assigned to subzonal boundaries in the manner of Jones (1992) by interpolation between the tie points of Harland et al. (1990) on the basis of equal ammonite subzones with an arbitrary two subzones being allowed where a zone is not currently subdivided into subzones. The relevant tie points are at the base of the Rhaetian, the base of the Sinemurian and the mid Bajocian. The subzonal schemes used are those described above for the Lower Jurassic, and that of Callomon & Chandler (1990) for the Aalenian

and Bajocian. It should be noted that this latter reference has critical addenda and corrigenda which may become detached.

The stratigraphic scheme for the Lower Jurassic and the numerical ages assigned to the subzone boundaries are summarised opposite.

	Stages	Sub-stages (European Usage)	Ammonite Zones & Subzones*	Age interpolated from Harland et al. 1990
Middle Jurassic			opalinum scissum opalinum	176.7 177.7 178.8
Lower Jurassic	Toarcian	Late Toarcian	levesquei moorei levesquei dispersum fallaciosum striatulum	179.3 179.8 180.3 180.8 181.3 181.9
		mld-Toarcian		
		Early Toarcian	bifrons talciferum tenuicostatum	182.9 183.4 183.9 184.4 184.9 185.5 186.0 186.5 187.0
			spinatum margaritatus	187.5 188.0 188.6 189.1 189.6
		Late Pliensbachian (Domerian)	davoiei ibex jamesoni	190.1 190.6 191.1 191.6 192.2 192.7 193.2 193.7 194.2 194.7
	Pliensbachian	Early Pliensbachian (Carixian)	faricostatum aplanatum macdonnelli raricostatoides densinodulum oxynotum simpsoni denotatus stellare obtusum turneri brooki resupinatum scipionianum lyra bucklandi rotiforme conybeari	195.3 195.8 196.3 196.8 197.3 197.8 198.3 198.9 199.4 199.9 200.4 200.9 201.4 202.0 202.5 203.0 203.5
		Late Sinemurian (Lotharingian)	oxynotum obtusum	
		Sinemurian		
		Early Sinemurian		
			angulata liasicus planorbis	
Triassic	Hettangian			205.0 205.8 206.5 207.3 208.0
	Rhaetian			209.5

* Cope, Getty et al., 1980; Ivimey-Cook & Donovan, 1983

● Tie points of Harland et al. (1990). The mid-Bajocian tie point (168.2Ma) is not shown.

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