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Sedimentological and Palaeoenvironmental Studies in
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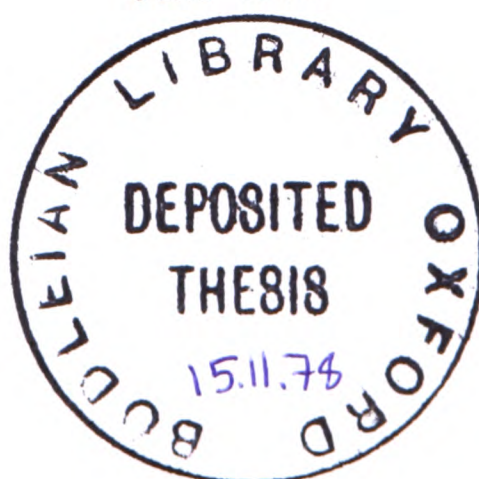
KOUROSH AMIRI-GARROUSSI, B.Sc.

St. Peter's College

and

Department of Geology and Mineralogy

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To Shahila

*"Knowledge of necessity need not be truth
and can only be a picture of reality"*

*Lao Tse
Tao
(The way of being)*

Sedimentological and Palaeoenvironmental
Studies in the Broadford Beds (Hettangian - Sinemurian) of
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K. Amiri-Garroussi

St. Peter's College, Oxford

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Abstract

Planorbis to Turneri Zone sedimentary rocks were studied at outcrops on Skye, Raasay, Applecross, Ardnamurchan, Morvern and Mull. A proposed litho-stratigraphy includes two Groups, four Formations and seventeen Members. Increased faunal diversity in Mull and Skye reflects environmental stability during the Angulata and Semicostatum Zones; low diversity in the Bucklandi and Turneri Zones signifies lower stability. Species abundance in Mull and their decrease in Skye during Angulata-Bucklandi Zone times signifies a northward increase in environmental instability.

Smectites are present in the Planorbis-Bucklandi Zone shales. The Semicostatum-Turneri Zones are composed of non expanding clays; the smectite formed by erosion of previously exposed alkaline igneous rocks under warm, alkaline, low rainfall conditions. Progressive transgression covered the source area and a change to more acidic conditions with higher rainfall also promoted the formation of "ironstones" and chlorites.

The Angulata and Bucklandi Zones comprise bioclastic, oolitic, carbonate and siliciclastic sediments signifying shallow marginal marine semi-enclosed basins. Thin laterally impersistent beds of coarse, poorly sorted pebbles signify an unstable hinterland; during the Semicostatum and Turneri Zones fully marine conditions were established.

Ferruginous beds are oolitic in Skye. They probably formed as "mud balls" and due to acid leaching of feldspars. Phosphatic nodules are calcium-hydroxyapatite.

The limestone/shale "rhythms" are explained by a combined primary and diagenetic origin. The limestones consist of low-Mg calcite. Both ferroan and non ferroan calcite are present. Tectonic stresses and diagenesis deformed crystals and formed veins.

Sr^{+2} values show a decline as the carbonate fraction increases. Three different ranges are distinguished for three facies and ages. Sr^{+2} concentration is bimodal and probably inherited from the original sediments.

The northern and southern basins evolved separately, sedimentation was controlled by differential tilting, uplift and subsidence. Palaeoslopes were gentle, to the west-northwest and south-southwest. Local topography (e.g. Central Strath High) modified the slopes; no movement occurred along the main faults of the area. Sedimentation relates to the "taphrogenic rifting stage" proposed for the North Sea Mesozoic.

ABSTRACT

Lower Liassic (Planorbis-Turneri Zone) sedimentary rocks were studied in the Inner Hebrides Basin of northwest Scotland. The main sections were in Skye, Applecross and Raasay, located in the north of the Basin, and Mull, Movern and Ardnamurchan in the south.

The nomenclature relating to the "Broadford Beds" is explained and it is proposed that they should comprise beds which overlies Triassic strata and underlie a disconformity marking the late Turner Zone.

A formal lithostratigraphy is proposed which includes two Groups, four Formations and seventeen Members; twenty two different sedimentary facies are recognised. The principal units are the Broadford Beds Arenaceous Group in the north and the Broadford Beds Argillaceous Group in the south. The Milton and Strath Formations correspond to the Lower and Upper Broadford Beds respectively in the Skye area, while the latter are represented by the Loch Aline and Leacach Formations in Movern.

The already described ammonite fauna makes it possible to date each Member with respect to the standard zonal scheme.

Published faunal lists are presented with new quantitative data. Increased faunal diversity recognised in both areas in the Angulata and Semicostatum Zones is interpreted as signifying an increase in environmental stability and "transgression" whereas decrease in the Bucklandi and Turner Zones signifies decreased stability and "regression". These results are independently supported by sedimentary evidence.

The ammonite fauna shows a relative increase ^{in diversity} from the Planorbis to the Angulata Zone in the southern area of study whereas it decreases in diversity during the Bucklandi Zone. In the northern area no ammonites older than Angulata Zone were found. A relative increase is seen from the Angulata to the Bucklandi Zone which contrasts with the situation further south. This may be due to increased environmental instability during the Angulata Zone in Mull

whereas progressive deepening of the sea in the Skye area occurred. Ammonite diversity in the north increases from the Bucklandi to Semicostatum Zone; this together with other faunal evidence relates to two periods of transgression, one during the Angulata Zone and the other during the Semicostatum Zone.

The relative abundance of faunal species in the Mull area and their decrease in Skye during Angulata and Bucklandi Zones is interpreted as signifying a consistent northward decrease in environmental stability.

While the number of approximately 24-30⁰/oo salinity tolerant species increases from Mull to Skye during the Angulata Zone, the faunal salinity tolerance is much lower and also decreases from Mull to Skye during the Bucklandi Zone. In both areas euryhaline fauna decreases from south to north during the Angulata and Bucklandi Zones, signifying consistently more marine depositional conditions in the south.

The lack of cemented and boring bivalves in the Lower Lias indicate the scarcity of firmly compacted sediments; "hardgrounds" are lacking.

Facies 6 and 8 of the Milton Formation are reminiscent of the "reefal association" of Hallam (1976) whereas the beds of the Strath and Loch Aline Formations characterise the "nearshore marine association". Parts of the Strath together with the Loch Aline and Leacach Formations represent "marine basinal associations". "Lagoonal associations" are recognised in the Skye area where the oolitic facies 5 represents lower brachyhaline (ca. 16 - 23⁰/oo) conditions; upper brachyhaline (24 - 30⁰/oo) fauna is well represented in the various facies of Skye, Ardnamurchan and Mull.

The shales in the Milton and Loch Aline Formations contain up to 30% smectites, whereas the Strath and Leacach Formations contain illite, kaolinite and subordinate mixed-layer minerals together with chlorite. ?Rhaetic red mudstones exposed in western Mull contain up to 10% smectites in contrast to Triassic red mudstones which contain only illite. The smectite was probably derived from the weathering of basic igneous rocks previously exposed in the sediment source area. By the time the Leacach Formation was being deposited

in the south and the Strath Formation in the north, the igneous source was eroded/transgressed by the early Liassic sea.

Smectites probably formed under warm, alkaline climatic conditions with low rainfall, which prevailed during Planorbis to Bucklandi Zone times. Little kaolinite and a higher proportion of illite was deposited in the southern area as compared to the Skye area signifying more offshore depositional environments. The clay mineralogy of the Bucklandi Zone shales of Loch Aline signifies fluctuating environmental conditions and possible near-shore deposition. In Skye a higher kaolinite content coupled with other evidence signifies near-shore deposition.

The onset of high rainfall, humid, acid climatic conditions from the Upper Bucklandi to the Turneri Zone and an increased coverage of land areas by sea was responsible for the abrupt decrease in smectites and the abnormal rise in kaolinite content of the beds. The change of climate contributed to the formation of iron-rich beds during the Semicostatum Zone.

Sedimentation took place in transgressive-regressive, shelf-shoreface sequences during the Lower Lias. Three broadly defined "transgressive" sequences were recognised whereas four showed regressive characteristics.

The sporadic erosion of small upstanding isolated but resistant blocks of older strata produced local subaqueous conglomeratic debris flows in southeastern Skye and Facies 1, 2 and 3 developed. As the land surfaces in the north were inundated by the transgressing (?Angulata Zone) sea, a lagoon-barrier (?spit) complex developed (Facies 4, 5). Sediment was supplied from southeast Skye for the development of the sandy body which effectively came to isolate an embayment with lagoonal mud to the northeast of Applecross. The embayment was fringed on the landward side by a mixed facies and mud-flat belt. With continuation of the transgression and reduced sediment supply, breaches occurred in the "barrier" system. Shallow "tidal" channels with characteristic oolite lobes developed and the barrier migrated landward to overlies the lagoonal shales. The barrier was finally drowned and the sand was redistributed.

With the still continued transgression, marine conditions were established in the northern area and coral beds of Facies 6 came to directly overlie the sands and oolites. Facies 7 indicates a rapid regression which affected the Skye area. The result was a development of lagoons in the vicinity of Applecross, which were actively protected by a sand body continuously supplied by southeastern source areas.

Shallow subtidal deposition continued in eastern Raasay. The beds are bioturbated and Thalassinoides burrows have controlled the pattern of nodular limestone formation here. In southwest Raasay (Suisnish) and northwest Strath (Sligachan), shallow water detritus-free carbonate deposition continued. In Applecross oomicrites and lagoonal shales were deposited.

With the progressive transgression of the sea, oolite shoals and inter-shoal muddy channels developed which were eventually overlain by skeletal and sandy patches with a substantial development of oolitic sand bodies.

Continued rise in sea level drowned the embayment and Facies 8 containing Thecosmilia martini succeeds Facies 7; the corals grew upward to compensate for land subsidence or sea level rise.

Facies 9-12 represent a prograding shoreline sequence indicating a large scale regression; the sediment source area was situated somewhere to the north or northeast of Applecross. These beds are compared with conglomeratic shoreline deposits and offshore (Facies 9), lower shoreface (Facies 10) and upper shoreface (Facies 11 and 12) depositional environments are distinguished. Although isolated rootletted horizons and thin coal beds were seen to overlie beds of Facies 11 and 12 in northern Strath, the general features indicating foreshore and backshore deposition are lacking and were probably represented further to the northeast or eroded due to their low preservation potential. The foreshore belt areas probably extended from Applecross to Skye and into Ardnamurchan.

A large scale rapid transgression characterises Upper Bucklandi to Semicostatum Zone sedimentation in the Skye area. The overall deepening of

of the water is also indicated by the existence of planktonic organisms (calcispheres) in the beds of Facies 13.

The deposition of the different "cycles" of clay/silt/calcareous sand in Facies 13 represent shallowing upward events. The Upper Bucklandi-Lower Semicostatum sea which had transgressed the low lying land areas gradually regressed (stepwise) until at the culmination point, iron-rich beds formed in both areas of study (Facies 14 and 22). The ferruginous beds are oolitic in Skye and Ardnamurchan whereas ferruginous mudstones developed in Mull and Morvern.

X-ray diffractometry and powder photography shows that although the minerals in the rocks of Facies 14 do not behave as the ideal "chamosite", they show X-ray peaks typical of cronstedtite. It is proposed that the iron silicates occurring on these rocks should be ascribed in general to the chlorite group and the term "chamosite" should only apply to those beds which show the definite existence of the septechlorite, cronstedtite.

Major elements of the "ironstones" were determined using X-ray fluorescence methods and comparisons have been made with other Hettangian-Sinemurian "ironstones" of northwest Europe. The hypotheses that have been proposed for the genesis of ferruginous ooliths and pellets are discussed but are inadequate to explain the early Liassic oolitic "ironstones" and ferruginous mudstones of northwest Scotland.

A combination of feldspar kaolinisation and weathering under acid, warm and humid climatic conditions together with erosion of previously formed argillaceous sediment as "mud balls" can adequately explain their formation.

Phosphatic nodules contain as much as 22% P_2O_5 ; they are regarded as resulting from a combined primary and diagenetic precipitation. The phosphorus is in the form of calcium hydroxyapatite.

Facies 15 marks a minor transgression which covered the beds of Facies 14 as a result of a rapid subsidence or sea level rise. The beds are characterised by small scale shallowing upward cycles indicating a slow stepwise

uplift (sea level fall). The general upward increase in the clastic components of the beds together with cross lamination and increased trace fossil activity indicates deposition in waters with periodic, gentle wave and/or current activity; erosion surfaces were not found.

Facies 16, 17 and 18 exhibit an overall regression during the Turnerian Zone. The beds of Facies 16 resemble thin sheets of marine sand waves which show an upward shoaling with increased current and wave velocities. The depositional conditions of Facies 17 are not fully understood. They were probably formed as sand waves and dunes in response to higher shoaling energies.

These sand bodies are cut by channels of coquinoïd sandstones. The beds of facies 18 indicate persistently recurring periods when deposition under calm, deeper marine conditions gave way to shallower more turbulent environments. The characteristics of this facies are similar to the lower parts of shoreline sequences and also resemble offshore marine sand bodies in their upward shoaling characteristics. They cannot be regarded as parts of marine bar sands or barrier islands.

The extremely variable flow directions recorded from the various facies, formed due to shallow marine processes involved in distributing the sediments in shoal areas of shallow shelf seas.

It is argued that lowering of sea level combined with initial bottom topography produced the observed features. The top of Facies 19 is marked by a late-Turnerian-early Obtusum disconformity.

Beds representing the Planorbis and Angulata Zones are absent from the Craignure area (eastern Mull) but can be seen in western Mull and Morvern.

Facies 19 was deposited under less turbulent marine conditions in western Mull (Wilderness) but further north near Gribun evidence of abnormal salinity and crushed driftwood and plant debris suggest proximity of land areas.

Facies 20 shows evidence of coalified plant fragments and mound-like escape features (cf. Callianassa islagrande) which suggest muddy nearshore

depositional environments. The beds of this facies are only seen in Morvern.

Facies 21 suggests deposition under offshore-marine conditions and the local presence of laterally equivalent facies is evidence for a larger lateral extent of the sea during this time. The reddish-weathering calcareous ferruginous crust on top of the beds of this facies (Leacach Ochrous Member) signifies a major period of erosion during the Semicostatum Zone.

Facies 22 indicates deposition under marine conditions below effective wave base and far from land areas. Ferruginous mudstones were deposited in Mull while micaceous shales were formed in Loch Aline.

In the latter area, the top of these beds is shown by an erosion surface, with the ?Brooki and Birchi Subzones absent, and marked by the Leacach Pebble Member.

A combination of primary and diagenetic changes is preferred for explaining the origin of the limestone/shale "rhythms" found in the beds under study. Probably more CaCO_3 was precipitated in shallower water sediments (Skye) while in the deeper parts (Mull) the muds were secondarily enriched in CaCO_3 . After deposition, the solution and reprecipitation of parts of the limestones began; CaCO_3 segregated into thin beds of limestone within the shales.

The Liassic beds under study were examined petrographically and the following conclusions were reached:

- i) The stable diagenetic carbonate phase is low-Mg calcite; no aragonite is present.
- ii) The transformation of aragonite to calcite occurred early in diagenesis and involved "wet" solution precipitation.
- iii) Consolidation and compaction was completed early and resulted in the loss of pore space; cement of both ferroan and non-ferroan calcite is present.
- iv) During late stages of deep burial crystal enlargement occurred, which has in places completely destroyed the primary sedimentary textures and

produced a coarsely crystalline rock.

v) Tectonic stress and the weight of overlying sediments during deepest burial caused the formation of stylolites, veins and bent twin lamellae.

The Sr^{+2} , Mg^{+2} , Mn^{+2} and Fe^{+2} content of seventeen selected samples from three different Formations were studied by Atomic Absorption Spectrophotometry. The variation of Ca, Mg, Mn and Fe between the different samples is somewhat concordant whereas the Sr^{+2} variations do not follow the pattern shown by the other elements. The Sr^{+2} content of the samples shows a steady decline with increased soluble fraction. While the average Sr^{+2} content of the samples is in the region of 684 ppm and they show a more or less similar diagenetic history, the three rock populations show distinctly different ranges. The Sr^{+2} concentration is bimodal; a sequence of decreasing Sr^{+2} content may be seen from the deposits of the well aerated semi-enclosed, nearshore environments to the relatively more "offshore", less aerated shelf deposits. The bimodal distribution may be inherited from the original sediments. The Mn^{+2} and Fe^{+2} concentrations are much lower than predicted values. This reflects post-depositional changes.

The Lower Liassic beds are considered within the framework of seven sequences. During the Hettangian stage the areal extent of the sea was minimal and the Lower Sinemurian marks a widespread transgression. The northern and southern basins evolved separately. The pattern of sedimentation was controlled by tilting, differential subsidence and uplift on the Skye-Ardnamurchan, Morvern and Caignure Blocks. Although general slopes towards the northwest, west and southwest were present in the northern area, continuous subsidence of the Applecross area caused a thicker development of the Milton Formation, the slopes are locally modified by the existence of the Central Strath High. No evidence is present to suggest Liassic movement along the Camasunary or Great Glen Faults although in Skye a marked topographic difference (shallowing to the west) is seen across the Camasunary Fault; a similar topographic change is seen in the Caignure area (shallow

towards the east) across the trend of the Great Glen Fault. The tectonic style controlling the sedimentation pattern follows the "taphrogenic stage" already proposed for the North Sea.

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"... and so justly impressed with the importance of being ... that as I am in the way of describing rare specimens at any rate, I must refer to him among the rest as if he had been one of the minor carnivorae of a Skye deposit, - a cuttlefish that preyed on the weaker molluscs or a hungry polypus terrible among the Animulae".

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CHAPTER 1

INTRODUCTION

1.1 Location of Sections

Lower Liassic strata are found as small, widely spaced outcrops in two major areas of northwest Scotland, here termed the northern region (Skye, Applecross and Raasay) (Leel, 1920; Hallam, 1959) and the southern region (Mull, Morvern and Ardnamurchan) (Lee and Bailey, 1925; Richey and Thomas, 1930; MacLennan, 1953).

The location of each of the major sections is shown on Figure 1.1 together with the national grid references; other minor outcrops are mentioned in the text and identified by their appropriate grid reference.

1.2 Previous Work

Observations on Scottish Mesozoic rocks have been recorded from as early as the 16th century. These were mainly related to the discovery of local seams of coal.

Coal was mentioned in the Sutherland Charter.

"In 1529 (22 April) Alexander Gordone sold to John Terrell the sea coal found and to be found ..."
(Origin Parochiales Scotiae, Vol.2, Pt.2 - The Antiquities Ecclesiastical and Terrestrial of the Parishes in Scotland).

Since Martin (1703) reported that the Isle of Skye "is very High Land and there are seven high Mountains near one another", and that in the Isle of Mull "there is a great ridge of Mountains about the middle of the Isle, one of them very high", the whole of the western Isles of Scotland have been extensively studied by numerous researchers who have ventured "in order to acquire from actual observation, a knowledge of the mineralogy of the Scottish Isles" (Jameson, 1800).

Mesozoic strata were first recorded in the Hebrides by Pennant (1774)

who observed crushed ammonites in Trotternish, Skye during his tour. The Jurassic rocks of the western Isles were first studied by Macculloch (1819) and Murchison (1827) in their pioneering works.

The Lower Lias in this area has been studied mainly from a stratigraphic point of view. Macculloch (1819) and Geikie (1858) gave the first detailed descriptions for the Lias of Strath (Skye); however it is widely acknowledged that the foundations for the Geological Survey Memoirs on Lochalsh and southeast Skye (Peach et al., 1910), the pre Tertiary geology of Mull, Morven and Ardnamurchan (Lee & Bailey, 1925) and the Mesozoic rocks of Applecross, Raasay and northeast Skye (Lee & Buckman, 1920), were set by "The Summer Rambles" of Hugh Miller (1858) and the works of Wright (1858), Tate (1873), Bryce (1873) and Judd (1873).

Judd separated the Liassic rocks into Infra Lias (sparsely fossiliferous) and Lower Lias (with ammonites and Gryphaea of Bucklandi and Semicostatum Zone time).

Later, Woodward and Wedd (Woodward, 1897, and in Peach et al., 1910) divided the Lower Lias into the Broadford and Pabba Beds; the Broadford Beds were divided into four divisions at Lusa and it was generally accepted that all the Liassic Zones from Planorbis to Obtusum were present.

Buckman (in Lee & Buckman, 1920) and Spath (1922) were among the first to contribute to the palaeontology and zonation of the Scottish Lias for correlation purposes.

Although Judd introduced the term "Pabba Series", he designated a lower Middle Liassic age to them; however, according to Woodward's (1896) classification, they were deposited during middle Lower Liassic (Oxynotum-Capricornus Zone) times. As the Oxynotum Zone (and parts of the Obtusum Zone) are apparently absent from Skye, all the Liassic strata below these levels were placed in the Broadford Beds, whereas those occurring above the mentioned Zones were classified as Pabba Beds by Wedd (in Peach et al., 1910).

The same classification is used in Ardnamurchan, but since the Oxynotum Zone shales are present together with the top parts of the Broadford Beds (which are also shaly in nature similar to those of Skye), in a subsequent Memoir (Lee & Bailey, 1925) the Semicostat. and Obtusum Zones were grouped with the Pabba Beds. However, Richey et al. (1930) stratigraphically defined the Broadford Beds according to "that applied in Skye, the type district", i.e. Planorbis to Obtusum Zones.

Arkell's (1933) suggestion for the presence of the Planorbis and Angulata Zones at Lusa were challenged by Trueman (1942).

The succession in Mull and Ardnamurchan, like that of the northern region, has been described by many geologists. Lee and Bailey (1925) described the Carboniferous-Cretaceous succession in Morvern. Scott (1928) dealt mainly with the Carboniferous strata but recognized Liassic beds resting unconformably upon Carboniferous sandstones. Contributions have also been made by Lee and Pringle (1932), MacGregor and Manson (1934) and Richey (1935).

MacLennan (1953) followed Lee & Bailey (1925) and included the Turneri and Obtusum Zones in the Pabba Beds at Loch Aline in Morvern. She recognized intra Liassic periods of erosion and related the Mesozoic disturbances along the west coast of Scotland to the north-south linear zone which was involved later, in the Tertiary volcanic eruptions.

Hallam (1959) introduced the important twofold division of the Broadford Beds in northwest Scotland. The Lower Broadford Beds (Planorbis-Bucklandi Zones) were deposited under marginal marine, semi-lagoonal conditions whereas during the deposition of the Upper Broadford Beds (i.e. Hallaig sandstone of Buckman (in Lee & Buckman, 1920)) fully marine conditions were established. He also observed a band of oolitic ironstone at the base of the Upper Broadford Beds.

Although Cheeney's (1962) work does not directly concern the Jurassic

strata, his two geological maps of the area immediately south of Loch Don (Isle of Mull) are helpful in the study of this area.

Kalander (1974) found hitherto unknown outcrops of ?Liassic age siltstones overlying Triassic conglomerates at Loch Slapin, Isle of Skye; Steel et al. (1975) related the presence of these siltstones to the basin-filling processes operating in the area during the Triassic/Liassic times.

Oates (1976) made a detailed comparison of the stratigraphy and palaeontology of the Mull and Skye areas, drawing attention to similarity of facies among the English Blue Lias and Scottish Broadford Beds. He excludes the Turnerian and Obtusum Zone deposits from the Broadford Beds in the southern region and the Obtusum Zone from these beds in the northern region. Jurassic movements along the Great Glen Fault are also postulated. Table 1-1 compares the nomenclature used for the northwest Scottish Lias since 1819.

1.3 Purpose of Study

The Hettangian-Sinemurian outcrops of northwest Scotland are the northwesternmost representatives of the European Lias, and mark the initiation of important changes in the sedimentary regimes and climatic conditions which were prevalent during the Triassic times.

The study of these strata, with due consideration given to recently available information on the geology of the continental shelf off northwest Scotland, will facilitate the identification and interpretation of the mechanism and control of such changes in the context of plate tectonics.

The northern Scottish Liassic deposits together with their counterparts in E. Greenland and eastern North America formed as a sedimentary response to the movements which preceded the main phase of the formation of the Central, Northern and Eastern North Atlantic; the study of these deposits will provide important insights into the early history and evolution of

the Atlantic Ocean.

1.4 Methods

A total of 21 weeks were spent on the Lower Liassic rocks of northwest Scotland during successive field seasons and standard field techniques were applied for data collection, a combined Jacob's-Staff and Hand-Level technique (Compton, 1962) was used together with tape-measure pacing methods for measuring stratigraphic sections; some areas were mapped geologically at a 6" to the mile scale.

A total of 620 samples were collected. These were obtained at regular stratigraphic intervals with additional specimens being taken when pronounced changes in lithology and lateral variations were seen in the field. Shale samples (0.5 - 1.5 Kg) were taken at approximately 60 cm. (and shorter) intervals, from a depth of at least 40 cm; they were placed and transported in resistant polythene bags to avoid mixing and further contamination; the top and bottom of each limestone or sandstone sample was marked in the field before it was transported to the laboratory.

Altogether 488 indurated samples were cut with a diamond saw and 342 thin sections prepared; half of each slide was stained by various techniques (Dickson, 1965, 1966; Friedman, 1959, 1971; Evamy, 1968) before covering. Each sample was studied and point counted with the aid of a standard petrographic microscope and point-counting stage, larger individual slabs were polished and photographed.

In order to determine the clay mineral composition of the shales by standard X-ray techniques, 132 shale samples were disaggregated in deionized water by mechanical means; after removing the carbonate content of the shales (Bodine and Fernald, 1973) a few drops of sodium hexametaphosphate were added to prevent flocculation of the clay minerals; the <2 μ m fraction of this assemblage was then separated by centrifuging and subsequently mounted

on ceramic tiles following the method proposed by Shaw (1972), these were dried in a desiccator.

For the purpose of SEM studies of the surface textures of quartz grains, 5 siltstone/sandstone samples were disaggregated in concentrated HCl for 48 hours, the released quartz grains were then washed in distilled water, dried and studied with a binocular microscope to determine possible surface features, iron oxide was removed by boiling the grains in a stannous chloride solution for 20 minutes. After selecting appropriate grains, they were mounted on metal plugs with the aid of UHU fluid, the edges of the metal studs were coated with silver paint for better conduction. The samples were then coated with a gold-platinum alloy vapour in a standard evaporator with a rotating tube (Krinsley and Doornkamp, 1973; Krinsley and Donahue 1968).

A Perkin-Elmer 306 Atomic Absorption Spectrophotometer was used to determine trace and some major elements of 20 selected samples, which were dried in an oven, powdered by agate and ball spinners, washed, dried and placed in separate polythene bags for later use.

A combination of colorimetric, titrimetric, flame photometric and X-ray fluorescence methods was used to determine the major elements of 9 samples obtained in phosphatic nodules and 2 samples obtained from the "Breakish" ironstone bed; these were powdered by agate and ball spinners, dried, weighted and stored in a desiccator. See Appendices 3, 4, 5 and 6.

CHAPTER 2

STRATIGRAPHY

2.1 Nomenclature Problems

Although the presence of different but contemporary sedimentary facies in an area, usually means that the application of stratigraphic terms must involve a plethora of rather meaningless local names, there has been remarkable general agreement over the application of two local names to the three divisions of the Lower Lias of northwest Scotland. However, considerable controversy has existed over the true Zonal limit of each division. This has been partly due to the lack of a standard stratigraphic framework of reference in the past and also partly because of the non-recognition of different but contemporaneous facies present in the area.

The work published on the strata of the Scottish Hebrides before 1819 are of little value to stratigraphy due to the fact that they are confined to systematic descriptions of the rocks without a clear mention of the exact locality or the establishment of a definite succession.

Even Macculloch (1819) had his doubts about the principles of correlation and superposition of strata; nevertheless he distinguished the "Gryphite Limestone" from the "White sandstone (of Scalpa)" and gave the order of the whole of the "Secondary Strata" as "gneiss, red sandstone, gryphite, white sandstone, limestone and shale" and although he does not attempt to place them in a comprehensive stratigraphic order he maintains that the "Gryphite limestones are the lowest in the Lias series".

Murchison (1827), comparing this succession with the oolitic formations of England, defines them as "Lias limestone" which formed part of the members of the "oolitic series of Scotland". Subsequently in 1829 he published a list of fossil species and their localities in Scotland, comparing them with the formations and localities in which the same species occur in England. He also

referred the strata in Mull to "the Oolitic Series".

Geikie (1858), placing the "Pabba shales" and "Scalpa" sands in the Middle Lias, refers to the "Broadford Limestone" and "Breakish Shale" in the Lower Lias and Wright (1858), while essentially disagreeing with him, published a useful table in which English, French and German stratigraphic subdivisions were compared and the stratigraphic levels of the Scottish strata were established accordingly. Bryce (1873) and Tate (1873) refer to the Lower Lias up to the E. raricostatum Zone (cf. Wright, 1858, and Geikie, 1858) as the "Broadford Strata" and to those from the Jamesoni to the Davoei Zones as "Pabba shales".

Judd (1878) divided the Lower Lias into Infra Lias (Planorbis-Angulata Zones) and Lower Lias Beds (Bucklandi-Obtusum Zones, / ^{namely} the Lima or Bucklandi beds of England), the Pabba Series (Oxynotum-Davoei Zones) and the "Scalpa Series" (A. margaritatus-A. spinatus Zones).

Woodward (1896) coined the name "Broadford Beds"; in the Annual Report of the Geological Survey for 1896, it is stated that "The Broadford Beds would include the Zones of A. planorbis, A. bucklandi and A. semicostatus". However in a subsequent Memoir (Peach et al., 1910) the Broadford Beds include the Turneri and Obtusum Zones. No reason is given by Woodward for this, and the previous, fourfold sub-division of the Broadford Beds is repeated. The O. oxynotum and P. davoei Zones are represented by the Pabba shales while the Scalpa sandstones are of A. margaritatus and P. spinatum Zone age.

Lee & Buckman (1920) consider the Broadford Beds to comprise the "Lower Lias Zones up to the Turneri Zone" (p. 3) whereas in page 7 of the Memoir it is stated about the succession in Applecross that "These strata can evidently be referred to the basal portion of the "Broadford Beds", a division of the Lower Lias of Skye thus named by Mr. Woodward, which includes the Lower Lias Zones from that of Psiloceras planorbis to that of Asteroceras obtusum", and on page 12 it is stated that "The Broadford Beds of Skye range from the basement Beds, which probably include the Planorbis Zone to the Obtusum Zone". The Pabba Shales and

Scalpa sandstones in the aforementioned Memoir have the same stratigraphic extent as was given by Peach et al. (1910).

In the Mull Memoir Lee & Bailey (1925) state that the Broadford Beds are

".....regarded as equivalent to the Blue Lias limestone of England. Which includes the zones of Psiloceras planorbis to Arnioceras semicostatum and part of the Obtusum Zone. In the present Memoir, however, on the score of local convenience, all above the Semicostatum Zone is referred to the Pabba Beds".

Arkell (1933) followed the same scheme but referred to the sandy shales of the Obtusum Zone of Loch Aline as the "Loch Aline sandstones"; this term is somewhat misleading and leads to their confusion with the Cretaceous greensand beds which immediately overlie them.

Although MacLennan (1953) follows the subdivisions of the Lower Lias in the Hebrides proposed by Lee and Bailey (1925), she includes the Semicostatum Zone in the Lower Broadford Beds, the Pabba Beds comprise the shales of Turneri to Obtusum Zone age.

Hallam (1959), adhering to the original suggestions put forth by Woodward in Peach et al. (1910), further subdivided the Broadford Beds into the Upper Broadford Beds and Lower Broadford Beds.

Oates (1976) made a detailed study of the stratigraphy and palaeontology of the Mull and Skye areas, drawing attention to similarity of facies among the English Blue Lias and Scottish Broadford Beds. He includes the Turneri and Obtusum Zones of the southern region (Mull, Morvern and Ardnamurchan) in the Pabba Beds whereas in the northern region the Broadford Beds include all the Zones below the Turneri. According to Oates (1976), the Broadford Beds of the southern region of the Scottish Hebrides are similar to the strata of the same age in southern England, therefore the name Blue Lias may also be applied to them; also the junction of "Blue Lias: Broadford Beds/Pabay Beds" is diachronous from north to south, and is taken at the top of the Turneri in the northern area while it is established at the base of the Turneri in the southern area. In the view of the present author, the above two distinctions are inappropriate for the following

reasons:

- i. The Broadford Beds in only two localities of the southern region are equivalent in time and ?facies to the "Blue Lias" of England and even these are 500 miles from the Dorset Coast; they are slightly more arenaceous in places and their mineralogy is altogether different.
- ii. According to Oates (1976), the Broadford Beds of Loch Aline (NM 694453) differ palaeontologically from the "Blue Lias" of Somerset and Dorset in that the giant species Schlotheimia pseudomoreana Spath which succeeds the S. similis fauna is absent from the Scottish sections.
- iii. The Turneri beds in both regions are products of the same process therefore it would be only natural to include them in a common group of strata to facilitate any palaeoenvironmental interpretations.
- iv. Important evidence to be cited later, of a progressive offlap (from north to south) due to a minor regression which affected the Hebrides during the Turneri Zone times, will be lost between the northern and southern areas of study, in the Broadford Beds.
- v. Mere "local convenience" (Lee and Bailey, 1925) which formed the basis of this subdivision, cannot be justified.
- vi. The non-sequences which define the subdivisional boundaries in Oates' work are of entirely different origin in the northern and southern areas respectively, therefore they should not be used as stratigraphic markers.

As seen in Table 1.1 the Hebridean Lower Lias has been subdivided and named differently by various workers since Macculloch (1819). This is mainly due to the great variety of depositional conditions which is so characteristic of the region.

The term "Broadford Beds" was originally applied to the alternating sandstone/shale/limestone divisions of the western Scottish Lower Lias. The principal features of the type district are present in all of the sections in the northern region and in Ardnamurchan of the southern region. However the beds which are

fully represented in Applecross, Raasay and Skye, are absent in Ardnamurchan. In Mull and Morvern the Lower Lias is represented by a limestone/shale facies, the similarity and stratigraphic equivalence of which to the Blue Lias of Dorset has led to many comparisons since Judd (1819). However it would be inappropriate to adopt the term "Blue Lias" to describe their equivalent strata in the Scottish Hebrides.

The Lower Broadford Beds in this study denote those strata which were deposited in northwest Scotland during the earliest Liassic Zones, up to Bucklandi, as a major transgression inundated the southern (Mull) region during the Planorbis Zone times, but which did not reach the northern (Skye) region before Bucklandi Zone times.

The Upper Broadford Beds were deposited during a further spread of the transgression (its maximum geographic extent being attained during Semicostatum Zone times), together with those strata which were deposited as the result of a minor regression, producing two non sequences in the northern (Skye) region at the end of Turneri Zone times. Regressive-type sedimentation continued in the southern region until the end of the Turneri Zone times.

The time is ripe for a precise lithostratigraphic nomenclature to be proposed (attention being paid to the facies present) based on the stratigraphic recommendations of the Geological Society (Harland et al., 1972). The lithologies in each unit are described together with their fauna. Where relevant, details of sedimentology and animal-sediment relationships are noted. Correlation with the standard ammonite Zones is also shown in Tables 2.1 and 2.2. In this study, the Broadford Beds fall within the stratigraphic boundaries applied by Woodward (in Peach et al., 1910), Richey (1930), Arkell (1933) and Hallam (1959) except the Obtusum Zone.

2.2 Biostratigraphy

A combination of poor exposure and unfavourable facies impedes faunal

collection for the purpose of an accurate zonal boundary determination for the Broadford Beds. The zonal determination and correlation for these strata is facilitated however, by collections made during decades of geological work, the most recent of which are those of MacLennan (1953), Hallam (1959) and Oates (1976).

The fieldwork for the present study was made according to the zonal classification put forward by Hallam (1959). Later adjustments were, however, made in accordance with the findings of Oates (1976). In the following section, stratigraphically significant fauna recorded from the strata of each locality are noted and their position within the sequence specified.

Applecross (NG 714434)

Ammonites found in the Broadford Beds of this locality are shown in Figure 2.1; these include a Schlotheimia sp. (Lee and Buckman, 1920) indicating the presence of the Liasicus or Angulata Zone at the 18 m. level. At 46 m., Thecosmilia martini which was considered by Hallam (1959) to be contemporaneous with those developed in Ob Breakish (N. Strath) is seen. On the basis of the occurrence of Schlotheimia sp. below the coral bed in Ob Breakish, Oates (1976) favours Trueman's (1942) conclusion that the bed with Thecosmilia martini is within the Angulata or near the base of the Bucklandi Zone and not within the Bucklandi as Hallam (1959) suggested.

Coroniceras coronaris (Quenstedt) was found at the 51.20 m. level by Hallam (1959), indicating the presence of the Rotiforme Subzone in this locality. Further proof for its existence is Coroniceras hyatti Donovan which was reported by Oates (1976). Therefore since the Breakish coral bed is of Hettangian age, the Conybeari Subzone in this locality is very thin.

Hallam (1959) discovered Plagiostoma and Parallelodon cf. hettangiensis which he tentatively accepted as evidence for the presence of the Johnstoni Subzone of the Planorbis Zone in the basal parts of these beds.

Among the fauna recorded by Judd (1878) is Coroniceras kridion (Zieten) which is an Upper Rotiforme Subzone indicator. However it was not found in the main section (i.e. Allt nan Breugh NG71954375), and the presence of this Subzone at the topmost part of the section can only be inferred. Liostrea sp. occurs at various horizons throughout the section; because of its limited stratigraphic range and as it is not associated with Gryphaea, it may be concluded that it is of Hettangian age (Oates, 1976).

Ob Lusa - Ob Breakish (NG 700250)

Detailed work on the stratigraphy of this locality was carried out by Peach et al. (1910), Spath (1922a), Hallam (1959) and Oates (1976). The study of Spath was based on the ammonites collected from the western Ardnish coast where the establishment of a succession is hindered due to considerable faulting.

The occurrence of Isastrea murchisoni 2 m. above the base of the Broadford Beds in Ob Lusa is in line with Trueman's (1942) intention that the base of the Broadford Beds in Lusa is of Angulata Zone age.

The bivalve fauna of this bed and the one immediately overlying it, e.g. the bed with Cardinia cf. concinna indicates an Angulata Zone age. The occurrence of Coroniceras aff. conybeari, which was reported by Hallam (1959) reported from the 9.55 m. horizon (Figure 2.2) signifies the Conybeari Subzone, the basal member of Sinemurian stage.

The beds rich in Thecosmilia martini and Liostrea hisingeri are interpreted as representing the Angulata Zone (or near the base of the Bucklandi Zone) at the 19 m. level; Oates (1976) found a single Schlotheimia sp. at about 3 m. below one of the coral outcrops.

Gryphaea arcuata assemblages, together with Charmasseiceras and Arnioceras spp. indicate the presence of a Semicostatum Zone age in the Breakish Peninsula; the Coroniceras reported by Hallam (1959) indicates the Reynesi Subzone at 33 m. above the base.

Piarorhynchia juvenis (Quenstedt) is found together with Chlamys calva; these and the ammonite evidence suggest a Semicostatum age for the oolitic ironstone bed (bed. no. 20 of Hallam) at the 33 m. level.

The sandstones and shales of bed 21 and 22 of Hallam at the 38 and 40.50 m. horizons respectively, are full of Arnioceras aff. semicostatum, Coroniceras reynesis and Coroniceras aff. parthenope. The Scipionianum Zone is represented by a sandstone (bed 27 of Hallam) at the 60 m. horizon which yields Agassiceras scipionianum.

The sandstones which constitute bed 28 of Hallam (1959) have yielded Euagassiceras resupinatum, which represents the Sauzeanum Subzone at the 60.50 m. horizon.

North of the pier in Broadford Bay, at the headland of Rudhanan Eireannaich, Hallam discovered Pararnioceras cf. alcinoe which indicates a Turnerian age for 3 m. of cross-bedded micaceous sandstones which are exposed. Also in the overlying shales exposed to the west of the headland, Asteroceras margaritoides and Promicroceras planicosta attest to the presence of the Obtusum Zone in this locality.

Loch Slapin (NM 586184)

The Broadford Beds unconformably overlies the Ordovician Durness limestone on the shores of Loch Slapin. On the basis of Hallam's (1959) finding of Arnioceras in the Broadford Beds 3 m. above the base, the Loch Slapin succession is regarded as representing the Semicostatum Zone, with the ?Euasteroceras spp. which is an Upper Semicostatum Zone indicator occurring towards the top (Figure 2.3).

Loch Eishort (NM 625163)

The Broadford Beds are well exposed along the northern shore of Loch Eishort. They are intruded by numerous dolerite dykes and faulting has disturbed much of the strata. Based on ammonite collection, Hallam established a coherent section identifying the various Zones which are present.

No ammonites have been found in the lower, more arenaceous and calcareous portion of the section. However, by comparison with the strata exposed in Lusa, northern Strath, a Bucklandi Zone age may be suggested for these.

Sandy micaceous shales with abundant Gryphaea arcuata and containing Arnioceras aff. semicostatum, positively show the presence of this Zone at the eastern Waterfall (NM 623163); the contact of these beds with the underlying beds is obscured by a fault.

Although Hallam's beds 8-17 (56 m. thickness) have yielded ammonites signifying the Semicostatum Zone, none are of the genera characteristic of the Subzones.

The ammonites Euasteroceras sp. and ?Caenisites were found by Hallam (1959) indicating a Turneri Zone age for bed 18, at the 12 4 m. level. This age is confirmed by Oates (1976) who also found ?Microderoceras near the base of the Turneri Zone sands.

The Turneri sandstones of Loch Eishort, 35 m. thick, are overlain by 7 m. of sandy shale which yield Asteroceras margaritoides, Promicroceras planicosta and Xiphoceras sp. indicating the Stellare Subzone of the Obtusum Zone (Hallam,

1959; bed 19). A specimen of Eparietites sp. found by Oates (1976) also confirms the presence of the succeeding Denotatus Subzone. A non sequence is evident from the lithological and palaeontological evidence. It possibly involves

the top of the Turneri Zone and the whole of the Obtusum Subzone at this locality.

This apparent non sequence was reported by Oates (1976) (see Figure 2.4).

Raasay (NG 595385)

The Broadford Beds here are disturbed by various intrusions, but faulting is less frequent than elsewhere; on the occurrence of Liostrea irregularis var. hisingeri in beds 6 and 13 of Hallam (1959) at the 12.72 and 26.15 metre intervals respectively and by lithological comparisons with the Lower Broadford Beds in northern Strath, an ?Angulata to Bucklandi Zone age may be assigned to these beds which are up to 29.50 m. thick.

The base of the Semicostatum Zone is taken at 30 m. above the Lias/Trias contact which is marked by two bands of Gryphaea arcuata (bed 17 of Hallam). The succeeding bed (no. 18 of Hallam) is compositionally similar to the oolitic ironstones found in northern Strath; it contains Piarorhynchia juvenis together with ferruginous pellets. The presence of Arnioceras spp. in the succeeding bed (no. 19 of Hallam, 1959) ascertains a Semicostatum Zone age whereas Agassicerias scipionianum found in the bed immediately above it (no. 20 of Hallam, 1959) marks the Scipionianum Subzone. Hallam's bed 21 yields numerous Euagassicerias resupinatum specimens which suggest a Sauzeanum Subzone here.

The Turneri Zone which is represented by beds 22 to 26 of Hallam (1959) is 26 metres thick, the base of which occurs at 71 metres above the base of the section. Hallam's (1959) dating involves the occurrence of Euasteroceras cf. brooki (bed 22), Euasteroceras aff. turneri (beds 24, 26) and Microderoceras birchi (bed 26). Oates (1976) maintains however that

"the range of Microderoceras birchi probably extends far below the Birchi Subzone, and the genus possibly even originated in the Semicostatum Zone".

He also recorded a specimen of Euagassicerias 2 metres above the base of the Turneri Zone established by Hallam and readjusted its thickness accordingly (Figure 2.5).

Arran

The Broadford Beds exposed in this island were dated as belonging to the Planorbis Subzone of the Planorbis Zone by Trueman (1942) on the evidence of Psiloceras planorbis. Tyrell (1928) found specimens of Schlotheimia ?angulata in some higher beds, thus indicating the presence of the Angulata Zone.

Ardnamurchan

Mingary Pier (NM 493627)

According to Richey et al. (1930), the Broadford Beds exposed in this locality, have yielded 'Ostrea' sp., 'Mytilus' sp. together with specimens of 'Cypricardia' porrecta Dumortier which is a species belonging to the Planorbis Zone in the Rhone Basin. Therefore a Planorbis Zone age for the strata exposed here, can be tentatively suggested. It should be noted that bivalves are not regarded as stratigraphic indicators (see Figure 2.6).

Mingary Castle (NM 506630)

In the Ardnamurchan Memoir (Richey et al., 1930), evidence for the existence of the Bucklandi Zone near the south extremity of Rudha a Mhile was very tentatively suggested based on the finding of Pleuromya sp. but the true age of the strata immediately overlying the Trias in this area is still uncertain and open to question, due to the lack of further detrimental evidence.

Based on the findings of an Ostrea of the group of O. arietis Quenstedt, Richey et al. (1930) suggested that "it may be allied to an unnamed form from the Angulatus Zone figured by Dumortier (1864)". The presence of Gryphaea arcuata Lamark in the strata on the shore south-east of Mingary Castle, probably points to the presence of the Semicostatum or the top of the Bucklandi Zone here. The section in this locality is extremely faulted and although good faunal collections were made by 'Mr. Manson', no attempt except Richey's was made to establish a definite section. The presence of Arnioceras in most of

the sections points to the existence of the Semicostatum Zone at Mingary Castle. The beds with arietitid of Obtusum Zone age which were reported by Spath (1922a) from this area were not found by the survey.

Oates (1976) succeeded in compiling a 6.50 m. section on the Castle foreshore. These are beds with abundant Arnioceras sp. and Euagassiceras sauzeanum, indicating the presence of the Upper Semicostatum Zone in the vicinity. In general it is considered that beds containing Gryphaea probably represent the Bucklandi or early Semicostatum Zone in this area (see Figure 2.7).

SWORDLE (NM 539710)

i. Ockle Point (NM 550717)

No fossils of definite stratigraphic value were found here. At the base of the series Richey (1930) reports the presence of Modiola hillana J. Sowerby and Perna infraliassica Quenstedt which he regards as representatives of a Hettangian or Sinemurian age; however bands of Gryphaea arcuata (?Semicostatum Zone) can be seen in the coastal cliffs which are inaccessible and cannot be related to the section in any way. The total thickness of the beds in this locality is 9 metres (see Figure 2.8).

ii. Garbh Rubha (NM 536706)

a. At this locality the base of the Lias directly overlies ?Triassic conglomerates and it is possible that a short period of erosion is also involved at this junction. No significant fossils were found.

b. The fairly ubiquitous occurrence of Liostrea hisingeri in the 20 metre thickness of this section signifies a Hettangian age for it (see Figure 2.8).

c. At the base of the section (NM 510710), at the 2 metre level, Liostrea hisingeri suggests a Hettangian age; this is further authenticated by Schlotheimia sp. found by Oates (1976) from the same beds. At a level 15 m.

above the base, a two metre thick bed crowded with Thecosmilia martini in different horizons is found; this may be compared with the Breakish Coral Member in northern Strath therefore an Angulata Zone age (?top Complanata Subzone) may be inferred. This horizon of corals was first noticed by Richey et al (1930). A horizon of crushed Pinna shells is seen at the 25.5 m. level together with a horizon of large Plagiostoma gigantea at 26.5 metres and another pinna bed at 27 metres; at 27.20 metres the first Gryphaea arcuata bed is seen. Oates (1976) regards these occurrences as 'reminiscent' of his beds +6 to +70 in the Morvern succession. They were dated as marking the Conybeari/Ratifforme Subzonal boundary. He has also reported Gryphaea arcuata occurring near the base of the Conybeari Subzone.

Craignure (NM 734313)

The basal part of the lower 6 metres of this section is unfossiliferous, except for a bed full of Lisotreia hisingeri, found by the author (1976) at the 2.50 m. level. At about the 8-10 metre level, indications for the presence of Gryphaea and Lima are seen (Oates (1976) believes the latter to be of Plagiostoma gigantea affinity). Gryphaea beds become numerous towards the 10-12 m. level. The ?Schlotheimia thalassica fragments found by Spath (1922a) as a possible evidence for the Bucklandi Zone, is the only definite Zonal indicator found for this lower succession to date however.

At the 25 m. level, the Survey have reported the characteristic Semicostatum Zone ammonites, Arnioceras bodleyi, A. hartmani and A. semicostatum.

Oates (1976) refers to a ?Coroniceratid at the 27 m. level which may be referred to the Sinemurian. Above this level Arnioceras spp. signifies a lower Semicostatum Zone age, in the absence of higher Semicostatum Zone indicators.

Turneri Zone indicators such as Promicroceras planicosta (28 m. level) Microderoceras (32.5 m. level), ?Caenisites sp. (44 m. level) have been

reported by Oates (1976); in the Craignure succession he has also found specimens of Promicroceras planicosta which is a valid indicator for the presence of the Obtusum Zone at least for the top 6 m. of this section (see Figure 2.9).

Loch Aline (NM 694453)

Allt Leacach

Lee and Bailey (1925) and MacLennan (1953) give accounts of the Liassic stratigraphy of this locality. More recently the detailed stratigraphy of this area has been worked out by Oates (1976, 1978); he recognised the Allt Leacach section to be one of the most complete of the Lower Lias sections in Morvern. The lower 12.75 m. here (beds -39 to +37 of Oates) are assigned to the Angulata Zone on account of the occurrence of Schlotheimia similis from various horizons. On the basis of its relation to Schlotheimia (an indicator for the presence of the Complanata Subzone), the presence of the top parts of this Subzone may be assumed in this section. The succeeding 7 metres are assigned to the Bucklandi Zone (Conybeari Subzone) on account of Vermiceras conybeari specimens found in the 5.62 m. level. The base of the Rotiforme Subzone is taken at bed 70 of Oates (1976) 80 cm. above which he has found a specimen of Coroniceras rotiforme. This Subzone is 3.85 m. thick and Coroniceras kridion has been reported from its top and specimens of Coroniceras rotiforme are found in various beds throughout the succession. The Bucklandi Subzone of the Bucklandi Zone is 5.30 m. thick, the top of which is very easily accounted for by a distinctly ironstained reddish surface. The bottom is taken at the bed immediately overlying bed 97 of Oates (1976); he reports Coroniceras latissulcatus from bed 98; Arnioceras sp. (bed 118 at the 27 m. horizon) and Oxytoma inaequalis are also present.

Here the total thickness of the Angulata and Bucklandi Zones is 28.90 m.; it should be mentioned that according to Oates the Sulciferites sp. (bed

no.30) and Vermiceras sp. (bed -6) are unusually early representatives of these genera which are normally associated with the Sinemurian Stage. At the 12.95 m. level, beds full of Gryphaea occur. The first occurrence of Gryphaea arcuata is recorded here and they are regarded as representing the earliest Bucklandi horizon, following Hallam's (1960) description of the English and Welsh successions. The only in situ specimen of Vermiceras conybeari, which is the index species for the lowest subzone of the Bucklandi Zone, was recorded by Oates (1976, 1978) in Allt Samhnachain (bed +61); in bed +76 of the Allt Samhnachain quarry he recorded Coroniceras (Coroniceras) of the Rotiforme group, therefore the Conybeari/Rotiforme boundary is taken between these beds, the most likely position being at a horizon of condensed shell material and ?ironstone (19.87 m. above the base of the section at Allt Leacach.

A specimen found at the 23.90 m. interval is a fragment of a closely ribbed ammonite which was also found by Oates (1976; bed 98) which he considered to be a Coroniceras (Coroniceras); as the Arnioceras sp. (whose earliest representatives are found in the Bucklandi Subzone) appear at the 27 m. horizon and the Oxytoma inaequalvis only 50 cm. above them, it is reasonable to consider these topmost beds in Allt Leacach as of Bucklandi Subzone age.

Although in most of the sections here, the Bucklandi Zone beds are directly overlain by Turner Zone dark shales; the identification of numerous specimens of Arnioceras sp. in the sequence WNW of Old Ardtornish House (NM 686435) signifies the existence of horizons in the Reynesi Subzone of the Semicostatum Zone (Oates, 1976). The ochrous-weathering bed (MacLennan, 1953) which separates the shales of Turner Zone age from their underlying Bucklandi-?Reynesi shales, is representative of a period of major denudation and erosion as indicated by an irregular surface and iron oxide concentration (MacLennan, 1953). However, Oates (1976) reports that in those

sections other than the Allt Leacach one (i.e. Allt Mor and the shore WNW of Ardtornish House) a small thickness of crinoidal limestone ("biosparite") has yielded arietitids of Coroniceras reynesi (Spath) type and Arnioceras falcaries. He has recognised a difference in age between the shales immediately underlying this "biosparite" in Am Moidart, Allt Mor and those on the shore WNW of Old Ardtornish House, hence he has attributed this to a minor period of disturbance and the possible existence of a non-sequence at this level, prior to the main unconformity which omits the Scipionianum, Sauzeanum and even parts of the Reynesi Subzones in this area.

The strata overlying the ochrous-weathering surface have yielded several specimens of stratigraphic value; these were reported by Oates (1976) whose findings established that the 8.40 metres succession immediately overlying the unconformity is not lower than the Birchi Subzone age on account of the occurrence of Microderoceras birchi at the 1.80, 2.70, 4.30 and 6.90 m. levels above the ochrous-weathering bed and Caenisites turneri at 0.50 and 3.00 m. level. Promicroceras has been reported from throughout the section except the lowermost parts and those overlying the Turner beds. The top of the Turner Zone is established at the horizon 8.40 m. above the ochrous-weathering bed on grounds of lithological difference. Oates reports an Aegasteroceras specimen at the 9.40 m. level and Asteroceras sp. (?A. obtusum) at the 9.70 m. level. Xiphoceras sp. (19.50 m. above the ochrous-weathering bed) and Asteroceras obtusum 18.20 m. above the same bed occur in this section as well.

There is no evidence for the presence of the Stellare and Denotatus Subzones in the Morvern section. However, studying the various sections it is evident that the total thickness reaches 50 m. (see Figure 2.10a, b).

Wilderness (NR 404287)

As reported by Lee and Bailey (1925), this locality (together with that of Arran)

is the only Hebridean representative of the Planorbis Zone which has yielded the index ammonite Psiloceras planorbis. The strata here are 10.50 m. thick and succeed the ?Rhaetic strata which are thought to be equivalents of the Contorta shales of England. The contact cannot be seen in the field for the Planorbis strata are found in a faulted block which lies adjacent to Triassic red clayey siltstones and cross bedded sandstones.

At their base these beds are marked by a prominent oyster bed (Liostrea hisingeri) well known from the Lower Hettangian of England and Skye. The succeeding beds of limestone which have yielded specimens of Psiloceras planorbis are indicated on Figure 2.11.

Carsaig (NR 540222)

At this locality, near Loch Buie, the Broadford Beds are extremely faulted and are poorly represented. The only representatives of possibly the stratigraphically higher portion of these beds can be seen at An Cuileim (NR 588227) on the coast about a mile west of Glenbyre farm where highly metamorphosed, vertically dipping strata show remains of Gryphaea arcuata and crinoidal debris. (Plate 2.1a-c).

Rubha na Fear (NR 623194, 63152000)

The "Broadford Beds" here are exposed on the shore opposite Frank Lockwood's Island. No ammonites were found in this section, however on the basis of the occurrence of Gryphaea, together with the identification of Triassic strata below and Pabba beds above them, a tentative Lower Liassic (?Planorbis to Lower Semicostatum age may be assigned to them. (Plate 2.2a-c).

2.3 Non sequences

The non-sequences here are defined as any surface of erosion or non deposition within a stratal sequence. It should also be borne in mind that

some of these horizons only indicate very short periods (perhaps some not more than a few thousand years) during which sedimentation was arrested; they are identified in the field not only on the basis of changes in faunal content and character of the sediment, but also by phosphatized shell beds, nodular pebbles and glauconite-rich layers.

During the period between Planorbis and Semicostatum Zone times no such non sequence can be recognised in the area of study. A major non sequence is identified in the Leacach section of Morvern (Lee^{and} Bailey¹⁹²⁵), 28.90 m. above the base of the section. Here, the topmost part of the Bucklandi Zone shales and limestones is represented by a 50 cm. thick bed of ochrous-weathering limestone with an irregular surface. This bed is at least 40 cm. thick in places and the iron-stained weathering of its top penetrates a few centimetres. It is underlain by four Gryphaea beds, each up to 12 cm. thick, with the intervening shales not yielding any significant fauna. The gryphaea beds are full of crushed shell material consisting of other bivalves; ?crinoidal material and occasional broken shells of Pinna; coalified fragments and sand-grade material are minimal. The Gryphaea beds form more calcareous units within the shale. These are overlain by the iron-stained limestone bed.

The shales with Gryphaea beds are dated by Oates (1976) as high in the Bucklandi Subzone and the shales which immediately overlie the reddish-weathering limestone bed are considered to be of Turner Zone age. These are silty and micaceous in the basal metre, the silt content decreases upwards at the expense of a clayey shale with limestone nodules (2 to 5 cm. thick); no other evidence in the form of biogenic or sedimentary structures are observable.

MacLennan's (1953) suggestion, that the sequence is transitional from limestone below the reddish-weathering (Bucklandi) zone into "hard limey mudstones in which there is a gradual disappearance of the fauna of the Broadford Beds" cannot be supported. It is evident that the sediments of Turner Zone age lie non sequentially over those of the Bucklandi Zone.

This non sequence involves in most cases the two upper Subzones of the Semicostatum Zone with a small thickness yielding Arnioceras sp., considered by Oates (1976) to represent a horizon in the Reynesi Subzone, present WNW of Old Ardtornish House (NM 686435).

It is also significant that the beds of shale underlying a 'biosparite' bed, yielding Coroniceras reynesi and Arnioceras falcaries in Am Moidart, Allt Mor and on the shore WNW of Old Ardtornish House are each of a different age (Oates, 1976), which may demonstrate minor periods of ?erosion and ?non sequence before the one already mentioned; however the duration of this minor period is uncertain and it almost certainly does not affect beds much earlier than those of the Reynesi Subzone.

The section in Craignure Bay (Mull) is extremely faulted and evidence of any non sequence/erosion surfaces is also lacking. However Oates (1976) indicates a top Semicostatum Zone unconformity in Figure 93 of his work. This is not explained in his text and therefore its existence is doubtful, but due to its proximity to the Loch Aline locality it may perhaps be correlated with the major Semicostatum unconformity of that area. In the Mull area it may involve only a minor thickness of the sediments of late Semicostatum age.

The Turner Zone is marked with indications of contemporary erosion in both the northern and the southern area. In Morvern the topmost 60 cm. in the Birchi Subzone of Loch Aline is marked by a sandy bioturbated limestone with abundant angular fragments of micritic pebbles up to 3 cm. in diameter which are phosphatic in places and abruptly overlie the shales underlying them. The succeeding Obtusum Zone age beds are of an entirely different character. This bed has been recognised by Oates (1976), but it is not represented on his final diagram.

In the Ardnamurchan area, owing to the paucity of exposure and also the discontinuous nature of the outcrops, it is not possible to distinguish

periods of erosion and iron-deposition.

In the northern region, non sequences are developed near the top of the Turner Zone and can be seen in southern Strath (Loch Eishort) and Raasay (Rubha nan Leac).

The Turner Zone is represented in Loch Eishort by 39 metres of cross bedded calcareous sandstones. Their contact with the overlying Obtusum Zone sandy shales is reported by Oates (1976) and also identified by the author (1975) as representing an erosional surface involving a short period of time until it was inundated again by the bioturbated, carbonaceous shales with sand lenses (flaser and lenticular bedding) of Obtusum Zone age.

The upper part of the Turner Zone is reported to be possibly absent with the Obtusum Subzone of the Obtusum Zone certainly missing (Oates, 1976). This evidence clearly signifies a period of erosion and possibly emergence of the Skye region during this time. At the topmost part of the sandstones (Turner) of Rubha nan Leac (Raasay), an unconformity embraces all of the Obtusum Zone and the Birchi Subzone (Oates, 1976). This may be regarded as the culmination point of minor periods of emergence and possible erosion identified by the coarsening-upward cycles in the Turner Zone sands, with the top surfaces of each complete set marked by a sharp horizon of iron oxide staining followed by the deposition of dark shales.

The significance of these erosion/non sequence periods will be discussed later.

2.4 Lithostratigraphy

As previously mentioned, the "Broadford Beds" have been studied by numerous geologists for well over a century. Their studies have normally been concerned with the palaeontology and stratigraphy of these beds with little detailed attention being paid to the facies present and depositional environments which they represent.

The major problems of facies analysis are:

- a. Scattered exposure.
- b. Intense baking and disturbances due to the Tertiary intrusions.
- c. Frequent paucity of index fauna.
- d. Marked lateral facies variations.
- e. Many episodes of erosion and/or non-deposition.

The name "Broadford Beds" has referred to the sequence of Lower Liassic sandstones, limestones and shales of northwest Scotland since Woodward (1896) and geologists have been content to rely on this terminology, basing their work mainly on chronological identifications. While ignoring the lithological differences which exist within this unit, many have tried to adopt more broadly based terminologies capable of being used over a much wider area. Repeatedly the Broadford Beds have been compared with the Blue Lias of England, but those recognising the local differences have refrained from adopting a more generalized nomenclature. This may also be the reason why the Broadford Beds have received only a cursory notice as far as facies and environmental relationships are concerned.

It seems desirable in this thesis to introduce a series of local lithological names to facilitate detailed facies interpretation while preserving the term "Broadford Beds" for more general use.

A major hindrance to such a classification has been the lack of a refined stratigraphic framework and control based on faunal occurrences.

Recent progress has been made in the recognition of the standard ammonite Zones represented by the Broadford Beds in northwest Scotland (Oates, 1976, 1978) and a formal lithostratigraphic proposal can be set up along the guidelines explained in section 2.1.

In the Minch Basin, during Lower Liassic times onshore/offshore sequences recur in the Skye area while condensed argillaceous offshore sequences occur in the Mull area.

The nomenclature involves two groups, four formations and twenty two members each of which has been correlated with the standard ammonite zone (Dean et al., 1961) as shown in Tables 2-1 and 2-2.

2.4.1 I - Broadford Beds Arenaceous Group

12-140 m. Type section = Strath district, Skye.

The 'Broadford Beds', as originally defined by Woodward (1896), are given a Group status.

I-a - Milton Formation syn. Lower Broadford Beds
(Hallam, 1959) 47.30 m. Type section = Applecross
(NG 714434).

I-a-i - TorrMor Member (new name), 1.50 m. Type section =
TorrMor (NG 630164), northeast of Loch Eishort main
section(Fig. 2.12).

This sequence is referred to in the Skye Memoir (Peach et al., 1910) thus: "... inconsistent conglomeratic beds of quartz pebbles in a calcareous and sandy matrix develop frequently in the lowest part of the Lias and are not always easy to distinguish from the Triassic conglomerate ... the Liassic limestone near Boreraig changes laterally into conglomerates of quartz and sometimes of limestone". These Liassic conglomerates have subsequently gone unnoticed; in this locality they are easily confused with the Triassic conglomerates underlying them but may be distinguished on the basis of Torridonian clasts of red arkose which exist in the Triassic conglomerates but are absent from the Jurassic ones.

The Liassic conglomerates clearly overlie the orthoquartzitic sands of ?Rhaetic age and contain numerous fragments of minute shells of gastropods and bivalves.

The matrix-supported conglomeratic beds are up to 0.80 m. thick and contain angular quartzite fragments, up to 9 cm. in diameter, smaller chert

fragments (grey) are also seen; sorting is very poor and the cement is mainly blue limestone. The conglomerates alternate with lenses of cross laminated pebbly calcarenites with shell beds and wavy shale partings. The size of the pebbles decreases as the outcrop is traced laterally and also infrequent coalified plant material is seen. Pebbles are frequent in the laterally equivalent beds.

Age - Post Rhaetic - ?pre Angulata Zone time.

I-a-ii Applecross House Member (new name) 6 m.

Type section: Southeast of Applecross House (NG 714434), see Fig. 2.13.

This Member may be divided into three distinct types of beds. Immediately overlying the Torr Mor conglomerate in the type section 1 - 1.5 m. of cross bedded pebbly sandstones with shaly partings are developed; at Applecross the conglomerates are not well developed but cross bedded pebbly calcareous sandstones occur containing a bivalve and gastropod fauna together with plant remains. These are succeeded by 5 metres of alternating shales and pebbly calcareous sandstones; faint evidence of current activity is seen in the sandstones which are lens shaped (their thickness varying from 20 cm. to 1.5 m.) and contain pebble bands. The shales, which are of a much greater thickness than the correlative development in Skye, are dark grey with dark black-brown, flame-like (wavy) features. They are moderately plastic and mostly contain broken remains of thick-shelled bivalves. X-ray work on these shales has shown them to be rich in smectite. There follow four metres of grey-weathering oolitic calcarenites. These beds have been an important source of lime in the past and can now be examined in the abandoned quarry (NE 759443).

Fauna: Limited; Antiquilima succincta, Parallelodon hettangiensis, Liostrea sp.

Age: No ammonites, but evidence for a Hettangian age is based on the existence of the following fauna: Liostrea sp.: (Hettangian); Lima succincta, Parallelodon hettangiensis: (?Lower Liasicus Zone). See Fig. 2.13 .

I-a-iii Lusa Coral Member syn. Lussay Coral Bed (Woodward, 1896) 0.50 m.

Type section: Lusa Bay (NG 700250)

Corals are found in microsparite lenses 5 to 10 cm. thick and 25 to 30 cm. wide. The lenses are separated by paper-thin shales and in places the corals seem to have been disturbed and inverted from their life positions prior to lithification.

The nodular limestone comprises the compound coral, Isastrea murchisoni and inconspicuous bivalves. The intervening thin (2 to 5 mm.) shales are too altered for any meaningful clay mineral determination. The development of this Member is seen at Lusa and elsewhere in the Strath district (NG 210648, NG 210647). Tracing this bed southwards into Heast, it is replaced by beds of blue limestone with shale partings and can easily be confused with the "Breakish Coral bed". It can also be seen in Applecross. See Fig. 2.13.

Age: Top Liasicus or basal Angulata Zone.

I-a-iv Lower Sandstone Member (new name) cf. division 1 of Woodward (1896)
~18 m.

Type section: This Member represents three facies each of which show lateral and vertical variations; a type section for each is given below:

I-a-iv- α Applecross

I-a-iv- β Lusa

I-a-iv- σ Loch Eishort

I-a-iv- ζ Raasay

I-a-iv- α Applecross (NG 714434)

Type section: Applecross cf. beds 9-11 of Hallam (1959) 10.70 m.

The first appearance of shales overlying bed no. 8 of Hallam (1959) at 13.60 m. above the base of the Triassic/Jurassic contact is considered to be the first representative of this Member here; 1.5 m. of plastic, smectite-rich green grey shale beds 20-30 cm. thick, interbedded with thin beds of sandstone are succeeded by 2 m. of nodular oolitic limestone with very thin shale partings

and sandstone beds. Most of the nodules are draped with clays. The shale beds thicken towards the top and are overlain by a 1.20 m. development of calcareous pebbly sandstones. The overlying six metres is a coarsening upwards sequence grading from silty oolitic limestone at their base to marly sandstones at the top.

The alternations indicated above are seen together with a high density - low diversity faunal assemblage. Laterally impersistent, nodular beds of oolitic limestone with thin shales and pebbled beds are frequent.

Thalassinoides burrows are also present.

Age: ?Lower Angulata Zone.

I-a-iv-β Lusa (NG 700250)

Type section: Lusa Bay cf. beds 3-8 of Hallam (1959) 11.75 m.

Immediately overlying the Lusa coral bed is a sandstone unit with mound structures of problematic origin together with thin parallel cross laminated argillaceous beds containing scattered pebbles; these sheets of sandstone are laterally persistent and show more or less sandy parts with a high proportion of coalified plant material and broken shells. Some sedimentary structures suggest the presence of ?rill marks.

The fauna consists of low diversity - high density assemblages of Cardinia cf. concinna, Liostrea irregularis var. hissingeri together with thin-shelled gastropods also present; previous reports of marine ammonites (Hallam, 1959) cannot be confirmed as the present author was not successful in obtaining any.

Age: No determination, ?middle Angulata Zone.

I-a-iv-σ Loch Eishort (NM 625163)

Type section: Loch Eishort cf. beds. 3-6 of Hallam (1959) 14 m.

Overlying the alternating calcilutites and sandy calcarenites of the previous Member, these are poorly sorted calcareous sandstones. The top

surfaces of the beddings are uneven and they are overlain by an alternating development of wavy bedded sandy silty limestones; oolitic beds are also common. The top metre consists of coarse orthoquartzitic sandstones with calcite cement. These are all recognized in the field as very low ledges.

The succession is ripple cross laminated at its bottom portion with tabular and trough cross laminations present, uneven surfaces are very common and the cross laminations are lined with coalified material together with brownish-oxidized material. Each bed with ripple cross lamination is truncated at its top by a parallel laminated sandstone sheet; very small pebbles are also found in the beds and on their top surfaces. The overlying beds are faintly cross laminated in the sandstones but the limestones are oolitic and show no evidence of current activity, no sedimentary structures can be seen either; their contacts are wavy. The topmost white ortho-quartzitic sandstone is strongly cross-laminated, with planar and trough cross laminae present. The sediments are clean and well sorted, showing no evidence of biogenic activity or the presence of driftwood.

Age: ?Angulata Zone

I-a-iv-ζ Raasay (NG 595385)

Type locality: Rubha nan Leac cf. beds 6-11 of Hallam (1959) 9 m.

The equivalent of the Lower Sand Member in this area is a succession (up to 9 metres) of nodular calcilutites with shale partings and a minimal sand or silt content. The nodular limestones here were produced by the burrowing activity of crustaceans and resemble Thalassinoides burrows.

They comprise Hallam's beds 6 to 10; oysters and gastropods occur, the common species is Liostrea irregularis var. hisingeri. Pectinids are present and some are covered with encrusters.

Age: ?Top Angulata Zone.

I-a-v. Ob Breakish Coral Member syn. Breakish Coral Bed, Woodward (1896)

0.60 m.

Type locality: Breakish (NG 682241)

Lithologically, these beds comprise up to 3 metres of blue calcilutites and shale partings with a moderate smectite content. The thickness varies from 10 to 25 cm. in northern Strath and the beds tend to form impersistent lenses (Fig. 2.13). At Lusa, three units of limestone may be distinguished containing the slender, branching coral Thecosmilia martini and a well developed mollusc fauna including Ptychomphalus, Gryphaea arcuata and Lima gigantea together with Isocrinus, Calcirhynchia calcaria and Zeilleria perforata (Hallam, 1959). The corals seem to terminate at their junction with the shale partings.

Age: Top Angulata Zone (base of Conybeari Subzone).

I-a-vi. Upper Sandstone Member: Division 3, 'sandstones' of Woodward
(1896); Hallam (1959), beds 10, 11, 12

Type locality: Ob Lusa (NG 700250) 4.5m.

At the base a thin (15-20 cm.) moderately well sorted calcareous sandstone is succeeded by a development of sandstone sheets interbedded with silty limestone lenses. The sandstones contain abundant coalified wood fragments and at the junction of the beds a peculiar ?coalified material appears to "ooze" out. (See Fig. 2.14).

The lowermost sandstone beds are sheets with planar cross lamination whereas towards the top the sheets become distinctively trough cross bedded; ultimately they become a succession of wavy sandstone sheets.

Bivalves and gastropods are very common and include ?Pinna sp.; Thalassinoides is also seen in the Applecross section which shows the same lithological development. In Loch Eishort however these beds are represented by marl/nodular limestone alternations, the marls being replaced by sandstones towards the top. In Raasay the same type of nodular limestone development is

also seen.

Age: The report of Coroniceras coronaris, Coroniceras kridion and Coroniceras hyatti by Oates (1976) together with Coroniceras rotiforme by Hallam (1959) gives an Upper Bucklandi Zone (Rotiforme Subzone) age for this Member.

I-a-vii. Breugh Pebble Member (new name) 0.5 - 6 m.

Type section: Allt nan Breugh, Applecross (NG 714434)

At its type section this Member is a cross laminated sandstone with a laterally variable thickness, consisting of well rounded pebbles of white, pink and red quartz with chert; it grades upwards into finer calcareous sandstones. (Fig. 2.14).

The bottom parts of these beds exhibit planar and parallel cross bedding which develops vertically and laterally into trough cross bedding with the troughs lined with pebbles of variable sizes. As these grade upwards, they are represented by ripple cross-laminated, well sorted yellowish sands. The topmost 0.80 m. of this Member is represented by an intensely ripple cross-laminated unit with poorly sorted but rounded pebbles. The fauna consists of broken and reworked bivalve shells which are very scarce.

Age: Bucklandi Zone (Rotiforme Subzone).

I-b. Strath Formation (new name) syn. Upper Broadford Beds (Hallam, 1959)
98 m.

Type section: Strath district, Skye.

I-b-i. Lower Teampull Chaon Member (new name) Beds 8, 9, 10, 11 and part
of 12 of Hallam (1959) 21.50 m

Type section: Loch Eishort, on the shore east of the site of Teampull
Chaon (NG 622162)

The lithology comprises a series of alternating sand, micaceous shales

and sandy limestone couplets, each representing a coarsening upward cycle (fig.2.15). Towards the top of this unit sandy limestones develop at the expense of shales, which are very micaceous and contain up to 20% silt size quartz, with a carbonate content as high as 35%. The sandy limestones contain up to 65% CaCO_3 and 25% silt; this pattern repeats itself throughout the sequence. The thickness of the shale units at the bottom is up to 4 m. but the total thickness of each decreases towards the top of this Member and a series of ledges, each not more than 80 cm. higher than the previous one are seen. The frequency of the coarsening upward cycles increases towards the top of the sequence.

The fauna is varied consisting mainly of Piarorhynchia juvenis, Pinna hartmani and Gryphaea; the trace fossils Diplocraterion, vertically regressive Rhizocorallium and Thalassinoides are also present. Thalassinoides and Diplocraterion almost invariably occur at the top while Rhizocorallium is ubiquitous to the base of the cycles.

In the lower parts, beds entirely composed of Gryphaea shells sometimes constitute the limestones but towards the top of the cycles they diminish. The trace fossil Kulindrichnus langi (Hallam, 1960a) is also seen with replacive phosphatic material.

Age: Arnioceras aff. semicostatum; Coroniceras sp. and Coroniceras reynesi Hallam (1959); Oates (1976) suggest a Reynesi Subzone age, but the occurrence of Schlotheimia and Charmasseiceras in the Ob Breakish succession of the same lithology suggests a Bucklandi Subzone age.

I-b-ii. Breakish Ironstone Member syn. Ardnish Ironstone bed (Hallam, 1959)

Type locality: Ob Breakish (NG 684246) 0.6 m.

Brown, reddish weathering, oolitic, nodular ironstone bed. The various aspects of ironstone development are discussed elsewhere. The fauna consists of Piarorhynchia juvenis and bivalves including Chlamys (?) calva. It can also be traced in Loch Eishort (at the 58 m. level) and in Raasay

(at the 60 m. level) where only washed-out oolites can be found. (See Fig.2.15).

Age: Bucklandi Zone (Reynesi Subzone).

I-b-iii. Upper Teampull Chaon Member

42 m.

Type section: Loch Eishort, West of the site of Teampull Chaon.

This Member comprises marls, shales and sandy limestones; the marl beds are never more than a few centimetres thick, while the shales are up to 60 cm. and the limestones at least 35 cm. thick. The shales are very micaceous and in general all three lithologies are very sandy. The three together constitute groups, the frequency of which increases towards the top of the Member. (Fig. 2.15).

This grouping of the three lithofacies constitutes coarsening upward cycles, the existence of which is indicated by the upward increase in silt together with changes in the trace fossil content. Some of the top beds (sandy limestones) are wholly made of broken shell material. The intensity of the burrowing of these beds increases towards their tops and they frequently show digested remains of various ?crustaceans. Kulindrichnus langi is present and phosphatic replacement of the shells and other remains is very common. The clay mineral content is comparable to that of the Lower Teampull Chaon Member since it contains traces of chlorite, no smectites, a high proportion of illite and very low amounts of kaolinite. The CaCO_3 content increases from the marls to the shales and into the micaceous sandstones. Possible traces of Zoophycos are found in Raasay.

Age: ?Lower Bucklandi-Semicostatum Zone.

I-b-iv. Dun Boreraig^{Sandstone} Member syn. Hallaig Sandstone, Lee & Buckman (1920) 35-50 m.

Type section: Below Dun Boreraig, Loch Eishort (NM 625163).

The succession consists of varicoloured calcareous sandstones, reddish, medium bedded at the base, medium-fine grained, which are well sorted, and contain bioclasts in places. A greenish massive unit on top with bed

thicknesses up to 1 m. is seen containing glauconitic grains and ferruginous nodules, some with oxidation rings. This lower unit is succeeded by reddish lenses of calcareous sandstones with beds of pebbles and coquina. The uppermost unit comprises fine grained yellow calcareous sands with colour banding. Coalified material are common and the beds show a characteristic sugary weathering surface.

This unit is parallel cross laminated at the base and planar cross laminated in the middle part. The greenish unit however shows only faint indications of cross laminae and so do the reddish lobes; at various horizons the latter show beds full of broken, reworked, silicified shell fragments. The top of the unit is characterised by abundant trough and large scale festoon cross lamination. (see Fig. 2.16).

These beds show a number of ?erosion surfaces and the trace fossils include Chondrites, Thalassinoides and horizontal Rhizocorallium, together with indistinct vertical and horizontal burrows. The topmost unit of this Member exhibits coarsening upward cycles which consist of mottled limestone (linsen and flaser structures) in the lower part with a moderate clay content. The clastic content increases in proportion, with the maximum grain size increasing also towards the top which is marked by bands and surfaces exhibiting ferruginous oxidation.

The fauna consists of bivalves and gastropods while Spiriferina walcotti is particularly common together with various oyster beds.

Age: The presence of Caenites brooki, Microderoceras and Euasteroceras establish the presence of the Brooki Subzone. The unconformity here involves the whole of the Birchi and Obtusum Subzones.

II. Broadford Beds Argillaceous Group syn. Broadford Beds (Woodward, 1896)

Type section: Mull and Morvern.

II-a. Loch Aline Formation

Type section: Loch Aline (NM 694453)

II-a-i. Wilderness Shale Member Blue Lias type 1 of Oates (1976) 10.50 m.

Type section: Wilderness (NR 404287)

The lithology is alternating dark shales (25 - 50 cm.) and yellowish-brown weathering limestones 20 - 30 cm. thick. Pyrite is present, replacing most features. Smectite is abundant and the CaCO_3 content of the shales is very low (5-10%); the limestones, on the other hand, have up to 75% CaCO_3 .

No sedimentary structures are seen, except for a collapse structure in bed 3 (Fig. 2.17).

The trace fossils are Ophiomorpha-like features which have their tubes infilled with pyrite together with Rhizocorallium which is abundant in the lowermost marly bed; Liostrea hisingeri is also present in the mentioned beds.

Age: Planorbis Subzone of the Planorbis Zone.

II-a-ii. Leacach Nodular Member (new name) Blue Lias type 2 at lower parts and type 3 at top (Oates, 1976)

Type section: Allt Leacach, Loch Aline (NM 694453) 28.37 m.

Nodular limestones alternating with shales are seen; the thickness of the limestone units, which are invariably packed with thick-shelled bivalves (in the upper 75 cms.), increases towards the top of the unit while in the lower parts the shale units are up to 35 cm. thick and contain various shells, mainly Gryphaea (Fig. 2.18).

The silt content is minimal and these beds are readily comparable with the 'Blue Lias' of Dorset. At the base of this Member, escape structures are seen together with coalified plant remains. The fauna consists of abundant Pinna sp., Cardinia ovalis and less abundant Liostrea sp., Pholadomya sp. and Modiolus sp.

Age: The lowermost 12.28 m.: Angulata Subzone	) <u>Sulciferites</u>) <u>Schlotheimia</u>) <u>Vermiceras</u>
The following 7 m.: Conybeari Subzone	) <u>Vermiceras conybeari</u>
The following 3.87 m.: Rotiforme Subzone	) <u>Coroniceras rotiforme</u>) <u>C. bridion</u>
The following 4.29 m.: Bucklandi Subzone	) <u>Arnioceras</u>

II-a-iii. Leacach Ochrous Member (new name) Top of Blue Lias type 3 of
Oates (1976) ~0.50 m.

Type section: Allt Leacach, Loch Aline (NM 694453)

These are mostly limestones with ochrous weathering and uneven surfaces; regarded as a significant marker of an erosion period (Fig. 2.18).

Age: Top Bucklandi, ?lowermost Semicostatum Zone.

II-b. Leacach Formation

Type section: Allt Leacach (NM 694453) 88 m.

II-b-i. Ardtornish House Member discovered by Oates (1976) variable thickness, up to 1 m.

Type section: Ardtornish House (NM 686435)

Lithology: Limestone and silty shales, occasionally overlain by a small thickness of crinoidal biosparite, with a sparse fauna of the Bucklandi Zone.

Age: Reynesi Subzone (Arnioceras sp.), however the shales beneath this bed are of a different age in different localities therefore a minor period of erosion is inferred (Oates, 1976).

II-b-ii. Leacach Shale Member 4-7 m.

Type section: Allt Leacach (NM 694453)

Occurs immediately overlying the Leacach Ochrous Member as black, silty, indurated, fissile and moderately micaceous shales with bands of nodular limestone up to 2 cm. thick. These nodular bands are laterally very

persistent and become thin towards the top; they weather with a reddish colour. The silt content is minimal, but the lower metre is somewhat silty.

The shales are of very low lime content and contain approximately 3% smectite, 40% illite and 25% kaolinite. The fauna consists mainly of gastropods and brachiopods, with occasional crinoid stems (Fig. 2.19).

Age: Turneri Zone no later than Birchi Subzone (Oates, 1976); based on the occurrence of Caenisites sp., Promicroceras, Microderoceras and Caenisites turneri.

II-b-iii. Leacach Pebble Member 0.80 m.

Type section: Allt Leacach (NM 694453)

Abruptly overlying the previous Member is a bed of silty shale with numerous micritic and phosphatic, angular pebbles which are up to 6 cm. in diameter. They show similarities to the nodules that are present in the shales and do not show signs of extensive reworking.(Fig. 2.19).

Age: Oates (1976) maintains that the abrupt transition up into this bed is probably a non sequence; this view is accepted here and the bed is placed at the base of the Obtusum Zone.

2.5 Local correlation

2.5.1 Milton Formation

The Milton Formation and its representative Members are found in all of the principal outcrops of the northern area. It can be seen that in Figure 2.20, this Formation progressively thickens from the southwest to northeast, the sections at Raasay and Applecross being those with the maximum thickness.

Only a rough correlation may be established between the individual members, due to the lack of stratigraphic marker horizons. However, the basal and topmost conglomerates, together with the two coral beds are of invaluable help in this region.

i. TorrMor Member: This member and the overlying cross bedded calcareous pebbly beds are wholly or partly present in all sections except in that of Raasay. The total thickness increases towards Applecross while the pebble content and size decreases. Cross bedding is only observed in the sandstone beds of TorrMor and is faintly present in the Ob Breakish section. Coalified plant fragments are present in all localities and are best preserved in Applecross. The shale beds which alternate with the cross bedded sandstones in TorrMor are not seen in Ob Lusa, but in Applecross they are thicker and the sandstone interbeddings lose their cross lamination. In Applecross towards the top of this unit, beds of sandstone which were so well developed in TorrMor (and to some extent in Breakish) are replaced by oolitic calcarenite. The total thickness of this Member changes from 1.5 - 3 m. in TorrMor to 1.20 m. in Breakish and 1.6 m. in Applecross.

ii. Lusa Coral Mem.: This Member is particularly well developed in Skye and

Applecross, but it is hard to estimate the variation of its thickness mainly perhaps because of its original patchy development. In Loch Eishort a particularly well developed oolitic, skeletal calcilutite is seen at the same stratigraphic level, but there is no indication for the presence of this coral bed. It occurs however, in various localities in central Strath (NG 64852090; NG 64612010).

iii. Lower Sand Member: Not only does the thickness of this unit increase from southern Strath to Applecross, its composition varies as well. The gap observed in the section at Applecross may or may not represent strata of the same type. In either case, the thickness of this Member is greater at Applecross. In the Loch Eishort section it is represented by a ripple cross laminated unit which grades up into a distinctly cross bedded, white, very fine siltstone. However, its equivalent at Ob Lusa is a brownish parallel laminated, thin bedded muddy calcarenite with abundant coalified plant remains and ?rill marks. Laterally these grade into nodular oolitic, sandy calcarenites with shale partings at Applecross, where their sand content increases towards the top of the unit.

iv. Breakish Coral Member: Well developed at Applecross and Breakish, this Member is particularly valuable in this northern region since it represents the first member of the Applecross Formation which can be used as a relatively accurate local marker horizon. As previously indicated, it may be placed at about the ?Angulata-Bucklandi Zonal boundary. However, its use here is restricted to very local correlations within the northern area.

v. Upper Sand and Breugh Pebble Members: The Upper Sand and the Breugh Members, while representing the Conybeari and Rotiforme Subzones of the Bucklandi Zone respectively, show an overall, gradual increase in thickness from the northeast to the southwest (i.e. 12 – 13 – 12.25 m.), but a decrease in the Breugh Pebble Member's thickness to zero may be followed along the same trend (i.e. 1.8 – 2 – 0.5 – 0 m.).

It is hard to establish confident correlation horizons among the members

of the Milton Formation present in Raasay and those of the other localities.

The presence of Liostrea hisingeri suggests the existence of the Hettangian stage and the lowermost 10 metres of silty limestone with shale shreds together with limestone and shales may be equivalent to the TorrMor, Applecross House and the Lusa Coral Members present elsewhere. The succeeding Members of the Milton Formation are represented here by a 15 metre thickness of nodular limestone which is only interrupted by 2 m. of oolitic limestone.

Any correlation would be extremely tentative and approximate. The Milton Formation decreases in thickness from Applecross (64 m.) to Loch Slapin (0 m.) (Fig. 2.20).

2.5.2 Strath Formation

i. Teampull Chaon Member: As represented in Figure 2.21 a good bio- and lithostratigraphic correlation can be established within this Formation. The Breakish Ironstone is particularly well developed in the Ardnish peninsula and its equivalent beds may also be traced in Raasay and Loch Eishort, where however, these are not as well represented. Also horizons rich in phosphatic nodules occur at roughly similar intervals above the ironstone. The total thickness increases from the west (Loch Slapin, 8 m.) to the east (44 m.), although there are no representatives of this Formation to be found in Applecross (possibly due to erosion), the thickness changes from south (Loch Slapin and Loch Eishort) to north (Raasay, 53 m.) with a maximum thickness in Breakish (55 m.). The top Bucklandi Subzone is represented in the lower 8 metres of the Breakish section. It is not represented in any other section and the Reynesi Subzone predominates in all others. In Breakish, however, the thickness of the Reynesi Subzone beds varies from place to place. Within the northern region, in Breakish it is 41 m., in Loch Eishort 44 m. and in Raasay it attains the thickness of 48 m.

In Loch Slapin the Reynesi Subzone is represented by only 1 m. of silty

shales. The succeeding two subzones are not represented in Loch Eishort, but being present in other sections, the thickness varies from 5 m. in Raasay to 5.5 m. in Breakish and 6.5 in Loch Slapin; a possible overall increase in thickness from the north to south of the region can be invoked if the 9 m. gap above the succession in Loch Eishort is believed to represent (at least in part) the Reynesi Subzone of the Semicostatum Zone.

ii. Dun Borerraig Member: Although well developed in the southern region, this member is poorly represented in localities other than Loch Eishort in southern Strath and Raasay to its north. It is considered to represent the Brooki Subzone of the Turneri Zone, the top of which is marked by a disconformity involving possibly the Birchi and Obtusum Subzones. The unit is 57 m. thick in Raasay and 48 m. thick in Loch Eishort. The basal 9 metres of silty, bedded, micaceous shales may be correlated with the lower 5 m. of thin bedded fine grained sandstones of the basal Turerni in Loch Eishort and the 2.5 m. of northern Strath (Rubha nan Eirreannaich). It is difficult to correlate the succeeding 17.5 m. of even bedded, finely cross laminated greenish calcareous sands of Loch Eishort with any other succession in the immediate vicinity. The seven metres of bioturbated, yellow, calcareous sandstone lobes with coalified material in Raasay is represented by 9 metres of purplish, cross laminated, calcareous sandstone of the same type which are cut by at least two beds 30-35 cm. thick of coquina limestone full of pebbles. The succeeding 45 m. of coarsening upward cycles of silty carbonaceous shales and calcareous sandstone couplets in Raasay are represented by 14 m. of coarsening upward cycles of mottled, carbonaceous calcareous silts/festoon cross laminated calcareous sandstone components in Loch Eishort (Figure 2.22).

2.5.3 The Southern Region

Due to the lack of proper exposure, it is difficult to establish a satisfactory local correlation for the various members of the Broadford Beds

Argillaceous Group in the Inner Hebrides.

2.5.4 Loch Aline Formation

- i. Wilderness Member: Exposed along the shore of Aird na h-Iolaire (NM 405285), this Member may be correlated with the 4.50 m. succession of ?Planorbis Zone age directly overlying the Triassic conglomerate at Mingary Pier (NM 479636). As the age of the latter succession is based on the identification of Ostrea sp., Mytilus sp. and Cypricardia porrecta (Richey, 1930), none of which are fauna of index value, this correlation is at the best tentative (see Figure 2.23).
- ii. Minor Exposures: Within the Ardnamurchan area, many isolated exposures of the Lower Lias exist; these are sparsely fossiliferous and lack definite zonal index fossils, however, on the basis of lithological and general faunal similarity to other Liassic exposures in the immediate neighbourhood, they are regarded as representing the Loch Aline Formation. As it will be shown later, these minor exposures are of true importance to a palaeoenvironmental interpretation.

In the Loch Mudle locality (NM 546660), sandstones containing coalified wood fragments, representing the Loch Aline Formation, directly overlie Moinean schists. This succession also contains a 70 cm. limestone bed which shows impressions of ?Isastrea which, if present, might represent the Lusa Coral Member of the Broadford Beds Arenaceous Group (Milton Formation) in Ardnamurchan. The total measured thickness of this succession is 8 metres (see Chapter 7).

In the Ben Hiant area, Lower Liassic rocks of the Loch Aline Formation partly overlie Moine schists and partly Triassic conglomerates. At the 11.50 m. level a limestone bed containing impressions of the coral ?Thecosmilia martini were found. The pebbly sandstone beds of the 16-17 m. horizons provide good means for correlation with the pebble beds found in

the succession at Garbh Rubha Bay. At Garbh Rubha (east), Port an Eilean and Ockle Point, the representatives of the Loch Aline Formation directly overlie Triassic conglomerates. Fauna representing a definite age is lacking from these rocks and Pleuromya, Modiolus and Liostrea are the main faunal elements.

At the western edge of Garbh Rubha (NM 539710) a 5.50 m. thickness of cross bedded calcareous sandstones containing large pebbles, coalified wood remains and strongly abraded bivalve shells cannot be successfully fitted in with the other Lower Liassic exposures in the area, and based on the lithology alone the suggestion of Oates (1976) that they "pre date higher parts of the Semicostatum Zone" in this area cannot be supported.

As shown on Figure 2.19, the Leacach Ochrous and Leacach Nodular Members of the Loch Aline Formation are well represented in Loch Aline, and also on a very local basis in the various streams in Loch Aline, to the extent that minor thickness variations may be recorded. This pattern does not exist in the Lower Lias succession of Craignure Bay (NM 730360), where it is not possible to define the Members of the Loch Aline Formation which are so well represented only 8 kms. to the northwest of this area.

In general it can be seen that the Loch Aline Formation attains its maximum thickness (28 m.) in the type area whereas in Mull it attains a total thickness of 10 m. In Ardnamurchan it is represented by 42 m. of strata.

The Leacach Nodular and Leacach Ochrous Members of this Formation can only be found in the Loch Aline district and local minor thickness variations may be established among the exposures of Allt Leacach (NM 693459), Allt Smahnachain. (NM 694460), Larachbeg (694486), Loch Aline Lime Kiln (NM 693973), Aoineadh Achad Rainich (NM 700466) and Allt Buaile (NM 693455). The total thickness of each section is uncertain due to vegetation cover and faulting. Based on the occurrence of Coroniceras kridion Oates (1976, Fig.64) has shown that the total thickness of the Leacach Nodular Member below the C. kridion

horizon is greatest at Allt Leacach (5.44 m.) and least at Allt Buaille (5.08 m.). The thickness of the strata between the Loch Aline Ochrous Member and the C. kridion horizon also varies, with a maximum development at Allt na Smahnachain (10.77 m.) and a minimum developed at Larachbeg (3.19 m.).

Identification and correlation of Plagiostoma and Pinna-rich horizons together with beds of 'ironshot' limestones is less reliable but nevertheless, are used by Oates (1976) to demonstrate a progressive thinning of laterally equivalent beds from 1.7 m. at Larachbeg to 10 cm. in Allt Leacach.

2.5.5 Leacach Formation

The total measured thickness of this Formation varies from 8.8 m. in Morvern to approximately 38.50 m. in Craignure Bay. The Ardtornish House Member together with its topmost "crinoidal biosparite" bed is of variable age and thickness in Morvern, furthermore it is not possible to demonstrate the true thickness of this Member and its local variation.

The thickness of the Leacach Shale Member and the overlying Leacach Pebble Member changes from 8.40 m. in Allt Leacach to 5.30 m. in Allt Mor. The correlations established among these Members in Loch Aline and their equivalents of Craignure Bay are extremely tentative (Figure 2.23).

It is significant that the Birchi Subzone is absent from the Strath Formation in the northern region (the whole of the Dun Boreraig Member is of Brooki Subzone age), whereas it forms almost the whole of the Leacach Formation.

During this time a progressive offlap from the northern to the southern region can be recognised in Figure 2.23.

i. Obtusum Sandy Shales: In Figure 2.23, the Obtusum shales and sandy shales are seen to overlies the Broadford Beds Arenaceous Group in Loch Eishort and the Broadford Beds Argillaceous Group in Mull and Morvern.

The north-south thickness variation of these beds is remarkable, the

whole of all three Subzones of the Obtusum Zone are regarded as being present in the northern region whereas the 9.50 m. succession of Loch Eishort represents the Denotatus and Stellare Subzones, the Obtusum Subzone is omitted due to a minor disconformity. The maximum thickness reached by these beds is 50.40 m. in Loch Aline (Allt Mor NM 700435), where its thickness varies considerably on a local basis; it is represented by 21 m. of sandy bioturbated shale in the Allt Leacach section (NM 694453). The relation between the already described Members and the different facies (described in later chapters) is shown in Fig. 2.24.

In Mull the Obtusum Zone sandy shales are exposed on the eastern flank of the Craignure anticline (NM 730360) and according to Oates (1976) on the eastern shore of Duart Bay northwest of Torosay Castle (NM 722359).

The sequence above the Turneri Zone sandy shales, as exposed on the flank of the Craignure anticline, is 20 m. The thickness of these beds cannot be determined with much certainty due to Tertiary disturbances and the presence of a stratigraphically long ranging ammonite fauna. In Morvern the Obtusum strata are ca25 m. thick.

Particular mention should be made of a sequence of sandstones exposed on the shore of Duart Bay which stratigraphically overlies the Obtusum sandy shales but are separated from them by a craignurite sill. This sequence comprises a lower faintly cross laminated unit with an interference-rippled top surface showing Gyrochorte-like trace fossil markings (Plate 2.3 and 2.4). A 6 m. thick ripple cross laminated calcareous sandstone unit with very large calcareous doggers overlies this unit; Lee and Bailey (1925) considered them to be equivalents of the Scalpa Sandstone beds, but they have been considered to be of late Obtusum to early Raricostatum Zone age by Oates (1976) for the following reasons:

- a. The immediately underlying shales are probably of Obtusum Zone age.
- b. Large sandy doggers are seen in the top of the sandy shales of Morvern.
- c. None of the lowest Scalpay Sandstone sequences seen in Abhainn Lirein (NM 718326), Port Donain, or Carsaig Bay resemble that seen at Torosay.
- d. Beneath the Raricostatum Zone shales of Carsaig Bay at least 2 m. of

ripple marked sandstones with large calcareous doggers have been recorded and almost certainly correlate with a similar unit in the An Garradh section.

- e. Jamesoni Zone shales of normal (but baked) lithology were revealed in the slopes of Maol nam Uan in 1975 only 1 km. to the west.

The above reasons cannot support the suggestion that the sandstone sequence is of ?Late Obtusum Zone age, the following reasons are given by the present author:

- a. There is no fossil evidence for establishing even a tentative age.
- b. Faulting is very common in this locality.
- c. The sandy calcareous doggers of Morvern are in no way like those of Torosay, they are not cross bedded and lie within silty shales.
- d. The Scalpa Sandstones of Port Donain do not resemble these due to the fact that although they are of the same age, the Port Donain succession represents a lagoonal facies while the Torosay beds represent a tidal inlet delta facies (McCallum, 1971).

In addition to the above reasons, it would be logical to expect a thinner development of the Obtusum Zone in the Mull area due to the palaeotopography of this region which will be further discussed in later chapters.

CHAPTER 3

SEDIMENTOLOGY

Description and field occurrence of rock types

3.1 Limestones

For the purpose of nomenclature and classification of carbonate rocks, a number of effective and meaningful schemes have been proposed (Folk, 1959, 1962b; Dunham, 1962; Embry and Klovan, 1971). For this study^a terminology has been sought which explains both particle type and depositional or diagenetic texture; thus compound abbreviations are constructed which combine grain composition with textural nouns (cement type) (Folk, 1959). This name is then followed by one of Dunham's (1962) appropriate terms to depict the textural arrangement (packing). The minimum number of categories sufficient to depict the major differences are used in this classificatory system and although the nomenclature is undoubtedly cumbersome, its usage is advantageous in that it relatively accurately delineates the microscopic characters of the rocks.

For example, silty oolitic microsparite wackestone has the advantage of depicting:

- i. Size of the quartz grains.
- ii. The presence of bioclasts.
- iii. The type of calcareous grains present (in order of abundance).
- iv. Type of carbonate cement.
- v. Packing arrangement of the rock constituents.

In this study the following definitions are followed in describing the constituent particles.

3.1.1. Terrigenous constituents (Folk, 1959, 1962b)

These include all particles which are formed at sources other than their basin of deposition (e.g. quartz, silt, sand, clay minerals, feldspars and

heavy minerals).

3.1.2. Allochems (Folk, 1959)

Refer to grains formed within the basin of deposition which have been little transported.

i. Intraclasts (lithoclasts of Folk, 1959)

These are somewhat large clasts which are formed by dessication, breakage or burrow disruption of penecontemporaneously deposited carbonate sediments; those lithoclasts which are derived from older lithified rocks (outside the basin of deposition) will be indicated separately.

ii. Ooliths

These are spherical, multiple coated particles; the coatings show either radial or concentric structures.

Proper ooliths

These are defined as a mineral or organic grain surrounded by at least two concentric layers (Carozzi, 1960); the nucleus is enveloped by calcite or aragonite with a more or less finely concentric or radial texture.

Multiple ooliths

These are defined as ooliths which are cemented together and surrounded by a common envelope, building composite individuals of variable shape and size (Cayeux, 1935).

Superficial ooliths

Grains with only one layer of coating (radial or concentric) are referred to as superficial ooliths in this study.

Pseudo-ooliths

See section 3.1.2.iv on pellets and peloids.

Asymmetric ooliths

These are characteristic of less turbulent environments and show

protruding nuclei and oolitic rings which abut against them; they are poorly sorted and have low to moderate sphericity (Freeman, 1962; Davis, 1966).

Pisoliths

These are concentrically laminated particles of CaCO_3 ranging from 1 - 10 mm. in diameter (Blatt et al., 1972). Their origin is considered to be similar to that of algal oncoliths (Dunham, 1969b). Pisoliths in the Applecross Formation do not show inverse grading or polygonal boundaries of the grains, therefore are not considered to be of diagenetic origin.

Sediment particle inclusions are seen within the pisoliths which are attributed to adhesion resulting from the rolling of ?algal mud on a silty floor. This algal mud may or may not form as a coating around a nucleus, depending upon the availability of detrital particles and the amount of winnowing activity.

iii. Bioclasts

Fragments of tests and shells of fossils, for the identification of which standard works (Johnson, 1951; Horowitz and Potter, 1971; Hudson, 1962, 1968) are used, will be referred to as bioclasts.

iv. Pellets and peloids

In this study pellets refer to rounded, spherical to elliptical or ovoid aggregates of microcrystalline calcite (micrite) which show no internal structures (Wilson, 1975). Peloid refers to all grains that are composed of an aggregate of cryptocrystalline carbonate irrespective of their origin (McKee and Gutschick, 1969).

a. Faecal pellets

These are rounded, spherical, elliptical or ovoid elongate bodies of microcrystalline calcite (micrite) without any internal structure; they range in size from 0.03 mm. - 0.15 mm. (most common sizes are in the range of 0.04 - 0.08 mm. (Hatch, Rastall and Greensmith, 1973)). These are also

reported from present day environments; fresh faecal pellets contain an ill-sorted aggregate of fine particles of various kinds (Illing, 1954) which change with time to a uniform micrite (Purdy, 1963a). It should be noted that fossil faecal pellets must have been rigid bodies because their shapes were maintained after burial (Bathurst, 1971), contrary to terrigenous mud faecal pellets (micritic pellets, peloids) such as those described by Moore (1939).

b. Micritic pellets of unknown origin

These are roughly sphaeroidal or irregular peloids (Bathurst, 1971); they are referred to as "Pseudo-ooliths" by Cayeux (1935). The term oolith carries certain connotations regarding the structure, morphology and mode of origin of grains none of which are seen in what Cayeux (1935) called "fausse oolithes". The term peloid refers to grains which may be micrite-replaced skeletal particles (Purdy, 1963a; Bathurst, 1966); the pelsparites in which peloids merge and produce a flocculent structure (Beales, 1956) are common in the Milton Formation. These are what Cayeux (1935) termed "structure grumeleuse"; such coalescing peloids show patches of spar among the rocks.

v. Grapestones

Grapestones (Illing, 1954) or aggregate lumps are cemented clusters of peloids and ooids which may be coated (Wilson, 1975). Usually limited numbers of grains become firmly cemented together (Taylor and Illing, 1969).

3.1.3. Orthochems (Folk, 1959)

Orthochems are generally considered to encompass all essentially normal precipitates which form within the rock itself, i.e. cements (both diagenetic and primary) and diagenetically formed particles. Two different types of cement are recognised within the rocks:

i. Micrite (microcrystalline ooze of Folk, 1959)

Carbonate muds are inorganic precipitates or products of organic

breakdown formed of grains 1 - 4 μm in size. In modern sediments most carbonate mud is composed of individual crystals of aragonite (average dimensions 3 μm length and 5 μm width). It is presumed that the aragonite needles which form the carbonate muds of today represent the direct analogue of the microcrystalline components which constitute the cementing material of ancient carbonate rocks (Blatt et al., 1972; Bathurst, 1971).

ii. Sparry calcite cement

These are cements (or neomorphic spar) with calcite grains 10 μm in diameter, they include "microspar" and "pseudospar" (Folk, 1959, 1962b, 1965). Microspar is referred here to a neomorphic crystal mosaic with crystal diameters of 4 - 10 μm .

It should be noted that differentiation between the three various sizes of orthochem cements is very subjective in borderline cases, as pointed out by Folk (1959); in such cases the relative clarity is used as a distinguishing feature.

3.1.4. Micritic limestones

Pure micritic limestones are very rare in the two Groups of the Broadford Beds. Biomicrites and silty biomicrites are found in the lowermost beds (?Rhaetic) of the two areas of study. In the southern area it is difficult to distinguish among the micritic limestones and those with microsparite cement. The limestones of the northern region mainly consist of sparitic cement, while those of the southern region mainly consist of microsparitic cement; therefore the occurrence of micritic limestones are found but not confined to the Ardnamurchan and Mull areas.

3.1.5. Lithoclastic, bioclastic and oolitic limestones

Sparitic and microsparitic limestones frequently occur throughout the various Members of the Milton Formation. Coarse grained (sparitic) limestones are less common and confined to purely oolitic beds together with

cavity fillings and body replacements of the coralliferous beds. The oolites are most common in the Milton Formation and although they also reportedly occur in a limestone bed of the Leacach Formation (Oates, 1976) in Loch Aline, they are absent from the southern area of study. The various types of oolite are present and respective environmental conclusions will be suggested in the text. Oolith sizes range up to 0.4 mm., with 0.25 - 0.35 mm. average size. The oolith nuclei consist mostly of detrital quartz but shell fragments also commonly form their centres. Large (1 - 1.5 mm.) rounded objects with faint oolitic (concentric) structures, showing possible algal remains are referred to as pisoliths. These occur only in Applecross and in the corresponding oolites of the Milton Formation in Raasay. Bioclastic limestones (Lumachelle, shell hash (Flügel, 1972) or Coquina) are seen at various horizons in the limestones of the two Groups of the Broadford Beds.

They are exclusively composed of worn, abraded, bivalve shells in the Milton Formation whereas the coquina of the Strath Formation and those occurring in the southern area of study, are mostly composed of echinoid and brachiopod debris.

Lag grainstones (Wilson, 1975) composed of mixed resistant particles of blackened, phosphatized and iron-stained lithoclasts and peloids together with worn, resistant, echinoid and bivalve shell fragments are seen at various horizons within the beds of Milton and Strath Formations.

The Dun Boreraig Sandstone Member (Turneri Zone) shows pebble-lag beds consisting of haematite and crushed, banded, crenulated diagenetic ?chalcedony pebbles.

Biomicrosparites containing totally altered shell material showing no obvious signs of extensive abrasion, floating in a calcite cement, are common in the Strath and Loch Aline Formation. The beds of the Strath Formation are very silty whereas their silt content diminishes in the Mull and Loch Aline areas.

Micritic subrounded to rounded pelletoids are ubiquitous in the limestones

of the Broadford Beds Groups; these pelletoids are up to 2 mm. in diameter and frequently show an oolitic rim. Algal borings on the surfaces are also seen mainly in the Milton Formation.

These pelletoids could also be micritized debris of calcareous organisms as described by Purdy (1968).

3.1.6. . Microsparites

The term 'Microsparite' was first used by Folk (1965) and is used here for describing calcite crystals with diameters ranging from 4 - 10 μm . or even up to 30 μm . These normally form crystal mosaics in the limestone beds of the Strath and Loch Aline Formations.

Silty biomicrosparites are ubiquitously found in the Lower Sinemurian shales of the Broadford Beds Arenaceous Group. They occur as broad undulose, nodule-like beds 15 - 25 cm. thick, but ranging up to 45 cm.; determination of the insoluble residue in 21 samples taken from the Semicostatum Zone shales and limestones of the Strath region shows that some limestones contain between 65 and 74% CaCO_3 whereas the shales contain 19 to 44% CaCO_3 . In the Mull area limestone/shale alternations are seen to persist throughout the period of Broadford Beds Argillaceous Group deposition.

The limestones of the Wilderness Member are considered as argillaceous biomicrosparites in this section and wet chemical analysis shows a CaCO_3 content of 48 - 54% whereas the shales contain as little as 3% CaCO_3 in places. The Leacach nodular Member consists of silty biopelmicrosparite beds with up to 80 - 85% CaCO_3 ; it is difficult to trace the exact amount of clay present in these rocks since it is obscured by the fine crystal mosaic of the calcite; coarse calcite forms fine ("drusy") rims around originally aragonitic molluscan shells and also occurs as coarse (blocky) replacement of shells and as patchy developments throughout the rocks. In these beds (together with those of the Strath Formation) shelly material (molluscs,

brachiopods, echinoderms) is usually inhomogeneously distributed and in the microsparitic beds of this Member, grains of glauconite are found. In the Leacach Formation, the entire thickness of the Leacach Shale Member is composed of micaceous calcareous shale with 20 - 30% CaCO_3 content, alternating with 1 - 3 cm. thick beds of laterally contiguous (lenticular in places) limestones with 85 - 90% CaCO_3 ; these beds show no evidence of bioturbation and are devoid of fossils. In both northern and southern areas, pyrite is present in small quantities forming:

- a. Finely disseminated particles.
- b. Framboidal pyrite clusters.
- c. Well defined cubic crystals.
- d. Patches of non distinct form.

Bituminous matter is seen occasionally and is more common in the shale beds. The existence of these minor rhythms or cycles of limestone and marl in the studied beds raises the question of their primary sedimentary or secondary diagenetic origin. This problem has been amply reviewed and discussed by Hallam (1960b, 1964, 1971c); the application of his arguments to the strata of the Broadford Beds Argillaceous and Arenaceous Groups are presented later .

3.2 Sandstones

Sandstone nomenclature and classification has been subject to disagreement and debate for decades. Krynine (1948) adopted the method of classification based on plotting the mineral composition of sand size grains in a "sandstone" within an equilateral triangle. The use of such "triangular" classifications in which the principal minerals are chosen to represent the poles together with an arbitrary internal subdivision of the triangles has amply been reviewed by Klein (1963) and Okada (1971).

Sandstones occur both in the Milton and the Strath Formations in the northern area of study. Calcareous siltstones also occur in the Leacach Formation of the southern region. In the Milton Formation, beds of cross laminated fine-medium sandstones occur, the lowermost calcareous sandstone bed contains abundant coalified plant material and consists of fine medium poorly sorted subrounded quartz grains, the cement is calcite spar and the grains consist of up to 10% polycrystalline quartz.

In southern Strath, white, well washed cross laminated siltstone beds with up to 88% quartz constituents occur. Sandstones with calcareous cement also succeed the Breakish Coral Member.

In the Strath Formation, the Dun Borerraig Sandstone Member comprises prominent, cross laminated, varicoloured sandstones which in many cases contain up to 75-80% quartz grains. These beds contain fine to medium grained moderately sorted, subrounded quartz grains. Bioclastic fragments in the form of fenestral echinoderm plates are common and the sands are not well sorted in places. In contrast to the sandstones of the Applecross House Member, the beds of the Dun Borerraig Sandstone Member are seen to overlies beds representing younger beds. In Raasay and northern Strath these sandstones transitionally overlies the shales of the Strath Formation, whereas in Loch Slapin and Loch Eishort (southern Strath), the Semicostatum/Turneri contact is faulted. Sandstones as defined above do not occur in the Mull area, and in Ardnamurchan their occurrence is confined to the lowermost Rhaeto-Liassic beds.

Chert rich sources are present in the area of study (Durness limestone) and the grouping of chert and quartz together places constraints on genetic interpretations; therefore, where present, chert is plotted and/or mentioned separately.

The presence and variability of the amount of polycrystalline quartz, number of crystals per grain and undulosity of the quartz grains in the various beds has called for the use of single diamond diagrams (Basu et al., 1975;

Young, 1976) to enable the identification of plutonic low and high rank metamorphic sources.

It should be noted that the somewhat different interpretation of polycrystalline quartz which was given by Voll (1960) enables the identification of two types of metamorphic origins for polycrystalline quartz.

i. Polygonized quartz -

Component crystals of the grains show straight and polygonal bodies which meet at 120^0 angles.

ii. Polycrystalline quartz -

Component crystals of the grains show sutured boundaries.

The above two varieties may be of importance in northwest Scotland where polygonized quartz characterises the Western Highlands and the sutured quartz is typical of the Moine thrust and much of the Dalradian.

Blatt and Christie (1963) demonstrated the broad overlap in character of undulosity of plutonic and metamorphic quartz; Blatt (1967a) pointed out that there are great differences between the proportion of polycrystalline quartz in the detritus derived from the source rocks and those of reworked sandstones, this discrepancy is also seen for undulatory quartz. Such differences were attributed to the mechanical instability of polycrystalline and strained quartz by the mentioned workers; the above criteria suggest that undulosity in quartz together with the determination of polycrystallinity are of little value in the identification of source rock and provenance for deposits older than those of the Recent, nevertheless such observations on the constituent grains of the strata, will provide meaningful results when used in conjunction with other geological data.

Blatt et al. (1972) maintained that the average number of crystal units in sand-size grains (in this study grain refers to a clastic fragment which may be composed of one or more crystals of quartz) varies depending on the source rock. Basu et al. (1975) demonstrated that the number of crystal units

in polycrystalline quartz is useful only in identifying contributions from low rank metamorphic sources. The conclusions of Basu et al. (1975) are valid in studying the medium sand fraction of "first-cycle", single-source, erosional products. The presence of a multicomponent source system, providing the Lower Liassic depocentres of northwest Scotland with both primary and "recycled", weathered detritus casts doubt on the reliability of the conclusions reached solely on evidence provided by the above mentioned technique.

The studies of grain size distribution by Mason and Folk (1958), Friedman (1961, 1967), Moiola and Weiser (1968), Visher (1969, 1973), Visher and Howard/Tillman (1971), Middleton (1973), Moiola et al. (1974) and Sagoe and Visher (1977) are thought to be of value for the interpretation of sedimentary processes in modern sedimentary environments, but the validity of such data is questionable in those ancient rocks which show evidence of diagenetic alteration (e.g. leaching of feldspars together with etch surfaces and overgrowths on quartz grains). Well cemented sandstones are difficult to disaggregate and therefore may not yield reliable grain size distribution curves due to the production of "fines" in the process of disaggregation (Blatt et al., 1972).

The major assumption made in the interpretation of grain size distribution plots, is that they reflect processes and/or physical character of a specific depositional environment; the grain size distribution curves obtained from Recent environments are then matched with those obtained from ancient rock samples (Visher, 1969), the inherent assumption is that there is a "typical" curve for each of the sedimentary environments leading to the conclusion that "if your wife does not precisely resemble Laura Antonelli, she is not a real woman!" (Folk, 1977); for relevant discussions see Freeman and Visher (1977), Picard (1977), Folk (1977), and Steidman (1977). Ruzyla (1977) has in addition, shown that size analysis studies on Holocene beach sands produce a strong contrast depending on where and when the samples were

taken.

The sandstones of the Lower Lias in northwest Scotland are:

- i. Well cemented with 5-30% calcareous and ferruginous cement.
- ii. Quartz grains are diagenetically altered.
- iii. Exposures are scattered and patchy, therefore the sampling procedure is mainly dependant on the size, surface-weathering, shape and location of the outcrop. Hence the precise relation of the samples to cross bedding, truncation surfaces, flat bedding, limestone lithosomes ... etc. cannot be given in all cases.

Due to the above (and previously mentioned) limitations, no attempt has been made at an exhaustive granulometric study of the Broadford Beds Groups sandstones.

Rock composition and grain size distribution data were obtained by point counting (>500 grains per thin section) following the Glagolev-Chayes method (Galehouse, 1971). It has been shown that sufficient correlation exists between various techniques of grain size analysis (Friedman, 1958; Kellerhals et al., 1975).

Due to the above mentioned uncertainties, the samples cannot be regarded as "true representatives" in absolute terms, therefore the statistical treatment of the data would prove to be meaningless. (See Appendix 1).

In this study the term orthoquartzite refers to sandstones with more than 75% quartz grains; the roundness scale of Powers (1953) is followed.

The heavy minerals zircon and epidote are present in the sandstone beds; they are both well rounded. The study of these accessory minerals (for summary see Pettijohn, Potter and Siever, 1973) would be of least value and their selective diagenetic elimination renders mere presence-absence data for these species uninformative; they constitute less than 0.1% of these rocks. A semiquantitative, nonstatistical approach is followed with emphasis mainly

given to the relative abundances of these two species of heavy minerals (Hudson, 1964; Blatt et al., 1972; Hallam, 1975).

Surface texture and diagenetic features of quartz grains in the sandstone beds were studied by scanning electron microscopy; certain types of surface features were identified and interpreted. The SEM photomicrographs are presented as Plates 3.1 to 3.6 together with their interpretations which follow the works of Krinsley and Donahue (1968), Krinsley and Doornkamp (1973), Pittman (1972), Baker (1976) and Marzolf (1976).

3.2.1 Cross bedding

The sandstones and sandy limestones of the Broadford Beds Arenaceous Group in the Skye and Ardnamurchan areas are cross bedded; some of the oolitic limestones are also cross laminated.

The lowermost beds of the Arenaceous Group are cross bedded in Torr Mor. The sandstones at Loch Eishort together with the sandstone/limestone beds which overlie the Thecosmilia Coral Member and in Applecross, the Breugh Pebble Member are cross laminated; cross bedded strata are also represented in Ardnamurchan. The Dun Borreraig Sandstone Member is also distinctly cross bedded; it can be seen that except for a few localities, the rest yield polymodal distributions in the form of simple, bimodal opposed patterns to complex nonsymmetric patterns. In such cases methods of analysis devised for unimodal vectoral data (Reiche, 1938; Curray, 1956) cannot therefore statistically be applied to the problem.

The bimodal, opposed patterns are analysed by a modification of the Tokey χ^2 test proposed by Middleton (1965; 1967). Complex polymodal distributions are analysed by a method developed by Tanner (1955). Apart from the one Member which shows unidirectional palaeocurrents (viz. the Breugh Pebble Member), the palaeocurrents obtained from the Broadford Beds Arenaceous Group should be considered in terms of the types of currents

operating in the modern shallow marine environment which are semipermanent, tidal and meteorological (Swift, Stanley and Curray, 1971). Such a consideration is consistent with the overall conditions of deposition of the Broadford Beds Groups. It should be noted that due to the variable nature of the currents in shallow marine environments each palaeocurrent direction reading may represent a unique affect, therefore the clustering of data in order to produce statistically meaningful results would most probably obliterate the imprints of the various minor processes which were operative, and thus the palaeocurrent data are also considered individually. Due to the faulted nature of most of the Jurassic outcrops and their possible tilting, all palaeocurrent data are corrected for post-depositional re-orientation following the method proposed by Potter and Pettijohn (1963). (See Appendix 2).

3.3 Conglomerates

This lithology is least represented in the Lower Liassic deposits of northwest Scotland. The beds in question are consolidated, poorly sorted deposits with mean sizes in the range of granules, pebbles and cobbles.

Conglomerates of the cobble size range occur at the base of the Milton Formation; these are matrix supported (calcareous), and thin out laterally forming thin beds of rounded pebble conglomerates. Elongate clasts are not common, so that clast imbrication is almost impossible to determine. These may be termed paraconglomerates (maximum clast size 9 cm.) (Blatt et al., 1972).

Pebble conglomerates form the topmost beds of the Milton Formation in the Skye and Ardnamurchan areas. These are clast-supported, fine upwards and contain some calcareous matrix. In both regions, clasts are rounded and imbrication is not determinable.

3.4 Argillaceous Beds

- i. Shale - In the present study this term is applied to those argillaceous sediments which show some fissility and are not plastic; it is qualified by "micaceous", "silty", or "calcareous" according to the composition of the subsidiary components.
- ii. Clay - This term is applied to unconsolidated, plastic, argillaceous sediments. Their carbonate content is low and they can be carved by a knife.
- iii. Mudstone - Refers to argillaceous sediments which are not plastic or fissile; these may be "micaceous", "silty" and/or "calcareous".

The different varieties of shale occur in the Strath and Leacach Formations; shales of a higher silt content which are very micaceous comprise the beds of the Strath Formation. The CaCO_3 content of the shales of the Strath Formation is comparatively higher than those of the Leacach Formation.

Shales of both areas contain a marine fauna and clays are only found in the Milton Formation in Applecross; they are interbedded with thin calcareous sandstones. Mudstones are common in the Loch Aline Formation in the southern area of study. Clay mineral analysis of the argillaceous beds shows smectites to be the main constituents of the mudstones and clays, whereas illite and mixed-layer clay minerals are dominant in the micaceous shales.

3.5 Ferruginous Beds

Iron-rich beds are only of local importance in the Lower Lias of north-west Scotland. They occur as thin horizons within the Strath and Leacach Formations and their study will hopefully provide a greater insight into the palaeoenvironmental conditions which prevailed during the formation and deposition of the beds of the Breakish Ironstone. The mineralogy of the "Breakish Ironstone" together with the "iron rich" and phosphatic beds of Craignure Bay (Leacach Formation) and Ardnamurchan are presented and discussed in Chapter

9.2; the presence of beds with appreciable amounts of iron in the Lower Liassic succession of Mull (Leacach Formation) were hitherto not recognised and in view of their lateral equivalence with the Breakish Ironstone Member, the different facies represented by the described Members will be considered together.

The ferruginous strata which occur in the lowermost beds of the two Groups of the Broadford Beds are confined to the Strath and Leacach Formations. The examination of these iron-rich beds both in hand specimen and thin section, reveals considerable heterogeneity in terms of textural and mineralogical detail. Therefore a total of 33 thin sections were prepared of numerous samples taken from the ironstones.

i. Classification and nomenclature

The classification of Welsh ironstones used by Pulfrey (1933) was based on their textural attributes, i.e. the presence and absence of ooliths or pisoliths; the mineral composition of these constituent grains was also taken into account. The present classification not only accounts for the mineralogical composition of the grains, it also takes into account the mineralogy of the rock as a whole.

Taylor's (1949) original system of nomenclature is modified here and following Weinberg (1974), the major mineralogical components of the rock matrix are given in adjectival form while those forming the ooliths are mentioned in substantive form.

A rock consisting of ooliths of chlorite and goethite set in a matrix of siderite would be described as a sideritic chlorite goethite oolite. The presence of different sizes of quartzitic material in the rock is indicated by the addition of adjectives such as silty or sandy; the presence of clay grade quartz is indicated by the term quartzose.

Spherules of different diameters are termed ooliths if they show characteristic internal concentric features while those without such features are termed pellets. Pisolith size grains (diameter > 2 mm.) (Twenhofel, 1950), are absent in these beds. As in the classification of limestones in the two Groups of the Broadford Beds, the relative abundance of the mineral and rock constituents ordains the arrangement in which the adjective or noun is given.

The terms chlorite and chloritic used in this study refer to iron silicates but this does not preclude the possibility that they may be septéchlorites. The mineralogy and precise nature of the closely related green, ferrous silicates present in various forms in these ferruginous beds creates a problem. Chamosite was named after the type locality before the techniques of X-ray diffraction were well established (Tschmak, 1891; Orcel, 1927; Hallimond, 1938; Orcel et al., 1949). Since then "chamosite" has been restricted to describing an iron-rich septechlorite (Deer et al., 1966) having basal X-ray reflections with a $7\text{\AA}$ periodicity. Chlorite minerals are chemically similar to chamosites but show $14\text{\AA}$ basal reflections. As the type mineral is now known not to be a "chamosite" (Weinberg, 1974), the green iron silicate minerals seen in the Strath and Leacach Formations are ascribed to the chlorite group throughout this study and the general term "chlorite" is used except where a more precise usage is specified.

In recent years the name chamosite has been used to describe different materials with neither chemical nor structural similarities, thus it is widely used to refer to a very poorly defined mineral. As no conclusive data were obtained from X-ray diffraction work carried out during the course of this work, it is not possible to state whether the several varieties of iron silicates of different shades of green seen in these rocks have chamositic or chloritic affinities; for this reason these iron silicates which are

chemically similar to the chamosites and chlorites described in the literature (Foster, 1962) are all referred to as chlorites; thus relating to a large group of minerals rather than to any one species in particular.

CHAPTER 4

CLAY MINERALOGY

Although the Lower Liassic strata of the Milton Formation in northwest Scotland were mostly developed in different carbonate and siliciclastic depositional environments, significant amounts of argillaceous sediments were laid down during the development of the Strath, Loch Aline and Leacach Formations. A systematic investigation of the type of clay minerals present in the argillaceous strata of the two Broadford Beds Groups was carried out in order to obtain further clues related to the depositional and/or diagenetic conditions responsible for their formation.

Standard X-ray diffraction techniques were employed, the details of which are given in appendices 3 and 4.

4.1 Previous Work

The clay mineralogy of Lower Liassic shales of England, Scotland and Wales has been investigated by various workers (see review by Perrin, 1971). In virtually all cases they are reported to be dominated by illite with subordinate kaolinites and chlorites are also present. Hallam (1960b) determined the clay mineral content of nine samples of insoluble residues obtained from different parts of the Blue Lias succession in Dorset and Glamorgan. All of the specimens contained between 90 and 100% illite and up to 10% kaolinite, four samples contained traces of vermiculite and only one showed faint traces of smectite.

Sellwood (1970) found that illite was the predominant clay mineral in the shales found in the Lossimouth borehole (80.4m to 19.8m depth). Lewis (1972) using whole-rock powder mounts, found the Liassic shales of the Mochras borehole to contain mainly illite

together with subordinate kaolinite and chlorites. Cosgrove and Salter (1966) show the development of appreciable quantities of mixed layer minerals and illite-smectites in the Lias of southwest England, but their results should be regarded with some scepticism due to uncertainties of analysis and quantification. In a subsequent work (Cosgrove, 1975), the White Lias was found to mark the incoming of kaolinite to establish an illite-kaolinite-chlorite suite which persisted in the Lias and into the higher beds. Substantial amounts of smectites were found in the Liassic shales of northwest Scotland together with illite, kaolinite, mixed-layer illite-smectites and subordinate chlorites by the present author (1977; see also appendix 4v).

4.2 Present Work and Results

A total of 132 specimens were obtained at approximately 1m intervals in all major sections of both Broadford Beds Arenaceous and Argillaceous Groups, with additional samples being taken to determine local variations of lithology. Diffractograms obtained from the samples showed that in the northern area of study, shales of the Strath Formation contain only illite and kaolinite with some chlorite minerals; in contrast 26 analysed samples from the Milton Formation contain in addition to illite, kaolinite and mixed-layer illite-smectites, smectites in substantial quantities of up to 39% (see Table 4.1). Although the beds of the Milton Formation in Raasay are not represented in Figure 4.1, they also contain up to 26% smectites.

In the southern area of study, 82 Liassic samples were analysed; 31 samples from the Loch Aline Formation contain up to 39% smectites, together with abundant quantities of illite. Mixed-layer illite-smectites and kaolinite are also present in variable quantities. Detailed investigation of the clay minerals in shale specimens of the Leacach Formation shows an abrupt and substantial reduction of

the smectite content to less than 5%, above the Leacach Ochreous Member. Illite remains the dominant clay mineral, kaolinite and mixed-layer illite-smectites increase in relative abundance and some chlorite is also present.

Additional samples were obtained from the sandy shales (Obtusum Zone) overlying the Leacach Pebble Member.

Samples obtained from red mudstones (?Rhaetic) of Wilderness (NR 404287) contain up to 40% kaolinite and 30% mixed-layer illite-smectites, illite comprises 19% of the clay minerals, whereas smectite is present up to 10%.

Depending on the variations in the chemistry of smectite, its air dried 001 peak position is obtained at intervals between 12-14Å (Millot, 1970); the basal spacings also vary with atmospheric humidity (Grim, 1953). The detailed chemistry of the clay minerals found in this study has not been investigated. Clay minerals that show 12-14Å peaks which expand to 18Å upon glycolation and collapse to 10Å after heating up to 550°C, are generally referred to as smectites. The illite (10Å) and kaolinite (7Å) peaks are not effected by glycolation but heat treatment(550°C)destroys the 7Å peaks while the 10Å peak remains and is sometimes enhanced due to the large proportion of mixed-layer illite-smectites present.

4.3 Discussion

Smectites are known to form at present by the weathering of basic igneous rocks in an alkaline environment in the presence of Mg (Millot, 1970).

Because of the presence of Tertiary basic igneous intrusions which might be expected to weather to smectite, the possibility of weathering contamination of the adjacent shales must be considered.

A relatively unweathered doleritic sill 1m thick occurs about 50m above the base of the Lias at Applecrosss and a deeply weathered

doleritic sill was observed 2.50m above the Lias/Trias contact at the same locality (NG 725447).

Although a characteristic colour zoning (Hournung and Hatton, 1974) was absent, a yellowish-brown weathering friable core surrounded by a concentric layer of grey sandy shale-like material could be recognised in the latter locality. Igneous and metamorphic rocks which are weathered locally to depths as great as 40 ft. are known in many parts of Scotland (Fitzpatrick, 1963). Such deep weathering has been related to weathering during the preglacial period, the altered rocks have mostly rotted to a sandy material which can be easily dug into. The mineralogy of the products of such weathering however, depends on the composition of the intrusion and the prevailing climatic conditions (Mitchell, 1955; Basham, 1974; Kabata-Pendias, 1975; Dare-Edwards and Livsey, 1976; Millot, 1970; Ragg and Ball, 1964).

The clay mineralogy of samples taken progressively outwards from the core, representing the rotten remains of an olivine basalt sill which was intruded into the shales in Applecross, shows a systematic decrease of smectite and the formation of mixed-layer minerals outwards from the core; kaolinite is absent from the core itself.

The smectite content of the core is a product of in situ weathering and partial alteration of the dolerite to clay minerals. The epigenetic reworking of the erosional products by meteoric waters may have caused the breakdown of smectites in surface layers and the eventual formation of illites; illite-smectites develop in the intermediate stage (Roberson and Jonas, 1965; Hower and Mowatt, 1966; Suda and Shimoda, 1977). The enrichment of a thin zone underlying the two sills in Applecross in terms of smectites could therefore possibly be related to recent weathering contamination.

Although in the past 620,000 years highly silicic ash ejected from the Icelandic or Jan Mayen volcanoes has been deposited in sediments over an area of several million square kilometers in the North Atlantic (Ruddiman and Glover, 1972; see also Bain and Tait, 1977), the vertical mixing of the ash falls and their alteration to smectites (Millot, 1970), could not have affected the samples; the ash falls would only form thin veneers and could seldom penetrate the rock to depths of more than a few centimetres. Likewise, it is improbable that Tertiary volcanic ashfall covers together with the impact and alteration of volcanic emanations were responsible for the uniform occurrence of smectites in these rocks (only fresh, unaltered samples taken from at least 50cm depths were analysed).

Thermally induced chemical transformations in the immediate vicinity of igneous intrusions into shales are directly related to the composition of the intrusion and are the most complex sources of error in the determination of their mineralogy. Correia and Maury (1975), have shown that due to the intrusion of a basaltic dike into French Toarcian marls, mineralogical changes around their periphery (up to 60cm) include the disappearance of kaolinite, diminution of illite with the increase in relative 002/001 peak intensities, and the disappearance of mixed layer illite-smectites. Sauvin et al. (1975) show that chlorite, kaolinite and mixed-layer illite-smectites in shales are transformed into well crystallized illite-chlorite in the vicinity (up to 1.50m) of doleritic intrusions.

The heat to which the shales are subjected is also an important factor in the process; an association of quartz + albite + paragonite + muscovite + chlorite, is neoformed in smectite-bearing shales under controlled T/P conditions (Althaus and Johannes, 1969). In cases where fluids other than connate waters are introduced into the sediments, under 700°C $T/1000^{K_b}$ P, dissolution of silica occurs (remobilized from

the transformation of mixed-layer clays to illite and chlorite). The silica is reprecipitated in various forms above the sill roof as the temperature drops to 400°C.

Short-lived high temperature gradients present at the time of emplacement can create over pressures due to dehydration around the periphery of the sills. The result tends to be fissure emanations, and haphazard contamination of sediments above the intrusions; dikes with no surface impression are also potential sources of error.

With regard to the Liassic beds under study, a systematic disappearance of kaolinite together with the diminution of mixed-layer illite-smectites near the intrusions (Correia and Maury, 1975; Sauvin et al. 1975) was not observed.

Chlorite and interstratified chlorite dioctahedral smectites (Blatter et al. 1973) are totally absent and no evidence of metasomatic alteration is seen above and below the sills in Applecross. The overall clay mineral distribution pattern is persistent throughout the stratigraphic sections and there is no systematic enrichment of the sediments in certain minerals with respect to their distance from the intrusions.

This more or less uniform distribution of smectite in the Milton and Loch Aline Formations, whether intrusions are present or not, and its total absence in the Milton Formation with a correspondingly abrupt, substantial reduction in the Leacach Formation, calls for an explanation genetically related to the early Liassic depositional environment.

Two schools of thought dominate the interpretation of clay mineral data. Weaver (1958a, b) and Riviere (1953) consider the clay minerals present in Argillaceous rocks as reflecting the character of their source area and that they are only slightly modified in the

depositional environment (these may be termed inherited or detrital (Millot , 1970)). Grim (1953, 1958), Millot (1953, 1957), Keller (1956) and Hilterbrand et al. (1973) on the other hand believe that some clay minerals may suffer considerable diagenetic alteration, the extent of which depends upon the environment of deposition. The ease with which these transformations take place is not believed to be the same for all varieties of clay minerals and varies both with the nature of the environment of deposition and the material brought into it (these may be termed neoformed clays (Keller, 1970)).

Clay minerals may also reflect their source rocks rather than climate, relief or diagenetic effects (Morton, 1972; Potter et al. 1975; Parker, 1974; Latouch, 1975; see also Shaw and Bush, 1978).

Explaining the lateral trend found in the clay minerals of bottom sediments off the mouths of many rivers around the world has also renewed a controversy for years. To explain the lateral trends in the clay minerals of bottom sediments, two possible mechanisms apart from chemical alteration (Grim, 1968) have been offered, these are as follows:

- i. Differential flocculation (Snowden and Forsthoff, 1976; Whitehouse et al. 1960; Shiozowa 1970; Krone, 1962; Edzwald and O'Melia, 1975).
- ii. Size segregation (Gibbs, 1977).

Lateral trends found in the clay minerals of the Lower Liassic shales of northwest Scotland will be indirectly investigated in this chapter, but it should be mentioned that although the effects of chemical alteration can be determined more or less, it is impossible to determine whether the lateral clay mineral changes in ancient sediments have been formed due to differential flocculation or size segregation.

It is considered here that the clay mineral distribution in

sedimentary sequences results from a combined effect of the above mentioned processes; the problem of distinguishing among them in each case still remains (Millot , 1970).

4.4 Smectites

The possible origins of smectites have been adequately discussed in the literature (Perrin, 1971; Millot , 1970; Keller, 1970; Bradshaw, 1975). According to Brown et al., (1969) smectites may be formed through the dissolution of silica from opaline sponge spicules or flint, removal of alkalis and alkaline earths from glauconite, micas or feldspars and the dissolution of CaCO_3 .

Although feldspars are present in the Milton and Loch Aline Formations of northwest Scotland, together with mica, there is no trace of flint or sponge spicules and glauconite is rare.

Kounetsron et al. (1977) found that smectite formation in Togo vertisols developed on biotite-gneiss amphibolites.

Smectites may form by the post-depositional diagenesis of degraded micas in the presence of waters of low ionic strength (Perrin, 1971); this process may possibly be effective in porous sandstones but it has no relevance to the low-porosity shales and limestones of the Milton and Loch Aline Formations.

Smectites may form by the terrestrial and sub-aqueous alteration of basic volcanic material (Ross and Shannon, 1926; Brindley, 1957; Slaughter and Early, 1965; Hallam and Sellwood, 1968; Biscaye, 1965; Millot , 1970; Jeans et al. 1977; Perrin, 1971; Bradshaw, 1975). The presence of glass shards is the most direct and convincing evidence for the association of smectites with such igneous activity, and is reported from some Recent sediments (Millot , 1970). In the marine environment volcanic glass undergoes hydrolysis to give rise to smectite (Keen, et al. 1976), and the excess silica produced by such a reaction may appear in the sediment as cristobalite (Bridley, 1957),

remain amorphous or combine with alumina and alkaline earths to form zeolites. X-ray analysis of whole-rock samples has not revealed the presence of any of the above minerals in the two Liassic Groups; only 3.22Å and 3.26Å peaks have been recorded which are generally considered to be characteristic of sanidine. If present, sanidine may be related to acid volcanic activity and their corresponding tuffs, but more evidence is needed to confirm this mineral's presence in the studied strata. It should be noted that several authors have reported that potash feldspars are generated diagenetically while Pacific volcanogenic sediments partly alter to smectites. Based on the near-uniform distribution of smectites in all the sections, its low degree of crystallinity and the high proportion of illite-smectite mixed-layering present the possibility of the existence of true bentonites was rejected previously by the present author (Amiri-Garroussi, 1977).

Modern sediments derived from basic volcanics e.g. off the northwest coast of Cyprus (Millot, 1970; Blatt et al. 1972) contain up to 90% smectites in their clay fractions.

The Milton and Loch Aline Formations contain up to 30% smectites and the intrusion of Tertiary doleritic sills has not affected the smectite content of the shales. The presence of smectites may be related to the genetic environment of the shales. Older strata with considerable smectite content are absent from the region but it should be noted that Rhaetic mudstones taken from the Mull area contain up to 10% smectite; also Rhaetic sediments in south Devon show smectite contents of up to 20% (Henson 1973; Michael Mayall pers. comm, 1978). Rhaetic-Liassic sandstones in northeast Scotland and England also contain expandable smectitic clay minerals (David Jones and Andrew Hurst, pers. comm. 1978). Although Watts (1976) found smectites in the basal Permo-Triassic beds of Gruinard Bay (NG 900929), his investigation shows that smectites increase in the <0.2µm fraction and

are up to 95% in abundance at the base of the 90cm section; they decrease to <50% at the top. The decreased aridity of Permo-Triassic climatic conditions (e.g. formation of concretion stones) would explain the formation of poorly crystallized smectites in the Upper Triassic-Rhaetic calcretes and red beds of northwest Scotland (see Steel, 1974; Watts, 1976).

The lack of episodic, pulse-like increases in the smectite content (cf. Bradshaw, 1975) or more direct evidence of igneous detritus eliminates contemporary volcanism as a likely source. Smectites may form by the erosion of exposed basic igneous rocks under ineffective leaching conditions in alkaline environments where Mg^{+2} , Ca^{+2} , Fe^{+2} and Na^{+} are retained (see below).

A climatic change to more acidic conditions (increased rainfall, and humidity) accompanied by the more widespread coverage of exposed land areas by advancing seas could adequately explain the absence of smectites in the Strath and Leacach Formations; the formation of ironstones in the above mentioned beds supports this interpretation (see Chapter 9).

The geological environment of primary stage smectite was given by Keller (1970) as follows:

- i. Retention of Mg^{+2} , Ca^{+2} , Fe^{+2} and Na^{+} .
 - a. Evaporation > precipitation as semi-arid climate.
 - b. Ineffective leaching
 - Stagnant water and waterlogging
 - Ash in lakes and marine basins
 - Low effective permeability of rocks
 - c. Alkalinity
 - d. Fe^{+2} not combined with O^{-2} or S^{-2}
 - e. Silicates characterized by:
 - High Specific surface as volcanic ash
 - High susceptibility to hydrolysis

- ii. Retention of silica
 - a. Flocculated by Ca^{+2} , Mg^{+2} and other cations
 - b. Ineffective leaching
 - c. Clay size cristobalite
- iii. Retention of parent texture and mineralogy.
 - a. Shards or flow
 - b. Igneous-type minerals

Ample sources of basic igneous rock of various ages are present in the vicinity (e.g. Devonian volcanics in Oban) and may have been eroded under the suitable conditions mentioned above, to provide the Milton and Loch Aline Formations with smectites. In view of the discussion put forth by the author (1977; and Chapter 12 in this study) the possibility of Permo-Triassic volcanic activity in the form of minor sills and dykes cannot be ruled out. The possibility of such volcanic activity in northwest Europe is only recently being investigated and 'few and far between' occurrences have been reported (Faerseth et al, 1976; Klingspore 1976; Dr. I. Printzlau pers. comm, 1977). It may also be of interest to note that the large alkaline volcanic centre of Rosemary Bank in the northern Rockall Trough has a palaeomagnetic inclination consistent with an early to middle Permian age; the errors on this result are such that the date could lie between late Carboniferous and early Triassic (Scrutton, 1971; Dr. D. K. Smythe pers. comm. 1977).

The numerous doleritic sills and dykes which frequently disturb the normal sequence in the sedimentary rocks of northwest Scotland are summarily thought of as representing Tertiary volcanic events; detailed relative age determination of these minor igneous intrusions may prove interesting differences. Kalander (1974) reported: 'two doleritic sheets cut the Durness Limestone and the Triassic conglomerates but they seem not to cut the Jurassic, one of them seems to become thicker

near the Jurassic rocks' in the area of Camas Malag (NG 18655942) in Skye.

4.5 Mixed-layer Clay Minerals

A significant proportion of the clay minerals in the beds under study consist of mixed-layer illite-smectites.

Illite-smectite commonly forms from smectite through burial diagenesis (Kossovskaya and Drits 1970; Pollard, 1971; Sudo and Shimoda 1977; Ebrel, 1978). The nature of the constituent layers and the way in which they are stacked in mixed-layer minerals is not known and an arbitrary method of identification is followed (Weir and Reiner, 1974). Models for the constituent layers and interlayers of interstratified minerals are usually constructed by analogy with known mineral species (mica, smectites, vermiculites, chlorite) and modified in the light of the measured compositions of naturally occurring minerals (Hower and Mowatt, 1966) or analogues produced in the laboratory (Rich, 1968).

The suggestions of Bloss (1966), although useful could not be followed because of quantification uncertainties.

It should be mentioned that the diagenetic conversion of smectite to illite under conditions of deep burial was proposed by Burst (1959, 1967) who was followed by Powers (1957, 1959) and Weaver (1960, 1961); together with Perry and Hower (1970), Dunoyer de Segonzac (1970), Weaver and Beck (1971) and Van Moort (1971) they maintain that depth of burial and not stratigraphic level controls the proportion of expandable material present. Burst (1969) has since reinterpreted his data and concluded that smectite is compositionally unchanged by burial; the structural conversion involves only dehydration to one water layer (Perry and Hower, 1970).

The Liassic shales were studied at surface outcrops and there is no systematic change in the relative proportion of the clay minerals relative to the base or top of the sections and any such smectite -

mixed-layer conversions that may have taken place were probably independent of burial depth. It is readily seen that the above mentioned process could provide the studied beds with an original smectite content of up to 70%.

4.6 Illites

The term 'illite' was originally proposed as a general term referring to clay minerals belonging to the mica group (Grim et al. 1937); thus it is not a specific mineral name and many processes may lead to its formation (Keller, 1967, 1968).

Illite is widely distributed in different environments through space and time (Weaver, 1967; Muller 1967; Keller, 1964; Grim, 1962; Millot, 1970). It is found in the present-day marine environment.

Considerable controversy exists over the question of possible illite formation from smectites in the marine environment. The question is complex and the results are rather inconclusive. Keller (1970) explained that in the present-day environments little illitization of smectite takes place at or near the depositional interface of marine sedimentation. After deep burial following marine deposition, geologically large amounts of illite have been formed from smectites; the conditions leading to such a conversion have been studied and reported by many workers (Hower and Mowatt, 1966; Kastner and Bada, 1974; Heller-Kallai, 1975; Eberl and Hower, 1976).

Smoot (1960) showed that the distribution of clay minerals is essentially contemporary with deposition; the origin of different clay mineral zones in the sediment is related to the rate of sedimentation and the environmental conditions. The weathered minerals which were deposited slowly, migrated far out and were transformed by recrystallization into illite, which thus becomes the well crystallized mineral that one finds offshore.

Devine et al. (1973) found that illite increases away from the mouth of the Rio Grande river but maintained that it was not necessarily a marine indicator.

In the Atlantic Ocean, Biscaye (1964b) found that illite is the dominant mineral which constitutes more than 90% of the clay fraction, a similar pattern was observed by Griffin and Goldberg (1963) in the Pacific; it was regarded that the general smaller size of the illite particles as compared to the other clay minerals was the cause of differential setting which in turn influenced the clay mineral distribution. Illites are normally associated with sediments of marine origin in ancient strata (Krumm, 1963; Millot, 1970). This explanation may adequately elucidate the occurrence of illite in the Lower Liassic shales of the study area.

The high proportion of illite found in the Semicostatum beds of the northern area may indeed indicate a greater marine influence than for any of the other Zones; the illite content of the shales of Angulata Zone age in the southern area is much higher than the shales of the same age in the northern area (see frequency percent diagrams of Fig. 4.2).

4.7 Kaolinite

In recent marine environments, kaolinite is generally restricted to the nearshore areas (Griffin and Goldberg, 1963; Biscaye, 1964b; Griffin et al. 1968; Brooks et al. 1976). A large increase in kaolinite (to >50%) is seen in sediments adjacent to the large tropical rivers of Africa and South America. A similar pattern was observed by Griffin and Goldberg (1963) in the Pacific and by Devine et al. (1973) in the Deep Gulf of Mexico where kaolinite is mostly concentrated in the nearshore environments, near the mouth of the Rio Grande river. Griffin (1962) and Griffin and Parrot (1964) also regarded kaolinite

as a nearshore indicator. Kaolinite is a valid indicator of provenance at least in Recent sediments, (Keller, 1970).

Kaolinite in sediments may be derived as a product of weathering and soil formation in all rocks including volcanic ashes under strong leaching conditions. Its abundance relative to other clay minerals tends to increase with intensity of removal of silica and cations and increasing temperature; it decreases with the conservation of cations in arid soils (Perrin, 1971). Krumm(1963) found that kaolinite was associated with terrestrial coal-yielding beds in the Rhaeto-Liassic beds of east Franconia, Germany. Kaolinite is normally produced by the weathering and transportation of acid igneous rocks.

In contrast to mica, kaolinite is not affected by prolonged transport in waters of low ionic strength so that the kaolinite entering a basin of deposition, is essentially the same in type and amount as that leaving the terrain which is being eroded. The low surface charge of kaolinite causes it to remain unaltered when entering normal sea-water and its relatively large particle size accelerates its segregation from other clay minerals by differential flocculation and settling, thus it becomes concentrated in shallower water sediments. But it should be noted that abundant kaolinite is frequently reported from deep sea and oceanic sediments where it is most abundant in tropical regions (Blatt, et al. 1972). Kaolinite is not considered to be neo-formed in sea water. Its stability in the marine environment appears to continue after shallow burial and it has been suggested that it may even be further formed at this stage (Larsen and Chilingar, 1967).

With regard to the two Groups of the Broadford Beds, it can be seen that two broadly defined kaolinite suites can be recognised. The kaolinite content of the Milton Formation seldom exceeds 11% which contrasts the values of up to 47% obtained from the beds of the Loch Aline Formation. Ample sedimentological evidence is available for the

nearshore development of the cross-bedded calcareous sandstones and oolitic limestones of the Milton Formation. The general low amount of kaolinite in the shale beds may have been caused by one of the following factors:

- i. Source area
- ii. Offshore deposition
- iii. Chemical and climatic conditions
- i. Source area

Difference in source area(s) is an important factor in the concentration and distribution of kaolinite in the beds studied. Potential kaolinite-producing areas (e.g. metamorphic terrains) show a much greater outcrop area in the southern region of study than in the northern region. In some places the Liassic beds directly overlies Moine schists and have developed in the proximity of different Lower Palaeozoic lithologies which may have had an even more widespread outcrop during the Lower Liassic times and served as rich, kaolinite-producing source areas. In the northern area of study, the Milton Formation mostly overlies Triassic conglomerates, Precambrian quartzites and limestones and in some places Torridonian sandstones, none of which are significantly rich in kaolinite.

Although there is some evidence for a metamorphic provenance of the clastic quartz fragments (strained and polycrystalline quartz), they may have been of a recycled origin, in which case progressive winnowing could effectively have removed the kaolinite flakes.

- ii. Offshore deposition

Although some shales in the Milton Formation were deposited under restricted, nearshore marine conditions, most beds are indicative of periods of marine incursion. In contrast to the northern area where clastic-carbonate deposits characterize a turbulent depositional environment, uniform mud deposition in the southern area of study typifies

quiet-water depositional conditions. Thus although clastic carbonate shoreline and offshore deposits prevailed in the northern area of study, muddy shorelines developed in the southern area (these deposits are most probably direct reflections of the bed-rock types and their weathering).

iii. Climatic conditions

As previously mentioned, warm acid climates favour the formation of kaolinite. The geological environmental conditions necessary for the primary formation of kaolinite are given by Keller (1970) and presented as follows:-

- A. Removal of Ca^{+2} , Mg^{+2} , Fe^{+2} , Na^{+} , K^{+}
 - a. Precipitation exceeds evaporation
 - b. Permeable rocks.
 - c. Percolating waters.
 - d. Oxidation of Fe^{+2} to Fe_2O_3 or FeS_2
- B. Oxidation of H^{+}
 - a. Fresh water.
 - b. Acids
 - sulphur compounds
 - carbonic air and soil atmosphere
 - organic, living and dead organisms
- C. High $\text{Al}^{+3}:\text{Si}^{+4}$
 - a. Removal of Si^{+4}
 - Na^{+} and K^{+} silicates
 - Organic complexes
 - b. High concentration of Al^{+3}
 - Al^{+3} in acid solution
 - Al-OH polymers

Although there is no direct evidence to suggest the development of climatic or chemical zones in northwest Scotland during the Lias, the slight increase of kaolinite in the Loch Aline Formation as compared to the

Milton Formation (Fig.4.2-4.7) may not only be a function of distance from the shoreline but also more favourable chemical and climatic conditions in the southern area of study. 'Ironshot' limestones are found in the Loch Aline formation together with pyrite whereas pyrite and iron compounds are rare in the Milton Formation.

The substantial increase in the kaolinite content of the Leacach and Strath Formations at first sight seems to contradict the shallow, nearshore interpretation of the occurrence of kaolinite. It should be noted that although the Strath and Leacach Formations were deposited under fully shallow-marine offshore conditions, the intense acid leaching of low-lying schist and gneiss terrains under warm climatic conditions probably not only inhibited the further stability of smectites in the environment (see section 4.4), it enhanced the abundant formation of kaolinite and its transportation into the environment. Chlorite also becomes a minor constituent of the clay minerals of the Strath and Leacach Formations.

As explained earlier (iii.C), aluminium released by acid weathering of clay minerals in soils, is often deposited as complex hydrated ions in the interlayer spaces to form chlorite. The increase of chlorite and concomitant decrease of expanding minerals in the iron-rich sediments of the Strath and Leacach Formations therefore probably resulted from acid weathering in the soils of the subdued Precambrian and Palaeozoic land areas. In this weathering regime iron would be rapidly mobilized and then transported either in solution or as oxide coatings on abundant kaolinite and other clay particles (Carroll, 1958). The diagenetic formation of kaolinite may occur in porous sandstones (Shelton, 1965; Perrin, 1971; Almon, et al. 1976; Hancock and Taylor 1978; Wilson and Pitman, 1977) but its formation in low porosity limestones and shales is doubted.

4.8 Chlorites

Although not ubiquitous, chlorites occur sporadically in the beds of the Strath and Leacach Formations. Chlorites found in a sedimentary rock may be derived from one or more of the following processes (Perrin, 1971).

i. Chloritization of biotite in igneous rocks -

This process is unlikely to have been responsible for the chlorite in the studied beds because such an origin necessarily produce uniformly distributed chlorite within the lithologically similar beds; furthermore biotite is rather scarce in the studied beds.

ii. Partial weathering of biotite and other ferromagnesian minerals in soils of temperate, cool or cold climates where leaching is restricted; this process may have been responsible for the formation of chlorite in the studied beds (although no direct evidence of climatic conditions is present).

iii. Erosion of chlorite-bearing schists and older sediments may well have been a source of chlorite in the Lower Liassic beds of northwest Scotland.

iv. Neoformation in basins rich in magnesium and perhaps ferrous iron may also have been a potential source of chlorite.

v. Diagenesis of 'chamosite' in some iron ores (see Chapter 9) may also have been a contributing source.

vi. Diagenesis of degraded micas, smectites, vermiculites, or associated mixed-layer minerals and kaolinite are also important sources of chlorite in the studied beds.

In Recent depositional environments, the distribution of chlorite in sediments in the Atlantic ocean is nearly reciprocal to that of

kaolinite (Biscaye, 1965). Chlorite is relatively concentrated at high latitudes and diminishes at lower latitudes. In the north Pacific chlorite increases with increasing latitude in nearshore sediments (Griffin and Goldberg, 1963). It is known that provenance is exceedingly important in the distribution of chlorite in the modern sediments (Keller, 1970). However, chlorite and chlorite-like minerals might form from smectites transported to the ocean. This may be carried out by mixing of smectites with Mg^{+2} or Al^{+3} solutions which are suitably alkalized (Keller, 1970). It is thought that the latter explanation is probably the most likely origin of chlorite in the beds of the Strath and Leacach Formations.

CHAPTER 5

FAUNA

The stratigraphy and palaeontology of the two groups of the Broadford Beds is well known and the extent to which the Lower Liassic strata of northwest Scotland have been studied for stratigraphic purposes was pointed out in previous chapters. Among the faunal lists available, those of Hallam (1959) and Oates (1976) are sufficiently ordered to yield significant palaeoecological and palaeoenvironmental information (Tables 5.1a, b, c; 5.2a, b, c). Meaningful quantitative analysis of these data is greatly hampered by limited exposure and the inconsistencies resulting from the comparison of data obtained by more than one worker; this results from errors due to differential observational ability (Palmer and Palmer, 1977).

Microfaunal studies on the two Groups were undertaken by Clark (1969), apart from a few palaeoenvironmental inferences, his study was concerned mainly with stratigraphic conclusions. Of the 67 ostracod species comprising the Scottish Lias, 35 are restricted to the Planorbis to Obtusum Zones; the species Cytherella inaquata (Angulata Zone, Allt Buaile) is abundant, four species of Ektyphocythere are stratigraphically of more importance due to their restricted occurrence, these are:-

<u>Species</u>	<u>Zone</u>	<u>Locality</u>
<u>Ektyphocythere morverniensis</u>	Obtusum	Allt Buaile
<u>Ektyphocythere intermedia</u>	Turneri	Allt Buaile
<u>Ektyphocythere luxuriosa</u>	Semicostatum	Ardnish (Skye); Allt Buaile (Loch Aline)
<u>Ektyphocythere glabra</u>	Bucklandi	Applecross
<u>Ektyphocythere glabra</u>	Angulata	Applecross

Although the majority of the ostracod fauna in the Scottish Lower Lias is associated with other marine faunal elements, in two instances, i.e. Larachbeg Burn (Loch Aline) and the borehole at Lossimouth, non-marine

ostracods have been found (Clark, 1969), these are:

<u>Locality</u>	<u>Fauna</u>	<u>Age</u>	<u>Environment</u>
Larachbeg Burn	<u>Lymnocythere</u>	Rhaetic-Liassic	Freshwater, mesohaline (?oligohaline) in Recent times.
" "	<u>Allocythere</u> cf. <u>combrookensis</u>	Rhaetic-Liassic	-
" "	<u>Cytherella</u> (single valves)	Rhaetic-Liassic	Recent salinity toler- ance 25% ; marine (Sohn, 1964).
Lossimouth	<u>Darwinula</u> <u>Ektyphocythere</u> <u>cf acutocostata</u>	?Sinemurian	Marine

The palaeoecology of the benthic macrofauna in the Broadford Beds Arenaceous and Argillaceous Group is outlined in Tables 5.1b, c, and 5.2b, c. The majority of the genera listed are suspension feeders; present day forms of suspension feeders are most common in sandy or firm mud bottoms whereas deposit feeders are typical of soft muddy substrates (Rhoads and Young, 1970). Tables 5.1b and 5.2b are based mainly on Hallam (1960b, 1972), Hudson (1963), Barthel (1969), Kauffman (1969), Nichols (1969), Stanley (1970), Rudwick (1970), Fürsich (1974), Townson (1975) and Hudson and Palmer (1976).

The data obtained from the faunal lists and Tables 5.1b and 5.2b are presented in percentage form in Tables 5.1c and 5.2c; these data are graphically represented by Figures 5.1 to 5.4 showing variations in taxonomic abundance (diversity) together with their regional changes.

The relative abundances and regional changes in ecological groups are also represented (Figures 5.5 to 5.19). The Liassic Zones of north-west Scotland are suitable for such comparative studies. Although accurate density (individual abundance) parameters are impossible to obtain, their palaeontology and much of their stratigraphy is well known. The Broadford Beds Arenaceous Group is characterized by trace fossils such as Chondrites, Thalassinoides, Diplocraterion and Rhizocorallium which are associated with

a rich and varied shelly benthos; the Broadford Beds Argillaceous Group contains Thalassinoides and ?Rhizocorallium trace fossils and the shelly benthos tends to be more restricted in variety and are composed of smaller sized individuals.

5.1 Diversity

Diversity in a biota is commonly indicated by the number of different species that occupy a particular area (Ross, 1974); many theoretical generalizations are proposed in order to explain the diversity phenomenon (Sanders, 1968; Bretsky and Lorenz, 1969), two of which are as follows:

- i. Heterogenous physical environments contain more niches, and therefore can support more complex and diversified biota.
- ii. Stable climates and physical environments result in narrower niches and more species.

The debate over Jurassic biotic distribution and its various causes has been long-standing (Neumayr, 1872, 1883; Ortmann, 1896; Arkell 1956; Imlay, 1965; Stevens, 1967; Fleming, 1967; Hallam, 1969a, 1971c, 1975, 1976). Diversity can be analysed at any taxonomic level and may be interpreted in terms of the theories of diversity control. Various hypotheses have been proposed explaining diversity patterns (Fisher, 1960; McArthur, 1975; Pianka, 1966; Valentine, 1967, 1968, 1969, 1971; Sanders, 1968; Bretsky and Lorenz, 1969; Scott, 1975). Hallam's (1972) discussion of Jurassic diversity gradients is also applicable to the beds presently under consideration. The general, but sometimes slight, northward reduction in the diversity of the ammonites together with the other fossil groups in the strata of the two Groups of the Broadford Beds are taken into account. During Bucklandi Zone times, three brachiopod species are found in the Leacach Nodular Limestone Member (Mull), whereas contemporaneously deposited beds of the Lower Sand Member of Skye (see Tables 5.1a and 5.1b) contain two.

The concept of environmental stability (Sanders, 1968; Bretsky and

Lorenz, 1969; Woodwell and Smith, 1969) may also be applied here. Modern marine faunas show an increase in diversity from nearshore, shallow-water unstable environments to offshore deeper-water areas which have a high degree of environmental stability. Similar increases in diversity offshore from regions of environmental instability have been recorded recently in the Ordovician, Devonian and Cretaceous of the U.S.A. (Bretsky, 1970a; Walker and Laporte, 1970; Sutton, Bowen and McAlester, 1970; Stevens, 1971; Scott, 1975). This pattern is known to occur in the Lower Lias of northwest Scotland but has not been documented before. Sanders (1968), has shown that individual species diversities in the modern oceans diminished in an order of decreasing environmental stability and predictability in terms of fluctuating temperature and salinity. Physical stresses (energy levels), water depth and turbidity are among other influential factors. Faunas in the unstable environments are subject to high stress and accordingly restricted to eurytopic organisms of comparatively low diversity. Stenohaline taxa of cephalopods, brachiopods, echinoderms and gryphaeid oysters are excluded from the beds of parts of the Milton Formation; this, together with sedimentological data obtained, indicate salinity variations and fluctuating physical energy levels. It has been shown that diversity changes in response to increasing stress conditions (Parker, 1974), these conditions may be in the form of reduced salinity and/or resources in an environment.

Data obtained from the Baltic Sea (Sergeyev, 1957) indicates that species diversity is reduced from the Kattegat (35‰), to the Baltic Sea (25-30‰ salinity), which clearly demonstrates the effect of salinity changes.

The Lower Liassic faunal patterns of northwest Scotland may be explained adequately in terms of modern ecological concepts by postulating a northward reduction in environmental stability from the Mull area to the

shallower waters of the Skye area, into which at least one river debouched from the north or northeast. The sedimentary facies patterns indicate the influx of ?freshwaters from the northern or northeastern lands, which would effect some degree of salinity reduction close to the coast. An important factor may have been the general reduction in water depth since an extensive sheet of relatively shallow water would be subjected to greater and more unpredictable variations in factors such as salinity, temperature and possibly the content of dissolved oxygen than the deeper waters further offshore (Hallam, 1972). The increase of faunal density inshore, has been generally accounted for by suggesting an increased food supply. Ryther (1963), showed that phytoplankton concentrations are higher in the inshore zones, although in such cases more phytoplankton is expected to reach the bottom, this explanation fails to account for the fact that densities may reach extremely high values in the most marginal environments, i.e. hypersaline lagoons (Hallam, 1972).

Studying the faunal distribution pattern observed in the Sinemurian of Sao Pedro de Muel, Portugal (Hallam, 1971c), an explanation based on postulating contrasted adaptive recruitment strategies of benthonic populations in stable or unstable environments termed K and r selection respectively (McArthur and Wilson, 1967), was given (see also Hallam, 1972). This phenomenon may possibly be applied to the Hettangian-Sinemurian beds of northwest Scotland, where the first marine beds overlie less fossiliferous, Triassic-?Rhaetic strata containing a low diversity, high density bivalve and gastropod fauna. This gives way up the succession to a higher diversity fauna (Fig.5.1) but the density diminishes. Such changes may indicate the replacement of the fauna of a slightly hypersaline, semi-enclosed marginal marine environment, by one of a shallow, extensive sea with near normal marine salinities.

Since diversity can be analysed at any taxonomic level, and as the number of genera at any given time interval during the Lower Lias seldom exceeds

2 or 3, although generic data are thought to be more reliable (Hallam, 1971c, 1976; Stehli et al. 1967), species data is used in the plotting of various diversity patterns. Diversity patterns in species and genera differ little from one another (Stehli et al. 1967; Hallam, 1976) so no essential data is lost in the analyses by considering only species distinctions. In Figures 5.14 to 5.19, the numbers of species of gastropods, bivalves and ammonites (expressed in%) known in each area, are plotted to form a north to south diversity profile for the Angulata and Bucklandi Zones. The absence of strata of the Planorbis Zone in the Skye area and the lack of data for the Semicostatum and Turneri Zones for the Mull area prevent construction of lateral diversity profiles for the two regions during these intervals. It can be seen that during the Angulata and Bucklandi Zone times:

- i. Faunal diversity of ammonites, brachiopods and bivalves decreases from the Mull area (south) to the Skye area (north), while no appreciable change is seen to occur in the gastropod diversity pattern.
- ii. Euryhaline species, together with infaunal biota are more abundant in the southern area of study than in the northern area. While upper brachyhaline organisms also show the same diversity pattern during the Bucklandi Zone times, the reverse is true for the Angulata Zone organisms. The regional diversity pattern for epifaunal species remains constant for both areas during the Bucklandi Zone whereas they form a larger proportion of the fauna of the Mull area during the Angulata Zone.
- iii. The distribution pattern of attached, sessile, active, deposit and suspension feeding organisms is altogether very similar during the Angulata and Bucklandi Zone times. A greater diversity is seen in the southern area (Mull) whereas

the fauna in the northern area (Skye) are less diverse in terms of life and feeding habits.

iv. While the strata of the Mull area show an overall higher faunal content, the diversity pattern of both areas seems to change in a similar fashion (Fig.5.1). The Angulata and Semicostatum Zones show diversity 'highs' whereas low diversity values are recorded for the Turneri and Bucklandi Zones. This pattern is repeated for the bivalves and euryhaline fauna together with the infaunal biota (Figs.5.3, 5.11, 5.13).

v. Although the diversity of the ammonite fauna in the Mull area shows a 'high' during the Angulata Zone and lows during the Planorbis (and ?Bucklandi) Zone(s) in the Skye area, their diversity increases from the Angulata to the Semicostatum Zone which is followed by a Turneri low.

vi. In the Mull area, the diversity pattern of the gastropod species present shows an increase during the Planorbis-Bucklandi Zone period. In the northern area, the diversity pattern remains constant during the Angulata and Bucklandi Zones and is lowered during the Semicostatum and Turneri Zones.

vii. Attached fauna progressively increase from the Planorbis to Bucklandi Zone times in the Mull area. This pattern is also seen in the Skye area where the diversity of the attached fauna increases from the Angulata to Semicostatum Zone times and decreases during the Turneri Zone (Fig.5.5).

viii. Sessile fauna show an increase in diversity from the Planorbis to Angulata and to Bucklandi Zone times. In the northern area (Skye) the diversity of sessile fauna progressively increases from the Angulata to the Semicostatum Zone times, they are less diverse in the Turneri Zone (Fig.5.6)

ix. The 'active' fauna of the Mull and Skye areas increases during the Bucklandi and Semicostatum Zone times; their diversity remains constant for the Skye area during the Angulata-Bucklandi Zones and decreases during the Turneri Zone (Fig.5.7).

x. The deposit-feeding fauna become progressively more diverse from the Planorbis to Bucklandi Zone times in the Mull area and also for the Angulata and Semicostatum Zones in the Skye area (Fig.5.8); they diminish in the Turneri Zone of Skye.

xi. While the suspension-feeding fauna became more diverse during the Planorbis to Bucklandi Zone times in the Mull area, a pronounced decrease in their diversity (from the Angulata-Bucklandi Zone) is seen in the Skye area; in this latter region, the Semicostatum Zone shows a marked high diversity of suspension feeding organisms which diminish totally during the Turneri Zone (Fig.5.9).

xii. The epifaunal organisms show periods of high diversity in the Angulata Zone (Mull) and Semicostatum Zone (Skye); the Bucklandi and Turneri Zones are periods of low epifaunal diversity (Fig.5.10).

xiii. Although upper brachyhaline organisms are present in the strata of both regions (Skye and Mull) they show a marked decrease in diversity during the Bucklandi Zone (Fig.5.12).

5.2 Interpretations

Understanding the nature of control on the temporal and regional changes exhibited by the various faunal characteristics of the Lower Lias of Scotland will greatly facilitate an overall palaeoenvironmental interpretation. The ecological significance of the bivalve fauna is of great importance, although these have not been dealt with comprehensively. It is impossible to establish a relationship between the different types of organisms and the sediments of the Broadford Beds Groups since some species

are common in all the various lithologies. In such cases it is more instructive to consider associations of sediments with trace and body fossils (for example see Hallam, 1971b, 1976). Following Hallam (1975), the following points are considered:

i. Liostrea usually occurs in high density, low diversity assemblages and is especially characteristic of very shallow water, marginal marine environments. Thus the presence of Liostrea in the beds of the Milton Formation in the Skye and Ardnamurchan areas together with the basal beds at Craginure Bay indicate marginal marine, shallow water conditions, whereas the Loch Aline Formation shales represent a relatively deeper water facies as evidenced by the presence of stenotopic fauna such as ammonites. The occurrence of Liostrea at the base of the Angulata Zone at Loch Aline (bed no.-31 of Oates, 1976) immediately below a bed yielding a specimen of Sulciferites sp. is either an error or an unusual occurrence of this bivalve in association with stenotopic organisms. Liostrea irregularis var.

hisingeri has also been reported in association with ammonites in the Skye area (Hallam, 1959), beds containing Liostrea sp. also occur at the base of the Liassic section (immediately overlying the Triassic beds) at Craginure Bay (Mull).

ii. Gryphaea is more characteristic of slightly deeper water, more open marine environments where ammonites also occur.

The above general pattern of bivalve distribution is also seen in the Broadford Beds Groups although sedimentological evidence for the existence of hypersaline 'lagoonal' conditions is totally lacking. The following ecological control factors seem to have been of importance in the distribution of Lower Liassic fauna; although as Hallam (1976) pointed out, the most important influences might have involved a combination of fluctuating factors.

5.3 Substrate

Cemented or boring bivalves are rare in the Lower Lias of north-west Scotland, indicating that a firm substrate of rock, shells or firmly compacted sediment was either not available or scarce. The latter suggestion is favoured since attached fauna are present in both of the studied groups (Tables 5.1b and 5.2b); a slightly firm, muddy substrate would be needed for infaunal or semi-infaunal species. Interstitial cementation or moderate sand or silt admixtures in the clay could result in such a firmness. Although a 'soft and soupy' substrate was limiting to most suspension feeders (cf. Rhoads and Young, 1970), it is seen that in Figure 5.9 the suspension feeding fauna show an increased diversity in a north-south direction, during the Angulata and Bucklandi Zones, thus indicating the probable existence of harder substrates in this latter region.

Their low diversity in the northern area (Skye) is thought to be mainly a function of water turbulence and the instability of substrate (oolitic sands, sand waves, etc. as revealed by careful sedimentological considerations) rather than its non-firmness. The general distribution pattern of north-south epifaunal and infaunal organisms also suggests that during the Angulata Zone the southern area (Mull) supported a larger population of epifauna (? harder substrates) than the northern area (Skye), where infaunal organisms are better represented. This situation can be accounted for by considering substrate firmness and stability in the southern area and its instability in the Skye area, together with the environmental food-supply mechanism. During the Bucklandi Zone, the firm substrate became more similar in both regions (Fig. 5.10) and was able to support the same number of epifaunal organisms. The infauna is practically negligible (<10%) in the northern area and becomes rather significant in the southern area. Figures 5.14 and 5.15 indicate that during the Angulata Zone most of the food was available probably in the form of

suspension, whereas during the Bucklandi Zone it was mostly available in the form of fine detritus in the bottom sediments. The abundance of both types of available foods increased from the northern area of study (Skye) to the southern area (Mull).

5.4 Turbidity

The effects of turbidity are hard to disentangle from those of the substrate and (indirectly) salinity fluctuations near shore. It was possibly effective for reef dwellers and probably for some of the larger suspension feeders (Hallam, 1976).

5.5 Salinity

Possibly some genera could tolerate some abnormalities or fluctuations in salinity of a few parts per mil. (e.g. Liostrea of the Loch Aline Formation). There is a sharp diversity reduction at about 30‰ and brackish water genera are absent. Figures 5.18 and 5.19 show the regional changes in salinity tolerance for the fauna of the Broadford Beds Groups. It can be seen that while the relative abundance of the euryhaline species increases from north to south during the Angulata and Bucklandi Zone times (indicating possible slight salinity reductions of the waters of the northern area due to freshwater ?stream influxes), this area (Skye) supports a greater (24-30‰) salinity tolerant fauna than the southern area (Mull) during the Angulata Zone; this pattern is reversed during the Bucklandi Zone times.

5.6 Temperature

The water temperatures were consistently high in northwest Scotland during the Lower Liassic times, probably comparable to the present-day subtropics (Hallam, 1972; 1975), this is borne out by evidence provided by ostracods and corals. It should be realized that the (ostracod) palaeotemperature inferences are mainly based on extrapolation from the conditions in which present-day ostracod species survive (Sohn, 1962). Therefore, this assumption was the basic premise that the habitats of these genera

have not changed over the past millions of years. Kronicker (in Sohn, 1964) suggests that with gradual decrease in temperature, the ostracod Cytherella could acclimatise to new extremes of temperature. The temperature ranges given by Sohn (1962) for Recent and Upper Cretaceous Cytherella are $4-9^{\circ}\text{C}$ to 32°C (depth $<1-675$ fathoms), for Cytherelloidea he suggested a maximum temperature of 10°C (depth $1-46$ fathoms); these temperatures may apply to the Liassic species of these genera but they should not be accepted without a more direct method of checking, i.e. $\delta^{18}\text{O}$ analyses in order to ascertain if these genera have adapted to new extremes of temperature during the Lias and Cretaceous. Bowen (1961b) gave temperatures of between 23.8 and 31.7°C for ^{the}English Upper Lias, with the suggested limits of Sohn (1964) for the two genera; although these are close to the temperatures applicable to the Scottish Lias direct comparisons must await further geochemical investigations. From the ostracod data it may be tentatively suggested that the Liassic water temperatures were between $20-26^{\circ}\text{C}$ with depths up to 46 fathoms.

5.7 Oxygen content

Oxygen deficiency was a restriction for most bivalves during the Lower Lias, although a number of genera were able to tolerate oxygen starvation for long periods (Hallam, 1976); perhaps on account of having a respiratory pigment in the haemolymph.

5.8 Ecological associations

Some of the major associations identified for bivalve groups by Hallam (1976) may also be identified in the Lower Liassic strata of north-west Scotland. Naturally, intergradations exist among the various groups and each rock formation does not contain all the faunal elements cited by Hallam.

i. 'Reefal' associations

In the Lower Liassic rocks of northwest Scotland, although beds of lenticular coralline rock are present and are enveloped in bedded

biosparites, biomicrites and oosparites, the characteristic high diversity, large-sized, ornamented and/or thick shelled bivalve fauna representing a distinctive ecological association cannot be identified. The coral beds consist of Isastrea and Thecosmilia and the fauna associated with them is confined to: Plagiostoma, Cardinia cf. concinna, Gryphaea arcuata, Zeilleria perforata, Calcirhynchia calcaria; the gastropod Ptychomphalus expansus and other indeterminate gastropods, echinoid and Isocrinus debris are commonly found in them. The possible environment of deposition was one of low stress and high predictability, with warm, shallow seas within the photic zone, high oxygen content, low turbidity and the existence of some firm and hard substrates. Although the coral beds in the Broadford Beds Arenaceous Group can hardly be termed 'reefs', Facies 6 and 8 of the Milton Formation are reminiscent of the 'reefal association' described.

ii. Nearshore Marine Associations

The Strath Formation (excluding the Dun Borerraig Member) in the Skye area fits quite well into the framework proposed by Hallam (1976) for this type of association. The lithology consists of sandy shales, bioturbated sandstones with subordinate oolitic and fine-grained ironstones to fine and coarse-grained limestones with some associated marls. The Loch Aline Formation presents evidence for the existence of such an association in the southern area of study. The relatively high diversity bivalve fauna includes many large and small sized epifaunal and infaunal suspension feeders (see Figs. 5.10, 5.11 and Tables 5.2b,c). The bivalves are dominant over the other invertebrate fauna which are generally characterized by a variety of other marine groups of lower diversity, e.g. ammonites, brachiopods, gastropods and crinoids. Trace fossils include Diplocraterion, Rhizocorallium and Thalassinoides. A warm shallow-water generally well oxygenated environment is suggested which was periodically subjected to storms and to slightly fluctuating salinity and temperature because of its nearshore situation. The substrate ranged from firm to

moderately soft. The genera present in the Scottish Lower Lias representing this association are:-

Mytiloids: Modiolus, Pinna
 Pteroids: Oxytoma, Camptonectes, Chlamys, Pseudolimea, Plicatula,
Plagiostoma, Gryphaea, Liostrea
 Venetoids: Cardinia, Mactromya, Astarte
 Pholadomyoids: Pholadomya, Pleuromya

The relative abundance of burrowing infauna during the Semicostatum Zone times in the Skye area (Fig.5.11) indicates the prevalence of a muddy substrate; this can also be seen for the Angulata Zone of the Skye and Mull areas. The predominance of epifaunal elements over infaunal organisms during the Angulata and Bucklandi Zones in the northern and southern areas (Figs. 5.18 and 5.19) indicates a firmer, less muddy bottom environment.

iii. Marine basal associations

One of the two categories mentioned by Hallam (1976) can be recognised in the Lower Liassic strata of northwest Scotland; these are the non-bituminous bioturbated clay or 'marl' category. The predominant lithology is calcareous shales and richly fossiliferous limestones, together with dark, laminated, compact shales and calcareous clays. Parts of the Strath Formation (Upper Teampull Chaon Shale Member) in Skye, together with the Loch Aline and Leacach Formations in Mull may be regarded as representing this association, which contains a relatively more varied fauna. There is no definite distinction with the clays of the near shore association and most of the fauna are found among both strata.

The nudids are relatively less important among the burrowing infauna; this includes such genera as Pholadomya and Pleuromya.

Semi-infaunal and epifaunal byssate cemented or reclining genera include Modiolus, Chlamys, Plagiostoma, Pseudolimea, Plicatula, Gryphaea, and Liostrea; the specimen of Parallelodon cf hettangiensis reported

by Hallam (1959) from the lower most Liassic beds in Applecross was found in shallow water clastic sediments and cannot be regarded as representing parts of a 'marine association.'

iv. Lagoonal associations

The dominant facies representing this association is alternating clays and fine-grained argillaceous limestones. These patterns become more arenaceous in the Skye area where bioclastic, oolitic and micritic facies dominate together with (? freshwater) plastic clays. The bivalve fauna is characterized by low diversity and high density with monotypic (or almost monotypic) shell beds (oysters) becoming very frequent, the shells are generally small, perhaps because the species are small or because of a high proportion of juveniles. The fauna are mostly a mixture of infaunal suspension and deposit feeders, the shells of which tend to be smooth and thin with little to no ornamentation. Hudson (1963) related bivalve associations to salinity regimes, following his work and that of Tan and Hudson (1974), an attempt is made here to relate bivalve associations tentatively to salinity regimes; although it should be realized that based on palaeontological and isotopic data, Liostrea hebridica inhabited waters with a wide range of salinities (Tan and Hudson, 1974; Hudson and Palmer, 1976). and the principal diversity limiting factor may have been mostly the fluctuations rather than actual levels of salinity (Hallam, 1976).

Mesohaline (Ca. 5-10%) fauna are totally absent from the Lower Lias strata of northwest Scotland, this may be a reflection of preservational and outcrop conditions, and more probably due to environmental conditions.

Lower brachyhaline (Ca. 16-23%) fauna are represented by Liostrea sp. found in Applecross (oolitic Facies 5).

Upper brachyhaline (Ca. 24-30%) fauna consist of Liostrea, Camptonectes, Cardinia and Modiolus which are well represented in the

various facies of Skye, Ardnamurchan and Mull.

The other invertebrate fauna comprises a low density assemblage of ostracods and some gastropods; stenohaline groups such as corals, brachiopods, echinoids and ammonites are absent or extremely rare. The environment in which the above mentioned invertebrate associations flourished were evidently those in which high stress conditions were operative with variably fluctuating levels of salinity; on sedimentary evidence, other movements were substantial and the substrate varied from soft to firm.

*"We must rely upon the records of rocks
and vestiges of the past to infer that all
these changes should have taken place in
very very long times and under unknown
conditions of cold and heat"*

*Abu Rayhan Al-Biruni
12th A.D.
Ikhwan Al-Safa*

*(The Epistle of the
Brethren of Purity)*

CHAPTER 6

FACIES DETERMINATION AND INTERPRETATION

In this section detailed descriptions of the rock types are given together with other data relevant to a palaeoenvironmental interpretation. The facies represented by each Member are depicted and their sequential arrangements discussed (both in vertical profile and lateral distribution).

Two major phases of sedimentation are recognised; a transgressive phase followed by a short regressive pulse, each comprising variable facies with marked differences. Examples will firstly be described, then fitted into general sedimentation models. Relationships among the various Formation Members and the facies to be described are shown in Tables 6.1 and 6.2; Fig.224. Although the beds concerned have been previously measured and described (Hallam, 1959; Oates, 1976), sections have been measured independently for the purpose of internal consistency.

6.1 Torr Mor Member (NG 630164) 3 m.

This member is best exposed in southern Strath, $\frac{1}{2}$ km. to the northwest of the Liassic section of Loch Eishort; it is not well represented in the other outcrops of the Broadford Beds Arenaceous Group of this area. It may be divided into three facies from base to top. The total thickness of this Member cannot be determined accurately and the thickness of each unit is given at the individual localities.

As the beds of this Member represent the lowest unit of the Jurassic of northwest Scotland, marking pronounced changes of depositional conditions from those prevailing here during the Triassic, it has deemed important to investigate the nature of their contacts with the underlying strata.

6.2 Basal contacts

i. Applecross (NG 44757265)

The Triassic/Liassic contact may be seen at the northern bank of the stream running southeast of Applecross House (NG 44757205). Here Triassic red mudstones, siltstones and conglomerates are overlain by a 1 m. development of whitish, grey-weathering calcilutites with variable compaction and uniform (15-20 cm.) bedding. Opaque material is present as linings of some recrystallised shells and as rounded pellets. Rounded, micritic limestone pellets are present together with rounded, fine sand grade quartz (Plate 6.1).

ii. Northern Strath (Ob Lusa) (NG 700250)

The Triassic/Liassic junction is difficult to determine here. The typical Liassic silty bioclastic limestone beds overlie 2.50 m. of beige-yellow, light-green weathering silty marlstones which contain abundant rounded limestone pebbles. In places these are indurated and contain broken shell fragments together with coalified plant remains. The lowermost beds are bioturbated, poorly sorted, moderately sandy biopelmicroparites with few shale partings. The top surfaces of these beds show structures hitherto not reported from the Triassic beds of northwest Scotland, and which are also absent from the ?Rhaetic of the Skye area. These surface markings consist of a sub-spherical to spherical calcareous centre (3-5 cm. in diameter) with slender, outward radiating arms (15-20 cm. long, 10-15 mm. thick), composed of yellow-weathering limestone, which frequently branch out away from the central sphere, forming an interconnected network in places (Plate 6.2).

iii. Southern Strath (Loch Slapin) (NM 586184)

The Liassic exposure at Loch Slapin has long been regarded as a good example of an angular unconformity where Jurassic rocks overstep the

previously formed strata to rest directly onto Lower Palaeozoic limestones (Peach et al., 1910; Hallam, 1959).

Kalander (1974) reported a number of previously unnoticed Mesozoic rock outcrops over the Durness limestone and the Torridonian sandstone of the southwestern Strath area, and provided a full description of each locality supplemented by detailed maps. The main interest here is in the two successions reported from the Camas Malag area (NM 58701864) at his locality 14 (NG 584186). See also Steel et al. (1975, Fig. 3, locality 5).

The two successions are found as patches of very steeply dipping (up to 80°), bedded finely laminated dark grey indurated silty shales with lenses and wedges of cross bedded coarse, angular sandstones (paraconglomerates of Steel et al., 1975). The shales also contain thin (1 cm.) layers of rippled orthoquartzitic angular pebbles (Plates 6.3 and 6.4a).

Thin sections of the two parts shows the cross bedded coarse sandstones to be composed of angular polycrystalline quartz; the black indurated silty shales consist of silt-size angular quartz with wavy extinction, mica flakes are present and are aligned parallel to the fine laminations; a few echinoderm fragments are seen. Kalander (1974) reports the presence of tension fissures and minor (micro) faults cutting across the rock. The presence of aggregates of Durness limestone boulders in the "pelite" itself, as claimed by Kalander, cannot be supported here. These beds appear to form a thin cover over the Durness limestone and in most cases seem to fill the many irregularities in these lower Palaeozoic beds. The dark indurated "pelites" are regarded as being of definite Jurassic age by Kalander (1974) from echinodermal evidence and their vertical and lateral relationships with the surrounding strata.

In this locality certain outcrops are found which consist of Durness limestone boulders up to 1 m. across enclosed in a carbonate matrix which is referred to by Oates (1976) as a "grey, crinoidal micritic Lias matrix". This limestone paraconglomerate (Steel et al., 1975) overlies the Durness

limestone in some places and is seen to interfinger with the dark quartzitic silty shales. The paraconglomerates are composed of quartzitic fragments together with Durness limestone clasts in a carbonate matrix and occupy fissure systems within the Durness sequence (Steel et al., 1975).

The cross stratified, orthoquartzitic angular pebble wedges within the dark shaly siltstones show a dominant transport direction towards the south west. They appear to have been deposited concurrently with the indurated shales, as evidenced by the presence of lenticular shale clasts contained within the "orthoconglomerate" and also by the presence of micro "flame features" (Conybear and Crook, 1968), composed of finely crushed quartzite, at the interfaces of the two lithologically different beds. It can be seen that superficial layers of semiconsolidated, well crushed quartz fragments have been pulled up from the surfaces of the "orthoconglomerate", forming distorted lobes and wavy strings within the dark mudstone. Undoubtedly these structures were formed due to the drag effect of the movement of a sediment-laden current on a semi-consolidated surface. Probably while an extremely turbid immature sand mass was carrying its load down a steep slope it gradually dropped it, and while moving, this load deformed the hydroplastic to quasiliquid surface over which it moved by plucking up and streaking-out the upper layers. This process is seen to occur in flume-simulated turbidity currents (Kuenen and Menard, 1952).

The relative ages and significance of the conglomerates and dark shales are of considerable importance in this area. Steel et al. (1975) suggested that the two types of conglomerates and the shales are lateral equivalents of one another and are at least partly of Jurassic age for the following reasons:

- a. The paraconglomerates locally overlie the dark silty shales.
- b. The orthoconglomerates and paraconglomerates interfinger with the shales.

The above suggestions are supported here, based on personal observation and collection of field data. It is evident that the silty shales partly forming the matrix and surrounding the wedges of cross bedded orthoconglomerates are of definite Liassic age, however, the cross laminated orthoconglomerates themselves have a more ?Rhaetic affinity as evidenced by similar clast composition and lack of fossils.

Fine grained pebble conglomerates, the matrix of which contains echinodermal debris, can be found to overlies Triassic conglomerates only a few hundred metres to the southeast of Camas Malag (NG 585181). Here the Milton Formation is overlapped by the Strath Formation along a short interval (from NG 58401850 to NG 58881835, along the line indicated on Fig. 6.1); this shows that the Strath Formation directly overlies the Durness Limestone (at least in most parts) along the shore of Loch Slapin, and the Triassic conglomerates immediately inland. This finding differs from that of Hallam (1959), who suggested that the 'Lower Broadford Beds' (Milton Formation) "wedge out completely" towards Loch Slapin. However it is acknowledged that the Strath Formation (Upper Broadford Beds) becomes thicker towards the southwest at the expense of the Milton Formation (Lower Broadford Beds) in the Strath area.

It should be noted that the 5 metres of paraconglomerate reported to underlie the Strath Formation along the course of Allt nan Leac and to the east of the road to Suisnish (NG 58101864), is regarded by Dr. Steel as being unmappable and hard to define (pers. comm., 1975); careful examination in the field by Professor Hallam and the present author (1976) only revealed the existence of Durness limestone beds underlying the Liassic strata containing Gryphaea along the course of Allt nan Leac. No comment is added here, on Kalandar's Figure 3.21 (1974).

Probably the interstratal solution of the carbonates of the Durness limestones produced a modest palaeokarst landscape during early Liassic

times; this was preserved as "buried karst" (Quinlan, 1972) when inundated by the turbid waters of later Liassic seas. Eroded surfaces of upstanding Durness limestone "pillars" within the Liassic (Strath Formation), dark shale matrix (together with some Durness Limestone fragments) would appear as paraconglomeratic beds containing lenticular and rounded clasts.

The formation of "rock" as a result of tropical karstification (Blaszcs, 1968) together with pseudostructural fissure caves (Jackus, 1977) as a result of surface rock slides and the collapse of large blocks may also produce the features observed here.

In view of the existence of Liassic conglomerates and pebble conglomerates both at the base and top of the Milton Formation at Loch Eishort, Ob Lusa and Applecross, the existence of conglomerates at Loch Slapin is not regarded to be unusual. It will be shown that the Milton Formation thins out considerably in Loch Slapin and is partly but not completely overlapped by the Strath Formation; the silty shales with orthoquartzitic lenses at Loch Slapin may be partly related to the Breugh Pebble Member.

iv. Skulamus (NG 665205)

The nature of Torridonian/Jurassic contacts here is uncertain; on the Geological Survey 1" and 6" maps, the Liassic strata are shown to be in fault contact with the immediately underlying Torridonian sandstones. In places their relationship in the field is unconformable, the Milton Formation immediately overlying the Torridonian sandstones.

v. Loch Eishort (NG 626162)

Along the northeastern shore of Loch Eishort, lower Cambrian quartzites are, in places, directly overlain by Liassic biomicrosparites containing subangular quartzite pebbles (?dropstones) up to 3 cm. in diameter. In this area the blue limestones of Liassic age appear to be "welded" (Peach

et al., 1910) onto the quartzite.

vi. Raasay, Rudha nan Leac (NG 595385)

The basal facies of the Torr Mor Member is not represented in Raasay. These have probably been faulted-out. However the Triassic/Liassic contact at Rudha nan Leac is not faulted.

6.3 Conglomeratic Facies 1

At Torr Mor (NG 630164) a 50 cm. thick unit of coarse, poorly sorted, matrix-supported, polymodal, unstratified, ungraded conglomerate with an unordered fabric (Plate 6.5) directly overlies the ?Rhaetic orthoquartzitic sandstones and Triassic conglomerates; in places, they overlie the Cambrian quartzites exposed in this area.

These conglomerates are regarded as distinct from the underlying Triassic conglomerates for the following reasons:

- i. They do not contain clasts composed of Torridonian sandstones.
- ii. The cement contains minute bivalve and gastropod shells.
- iii. They overlie white coarse grained sandstones which are regarded as being ?Rhaetic elsewhere in the northern region.

The clasts vary from ca. 1 mm. to 9 cm. and are mainly composed of angular quartzite fragments and subangular cherts; the clast composition of this bed is given in Fig. 6.2.

Laterally these beds thin towards the north and northeast until they wedge-out at NG 639170. At Torr Mor this facies is seen to overlie 10-15 cm. of blue micritic limestone.

6.4 Pebbly Sand Facies 2

The total estimated thickness of this facies is 1-2 m.; it overlies the Conglomeratic Facies 1 and consists of pebbly sandy limestone lenses

(up to 5 cm. wide, 25-45 cm. long). These lenses are interbedded with very sandy shales 2-6 cm. thick containing occasional subangular quartzite pebbles. The sandy limestone lenses show sharp contacts with the sandy shale interbeddings and show wavy (?rippled) tops together with trough cross laminations which are in places lined with pebbles. The cross laminae are mainly straight and the surface beneath each set is non-erosional.

The cross laminae of each set are discordant with respect to the lower bedding surface and are lithologically homogeneous. These may be termed "Alpha-cross-stratification" (Allen, 1963). The pebbles which line the cross laminae are sometimes as large as 6 cm. in diameter. The outcrop of this facies was followed to the northwest laterally and at two successive stations (Fig. 6.3) palaeocurrent measurements were taken; these are represented in Figures 6.4 and 6.5 (details of all palaeocurrent data obtained together with the method followed are given in Appendix 2). Further inland, along the eastern outcrops of the basal Liassic beds, immediately overlying ?Rhaetic, Triassic or Torridonian strata (from Heast to Skulamus), this facies is represented by faintly cross laminated sandy limestones with very thin beds of rounded, sorted pebbles up to 4 cm. in diameter. The pebble composition of these beds was determined at different stations (for the whole of the Lower Liassic outcrops of Strath) and are presented in Fig. 6.6.

6.5 Calcareous Sandstone Facies 3

The estimated thickness of this facies is 2-3 m. (NG 637190). This unit is composed of faintly tabular cross laminated sandy limestone lenses with wavy sandy shale partings; the partings occasionally contain rounded quartzite pebbles (ca.4 cm. in diameter). The cross laminated beds show non-erosional wavy lower contacts with the underlying strata. The cross laminations have a discordant relationship with the lower boundaries of the beds.

The three facies of this Member are not fully represented at each of the main sections of the Milton Formation, however they may be conveniently separated from the other Members by establishing an upper limit; for this case this upper limit is represented by the Ob Lusa Coral Member in which the coral Isastrea murchisoni is present. The blue micritic limestones of the Ob Lusa Member are present at Applecross and Lusa; further inland towards Heast, the coral limestone is also present and can be observed at two localities as previously mentioned. In the Strath district, a blue micritic limestone containing the coral Isastrea murchisoni succeeds the Torr Mor Member at various localities, the base of the blue limestone was taken as the upper limit of the calcareous Facies 3 and its lateral extent followed. In general the exposure is very poor and complicated due to faulting; in many places the sequence can only tentatively be ascribed to the Torr Mor Member or, at times, even the Milton Formation, the criteria for this distinction in this latter case being an upper limit of calcareous shales and limestones containing beds of Gryphaea arcuata and a lower limit overlying white orthoquartzitic cross bedded sandstones of ?Rhaetic age (where present) and Triassic orthoconglomerates.

As mentioned previously it is difficult to establish a complete succession throughout the lateral exposures of the various pebble beds as determined and presented in Figure 6.7.

Considerable variation (except for a few cases) exists among these beds; significant lateral differences among the various outcrops are also seen in thin section. Although no attempt has been made to quantify the differences (due to uncertainties of exposure) a pronounced change is seen in terms of clast composition among these beds in Central Southern Strath from locality 6 to 14 (Fig. 6.7). It is seen that in southwest Strath (localities 10-14) this Member (and others of the Milton Formation), although distinctly different from those of the Strath Formation, becomes more compact and the

gastropod, bivalve and echinoderm content increases; lithoclasts are totally absent and also the quartz sand is mostly composed of normal rather than polycrystalline grains. Mud pellets present, are mostly elongated due to pre-lithification compaction and the oolite content is considerably diminished, while the limited amounts present are reworked and washed-in from other environments (Pl.6.6) . The carbonate content decreases toward the top of these beds and so does the amount of superficial oolites and pellets; the topmost bed is a calcareous sandstone (wackestone) with grains mostly in the fine sand range 2 - 3 ϕ .

In Applecross (NG 44757265), beds representing the conglomeratic facies erosively overlie ~ 90 cm. of greyish, silty diamicritic limestones of ?Jurassic age

This facies is represented by a 1.30 m. thick succession of pebbly calcareous sandstones alternating with fine grained friable sandstone sheets containing abundant plant remains. The pebbles are mainly up to 4 cm. in diameter and composed of pink quartz fragments, although chert is also present. The beds are 30 - 35 cm. thick, poorly sorted, show grading and undulose erosive lower contacts, the three alternating, plant-bearing friable sandstones transitionally overlie these beds and are each 15 - 20 cm. thick. These are beige-coloured, greyish weathering, laterally impersistent beds which assume a sheet-like form in places; they also contain mud pellets and iron oxide stains are common.

The total pebble and sand content decreases from the base to the top of this facies in Applecross.

6.6 Interpretation

i. Conglomeratic Facies 1

The polymict, matrix supported nature of the beds of this facies, together with the other characteristics mentioned, suggests the presence of subaqueous sediment gravity flows (Middleton and Hampton, 1973). Debris

flows and (in rare cases) grain flows being the main transporting mechanism. The existence of bivalve shells within the calcareous cement, lack of mud and scarcity of plant remains together with the progressive fining upwards ("dilution") of these beds, rules out the possibility of a subaerial origin for the conglomerates. Debris flow is taken here as meaning a rapid flowage involving clasts of various compositions in a viscous medium. Larger clasts moved in a matrix of finer particles and ?water, with the finer components (pebbles and coarse sand in bluish carbonates) probably acting as a single fluid with high viscosity. The inverse grading (Fisher, 1971) reported from such debris flows was not seen here. There is also a lack of horizontal or imbricated orientation and graded bedding (Bull, 1972). The larger clasts are distributed uniformly throughout the thickness of the flow. It is generally considered to be the product of a highly viscous flow with a laminar mode. In such masses the clasts are generally carried by a watery mud matrix. The mass moves slowly and when the downslope pull of gravity no longer exceeds the shear strength of the debris it will stop; in this manner no current structures will develop within the flows (Collinson et al., 1977). Megabreccia (Mountjoy et al., 1972) and debris flows represent marginal facies of basins bordered by carbonate banks and shelves or reef complexes. The occurrence of these deposits in the ancient record has been summarised by Mountjoy et al. (1972), Cook et al. (1972) and Srivastava et al. (1972). Some of the characteristics of these beds, as given by Mountjoy et al. (1972) are as follows:

- a. They occur immediately adjacent to carbonate (or any other) buildups.
- b. They form thin sheets (3 - 12 m. thick) and some lenses and channels.
- c. Most coarse clasts are subangular or angular.
- d. The matrix around the clasts is dark lime mud containing intermixed microfauna (and lamellibranchs).
- e. Clast size variation is between silt and cobbles 15 - 20 cm. across and are subangular to angular.

- f. Clasts were somehow mixed with lime mud and silt.
- g. The deposits were derived and transported under subaqueous shallow marine conditions.
- h. Although palaeoslopes cannot be reconstructed for the Liassic debris flow conglomerates, Mountjoy et al, (1972) found that in such beds transportation was as follows:
 - a. Clasts were transported down slopes of very low angles ($1-3^{\circ}$ or less); these were derived from buildups or banks with low relief. There is only circumstantial evidence for the existence of low angle slopes during early Liassic times, nevertheless, the above evidence is regarded to be sufficient.
 - b. The clasts were transported considerable distances into a shallow marine basin containing lime mud and very fine grained limestones.
 - c. The clasts were transported as a single event (indicated by the lack of bedding, stratification etc.).

The main genetic problems of studying such deposits was adequately discussed by Mountjoy et al. (1972) a summary of which follows:

1. Detachment of clasts

Subangular and variable sized clasts suggest the lack of prolonged reworking and relative proximity of the source. Several mechanisms have been suggested for clast detachment, e.g. earthquake shocks, storm waves, gravity acting on an oversteepened bank margin. Overburden created by the movement of large reptiles may also contribute towards the detachment of loose blocks, so are the simple mechanical erosional effects of day versus night temperature differences.

It is not possible to determine which of the above was the main contributor and/or if a combination of mechanisms was involved. It is probable that the detachment occurred due to earthquakes acting in concert with the effect of gravity (together with other factors). This effectively

created an instability of the seaward margin of the upstanding quartzite and Durness Limestone blocks which existed during the Trias in central Skye (Steel, 1974; Steel and Wilson, 1975) and which may have extended into the Lias.

2. Initiation of movement

The initiation of movement in submarine masses is poorly understood. The allochthonous deposits must represent submarine mass movements responding to gravity forces which were responsible for the mixing and transport of the debris somehow. Factors which reduce either the shear resistance between particles or blocks and the substrate or the concentration of solids will help to initiate movement (Cook et al., 1972). Important factors in this case are, increased pore pressure, instability of buildup margins, flooding, storm or wave activity, tsunamis (Coleman, 1968) and earthquakes.

Submarine gravity movements were divided into four categories by Dott (1963). A more theoretical interpretation of the transport mechanism of these sediments is also given by Bagnold (1954b, 1956, 1965, 1968). The beds of Facies 1 represent a type of "mass flow" as evidenced by:

- a. Disarticulated clasts surrounded by calcareous matrix (paraconglomerate texture) are seen.
- b. Wide variation in size of clasts exists.
- c. Boulders and blocks are predominantly angular to subangular.
- d. Complete mixing of variety of clast types during transportation has occurred.
- e. Lack of stratification and sorting.
- f. No disruption of underlying strata was observed.

3. Mode of deposition

The deposition of these beds took place as a unit rather than as a gradual upward deposition, as indicated by the virtual lack of traction or internal shearing features within the beds and lack of normal grading,

parallel laminations or current ripple lamination that could be formed when turbidity currents gradually deposit thin debris. This would be characteristic of deposition from a grain flow in the viscous regime. The beds of Facies 1 are massive, poorly sorted and always non-graded, all of which are suggestive of deposition as a single sedimentation unit.

Features observed in these submarine debris flow deposits are also characteristic of viscous, subaerial mud flows (Blackwelder, 1928; Shreve, 1968, Broscoe and Thompson, 1969), but the presence of disarticulated bivalve shells strongly rules out this possibility. Johnson (1970) suggests that in such deposits the flow is laminar and the shear stress within the flow is very low.

Although the lack of fine grained clayey material in the beds suggests that "olistostromes" (Abbate et al., 1970) are not present, the obvious age difference between the matrix (Lias) and clasts (derived from Cambrian quartzites and limestones) remains and justifies the use of this term. Northwest pointing palaeocurrent data (scarce) together with thinning of these beds towards the north and northwest provide a picture in which a viscous clastic body built its way subaqueously (to the west and northwest) into a shallow basin while progressively losing its "shear" strength and larger clasts due to the aqueous dilution effects, becoming finer grained. Lack of adequate exposure precludes the further study of the morphological features of these beds. Comparisons may be made with alluvial fan deposits (Davis, 1925; Blissenbach, 1954; Allen, 1965b; Hook, 1967, 1972; Wasson, 1974), the abundant debris supply being provided by the rapid dissection of highland areas which were separated from the lowlands by a break in slope (probably a fault scarp). Ephemeral flooding due to flash storms may have been the main transporting agent, carrying the sediments on slopes which were unable to support much vegetation.

Alluvial fans are produced and preserved as marginal facies of the basin of deposition (Reineck and Singh, 1973). Conglomerates of this facies

are not found elsewhere in the northern or southern areas of study and only Facies 2 and 3 are well developed and represented (Fig. 6.8). This isolated occurrence of facies 1 type deposits eliminates the possibility of extensive alluvial fan building processes operating during early Liassic times in northwest Scotland. In the Skye area sediment was carried from source areas in the east, northeast and north along gentle slopes into a main depocentre situated off northwest Raasay. This uniform pattern was interrupted by the existence of relatively small, isolated but resistant blocks of upstanding Precambrian quartzite (to the east) and Durness Limestone (centre), the rapid erosion of which produced channels that were subsequently infilled by minor subaqueous debris flows. Outcrops of Torridonian sandstones did not contribute much sediment to the depocentres (see later sections).

Immediately underlying the conglomerates of the Torr Mor Member the existence of whitish limestones with a pelleted mudstone microfabric (of ?birdseye origin) together with a very limited fauna suggests the formation of very shallow, cut-off ponds and lagoons in which very fine muddy sediment accumulated (Wilson, 1975). These sediments are commonly pelleted with rounded micritic grains showing internal cross-partitioned structures of ?algal origin which suggests derivation from environments where algae thrived; the fenestrules (?birdseyes) contain crystalline calcite with a void-filling pattern which may have resulted from their desiccation (Shinn, 1969) best found well developed in areas of tidal flats where persistent inundation and exposure are experienced (Fischer, 1964). The absence of any other indication for the existence of tidal flats, the obvious features being stromatolitic structures, mudcracks, linsen and flaser bedding with megaripples, herringbone cross stratification etc., (Reineck and Singh, 1973; Raaf and Boersma, 1971; Reineck, 1972) rules out the above possibility. As these beds overlie those of the Rhaetic and underlie definite Liassic conglomerates, their stratigraphic position may be

tentatively chosen as that of ?infra Liassic; these beds probably form the main source for abundant rounded brown micritic pellets which are found in the succeeding strata of Facies 2 and 3. These Facies are represented by silty limestones and calcareous shales in eastern Raasay, and contain a minimum content of pebbles. The existence of outcrops of Durness Limestone during Lower Liassic times provided an ample source of CaCO_3 which formed the matrix of the strata of Facies 1; this also contributed to the micrite content of small, shallow subaqueous coastal ponds into which coarse, poorly sorted quartzitic fragments built, forming paraconglomeratic masses. Periodic, short-lived phases of instability in the hinterland (possibly in the form of earthquakes) initiated the movement of these debris-laden paraconglomeratic masses which while responding to gravity forces, moved down relatively low slopes and built-out towards the basin of deposition.

The paraconglomerates of Facies 1 abruptly become calcareous pebbly sandstones, shaly sandstones with minor pebbles of quartzite, chert and limestone towards their tops and adjacent to Facies 2 and 3. The lenses of calcareous shale which become apparent towards the top of Facies 2 and 3, are seen mainly in areas where beds of these facies directly underlie those of Facies 3 and 4 (cf. Applecross). The dominantly clastic sediments were deposited by northwest and westward flowing currents in a belt extending eastward (inland) from the edge of palaeal and semi-enclosed lagoonal areas near the sea to a relatively higher hinterland and in places (e.g. Applecross) beds of facies 2 and 3 can be seen to underlie the ?lagoonal Facies 4 and 5; here they overlies Triassic-Rhaetic strata whereas in Torr Mor they overlies Cambrian quartzites. The pebbly cross laminated sandstones of Facies 2 and 3 resemble "coarse member deposits" reported from ancient alluvial sediments (reviewed by Collinson et al., 1977), but the presence of echinodermal debris etc. rules out any possible subaerial transportation and deposition. The sedimentation pattern of Facies 2 and 3 clearly indicates deposition under

highly turbulent conditions and deeper normal wave base which, together with the presence of reworked open marine fauna indicates transportation under turbulent, wave dominated conditions which mixed the faunas derived from further offshore positions with clastic sediments derived from near-shore areas.

ii. Facies 2 and 3

Overlying the conglomeratic beds of Facies 1, the presence of sandy limestone lenses showing alpha cross stratification (see Allen, 1963 for review) may indicate that these units were constructed in shallow water by the building of solitary banks with "straight or curving leading edges above slip-off faces". This mode of origin can be seen to occur at the present day. In modern braided rivers solitary banks are common, they also form in estuaries, on beaches and in shallows just off beaches (Allen, 1963). The presence of very thin impersistent beds of rounded quartzitic pebbles within the sandy limestone may reflect conditions of intermittent uplift and erosion operating in the hinterland, providing the depocentres with a non-permanent source of supply. The influence of wave motion in sands may be inferred by the recognition of long-crested ripples with rounded profiles.

The formation of combined flow ripples and wave ripples of Facies 3 indicates that the sandy limestones formed in an environment where currents and waves coexisted (Harms et al., 1975). As a response to an imposed oscillatory motion, the ripple crests have become longer, more sinuous and the profiles are rounder, though still asymmetric as compared to current ripples (Harms, 1969). The stratification produced by such ripples would be inclined in the downcurrent direction, the laminae are tangential to lower bounding surfaces and are not distinctly trough-shaped. Also, asymmetric, rounded ripples with a consistent migration-direction develop under shoaling wave action, wave ripples indicate that wave motion related to wave oscillation is dominant and current drift is negligible (Harms

et al., 1975). Regarding the dynamics of transport and deposition, the incipiently tangential contacts of foreset laminae suggest relatively higher transport velocities and bed shear stress, with a greater proportion of the particles carried in suspension, beyond the lee surface and deposited in the form of bottomset and topset (Jopling, 1965b; Reineck and Singh, 1973). The existence of tangential units with gentle dips suggests a high "depth ratio" (ratio of the depth of stream flow to the depth of the basin of deposition (Jopling, 1963, 1965b)); the indices of foreset laminae (Jopling, 1966) are useful guides in demonstrating the relatively high current strengths which were operative during the deposition of Facies 2 and 3.

- a. Maximum angle of dip of foreset laminae is less than the static angle of repose (about 20°).
- b. The character of contact between foreset and bottomset is mostly a tangential shape.
- c. The foreset laminae are less sharp and distinct, i.e. there are little textural contrasts between adjacent laminae.

Although conclusive evidence is lacking, Facies 2 and 3 may be regarded as representing the development of an incomplete submarine discharge system, building out into very shallow ?off-beach environments of an open-circulation shelf-lagoon (Facies 4). (See also Fig. 6.9).

The result of pebble composition determination and counts for these facies shows that the clasts have their origin in very local relief. However, this contrasts with their high degree of roundness maturity

which suggests distant source areas or, alternatively, longer periods of reworking. The latter explanation is preferred. It is likely that the pebbles represent reworked redistributed deposits of previously formed facies and may be termed "relict" (Swift et al., 1971). The occurrence of quartzitic pebbles (and pink quartz) together with chert fragments in the underlying facies, indicates derivation from exposed sources with quartzitic

sands and cherts. Large pebbles form a relatively small proportion of these beds, therefore the term "conglomerate" cannot be applied to them. Of the various pebbles found in the Breugh Pebble Member, only a few may be recognised in these beds.

1. Quartz

a. Metamorphic

Some of the pebbles found in these beds show a quartzose mosaic with a simple equiangular texture, with the uniform extinction typical of quartz. This pattern was referred to by Macgregor (1952) as a "tessellate mosaic". Most however, show elongated sutured grain boundaries and undulose extinction of quartz. These features were shown by Macgregor (1952) to be characteristic of the Moine thrust belt and the Core Schists of the Morar Anticline. Quartz pebbles with elongated sutured grains are very common in the ?Rhaetic beds of this area as well; the presence of such pebbles in the sandstones of the Great Estuarine Series is regarded as good evidence that Moinian Schists were exposed in the source area (Hudson, 1964). The existence of such a source area is also postulated during the Lower Lias (?Rhaeto-Hettangian) times in light of the evidence given above. Some pebbles with even-grained mosaics are also found. Pure quartzites are more common in the Dalradian than in the Moinian (Hudson, 1964). The evidence for this is not conclusive and Phemister (1960) mentions that quartz veins in the Moinian may have a deceptive similarity to quartzites. In the absence of accessory muscovite, zircon, epidote and apatite the true origin of these pebbles may not be recognisable. The intense elongate suturing suggests high-stress metamorphism.

b. Orthoquartzites

Pebbles with rounded quartz grains of a recognisably detrital origin and variable size are seen; silica cementation has not occurred and the outlines of grains are clearly visible. These pebbles resemble the Basal

Cambrian of northwest Scotland and may have been derived locally from their outcrops.

c. Vein quartz

The quartz in these pebbles contain bubbles (Hudson, 1964, first reported such ?fluid inclusions in the quartz pebbles found in the Great Estuarine Series) and are strongly deformed as evidenced by undulose extinction and granulation of the crystal boundaries. The presence of these pebbles in the beds is somewhat insignificant as far as the determination of source is concerned, since they may have come from any of the Scottish metamorphic rocks previously exposed in the area.

2. Chert

These pebbles are mostly dark grey and appear as being very curious. In thin section their main constituent is pale brown granulo-chalcedony (Smith, 1960) with an apparent grain size of 5-10 μm ; within this cement are areas of finer grained and darker chalcedony which may represent replaced pellet and oolitic structures. Some of these cherts may be compared with those of the Durness Limestone and those reported from the Great Estuarine Series (Hudson, 1964) which were supplied by the same source material. The chert pebbles are almost certainly of silicified pellet (?oolitic) limestone. The pellets are of fine grained brown chalcedony in a matrix of clearer chalcedony which coarsens away from the pellets. This phenomenon has been observed in the cherts of the Great Estuarine Series (Hudson, 1964), the conclusion being that the pellets were silicified before cementation; this texture is reported to occur in parts of an oolitic chert from Division I of the Durness limestone (Hudson, 1964). Many other textures have been identified in the cherts of the various divisions of the Durness limestone (Hudson, 1964; Kalandar, 1974). Their occurrence in the Broadford Beds Groups has not been met with and although

direct comparison has proved to be difficult, the Durness Limestone seems to be the most probable source for the cherts found in the beds representing Facies 2 and 3. The conglomerates of the Torridonian also contain a significant proportion of chert (Teall, 1907) and may have been a possible source, but the absence of reddish chert (jasper) which commonly represents the cherts of the Torridonian, together with the lack of pink quartz and reworked, rounded, red arkoses (typical of Torridonian arkosic sandstone pebbles found in the Triassic conglomerates underlying the Lower Liassic beds) in the beds representing this facies, indicates either subdued Torridonian topography during this time or the total disintegration of the arkosic pebbles prior to their deposition. The former explanation is favoured here since although the disintegration of the arkosic fragments would provide the necessary detrital quartz fragments, it would also release a considerable proportion of reddish ?iron oxide pigmentation of the beds together with pink quartz and jasper which would have remained within the basin along with the other constituents.

3. Source

The existence of the Moine rocks (metamorphic source) of the Northern Highlands as a source area is indicated for most of the pebbles, with major contributions of orthoquartzites which may have come from the Cambro-Ordovician rocks. Cherts from a Cambro-Ordovician (?Durness) source area, although a minor constituent within these beds, significantly points to the existence of Durness source rocks. No Lewisian and/or Torridonian fragments were seen in these beds and the rocks which do occur, are medium grained, siliceous and well cemented, i.e. possess properties which would render them resistant to chemical weathering. As pointed out by Hudson (1964), this may explain the absence of the Lewisian but cannot explain the surprising absence of the Torridonian clasts; the arkoses were most probably broken down during the process of weathering.

6.7 Applecross House Member

Two Facies are recognised within this Member:

- i. Shaly Facies 4
- ii. Oolitic Facies 5

i. Shaly Facies 4

The total thickness of this facies is 3.70 m., its lower 2.30 m. consists of dark grey sandy fissile shales which are partly altered at their base by the intrusion of a sill. These beds are very finely laminated in their topmost parts and they contain a high proportion of muscovite. The CaCO_3 content is moderate and as shown in Fig. 6.10 there is a high proportion of smectites present among the clay minerals. When fresh, the topmost parts of these beds is olive-green in colour and appears to be very plastic. The top parts are also laminated, with the laminae composed of layers of muscovite-rich silt, high in CaCO_3 content. These layers laterally grade into lensoid, micro concretionary-like masses composed of muscovite, silt and calcite. Broken-down, coalified material is abundant, especially in the topmost 20 cm. layer where shaly micaceous siltstone layers underlie the immediately succeeding sandstones.

They also contain minute, broken-up ostracod shells. At the 1.40 m. level, a 70 cm. thick bed of shelly limestone occurs, with several thin silt interlayers showing undulating contacts. This is a matrix-supported silty biopelsparite (up to 47% quartz), with plagioclase feldspars. Glauconite is present but not very common, and bivalve shells are common.

Above this band occurs a greenish-yellow, plastic clay unit about 40 cm. thick, which contains a thin bed of black calcareous shale containing remnants and fragments of thick-shelled bivalves; it contains less mica than the surrounding shales and occurs as horizontal, wavy forms within the plastic clays. Succeeding the shales is a 15 cm. thick bed of limestone followed by

another bed of clay of the same characteristics as the previous one. These beds are followed by a shaly calcareous siltstone containing abundant coalified material; their contact is undulose. The relationships described above are shown in Figure 6.11. It is also interesting to compare the clay mineralogy.

The smectite content markedly decreases in the dark, flame-like features, and the kaolinite content becomes significant. The opposite is true for the plastic clays, both smectite and kaolinite form a low percentage of the topmost shaly calcareous siltstones.

The topmost beds of this facies are 1.60 m. thick, and comprise massive calcareous sandstones with somewhat large sub-rounded quartz pebbles (1-1.5 cm. in diameter) on their top surface. The limited exposure of this bed along the stream does not allow a detailed description and only wavy, low-amplitude sheet-like features were seen. In composition, the bed is a silty bio-oolimicrosparite packstone with < 14% silt (3,4,5 ϕ). It is almost entirely composed of broken, rounded bivalve shells with a thin micrite coating. Asymmetric oololiths are also present together with rounded limestone pebbles and geopetal structures are common.

This facies contains a fauna of Cardinia sp. and thick shelled Liostrea are seen in the shales.

ii. Oolitic Facies 5

In Applecross, the total thickness of these beds is 6.40 m. (Hallam's beds 6, 7, 8). At its base this facies consists of 1.60 m. of dark olive-grey brownish weathering, faintly laminated clays containing a moderate proportion of CaCO_3 (15-20%) and a minimum sand content. These are followed by a 4.70 m. development of oolitic limestones which are entirely composed of shell fragments in places. At the base of this unit a 0.50 cm. thick bed crowded with Liostrea sp. shells was reported by Oates (1976). The oolitic beds are grey and indistinct in the field, but some differences among the successive beds may be distinguished in thin-section. The basal parts

(7.90 m. above the Lias Trias contact, Plate 6.7) is a poorly sorted (bimodal) oomicrosparite packstone/wackestone. The matrix is microsparitic and does not contain opaque material. The ooliths are perfectly rounded (3ϕ maximum diameter, 4ϕ minimum diameter) and composed of micrite, although no concentric or radial structure is obvious they consist of a nucleus of quartz (mostly strained) or broken shell fragments (mainly very thin and delicate bivalve shells). Commonly second generation ooliths are seen; these contain a perfect oolith (Carozzi, 1960) with an indistinct nucleus but definite concentric and radial structures surrounded by a thick, round coating of micrite. Evidence of compaction is seen and veining is common.

Beds 8.50 m. above the Trias/Lias junction (bed 6 of Hallam, 1959) at the lowest part of Allt nan Breugh (NG7195437), are 1.40 cm. thick and composed of bluish-grey oolitic limestones with a beige undulating, weathered top surface.

Abundant calcite-filled hair-thin veins are also seen. Beds are up to 25 cm. thick and in thin section the lower parts are matrix supported bio-oosparites with the ooliths mostly in the size range of $<2\phi$ and with strained quartz nuclei. Most ooliths show concentric banding along which ?coalified and other opaque material has been lined. Quartz fragments commonly show superficial oolitic coatings and are sometimes clustered with others to form multiple ooliths (?grapestones). Most ooliths have a second generation origin and echinodermal debris is commonly scattered throughout the rock; compaction is not uniform. Laterally the composition of these limestones varies considerably. A sample (Apc 2) taken from these beds is an oobiomicrite/sparite, the ooliths occur in isolated pockets which contain a microsparite cement, whereas elsewhere the limestone matrix is micritic with no sparite present. The ooliths show both concentric and radial structures and the concentric laminations show evidence of borings (?algal). Ostracods are present and the ooliths are mostly in the size

range of $< 2\phi$ (Plate 6. 8).

In places these beds are of an entirely different character (lateral variation within 2.3 m.). They become silty oobiopelmicrosparites with quartz fragments ($< 4\phi$) and the bioclasts mainly 3.2ϕ . High spired gastropods are abundant and echinodermal debris is also seen together with some ?algal material (Plate 6. 9). As the higher parts of this unit are reached (Hallam's bed 7), they become matrix-supported biosparites (1.25 m. measured thickness), composed of totally recrystallised, large rounded bivalve shells (up to 1 cm. long) and echinoderm debris; these are mostly lined with opaque ?carbonaceous and coalified material most of which is scattered throughout the matrix. Gastropods with micritic fillings are also seen.

The topmost 1.80 m. of the oolitic facies in Applecross (Hallam's bed 8) is a regularly bedded (15-25 cm.) oolitic limestone with undulose contacts (3-5 cms. amplitude). In their lowest parts, they are normally packed oobiosparites (matrix-supported) with 80% superficial ooliths, their nuclei being mostly bivalve shell material and a few whole gastropods. Most of the quartz fragments are present as polycrystalline grains; some micritic patches are present containing ooliths which suggest an original oobiomicrite.

Size ranges are $2-1\phi$ and in some higher parts within this facies (Apc 6) these are partially altered oobiopelmicrosparites (normally packed, matrix-supported). The matrix is again micritic in patches, the ooliths are concentrically laminated and are mainly 2ϕ in diameter. It should be noted (Plate 6.10). here that the ooliths within the micritic patches only show radial structures./

The topmost beds of this facies are matrix-supported oobiopelsparites (Plate 6.11). Most of the bioclasts are totally recrystallised and are coated with a micritic rim (Plate 6.12).

Although not fully represented, the equivalents of these two facies are also found in Raasay. Hallam (1959) recognised a variety of lithologies representing the Milton Formation (Lower Broadford Beds) in the bay to

the northwest of Rubha nan Leac (NG 594386). Careful examination by the present author suggests that the lowermost parts of this Formation reported by Hallam (1959) as beds 1, 2, 3, 4 together with beds 8 and 9 underlie a 6.90 m. development of nodular limestone. The oolitic bed (bed 11 of Hallam, 1959) directly overlies this nodular limestone/shale succession and in thin section resembles the basal parts of the oolitic facies of Applecross. This oolitic bed is in turn overlain by a 70 cm. thickness of dark indurated calcareous shales with 5-10 cm. limestones with echinoid spines.

The beds overlying these are nodular limestones with thin shale partings. These are exposed only in the lower cliff (NG 59503815) where their total thickness is estimated to be 7.10 m. (the topmost part is obscured and definitely faulted). The shore section here is faulted and covered by large fallen blocks, making it impossible to draw up a complete section. Beds 14, 15 and 16 of Hallam (1959) were not seen in the field and a 6.90 m. gap separates the nodular limestones from the overlying sandy shales which contain beds of Gryphaea.

In Raasay the shaly facies of the Applecross House Member is probably represented in the shales and sandy limestones of the lower 3.40 m. of the Liassic exposure immediately overlying the Triassic conglomerates in Rubha nan Leac. The lowermost beds here are composed of brown-weathering fine-grained dark coloured, veined echinodermal biomicrites with some brachiopod remains (2.30 m.) and with 1-1.5 cm. thick sandy shale partings containing Liostrea. These partings are undulose and their 10-20 cm. spacing becomes less as the higher parts of these beds are reached. In thin section, the limestones become poorly sorted silty biomicrosparite packstones towards the top of the sequence and a high proportion of ostracod shells together with gastropods is seen.

Elongate, rounded bivalve shells with thin micrite veins are very abundant. This pattern is more or less repeated above the sill (Hallam's beds 1 and 2), but a well sorted calcareous sandstone (3φ) overlies the

echinodermal biomcirites here. At the 7.50 m. level a 10 cm. silty biopelsparite bed with fragments of inoceramids and glauconite is seen to occur within a 1 m. thickness of fining upwards silty micaceous shales.

One metre of baked calcareous shales containing abundant oysters together with beds of broken shell material are seen at the 10.70 m. level (Hallam's bed 4). The CaCO_3 content of these shales increases towards the top (bed 8 of Hallam) and they are overlain by a massive limestone which is sandy and contains occasional irregular shaly partings which become abundant towards the top of this 1.70 m. interval (bed 9 of Hallam). The limestone is a silty biopelsparite and contains fragments of echinoderm plates; bivalve shells are mostly altered and can only be recognised by their remaining micritic coating which in most cases contain opaque materials and patches of ?organic origin. These beds are overlain by 20 cm. of calcareous shales with abundant fragments and whole shells of pectinids and oysters.

The upper part of this facies in Raasay is represented by 6.90 m. of thin to medium bedded nodular limestones alternating with calcareous shales. It is apparent that elongate, cylinder-shaped burrows (?Thalassinoides) which branch in many cases, have contributed to the formation of this sequence (Plates 6.13, 6.14). The nodules appear to be horizontal tubes (4-5 cm. in diameter), and show the following characteristics:

- i. they radiate outwards from a central, slightly enlarged sphere 5-8 cm. in diameter.
- ii. they have bifurcating arms.

The tubes are not internally laminated and are composed of calcilutites and crushed shell debris. The interbedded shales contain Liostrea and other small bivalves. These and the nodular limestone beds contain appreciable amounts of sulphur (?jarosite) although pyrite is not common in them.

The nodular limestone is present up to the base of the succeeding facies. These contain the least amount of quartz grains and pebbles are altogether

absent.

a. Raasay

At the base, this facies is represented by 1.90 m. of dark bluish weathering oolitic limestones. These are matrix-supported oolites with a bimodal size distribution. The largest oolites are .5-8 mm. in diameter whereas the smaller sized oolites, together with the micritic pellets do not exceed .2-3 mm.; most to all of these are composed of micritic mud, their nuclei being broken bivalve shells, gastropods and rare echinoderm fragments. Few multiple oolites are also seen. The oolites show a second generation origin since most of the nuclei show inner brownish organic remains whereas their outer coatings are of "clean" micrite without other inclusions. Evidence of pressure solution is also seen and definite lamination is lacking in the oolitic coatings. This bed is only found along the shore in Rubha nan Leac (bed 11 of Hallam, 1959) and may be traced into the lower cliff where it clearly overlies the nodular limestones of the previous facies. Overlying this oolitic bed is a 70 cm. thickness of calcareous shale, the CaCO_3 content of which increases towards its top where it is overlain by 7.10 m. of compact nodular limestone with shale partings. The nodular limestones overlying the oolitic bed are different from those underlying them, being compact and closely spaced. Their morphology is consequently less readily recognisable. The limestones show no internal structures and are petrographically similar to the calcilutites of the lower nodular beds. Although no systematic determination was carried out for sulphur, its presence in the shales and calcilutites of this facies is easy to recognise in the field due to the strong odour released by the freshly broken specimens. The shales of this facies contain up to 25% smectites and their clay mineral content is similar to those found in the same facies in Applecross.

It should be noted here that the variety of lithologies representing the Milton Formation in other parts of the northern study area are not represented here in Raasay. Beds 14, 15 and 16 of Hallam (1959) cannot be

recognised in the field and all fall within the nodular limestone group. It is possible that higher Members are also represented in parts of the nodular limestone succession. On the other hand in view of the faults present, it is possible that the representatives of other facies have been faulted and are now covered by scree and vegetation, creating the 6.90 m. gap between the Milton and Strath Formations here.

b. Northern Strath (Ob Lusa)

The beds of this facies are exposed above the greenish weathering ?infra Liassic beds at Ob Lusa, and are considerably thinner than their counterparts exposed elsewhere in this northern region. Their relation to this facies is inferred due to their occurrence beneath the Lusa Coral Member.

At their base they occur as a grey creamish-green weathering, thin bedded (5-10 cm.) silty marlstone. Thalassinoides are seen at different stages of their development. Petrographically these are matrix-supported silty sparites with at least 40% subrounded, sorted quartz fragments. The quartz are mostly strained and the low feldspar content is mainly plagioclase. The bed thickness is at least 1.20 m. and the succeeding 35-40 cm. bed is a well indurated calcilutite with an undulose lower contact. It is a matrix-supported, silty oobiopelsparite, the sorting of which is not complete but bimodal with one mode (60%) occurring at the 2.2ϕ diameter range and another (40%) which mostly comprises the pellets and ooliths, occurs around the 4ϕ diameter range. Asymmetric ooliths are common (Plate 6.15) and consist of both concentric and radial structures. Quartz pebbles are very rare but polycrystalline quartz grains are present together with well rounded epidote grains.

c. Loch Eishort

The lower and upper limits of the Applecross House Member cannot be

clearly defined at this locality. The Lusa Coral Bed (Hallam, 1959) which serves as a local marker horizon, is absent and no other faunal correlation may be drawn. The calcareous shale beds are also absent and petrographically these beds are related to the oolitic facies.

The lowermost 1.20 m. (Hallam's bed 1) of this Member forms a series of small ledges under a prominent sill which is exposed along the first promontory (shown in Fig. 6.12). These beds can only be examined at low tide and their lower contact with the Torr Mor Member cannot be seen.

The beds here are a series of interbedded, silty micaceous biosparites and silty biomicrites. The calcarenite beds are mostly in the 4 ϕ range and are matrix-supported, containing echinodermal fragments together with opaque material which often forms thin shreds and brownish rims around the quartz grains- the quartz is moderately sorted and subrounded. The biomicrites contain 6% silt particles and are veined (hair-thin), they contain thin brown, ostracod shells. Patches of sparry calcite have developed in these and give it a "birds-eye" appearance (Plate 6.16).

Above the prominent sill, 6.80 m. of oolitic limestone and calcareous marlstones are seen, at the base of which abundant quartz pebbles occur (Hallam's bed 2). At their base, these beds are silty biopelsparites, the quartz grains are in the 2 ϕ diameter range and comprise less than 4% of the total and are mainly strained.

In many cases superficial ooliths are formed due to the thin micritic cover of these grains. Shell fragments are reworked and show a superficial oolitic rim. Ooliths do not show a definite nuclei and are mainly made of concentric laminations although radial lamination is also present.

At the 4.70 m. level a distinctive, dark blue calcilutite bed with recrystallised delicate shells is seen which shows sparry calcite patches and has a "birds-eye" appearance. In thin section, these are seen as oo(bio)pel-micrites with abundant patches of sparry calcite (Plate 6.17). Large recrystallised bivalve shells are scattered throughout the rock and the size variation of the ooliths is not great (0.2 - 0.3 mm.). These show both concentric and radial laminae. Most ooliths consist of a large centre

(shell fragment or quartz) around which two or more "oolitic" layers are arranged. In places the matrix is microsparite and the micrite appears to be isolated clusters of small pellets.

This unit becomes silty towards its top where it is a silty limestone with shaly partings.

6.8 Interpretation of the Applecross House Member

i. Facies 4

This facies represents a period of deposition in a semi-enclosed, open circulation, shallow, coastal lagoon, under mixed (cal m./agitated) energy conditions. Examples of modern coastal lagoons are widely described in the literature (Warne, 1971; Reineck and Singh, 1972; Phleger and Ayala — Castanares, 1971; Oertel, 1973). In these the several aspects of lagoonal sedimentation control are considered to be influential for the development of the numerous depositional units (e.g. barrier beaches, tidal inlets, tidal delta, tidal channels, ponds, tidal gullies ... etc.). This is not the case with ancient sediments where only parts of the unit may be recognised and described. The dark shales with thin intercalations of silt and muscovite which frequently form micro-lensoid features in this facies are comparable to what Masters (1967) has described from the Mesaverde Group shales of the U.S.A.

It is not possible to derive environmental evidence from the bivalves (Parallelodon hettangiensis and Antiquilima), for their salinity tolerance and life habits are not well known (see Chapter 5), but their epifaunal, attached, suspension feeding habits may be an indicator of the existence of firm substrates and calm waters with an adequate nutrient content.

The relatively high illite content of the more plastic, olive-green coloured clays indicates a constant marine influence whereas the flame-like, dark kaolinite-rich shale bodies together with abundant coalified driftwood indicate deposition from the erosion of nearby lands (see Chapter 4).

The occurrence of a 70 cm. thick bed of shelly limestone with silt interlayers and occasional quartz pebbles on the top surfaces of the beds amongst the shales may represent coastal dune and beach sand, reworked in a shallow marine environment, mixed with broken rounded shell fragments and deposited as thin, lenticular beds. Such beds represent sediment bodies which rapidly migrated into the sea, prograding the coast and building-out into deeper water. Such deposits have been described in Recent environments (Shinn, 1973C) and have also been reported from the Lower Ordovician Platform composed of shallow shelf and intertidal deposits surrounding the Cambrian shield of North America (Wilson, 1975). Although this may be true for the lower sandy beds of this facies it should be noted that the significant points considered by Shinn (1973C) for distinguishing sediment accretion along the southeast coast of Qatar peninsula of the Persian Gulf are as follows:

- a. Absence of freshwater fauna.
- b. Absence of clay layers.
- c. Steep planar accretion dips which show similarity to aeolian bedding and have a uniform dip direction.
- d. Upward increase in grain size.
- e. Lack of channel deposits.
- f. At the foot of the accretion slopes where the exotic quartz sands mix with indigenous carbonate muds, extensive burrowing especially by echinoids occurs, also drifted seaweed and other marine detritus accumulates.

It is evident that not all of the above criteria can be recognised in the beds of this facies, furthermore the calcareous sandstone beds are found in dark shales and not "marine carbonates". Thus although a similar process as that proposed above may be invoked for the silty calcareous bed of Facies 4, a model depicting the progradation of a carbonate coast by migrating

quartzitic sand dunes which were moving into the sea, partially but not wholly satisfies the evidence found in the rocks.

Bioclastic and lithoclastic packstones and wackestones containing organic debris have been reported to form due to growth on top of slopes and mounds (Wilson, 1975), thence accumulating in wavy, foreset beds on depositional slopes. Much of such bioclastic debris are broken and coated; geopetals are common and sediments of various generations of deposition and consolidation are intermixed with finer, exotic debris. Considering the association of facies 4, 5 and 6, the beds of this facies may be considered as the 'flanking beds' (Wilson, 1975) of small coral groupings ("mounds") of Facies 6. Although the arial arrangement of the Facies (4, 5, 6) cannot be determined their vertical arrangement suggests that they are closely comparable to those described from around the swells formed by salt-dome islands in the Persian Gulf; three types of bathymetric highs and their related sediment distribution pattern have been described by Purser (1973), of which the "intermediate homocline highs" show a sediment distribution pattern comparable to that of the association of Facies 4, 5 and 6.

The "intermediate highs" vary in diameter from several hundred metres to more than 50 kms. They are either submerged below wave base, show crests which remain temporarily above the wave base or are totally emergent. The common feature among these is that they are surrounded by relatively shallow water, varying in depth from 36 m. on the outer parts of the "homocline" to less than 9 m. near the coast. Well developed "reefs" fringe the permanently "emergent highs" (on their windward side) and sediment produced by the continual breakdown of these will be distributed downwind as a series of "sediment tails" which accrete to the "highs" and probably begin to form during times of lower sea level (Purser, 1973); these may be traced some 10 kms. downwind in the Persian Gulf.

Although synchronous deposition is difficult to ascertain, the fine sandy limestones (Am 25) containing broken shell debris, alternating with

micritic limestones with "birdseye" features, ?calcispheres and very fine grained ? ostracod shells (Am 26) exposed at the base of the succession in Loch Eishort (NG 615162) represent the shallower water equivalents of this facies in Applecross.

The intertidal (and supratidal) origin of "birdseye" voids (Loferites, fenestral pores) was proposed by Fischer (1964) and confirmed by Shinn (1968b). Although a frame building agent for the birdseye limestones cannot be recognised here, the fine shells and other material are set in a matrix of micrite with planar and spherical birdseye structures (fenestral structures) filled with sparry calcite. The birdseye limestones (Laporte, 1967), Loferites (Fischer, 1964) and laminoid fenestral fabrics of Tebbut et al. (1965) are considered partly synonymous, and intergradational fabrics occur in parts of the lower limestones of Facies 4.

It is known that vugs form during the dessication of supratidal carbonate muds and gas bubble migration (Shinn, 1968b); the birdseye structures are formed due to the formation of rims of carbonate cement around the vugs and their subsequent infilling by sparry calcite (Shinn et al., 1969). Although parts of the overlying and laterally equivalent beds of this facies represent deposits analogous to those found where marine strata merge with non-marine strata, no evidence is provided by the regional setting for the development of tidal flats and no feature suggestive of the periodic action of flood and ebb tides is seen. Although the sequences forming Facies 4 are silty and dominantly muddy, the following features normally regarded as characteristic of tidal flat sedimentation were not seen.

- a. Flaser and lenticular bedding (Reineck, 1960b) which are characteristic of present day North Sea tidal flats.
- b. Small channels with low angle muddy and silty lateral accretion surfaces which are characteristic of the lower parts of tidal flats (Walker and Harms, 1975).
- c. Reactivation surfaces, bipolarity of current flow, or features indicating

tidal emergence runoff (Klein, 1970a).

- d. A fining upward succession is established, and broken coalified fragments a very common. But there is no similarity whatsoever to the expected sequence from a modern regressive tidal flat in which sand flats pass upward through mixed flats, mud flats and supratidal salt marsh deposits (laterally persistent coal beds are entirely lacking from this facies; Evans, 1965, 1975; Reineck, 1972, 1975; Straaten, 1954b). Thus it is seen that the sediment assemblage does not provide enough evidence of tidal influence during deposition.

The "birdseye" limestone found in Applecross and southern Strath may represent periods of emergence although strictly speaking a "supratidal" origin cannot be established. Desiccation structures such as polygonal mud-cracks and sheet cracks are absent; these together with the lack of authigenic dolomite, gypsum and anhydrite or calcite pseudomorphs thereafter suggests that evaporation never exceeded precipitation. Pellet limestones with cryptalgal and fenestral fabrics represent tidal flat facies (Read, 1975); they have been described from tidal flats in Shark Bay, western Australia (Hagan and Logan, 1974; Logan et al., 1974; Read, 1974). Fine laminoid-fenestral fabrics are seen to form beneath smooth mat communities in the lower intertidal zone while irregular fenestral fabrics are developed beneath pustular mat communities in middle and upper intertidal zones. Read (1975) maintained that in cyclic carbonate units, the first occurrence of cryptalgal and fenestral limestones capping the cycles, defines the approximate position of sea level during cycle deposition and represents a sea level datum; the evidence of a lithofacies below this datum closely approximates the water depths where sedimentation occurred under stable conditions. No distinctive facies of the type noted by Read (1975) have been recognised in Facies 4.

The birdseye limestones cannot be considered diagnostic solely of the emergent (supratidal) environment. They may also be produced by the releasing

of gases by organic decay, which would occur in limestones rich in algae (Zamarreno, 1975); thus Facies 4 may represent low intertidal to subtidal environments.

Although modern examples of these beds have been reported from Florida, the Bahama Bank, the Persian Gulf and Shark Bay, disrupted laminations, dark micritic laminations suggestive of algal mats and rip-up clasts, further suggesting a shallow intertidal or supratidal origin, are not observed. Of the three types of fenestrae observed by Read (1975) in the Shark Bay tidal flats (which are characteristics of the overlying algal mats present, the type of mat being controlled by position within the intertidal and supratidal zones) only the type showing irregular fenestrae (middle and upper intertidal) occur; the lack of associated definite subaerial features rules out a supratidal or intertidal origin. One may however describe them as beds representing deposition near a strand line. The lack of features indicating subaerial exposure and the absence of cycles (Read, 1975) in beds representing Facies 4 suggests little change of sea level or considerable reworking. If sea level fluctuations occurred, they were of short duration and left little or no evidence in the geological record. The equivalents of the beds of Facies 4 in Raasay are fine grained biomicrites (Plate 6.18) containing echinoid debris, probably representing deposition under relatively calm, deeper water conditions; they are interbedded with coarse grained poorly sorted calcareous sandstones which are probable equivalents of the sandstone interbedding in Applecross. The development of pebbly sandy marlstone of this facies is only seen in Ob Lusa and sedimentary structures other than calcite-replaced slender branching tubes (Plate 6.2; ?Thalassinoides) are absent; a mixed sediment derived from the breakdown of quartz-rich material and some echinoid debris has also developed. The presence of echinoid debris in the sandy limestones indicates a marine derivation for these beds and the feldspar content and polycrystalline quartz grains present, suggest derivation from granitic and metamorphic rocks in nearby areas. The absence

of cross-laminae and other features indicative of sediment movement suggests relatively calm waters. The same alternations are seen in the Raasay section (Rubha nan Leac) with the exception that the micritic beds contain large fragments and are full of mud-coated, calcite-replaced bivalve shells together with ostracod shells and ?calcispheres.

Although it is not possible to determine the transportation direction it is possible that the lower sandier beds (Apc 41) are representatives of the "sediment tails" building out into quiet waters and settling on the muddy bottoms accompanied by the influence of kaolinite-rich waters which drained the nearby lands. The finely reworked, coated, bioclastic material was derived from further offshore areas as the result of transportation and breakdown of shell materials associated with areas of possible organic buildup. The derived material forms the bioclastic-lithoclastic packstone found comprising the limestone bed of Facies 4, which built into a relatively calm shallow muddy basin; the presence of well rounded epidotes together with scarce quartz pebbles indicates the existence of recycled detritus within the basin.

ii. Facies 5

The oolitic limestones of Facies 5 were probably deposited under moderate energy conditions comparable with those of the modern subaqueous environments (Bahama Banks, Persian Gulf, Shark Bay); although near-normal marine conditions are indicated by the fauna, the dominance of Liostrea life assemblages in the basal outcrops reflect a slight restriction of the environment; the dominance of oysters may also indicate the shallowness of the water (Kauffmann, 1970).

The occurrence, development and growth of sand bodies, particularly oolite shoals along shallow carbonate bank margins have been widely recognised from the Bahamas and south Florida (Bathurst, 1971; Hine, 1977); they also occur in marginal seas which lack "shelf edges" comparable with

those of modern Caribbean carbonate provinces (see Purser, 1973).

It should be realised that in present day environments, oolitic sands are common features of the platform edges, there is a reported occurrence of oolitic sand forming today on the edges of the Florida Platform.

The oolitic sand belt lies between two distinctly different environmental areas, i.e. between the deep Florida Straits and the shallow, broad Great Bahama Bank in the Cat Cay area (Gray-Multer, 1971). Here, a discontinuous belt of oolitic, cross bedded lenticular sand can be found in the process of formation.

The oolitic facies (Purdy and Imbrie, 1964) is typified by an abundance of ooliths and an extremely low skeletal content. The analysis of constituent particles of samples from south Cat Cay-Brown's Cay area revealed an inverse correlation between depth and the abundance of oolitic grains. The absence of ooliths in the shoal-water bottom deposits of the east and west coasts of Andros Island and the west coast of Bimini shows that other factors beside depth should be investigated; the influence of increased tidal current velocity and bottom agitation in the direction responding to the decrease in depth from the marginal escarpments to the barrier rim of Cays and shoals is the main oolite-forming factor (Purdy and Imbrie, 1964).

Ooliths are reported to occur in many different, agitated environments in the Persian Gulf (Loreau and Purser, 1973). Most of the ooids are seen to constitute tidal deltas associated with coastal barrier systems, others are formed in tidal bars situated in wide channels between islands and the adjacent shoreline and on open tidal flats and beaches in exposed embayments; in addition ooliths forming within lagoons and in the less agitated lee coast Qatar peninsula have been reported. As seen in Plate 6.1, the size and shape of the constituents of the beds of the oolitic Facies 5 differ markedly from each other, in the form of lateral equivalents etc. The evidence provided by each sample is compared with that provided by studies on recent oolitic rocks in order to propose an appropriate genetic explanation.

In general, in the Recent environments of the Persian Gulf, aragonite seems to be associated with organic matter in the cortex of ooliths (Loreau and Purser, 1973). These are tangentially orientated with respect to the nucleus in agitated environments and are tightly packed and coalescent whereas in the more protected depressions the aragonite in the outermost layer of the cortex have haphazard or radial orientation and the fabric is loose; in protected lagoonal settings the radial orientation is well developed and the ooliths are often unusually large and irregular in shape.

The principal difference between the ooliths of the Persian Gulf and their counterparts of the Bahama Bank is in their regional distribution, in the former ooliths accumulate along the continental shoreline far from any platform edge and some form in quiet water tidal flats. In the beds of Facies 5, the presence of poorly sorted oolpelmicrosparite packstone/grainstones (Apc 45) poses some problems of interpretation. The well rounded character of the grains suggests prolonged periods of transportation while the bimodal nature of the sorting (clearly illustrated in Plate 6.1) is suggestive of limited agitation and winnowing action. The larger well rounded micritic grains are up to 0.5 mm. in diameter and show very fine concentric lamination which in places display a fenestral character and is perhaps of algal origin. The grains do not contain any form of opaque material but occasionally display a polycrystalline quartz nucleus. The sample is essentially a non skeletal oolite, the ooliths of which show poorly developed tangential laminae; most of the interior of the larger ooliths consists of cryptocrystalline, micritic calcite coatings. Although tubiform and fenestral features are present in the coatings, there is no proof that algae were responsible for the formation of all the coated shells. Although the well rounded ooliths indicate prolonged periods of formation their association with smaller sized, rounded micritic pellets and poor sorting points to a period of limited agitation.

The nuclei are composed mainly of micritic pellets; the lack of quartz

and other terrigenous grains is probably the result of the distance of the oolith-forming environment from the shoreline. This situation is akin to that described from eastern Abu Dhabi (Purser and Loreau, 1973) where in the vicinity of a coastal barrier (tidal delta, beaches and associated aeolian dunes), isolated by lagoons from the adjacent continent, the oolith nucleus is generally in the form of a carbonate pellet and the rareness of quartz, feldspars and detrital grains is attributed to the ineffective nature of transport towards the sea in that area.

Although it is impossible to establish the existence and recognise the morphology of any of the above mentioned environments for the basal beds of Facies 4 in Applecross, a similar mode of formation for the ooliths may be invoked. In eastern Abu Dhabi, the (tidal) delta itself is a complex system of banks and channels the morphology of which indicates that its oolitic sands are constantly influenced by a complex system of multi-directional currents. The lack of any indication of current activity in the basal beds of Facies 4 is considered to be mainly due to post-depositional diagenetic changes coupled with poor field exposure. However, the bedding surfaces of the beds representing Facies 4 are modified by ripples of varying dimensions.

Two suites of sediments were defined from the Abu Dhabi coast which may also be found in the Liassic section in Applecross (see Plate 6.7 and 6.8).

a. Localised within the tidal delta, a suite of pure oolitic sand with > 90% ooliths within any sample is seen with ooliths of sizes up to 2 mm; these are found along the edges of the (delta) axial channels.

The ooliths decrease in size towards the outer fringes of the delta and are progressively diluted with pelletoidal grains (cf. Plate 6.7).

b. Mixed ooid-pellet-bioclastic grains (Plate 6.8), containing 10-90% ooids occur both on the outer edges of the delta at depths of 2 - 5 m., within the axial channel and especially as beaches and (aeolian) dunes on a coastal barrier complex. The ooid fraction varies, being relatively high within the

beaches and dunes, but is strongly diluted by coarse bioclastic debris within the channels.

The oolite of sample Apc 45 may be regarded as one which was formed towards the outer fringes of a ?delta. The silty oobiopelmicrosparites of sample Apc 1 with 34% ooliths and only 6% bioclasts can be regarded as being formed within the outer margins of carbonate dunes and beaches. It should be noted that the total absence of argillaceous material in Facies 5 may be due to either the progressive transgression of Facies 4 and the establishment of a relatively offshore carbonate-dominated environment, or due to the formation of a sedimentary body which had effectively cut off the supply of argillaceous material. The former explanation is preferred here since the carbonates are free of argillaceous matter; the mixed clasts are mostly covered by a rim of micritic material with tubular features and although the classic asymmetric ooliths (Freeman, 1962; Davies, 1966) are not observed, their deposition under partially protected environmental conditions is favoured. The presence of at least 2% quartz (strained) with superficial micritic rims indicates that the relatively calm environments were either proximal to land areas or were protected by sedimentary bodies of similar composition (? spits or bars). Careful study of the ooliths indicates their dual origin. Plate 6.19 shows ooliths with concentric layers of at least two generations; the inner (first) generation contains opaque material (?finely powdered coalified material) scattered throughout the tangential laminae whereas the outer oolitic layers of the cortex, essentially showing a radial arrangement of the calcite crystals, contain no opaque substances. The ooliths formed in a relatively turbulent environment with an abundant supply of allochthonous material. They were then transformed to an environment with relatively less allochthonous debris and turbulence.

It should be noted that the globular and tubular forms seen in the various layers of the oolitic laminations resemble algal borings (Lebauer, 1965; Johnson, 1964).

Although samples Apc. 1 and Apc. 2 were taken from positions immediately adjacent to one another, it can be seen that sample Apc. 2 clearly shows fundamental differences from Apc. 1. It may be classified as an oobiomicrorite with the development of sparite pockets containing the main body of the oololiths. The oololiths (16%) are mostly of second generation and also show both tangential and radial arrangement of calcite crystals. Quartz is rare but some grains act as nuclei of oololiths. These sediments represent poorly sorted oololiths with well developed internal structures. Their size varies from 0.22 to 0.5 mm.; such an association indicates textural inversion (Folk, 1962b). These beds normally contain a high proportion of skeletal debris and were probably also affected by the activity of burrowing organisms. Such texturally inverted sediments occur as very thin sheets among the beds of facies 5 and are analogous to beach and storm deposits described by Ball et al. (1967), Hayes (1964) and Imbrie and Buchanan (1965). Their presence may be interpreted as signifying deposition on high tidal or supratidal flats by high velocity water movements (Wilson, 1968b). The fine silt size (0.02 - 0.04 mm.) quartz grains of succeeding beds (Apc. 3) clearly dominate the bioclastic constituents; it should be noted that , the bioclastic constituents are mainly composed of broken brachiopod shells, coral debris together with echinoderm fragments and gastropod tests indicating a dominantly marine derivation. These beds may be interpreted as products of complex erosional processes which intermixed offshore bioclastic debris derived from biological buildups with clean quartz clastic material (? aeolian) derived from onshore environments. Such complex inter-relationships are common in the present day environments of the Persian Gulf (Purser, 1973). The presence of coral fragments in the samples together with bivalves, echinoderms and gastropods confirms the above notion. The position of these beds underlying the beds of Facies 6 does not show characteristic features of (reef) "talus" (Purdy, 1974a; Heckel, 1974), therefore it is questionable

whether any offshore coral "banks" (Heckel, 1974) stood much above the surrounding deposits. The amount of bioclastic debris in these beds together with the presence of algae contributing to the formation and boring of the ooliths (of the lower beds) suggests deposition under relatively shallow water conditions with a constant supply of land-derived quartzitic material. The beds of skeletal debris (biosparite) immediately overlying the siltstones and laterally equivalent to them are probably representatives of interbank areas where skeletal debris accumulated and was well washed by variable currents. No definite cross lamination indicating direction of transport can be seen. The configuration and morphology of these areas was of very shallow wide-channel types. Fine sandstones with up to 45% calcareous cement form the lateral equivalents of the beds of Applecross in Ob Lusa. The high percentage of polycrystalline quartz in the specimens may be regarded as evidence for a metamorphic origin; in sample Am 62 the polycrystalline quartz grains with 2 to 3 crystals are of a higher percentage (3%) than those with more than 3 crystals per grain, therefore a high-rank metamorphic source may be envisaged as a probable sediment contributor for the mentioned strata (Basu et al., 1975; Young, 1976). These may have been derived from extensive Moinian outcrops of northwest Scotland, exposed during Liassic times.

The strata are thin-bedded and do not contain fossils or other fragmented biogenic material. The lack of coating on the quartz grains together with the homogeneity of the rock indicates a well-washed environment of deposition. The moderately sorted, subangular nature of the grains may be attributed to their formation in and transportation from source areas proximal to the nearshore depocentres.

Grains in the overlying beds show asymmetric coatings (Freeman, 1962; Davis, 1966) indicating quiet-water, protected depositional conditions. It should be realised that the asymmetric coatings on the grains are either

composed of micritic mud or tangentially arranged calcite crystals, therefore the energy gradient may be thought of as being higher than in those environments where only the formation of calcite crystals around undefined nuclei form ooliths. Poor sorting, subangular quartz grains and the existence of asymmetric ooliths suggest poorly agitated waters. The lack of coalified material and relative scarcity of micritic carbonate constituents may be regarded as evidence for a partially enclosed environment where most of the "impurities" eventually found a way out of the system, alternatively they were never introduced into the environment due to its increased distance from land, lack of sufficient vegetation and combinations of climatic conditions which prevented the land-derived material from reaching this environment.

In Applecross the beds underlying the development of Isastrea corals, while stratigraphically equivalent to those beds of Am 102 and 103, exhibit different characteristics. Sample Apc 5 is a normally packed oolite with mostly well-formed, multiple-coated (two different generations), spherical grains. The nuclei are composed of echinodermal and bivalve fragments (long and rounded) with occasional ostracod shells, quartz grains and multiple ooliths are present. The cement between the grains is blocky calcite and is probably of late diagenetic origin (implying meteoric waters in the phreatic zone).

In the present day depositional environments, well-formed ooliths are found mainly in tidal bars (Wilson, 1975; Purser and Loreau, 1973; Illing, 1954; Purdy and Imbrie, 1964). They are always cross laminated and the individual ooliths are commonly 0.5 to 1.5 mm. in diameter.

The beds found in Applecross are not cross laminated and the relatively higher proportion of well rounded micritic pellets and also multiple pellets of rounded oobiomicrocrites suggests that these beds formed near active currents but at the same time possibly further out from the central axis of the channels where currents are more active and the process of particle breakdown

and oolite formation more vigorous. There is no evidence for the possible reversibility of the currents which were operative during the formation of these oolites. It is apparent that most of the oobiomicrorite pellets are relict and were derived from previously formed beds and as there is no evidence depicting the possible morphology and/or orientation of this oolite, it is hard to decide upon the environment in which it formed. It is also puzzling that the corals of Facies 6 directly overlies calcareous sandstones with asymmetric oolites in northern Skye whereas in Applecross they succeed biopelagites with ~ 1% rounded, silt size quartz grains and rounded recrystallized microtic pellets together with bivalve shells coated by microtic mud (with ?algal tubes). The lack of definite oolitic coating on the various fragments and grains of sample Apc 6 suggests moderate agitation-

It is possible to explain this by postulating the existence of extensive mud-banks fringing shallow, coastal embayments (faunal evidence for postulating truly lagoonal conditions is lacking). The microtic mud pellets and rounded pelletoids were eroded from the mud banks and transported into the coastal embayments. Finely crushed shell material was also caught up in the microtic mud, transported and formed calcareous sand bodies within the environment probably similar to those found in the present day coastal embayments and lagoons of the Persian Gulf (Loreau and Purser, 1973; Purser and Loreau, 1973). The pelletoidal gastropod sands characterising the "intertidal zone" of the Jebel Dhanna lagoon are represented in Facies 5 by the shallow, extensive mud and pellets banks (?intertidal) which in places contain oobiomicrorites, the oolites/being mostly relict and derived from previously formed environments, giving the rock an "inverted texture" (Plate 6.20 and 6.21).

The "lagoon, infratidal zone (muddy pelletoidal gastropod sand)" environment received the reworked elements of various channels which are now represented by beds from which sample Apc 4 was taken. Mixed bioclastic elements (gastropods, echinoderms, bivalves, brachiopods) and rounded microtic pellets with black (coalified) inclusions are found in a clean, coarsely

crystalline cement; the crystals of the recrystallised shells are mostly coated with pyritic material. These environments were probably protected from open marine activity by coastal spits and calcareous quartz sandstone bodies, parts of which are seen in Ob Lusa. Shallow channels with vigorous current activity (sample Apc 5) linked the coastal embayments with open marine waters. Evidence for the existence of the shallow mud banks is also found in Loch Eishort (samples Am 40, 42, b28).

The above mentioned relationships are illustrated in facies maps of Figs. 6.13, 6.14 and 6.15 from which a progressive transgression is also observed to have taken place from west to east as the corals of Facies 6 approach their present positions.

It should be noted that other than the evidence found in northern Skye (Ob Lusa) there is no indication of quartz sand bodies forming laterally accreting spits from the assumed shorelines. On the contrary, samples taken from Applecross seem to portray the formation of bioclastic sands. This may well have been the case as the conditions were becoming favourable for the full development of organic and coral buildups of Facies 6 in this area.

The distinction made between bays and lagoons along the Texas coast by Fisher and Brown (1972) is applied to the sediments representing the different facies in the beds of the Milton Formation. Bays are regarded as representing the lower reaches of drowned valleys and as such, their long axis forms at a high angle to the strand. Lagoons are bodies of water trapped behind barrier islands and therefore have their long axis parallel to the strand-line.

The characteristic facies and depositional conditions representing each of the above environments are summarised as follows (after Fisher and Brown, 1972):

	Lagoons (restricted bay)	Bay (open bay)
Sediment influx	Low	High
Water exchange and influx	Low to non	High
Distinctive sedimentary features	Non	Bay-head delta
Fauna	Low density - low diversity	High density - low diversity e.g. oyster reefs
Salinity	Hypersaline	Near-normal, low, fluctuating
Sedimentary character	Laminated mud	Burrowed, reworked, inter-reef mud channels
Sediment origin	Mainly windblown, washover fans	Terrigenous, storm generated

6.9 Lusa Coral Member

In the northern area of study, the beds of this Member are represented by Facies 6. It occurs in two of the principal sections (Applecross, Ob Lusa) and as previously explained, may also be laterally traced inland from northern to southern Strath. The total thickness of this bed in Applecross and Raasay does not exceed 0.50 m. It consists of lenticular beds (3 - 6 cm.) of the compound, massive coral Isastrea murchisoni (Plate 6.22) which alternates with very thin (5 mm.) beds of calcareous shale. In this coral bed no trace of individual epithecal walls can be seen and the septa from separate corallites are confluent. In places ?cemented oyster shells directly rest on top of these beds. The corals terminate at the shale/limestone interfaces in Ob Lusa. The shales are considerably altered and show illite to be the main constituent, with trace amounts of smectites present. The shale beds are not present in this facies at Applecross.

Petrographically these beds are biomicrites. The epithecae of the corals are entirely recrystallised and consist of sparry calcite whereas the

matrix is microsparite in places. The matrix consists substantially of recrystallised bivalve shells in northern Strath whereas in Applecross these beds are partly biomicrites and partly biopelmicrosparites. They show extensive signs of reworking and burrowing. Beds representing this facies were not found in Raasay. In southern Strath (Loch Eishort) it is possible that these beds are represented by a 0.70 m. coarsening upward bed which consists of a 0.30 m. marlstone with thin dismicrite beds at the base (Plate 6.23) and a matrix-supported silty biopelsparite (quartz 7%) at its top. Close examination of the relatively thin, lens-shaped features of the coral-bearing beds shows that stylolites are common and the coral structure is only well preserved within the outer few mm. of the thin beds where interstitial sediment fills the corallites, the interiors of which appear as masses of sparry calcite with a saccharoidal texture to the naked eye. Under the microscope, original skeletal material (?aragonite) is seen to be replaced by coarse textured sparry calcite which shows an increase in size towards the inner parts of the corallites. Dog-toothed drusy rim crystals are less common probably due to the existence of more than one cycle of diagenetic alteration. The matrix sediment is micritic to microsparitic limestone with scattered recrystallised shell fragments and little echinodermal material; also with a small percentage of silt and sand grade detrital quartz. None of the three grades of progressively more complete preservation of coral structures identified by Talbot (1972) from the Oxfordian Coral Rag of southern England could be identified in the coralline beds of the Broadford Beds Arenaceous Group. The coral structures show deformed, elongate inclusions and the preservation of the coral can possibly be related to Talbot's type 3 preservation with micritic calcite replacing the original aragonite in places.

6.10 Interpretation of the Lusa Coral Member

The maximum width of the beds containing Isastrea corals is 5 - 10 m., with thicknesses up to 20 cm. (Ob Lusa). The texture of the beds is mostly destroyed during diagenesis but in various areas evidence for vigorous bioturbation and mechanical destruction is seen. Immediately underlying these beds in Applecross transported, reworked and mixed shell fragments are seen in the samples whereas the corals directly overlie calcareous sandstones in Skye.

Internally the coral beds are not homogeneous, showing various irregularly arranged burrow channels and clay sheets which are not necessarily parallel to the bedding surfaces ; pectinids occur on the surfaces of the corals in places. In the absence of evidence for possible relief (e.g. skeletal talus), the term "Bank" (Heckel, 1974) may be used for their description. The corals overlie beds consisting of skeletal calcarenite which are oobiomicrorites in places. These do not resemble "talus" deposits and it is possible that they originally formed as accumulations of organic material, progressively contributing to the moderate buildup of substrate. The shallowness of the water, however, limited the growth of the buildup and with possible changes of water depth, the corals grew both laterally and vertically. The very thin shale seams interbedded with 2 to 5 cm. thick coral beds indicates that the sediment was able to spread over the coral beds due to a steady rise in sea level or periodic storm events, thus inhibiting their growth by possibly "choking" them. Continued growth would therefore be established in patchy areas where the corals were not totally covered by clayey material or where they were able to remove the sediment from their surfaces. Modern corals show an ability to remove sediment from their surfaces (Hubbard and Pocock, 1972). Discontinuous bottom subsidence (or sea level rise) and the rapid influx of terrigenous sediment disturbed their steady growth; also it is probable that the buildup stopped because its

breadth, height and slope relations retarded upward growth to the point that it lagged sufficiently behind bottom subsidence (or sea level rise) and was eventually covered by the sediment influx (Heckel, 1974). The existence of encrusting oysters on the top of some beds indicates that the surfaces of the growing buildups required wave resistance to continue developing within turbulent surface waters (Stanton, 1967a) indeed the beds of Facies 6 may have formed large resistant bodies in places. (See Fig. 6.16).

6.11 Lower Sand Member

In the northern area this Member is present above the beds which contain the coral Isastrea murchisoni. As indicated on Table 6.1, two facies represent this Member at each of the three main sections.

i. Sandstone Facies 7a₁

This facies is exposed along the shore at Loch Eishort, as indicated on Fig. 6.17 (Hallam's beds 3, 4 and 5). The total thickness is 12 m. and the beds are mainly composed of cross-laminated calcareous sandstones. At their base here, a 1.20 m. sandy limestone containing disarticulated shell fragments together with wavy sandy sheets are seen. These sheets show no internal structures and appear to be less calcareous than the surrounding rocks. Petrographically the limestones are matrix-supported silty biopelsparites with 5% quartz; an indeterminate opaque material (< 2%) is also present. Shell fragments are mostly reworked and are coated with a thin micritic rim. Towards the top of this unit, subrounded quartz fragments (2 - 4 mm.) are seen. Transitionally overlying this unit, are grey medium bedded, cross laminated sandstones 3.10 m. thick; this unit shows wavy bedding towards its top whereas at its base, the beds are more even. The wavy beds (uneven, up to 15 cm. thick in places) are laminated and show erosional contacts with the cross laminated sandstone lenses and wedges (20 cm. thick, 86 cm. long). The sandstone lenses show long, curving toesets (convex downwards) and at

their base, coalified driftwood and iron oxide accumulations are seen. Rounded quartz fragments (2 - 4 mm. diameter) also occur within the laminae which dip up to 25° (Fig. 6.18a).

In places, faint trough cross laminations are formed which are non-erosively overlain by faintly laminated beds (Plate 6.24 a-c; Fig. 6.18), 5-7 cm. thick; the above mentioned bedforms are frequently overlain by small-scale oscillation ripples.

The cross bedding orientation data obtained from these beds (Figs. 6.18a and 6.18) show that while a polymodal orientation is present, the majority of the readings point in the $260-320^{\circ}$ region. The laminated sheets show a predominant mode in the $180-190^{\circ}$ region.

Petrographically these beds are moderately sorted, well rounded grain-supported calcareous sandstones. The quartz is mostly (27%) composed of polycrystalline grains (> 3 crystals per grain) and the feldspars are mainly plagioclase with size ranges in the region of $3-2\phi$. Towards the top, these beds become matrix supported silty biopelsparites with polycrystalline quartz making up to 12% of the total rock. The grain size decreases towards the top of this unit and the beds become poorly sorted; the polycrystalline quartz content of these beds remains constant throughout (10%). Rounded quartz grains 3-4 mm. in diameter are seen at the top of this unit.

The "sandy limestone" unit (Hallam's bed 3) overlies the cross-bedded sandstones with a sharp contact, the total thickness is 6.14 m. and it consists of medium bedded finely indurated silty limestones with abundant shell fragments alternating with oolitic limestone beds. These beds have undulose contacts, show very faint evidence of cross lamination and are 15-20 cm. thick; overlying them is a 50 cm. thick, white, cross laminated siltstone bed. The cross lamination data obtained from these beds is represented in Fig. 6.19 and it is evident that two modes are present, one in the $10-40^{\circ}$ region and the other in the $60-100^{\circ}$ region. Both tabular (Plate 6.25) and trough (Plate 6.26) cross laminated beds are present. Petrographically these are orthoquartzitic siltstones with 2-3% plagioclase

and <5% polycrystalline quartz. The grains are 4 ϕ in size, well sorted and subangular (Plate 6.27).

ii. Oolite/Sandstone Facies 7a₂

This facies is represented by a 4.40 m. development of thin bedded 10-12 cm. thick, faintly cross laminated oolitic limestones alternating with 10-12 cm. thick calcarenites with undulose contacts and recrystallised shell fragments .

Petrographically the oolitic limestones are matrix supported oosparites, the ooliths, which are in the 2 ϕ size range and medium sand sized, mainly consist of multiple generation coatings, the second generation being represented by a single superficial oolitic rim around "normal" ooliths. Quartz fragments are not common as nuclei and ooliths mostly contain a micritic pellet or shell fragment for nucleus. Polycrystalline quartz (2-3 ϕ , fine sand) are common, and are mostly coated by a pseudo-oolitic rim of micrite; the matrix is well washed, with no indication of organic matter or other opaque material, long ooliths are not common and compound ooliths are mostly composed of not more than two individuals (Plate 6.28).

These beds are interbedded with matrix supported, silty oopelsparites which contain a high proportion of pseudo-ooliths (18%); superficial ooliths are also present (10%) and small pellets of micritic mud are common (Plate 6.29). Recrystallised, long, rounded shell fragments with thin micritic rims are common. The quartz content of these beds (17%) is higher than that of the oosparite beds, and the grains are mostly covered by a thin pseudo-oolitic rim of micrite.

The quartz content of this facies increases towards the top and polycrystalline quartz dominates over normal quartz. The beds become thicker (20-25 cm.) and interbeddings of very poorly sorted silty biosparites are seen, in which intraclasts of oolitic limestones identical to those underlying them can be found.

The grain size is that of fine sand and it should be noted that towards the upper parts of this unit superficial ooliths become dominant, with 0.14 mm. nuclei mostly composed of micritic rounded pellets and a single oolitic rim.

iii. Limestone/Oolite Facies 7b₁

This facies is exposed in northern Strath at the Ob Lusa section (beds 4, 5 and 6 of Hallam). The total thickness of this facies is 12.30 m., at the base of which is a prominent calcareous siltstone bed crowded with Cardinia cf. concinna (Pl.6.29) with well sorted subangular quartz grains 3-4 ϕ in size. This bed shows a wavy top surface and is overlain by a thinly bedded calcareous siltstone (1.20 m. thick). The lowermost beds show very faint cross laminations and at least two bands of broken, unsorted bivalve and gastropod shell fragments are seen in the basal 50 cm. beds. Mound-like features with an average height of 10-12 cm. and a diameter of 25-26 cm. occur, also large coalified plant remains can be found on the top bedding surfaces. The beds of this facies are thin, undulating, laterally impersistent (in places), alternating sandstones and limestone, the quartz content of which increases towards the top of the facies. Petrographically these consist of more or less silty oobiopelsparites. The following samples were taken progressively towards the top of this section.

Am 109 - Totally recrystallised oobiopelmicrite, the patchy development of the clusters of ooliths, pellets, silts and the distribution of the matrix can be seen ; the quartz grains are <4% and poorly sorted.

Am 110 - Poorly sorted, rounded calcareous siltstone (matrix supported). The size distribution is bimodal with one mode (30%) at 3 ϕ and one (80%) at 4 ϕ ; rounded epidote grains occur.

Am 111 - Well sorted matrix supported silty oobiopelsparite (Plate 6.30). The ooliths are mainly superficial; here and covered by a substantially thick layer of radially arranged calcite crystals; the ooliths

are mainly 3ø in diameter whereas the quartz fragments are mostly in the 4ø size range. Asymmetric ooliths are also present.

Am 112 - A grain supported silty biopelsparite. The compaction is mainly diagenetic and due to the secondary growth of the matrix spar. Ooliths are very common (26%), but most (22%) are superficial, with asymmetric ooliths also occurring. Pellets are present (10%) and together with the ooliths have a diameter of 0.16 mm. The bivalve shell fragments found here are mostly recrystallised and rounded, with a thin micritic coating (samples Am 113, 115, 116, 117, 119).

Towards the top, 12.50 m. above the Lusa Coral Member, beds of this facies become well packed silty biopelsparites with abundant shell fragments which do not show any type of coating. Micritic limestone pellets are present and the silt content seldom exceeds 8-10%; these are 4ø in diameter and smaller than the rounded micritic limestone pellets which are ~0.1 mm. In some sections larger pellets are seen exhibiting evidence of algal boring activity. In some cases the limestone pellets seem to exhibit very small scale cellular structures of possibly algal origin. These beds commonly contain coalified plant and wood remains (Plate 6.31).

iv. Sandstone/Oolite Facies 7b₂

Overlying the beds of Facies 7b₁ in Ob Lusa, is a 5.50 m. development of interbedded limestones and sandstones. The lowermost 2 m. is composed of thin (2-3 cm. thick), bedded yellowish weathering grey calcareous sandstone sheets. These are laterally persistent and show wavy contacts. Petrographically the calcareous sandstones are matrix supported, poorly sorted with some subangular grains of quartz (strained and polycrystalline). The feldspars present are mostly orthoclase with a size range of 2-3ø. These beds become thicker bedded with less wavy contacts towards the top of the unit. Bed surfaces at the lower parts of this facies show markings which are up to 50 cm. long, irregular, meandering and V-shaped resembling channels

with upturned edges.

Towards the top of this unit the sandstone beds are interbedded with limestones. In their lower parts, the limestones are matrix supported silty oolipelsparites, moderately sorted and contain up to 10% quartz fragments which are commonly unstrained (3ϕ size range). Oololiths are rare but superficial oololiths are seen (8%); rounded micritic pellets constitute up to 21% of the rock (size $>3\phi$). At the top of this facies the limestones become more common at the expense of sandstones, i.e. the limestone/sandstone ratio increases.

Oololiths constitute up to 10% of the total rock and the quartz grains are mostly polycrystalline; rounded micritic pellets are common together with asymmetric oololiths; quartz grains are in the 2ϕ size range. It should be noted here that the topmost part of this facies immediately below the "Coral Facies 8" is composed of a medium bedded, matrix supported silty oosparite with very large (0.32 mm., $<2\phi$) grains of polycrystalline quartz (> 3 crystals); quartz constitutes $\sim 5\%$ of the total rock. Oololiths are common and pseudo-oolitic micrite rims were formed around the rock constituents. Micritic patches are also present suggesting that the rims are merely by-products of diagenetic alteration. Most oololiths are not "normal" and are formed by a single oolitic rim around a nucleus.

Although epidote and ?zircon are common in this facies (and those occurring below it), they rarely serve as nuclei for oololiths perhaps due to insufficient to and fro motion necessary for oolith formation.

v. Shale/Oolite Facies 7c₁

This facies is exposed in Applecross (Hallam's beds 11, 12, 13). Here, a lower sequence of black shales followed by oolitic limestone is overlain by a unit consisting of black shales at the base and calcareous siltstone above.

The base of this facies is represented by the shales immediately overlying

the "Coral Facies⁶" at Allt nan Breugh. These are 90 cm. thick, show up to seven nodular calcilutite beds in Applecross (each 6-12 cm. thick) and are strongly bioturbated exhibiting Thalassinoides patterns. The shales are dark and micaceous with a low proportion of CaCO_3 (10%); flattened bivalves are common. The shales are somewhat silty (<5% silt) and contain up to 38% smectites.

The overlying strata are 2 m. thick and consist of calcareous micaceous shales. The oolites are matrix supported and contain reworked micritic fragments which are mostly 1-2 ϕ in diameter. The ooliths are mostly 2 ϕ in diameter and rounded shell fragments with a micritic rim are present. The oolitic limestone beds are silty oobiomictites at their bases (immediately overlying the thin shale beds). Organic matter is common forming the films around the ooliths and other constituents. The bioclastic material is crushed, and coalified plant material is common. The ooliths are mostly second generation and coalified material is commonly present in the first generation coating; scarce, very poorly sorted compound ooliths are also seen. In

the field, very small scale mound-like features are seen, the surrounding hollows being infilled by crushed shell fragments and finer material. The top parts of the limestone beds contain micritic pellets of the lower portion containing ooliths, some of these parts appear as shale beds in the field. Close examination however, shows that they are entirely composed of ooliths in a micrite matrix (Plates 6.32, 6.33).

The overlying beds are composed of 1.50 m. thick calcareous (35%) light grey shales with up to 33% smectites. These do not contain fossils but have concretionary beds, becoming more sandy and micaceous towards the top.

The concretions are mostly calcareous siltstones with up to 47% CaCO_3 ; they are very fine grained (3 ϕ) and contain rounded limestone fragments of the same size. Muscovite is common but not abundant. The 1 m. calcareous sandstone bed representing the top of this facies is a packstone (silty

oobiopelmicroparite) with ~48% calcareous matrix, and contains (12%) rounded limestone pellets.

It is bioturbated and poorly sorted; rounded quartz grains (up to 3mm. in diameter) occur and coalified plant material is abundant.

vi. Limestone/Oolite Facies 7c₂

Beds representing this facies in Applecross are 5.90 m. thick and composed of oolitic limestone at the base. These are thin bedded (8 - 10 cm.) and are interbedded with sandy/marly beds towards the top of the succession where they also become progressively siltier. The beds identified by Hallam as "Ferruginous Oolite" (bed 14) were not found.

Beds 17, 18 and 19 of Hallam (1959) are difficult to follow in the field because part of the section is faulted and consequently some beds are repeated. In the lower 0.80 m. of this facies there is a limestone unit containing large (1ø) ooliths with thick micritic coatings. The lithology is matrix supported oobiomicroite/microsparite.

Most of the ooliths have a second generation origin, the later generation being represented by a thin oolitic coating, showing a distinct textural difference from the original structure. Shell fragments are present, contributing to the formation of the more elongate ooliths (16%) (most of the ooliths are well rounded); the matrix contains very fine shell debris as well (Plate 6.34).

The top parts of this basal limestone unit is a matrix supported, poorly sorted silty biopelmicrite (Plate 6.35), containing totally recrystallised delicate bivalve (8%) and gastropods (2%) together with abundant ostracod shells (10%) and a few echinoderm remains (2%). These show a sequence of interbedded calcareous sandstones and oolitic pelletoidal limestones. The sandstones are silty oobiopelsparites and show moderate packing; rounded limestone pellets comprise 9% and quartz silt grains, 17% of the rock.

Gastropod and echinodermal debris is present but not frequent. Most

shells show a dark micritic rim around them and muscovite is scattered throughout. The oolitic pelletoidal limestones are poorly sorted silty biopelmicrosparites with large bivalve and other shell fragments. Large rounded pellets of micritic limestone (3 mm. in diameter) are seen and most bioclasts show a thick micritic coating. The beds become poorly sorted, matrix supported oobiopelmicrosparites towards the top. Multiple and second generation ooliths are very common, their maximum sizes reaching 1-2 mm. The matrix contains 3% silt together with reworked, fine shell material (Plate 6.36).

vii. Facies 7D

This facies is present in Raasay and is thought to be represented among the nodular limestones with shale interbeddings which have already been described.

6.12 Interpretation of Facies 7

This facies represents the re-establishment of shallow, restricted marine conditions over the northern area of study following the open marine conditions which produced Facies 6. Although its development is somewhat similar to that of Facies 5, some differences (namely source of supply, agitation and proximity of mud flats) can be recognised. The term "marginal marine" in this study refers to nearshore environments such as bays, sounds, lagoons and estuaries which may share parts or most of the following characteristics:

- a. Relative concentrations of the principal constituents of sea water are not constant or predictable.
- b. Detrital sediments are actively transported and deposited.
- c. Boundaries of environmental zones are sharp.
- d. Organic production and benthonic fauna are most probably affected by seasonal changes in river discharge and biological activity .

i. Facies 7D - (nodular limestones)

The beds of this facies are represented in Raasay by blue nodular limestones with alternating shales, which are taken to be equivalents of Facies 4, 5, 6 and 7 elsewhere in the northern area. The nodular limestones representing Facies 4 and 5 are separated from the higher beds representing Facies 5 and 6 by a bed of oolitic limestone (Ra 9, 10) which is identical to those found in Applecross (Apc 45) and regarded as parts of shallow channels which represent Facies 5. The nodular limestones below the oolite are identical with those of Facies 7D therefore they are both considered in similar contexts here.

The nodular beds almost certainly owe their origin to the presence of abundant burrows of the ichnogenus Thalassinoides (Plate 6.14). The occurrence, relationship to facies, bathymetry and diagenetic history of the trace fossil Thalassinoides has been dealt with by Fürsich (1973a, 1976) who related the formation of nodular limestones in the English Upper Oxfordian limestones to this trace fossil and suggested a mechanism for nodule formation based on the activity of burrowing crustaceans. In modern environments, Callianassaburrow networks which bear the most resemblance to Thalassinoides have been reported by various workers (Shinn, 1968a; Farrow, 1971; Braithwaite and Talbot, 1972); the burrows of lobsters and crabs such as Nephrops norvegicus (Lamarck), Genophax rhomboides (Lamarck) (Rice and Chapman, 1971) together with those of Alpheus (Farrow, 1971; Shinn, 1968a) also show a similar morphology. The mud-linings characteristic of these burrows are not seen in the beds of facies 7D; the absence of vertical shafts in the burrow systems of this facies may be explained in line with Fürsich's (1973a) interpretation.

a. The Liassic burrows were probably situated near the sediment/water interface, eliminating the necessity for vertical shafts, i.e. the entrance tunnel to the burrow penetrating the sediment obliquely (cf. burrows of

Gonephax rhomboides and Nephrops norvegicus (Rice and Chapman, 1971)).

- b. The thin linings or their absence may have caused the obliteration of the vertical shafts.
- c. The vertical shafts are characteristic of hardground associations (Kennedy, 1975), their absence indicates the limited or non development of hard surfaces.

The presence of branched nodules and their cylindrical shapes allows their interpretation as cemented infillings of Thalassinoides networks; irregular nodules are also seen with cylindrical tubes (4 to 5), radiating from a central, enlarged lensoid body which are clearly cemented infillings.

Interconnected dwelling structures made by hemisessile suspension feeders, with distinctly smooth, well constructed walls were also reported by Howard & Dorjes (1972). Such features which resemble Callianassa major burrows are equally common in offshore and nearshore sediments, being especially characterised by thick pelletoidal burrow walls. Howard and Dorjes (1972) realised that such burrow walls were absent in laboratory-simulated experiments. An explanation given for this was that such thick, pelletoidal walls were an adaptation for combating loose, shifting sediments in the natural environment, and was not required in quiet waters and stable substrates. This may be true for the Thalassinoides burrows seen in this facies. The lack of burrow linings signifies either of the following:

- a. Original substrate was stable.
- b. Lack of turbulence.
- c. Post depositional diagenetic infilling, replacement and growth of the burrows (Fürsich, 1973a) with the consequent removal of the original shell wall features.

In this facies, the burrow systems consist of horizontally connected cylinders (with only occasional vertical shafts) filled with predominantly argillaceous (?carbonaceous) sediments. The Y-shaped branching patterns are

typical of these beds and the burrows surfaces do not show any indication of scratch marks (or ridges). Although the sequence was well studied, no indication for the presence of other burrow systems were seen. Thalassinoides dominates the trace fossil assemblage; some evidence for the presence of encrusting bivalves is seen but there is no evidence pertaining to their association with the development of hard substrates (hardgrounds). The burrow tubes and cylinders constitute whole limestone beds which alternate with smectite-rich calcareous shales and contain a high percentage of organic matter. The limestones are micritic and commonly show reworked, broken bivalve shell material together with whole shells assimilated within them. The shales are crowded with bivalves and are disturbed by the development of the limestones in places. The predominance of a horizontal Thalassinoides network over the vertical elements also indicates their formation in "soft" rather than "hard" grounds (Kennedy, 1975).

Thalassinoides is a poor facies indicator and although it is known to occur in littoral and sublittoral deposits (Cruziana facies of Seilacher, 1967), it is found in a variety of rocks ranging from muddy calcarenites to oolites, siltstones and fine sandstones; it is also of restricted bathymetric value (Bromley, 1967).

A variety of Thalassinoides species are found in the Jurassic limestones of Europe, which mostly resemble alpheid traces described by Farrow (1971). In contrast to the Lower Liassic Thalassinoides burrows of Raasay, those reported by Kennedy et al. (1969) and Sellwood (1970a) are packed with faecal pellets.

Although compositionally different, the beds representing Facies 7D resemble to some extent the beds of the Osmington Oolite Series (Middle Oxfordian) found at Bran Point (Dorset). The oolites of the Corallian sequence in southern England signify an intertidal and partly supratidal environment according to Wilson (1968b). They are preceded and followed by slightly deeper water facies represented by the "Nodular Rubble" (Arkell,

1936) and the "Littlemore Clay Beds" (Arkell, 1927) respectively.

The nodular limestones of Facies 7D represent muds deposited in slightly deeper, more tranquil conditions than the underlying oolites; the nodular appearance may be explained as being due to intense bioturbation which caused the mixing of the various abundant shells. The pattern was enhanced during later stage diagenetic segregation of CaCO_3 . A quiet, shallow subtidal environment of deposition may be envisaged for the nodular beds of Facies 7D, the relative sparsity of the macrofauna suggests slightly unfavourable living conditions, but this did not affect the trace fossil-forming organisms. Although the lateral extent of the beds is difficult to establish it is possible to suggest that they were laid down in channels and lagoons between patchy coral "banks" (see Fig. 6.20 and 6.21).

ii. Facies 7(a, b, c)

The development of cross laminated, wavy bedded calcareous sandstones at the base of this facies in southern Strath (samples Am 46, 47 and 48) and at Ob Lusa (samples Am 106 and b 66) is regarded as representing a quartz-sand body which actively protected areas situated toward its eastern and north-eastern sides against open marine conditions. The size ranges of the sandstones varies from very fine to fine sand in southern Strath to very fine sand at Ob Lusa. No indication is seen for the development of this sandstone body in Applecross where oomicrites and lagoonal shales were being deposited (samples Apc 10, 11 and 12). The presence of coalified material lining the cross laminae in southern Strath together with the presence of coalified plant stem fragments in the succession of northern Strath indicates the proximity of vegetated land areas while the high proportion of polycrystalline and undulose quartz grains may be regarded as indicators of a predominantly metamorphic source area. The palaeocurrent directions obtained from the beds of southern Strath show a dominant flow direction towards the northwest. Frequently the cross laminae are overlain by thin beds of parallel

laminated sandstones (Plate 6.24c) which show predominant flow directions towards the south (Fig. 6.18).

Although the depositional conditions in northwestern and southwestern Skye are not known during this time (due to the lack of exposure), it is probable that deeper marine conditions prevailed (Fig. 6.20, 6.21). A possible interpretation for this facies is the formation of a "spit" which may have formed due to the abrupt landward turn of the coastline. The littoral currents which followed the coastline and held their course, passing from shallow to deeper water (Gilbert, 1890; Evans, 1942; Kidson, 1963), formed what may be regarded as an extension of a beach. It should be pointed out that no feature indicative of definite subaerial exposure was seen in the beds of Facies 7a₁. The presence of features such as those which are interpreted as ?rill marks together with the existence of well preserved coalified plant fragments cannot be held as evidence for the probable emergence of the concerned beds.

The sands also contain a high proportion of ooliths and rounded micritic limestone grains; grain size decreases markedly from Loch Eishort to Ob Lusa.

The term "offshore bar" or "barrier sand" (Price, 1951) cannot be applied to this sand body for the following reasons:

- a. Although the sands were deposited subaqueously, the external morphology of the body is not determinable and an elongate form characteristic of "bars" is not seen.
- b. Barrier islands are normally separated from the shore by a coastal lagoon; although evidence for a restricted depositional environment is seen in Applecross (eastward of the Skye outcrops), there is no such evidence in southern Strath.
- c. The various criteria for the recognition of barrier environments (Davies et al., 1971) and "intertidal sand bars" (Klein, 1970a) are mostly absent and do not fit into an overall framework.

Spits are free forms which are attached to the coast or an island at one point (King, 1972); they grow in the direction of predominant longshore sediment drift and are often a continuation of the beach that is adjacent to the coast (Komar, 1976).

It is difficult if not impossible to distinguish between littoral sandy deposits formed as bars and spits in the fossil record; this is because the structures and the processes forming them are almost alike. The features of the bayward side of the offshore bar or spit consists of layers of short, steeply inclined foreset laminae interstratified with layers of long, gently inclined top-set laminae which also dip bayward (Thompson, 1937). The above situation may be seen at Loch Eishort where cross lamination dips of the individual beds are 20° and 30° in most places; also beds with horizontal stratification with laterally traceable individual laminae are seen.

It should be noted that because these beds are most probably tilted due to various movements (minor faults, dyke intrusions ... etc.) it is hard to establish with certainty, a horizon which was horizontal at the time of deposition; such high angle stratification occurs on both the upper and lower foreshore of modern beaches (Hoyt, 1962).

The landward migration of sand bodies in nearshore beach environments is indicated by steeply inclined foreset laminations dipping shoreward, produced at the shoreward edge of the sand body as it migrated (Hayes, 1972; Wunderlich, 1972; Komar, 1976). Although the external morphology of the sand body of Facies 7a₁ in Loch Eishort is not clearly known, the exposed parts show that the succession was formed by the association of medium-scale cross stratified units alternating with remnants of flat bedded deposits. The possible mode of formation of such beds is hypothesised by Reineck (1960a, 1963a) who mainly formulated it to account for the preservation of this type of cross laminated unit within the interiors of intertidal beach bars. Medium to large scale ripples are built up on the bar whilst the tide is high; the ripples advance towards the land with high-angle stratification ($25-35^{\circ}$)

and form erosional cross-laminated sets beneath themselves. With the ebbing of the "tide" off the "bar", the ripples become planed down and the hollows of the beds are filled up with gently inclined or flat strata. The repetition of this process, i.e. the unsteadiness of the flow system at any rate (Allen, 1968) would be responsible for the creation of the above mentioned sand body. Fluctuating energy conditions may be envisaged during the formation of the sand body of Facies 7a₁. However, it should be stressed that the palaeocurrent data from the steeply inclined cross laminae suggest currents flowing towards the northwest, which according to the Reineck (1960a, 1963a) hypothesis should indicate the existence of land areas to the northwest; this is not consistent with the facies pattern observed in Applecross and Raasay. In the vicinity of modern coastal sand bodies a tripolar or quadripolar current system is to be expected depending on the relative strength of the various currents (Allen, 1968). This is evident in Fig. 6.1b which displays the palaeocurrent data obtained from the rippled, calcareous sand bodies of this facies. A continuous outcrop of sandstones from Loch Eishort to Ob Lusa overlies the patchy development of the beds with Isastrea. This may be due to an original bar-like, elongate morphology along the eastern edge of the synclinal outcrop of the Broadford Beds Arenaceous Group in the Strath area. Although the above discussion mainly concentrates on nearshore bars characteristic of beaches with tidal ranges, it is realised that King and Williams (1949) pointed out that barred beaches typical of tideless seas are very common in the Mediterranean and Baltic seas. They distinguished bars with two different patterns:

- a. Several parallel bars (not more than four), comparatively straight and forming parallel to the water line.
- b. Sand "crescents" with inshore facing points.

In both of the above cases, the "bars" were considered to be permanently submerged below a relatively constant sea level. Such "bars" have been observed in the Caspian Sea, Sea of Azov and the Gulf of Baykal (Zenkovich,

1967) and the western Black Sea (Zenkovitch, 1969).

Although the external morphology and occurrence of these bodies is well documented, details of their internal structures are less thoroughly dealt with.

McKee and Sterrett (1961) investigated the effect of bottom slope, wave intensity and sand supply on longshore "bars". According to their work these features are produced at the point of wave-break. After establishing a close similarity between the beach structures formed during their laboratory experiments and those formed on modern beaches at several localities, they found that large parts of modern longshore "bars" were formed of strata dipping shoreward at 16° - 20° whereas in both actual and experimental beaches the strata dip gently seaward with long, even slopes. The structures produced on laboratory-simulated, modern shoreface terraces show strata dipping 27° - 31° seaward (these dips are somewhat greater than those formed in nature).

This experiment was carried out under deep water conditions (steep floor slope) with strong wave action and continuous feeding of sediments at the base of the shoreface terrace intended to create conditions of sand introduction by longshore and rip currents. Under the resulting conditions, advancing waves moved the sand shorewards to meet other sands carried seaward by backwash, and a landward-thickening wedge was formed comprising shore-derived sands (dipping seaward at 12° - 15°). Seaward of this feature, newly introduced sands advanced landward forming a complicated ripple, scour and fill structure; this deposit would build-up as rapidly as the sand supply permitted. The new sand is also carried shorewards ultimately, contributing to the construction of foreset beds as the backwash redeposits this sand in the form of a shoreface terrace.

Experiments carried out under deep-water (steep floor) and moderate wave action conditions with the sand supply cut off, produced a sand body with a thick series of top set beds which together with the foreset beds show relatively steep dips of 11° - 13° and 28° - 31° respectively. As the

information from the sandstone beds of Facies 7a₁ does not permit detailed morphological observations (e.g. shape, height and width), suggestions towards its mode of formation can only tentatively be made through comparisons with the data provided by modern environments. The relative high proportion of land-derived material (broken, coalified plant remains, polycrystalline quartz fragments and rounded micritic sand grains) together with very well rounded heavy minerals (probably relict) supplied by offshore areas, considered with the cross lamination and minor sedimentary structures seen in Ob Lusa, seem to indicate conditions akin to those operating while sedimentation was taking place in a quiet to moderately agitated, subaqueous environment. The sand supply from the onshore areas (onto a steep floor slope) was moderate and some mixing with offshore-derived, "relict" material is common; no evidence is seen for the existence of an onshore prograding sand body in the lower parts of these beds. The variation in grain size from 0.2 - 0.1 mm. in Loch Eishort to 0.083 mm. in Ob Lusa should be noted, together with the occurrence of low-energy asymmetric ooliths (Freeman, 1962) in the sandstones of the latter locality. Present day stretches of beach demonstrate systematic variations in grain size in the longshore direction (Bascom, 1951; Carr, 1969), as well as across the profile (Krumbein and Griffith, 1938; Krumbein, 1938; Miller and Zeigler, 1958; Greenwood and Davidson-Arnott, 1972). It is impossible to show the latter in the beds of Facies 7a₁, nevertheless a reasonable measure of certainty can be achieved by considering the former situation. A roughly north northwest to south southeast trending sand body may have existed between Loch Eishort and Ob Lusa. The variation of wave energy along its length would have been reflected in the form of a progressive change in particle size, the finer sand accumulating in the low energy, sheltered areas existing further in the north. Examples from modern environments are provided by the Halfmoon Bay, California (Bascom, 1951) which shows longshore sorting due to longshore variations in the wave energy level. Although there is no proof to show the existence of protective

headlands in the vicinity of northern Strath, in order to refract onshore waves and to reduce their energy, this may have been achieved by the combined effect of relative local deepening and abrupt change of the coastline in northern Strath. It should be noted that the equivalent beds of the Milton Formation examined at Sconser (NG 520320) and Suisnish, Raasay (situated to the northwest of Ob Lusa), are entirely composed of blue limestones crowded with broken, fragmented shells of gastropods and bivalves. The absence of terrigenous sediments in this case may signify very shallow, unstable conditions under which terrigenous material could not accumulate. The well washed tabular cross laminated, white siltstone beds marking the top of these beds, may represent partly wind-transported material due to their very fine grained nature (true aeolian features such as concave downward laminations (McKee, 1966b) and reactivation surfaces are totally lacking). The mentioned wedge-shaped, cross laminated sets may also be regarded as characteristic of high energy environments; they possibly signify the action of partly longshore-driven currents which transported the sediment towards the north and northeast in general (see Fig. 6.19 for palaeocurrent orientations).

Typical low-angle, seaward-dipping, wedge-shaped sets of evenly laminated fine sand were described by Howard and Reineck (1972a) from the upper shoreface (beach) environment (0 - 1 m. depth). Beach surfaces dip gently seaward with long, even slopes; the measurements of McKee (1957b) and McKee and Sterret (1961) showed that the angle of dip of beach laminae range from 2° to 5° on the Texas Gulf Coast and from 7° to 10° on the coasts of California. Laminae in the beds under consideration here (top of Facies 7a₁), mostly show dips of 20° to 30° .

Longshore bars often develop along the foreshore; these are made up of wedge shaped layers with tabular cross laminae which dip shoreward at angles of 10° to 30° (Thompson, 1937; McKee and Sterret, 1961; Hoyt, 1962; Bigarella, 1965; Psuty, 1966). Although trough shaped units were reported to be absent, Reineck (1963a) observed festoon-shaped cross bedding associated with

megaripples on the longshore bars; the megaripples were oriented at an angle to the bar and the shoreline. Moreover it was reported that to the landward side of a longshore bar, a channel (trough) was present in which megaripples and/or small current ripples would originate, oriented in the direction of flowing water more or less at right angles to the shoreline (Reineck and Singh, 1973).

Although the exact morphology of the beds under consideration is not well known and as further detailed information was unobtainable from the outcrop, it may be inferred that the top beds of Facies 7a₁ represent a part of an incompletely developed longshore bar.

The sandstones of Facies 7a₁ show bedforms and sedimentary structures commonly reported from barred coasts in shallow water wave dominated environments of the Kouchibouguac Bay (Davidson-Arnott and Greenwood, 1974, 1976) and may be compared with the sequence found in the "trough" and "beach-face" zones; the terminology follows that of Clifton et al. (1971).

Although any interpretation can only be tentatively suggested due to the lack of proper exposure, it can be seen that the beds of Facies 7a₁ resemble those found in the "seaward slope" and "bar crest" facies described by Davidson-Arnott and Greenwood (1976). The unit underlying the top planar cross laminated beds consists of two possible principal types namely ripple and plane beds which are mainly generated by shoaling waves.

The seaward slope facies of the inner and outer bar systems in the Kouchibouguac Bay are characterised by sets of small scale ripple cross laminae which dip predominantly landward interbedded with seaward dipping, low angle, plane bedding; the above association was shown by Davidson-Arnott and Greenwood (1976) to form composite bed sets (described by Campbell, 1967) as "plane-to-ripple" bedding.

The exact determination of the direction of flow in the beds under study is not possible, but in general, bedforms representing the "bar crest" facies are much better developed than those of the "seaward slope" facies.

The basic representatives of the "bar crest" facies are represented in Plate 6.24c; the principal bedforms are "plane beds" and what Clifton et al. (1971) termed "lunate megaripples". Small scale oscillation ripples which may have been generated during periods of low wave activity are present on top of the above mentioned bedforms. These bedforms were probably controlled by wave action breaking on the bar but were also influenced by the interaction of waves with currents flowing seaward across the crest.

The above mentioned characteristics of the lower parts of Facies 7a₁ fit in well with those of the "bar-crest facies"; they consist of sub-horizontal plane beds developed in somewhat coarser sediments interbedded with cross stratified sets possibly produced by the migration of lunate megaripples. The individual sets of plane beds do not exceed 10 cm. in thickness but show a considerable lateral extent. Davidson-Arnott and Greenwood (1976) indicate that the dip of these bedforms varies from gently seaward to gently landward which is controlled by the bar slope at the point of deposition. The "plane beds" under study were probably formed under northerly flowing palaeocurrents (Fig. 6.18).

The "lunate megaripple cross stratified" units are 15-20 cm. thick and extend up to 86 cm. laterally, the long curving toesets and wedge-shaped units characteristic of this type of cross stratification (Clifton et al., 1971; Davidson-Arnott and Greenwood, 1976) are also seen.

The directional data obtained from these units probably reflects the effects of a wave-dominated environment where a seaward flowing (northeast-southwest) unidirectional current is superimposed on the pattern produced by other, wave generated currents. Some of the individual foreset laminae are lenticular and probably reflect a pulsating sediment supply and/or lateral transport under very turbulent conditions (Davidson-Arnott and Greenwood, 1976). These fluctuations in wave strength and seaward/landward flowing currents would result in a complex interbedding of landward

and seaward dipping sets in the present day environment, hence it is hard if not impossible to interpret the current orientation data obtained from fossil rock examples in the absence of proper, three dimensional exposures.

Probably the regional coastline faced obliquely into east-northeast winds with its associated waves and surface currents; this area of the palaeo-coastline may have been partially exposed to the activity of maximum fetch due to the combined effect of palaeowind directions and the position of emergent areas (probably to the west of present day Liassic outcrops) which only effectively protected the areas north of Loch Eishort. These factors may have combined to make this area one of maximum water agitation and effective longshore transport; the development of longshore sand bodies is thus hampered as the result of non-uniform conditions along the coastline.

In Ob Lusa (Facies 7b₁), very fine calcareous sandstones (0.083 mm. diameter) are seen. These are moderately rounded and contain the minimum amount of coalified (land-derived) material. The quartz fragments are mostly composed of metamorphic grains with undulose extinction. The decrease in grain size together with the relatively well washed nature of these beds (samples Am 106, b 66) indicate a progressive winnowing, also the presence of trace amounts of epidote in the Ob Lusa area is suggestive of a localised occurrence. The biomicrite sample Apc 12 (from Applecross, Facies 7c₁) contains gastropod and bivalve fragments. The ooliths are undoubtedly derived from areas with greater turbulence and the rock shows an inverted texture (Folk & Ward, 1957); they were probably derived from extensive oolith shoals which probably formed in lesser protected areas further north.

The biomicrites may represent the existence of a micrite-mud channel separating the sand bodies (Facies 7a₁) in the southwest from ooid shoals to the northeast (Facies 7c₁) shown in Fig. 6.20. The existence of bored micrite pellets in the samples taken from Ob Lusa (Facies 7b₁; Am 108, 109) together with mud-coated quartz grains with undulose extinction is taken to signify the proximity of muddy land areas. A sand body built out from the

southern Strath area towards Ob Lusa under the influence of longshore currents which distributed sediments probably supplied by a large river draining extensive metamorphic lands situated to the south of Loch Eishort, humid climatic conditions most probably facilitated the breakdown and disintegration of bedrock. The presence of ooliths in biomicrites of the Applecross area (Apc 12) is probably due to the proximity of oolitic shoals which existed landwards of the sand bodies. "Barrier bars" of large dimensions seaward of oolith shoals have been recognised in the Pleistocene deposits near Miami, Florida (Halley et al., 1977); as no "barrier bar" features are known to be associated with young, Bahamian Holocene oolith deposits (except in Joulter's Cay), they concluded that these features probably represent later aspects of oolith-shoal development, i.e. given favourable conditions, a maturing oolith-shoal complex will develop a significant seaward "barrier". The broken gastropod, ?ostracod fragments are probably products of prolonged winnowing in the shoal environment (some ooliths contain gastropod nuclei) and their transportation into the main "muddy" channel via minor intershoal channels is a possibility (Fig. 6.20).

The pattern of sedimentation which developed during the formation of sand bodies offshore of oolitic shoals ceased to continue through time, as evidenced by the formation of extensive bioclastic calcarenite sand bodies and oolites in the Skye area with mixed oolith/quartz/mud pellet and ?algae coated grains in Applecross.

The sand body (? "barrier bar") was probably submerged along with the "back barrier" semilagoonal, muddy channels. In Loch Eishort, sandy biopelsparites (Am 49) with rounded micritic grains are seen with elongate, mud-coated, rounded bivalve "ghosts". The quartz fragments show undulose extinction and are mostly polycrystalline. It is possible that the sand body shown in Fig. 6.20 receded to the position shown in Fig. 6.21 due to its rapid submergence which also partially drowned the fringing, "mixed" depositional belt; as a result normally packed ooliths with mostly well formed

multiple-coated spherical grains formed in southern Strath (Loch Eishort; Am 50). The ooliths are medium sand sized (0.32 mm.) and the nuclei are composed of echinodermal fragments with rounded micritic pellets and undulose quartz (< 5%), thus open marine deposition on shoal areas is invoked. In present-day environments such well formed ooliths are found in tidal bars but there is no morphological or physiographical evidence for the existence of such features in Facies 7. The presence of rounded micritic grains and mud-coated, stretched ?pisolitic grains (Plate 6.38) is probably due to the proximity of mud-flat areas. Second generation ooliths are present with micrite-coated oolitic nuclei indicating increased environmental turbulence. In the Ob Lusa area, poorly sorted silty oobiopelsparite wackestone (Am 111) are seen with asymmetric ooliths the single-layer radial (often asymmetric) coating of the micrite-covered nuclei together with the poorly sorted nature of the sample, signifies poor winnowing. The oobiopelsparites of sample Am 112 represent deposition under moderately turbulent conditions; some asymmetric ooliths are present but most are in the process of transforming into proper ooliths. Samples Am 113 to 116 represent a complex system of skeletal, pelletoidal sand bodies which were constantly reworked in a turbulent environment. Skeletal sand bars are strung along the inner edge of the Great Pearl Bank in the central parts of the Trucial Coast on the southern shores of the Persian Gulf (Purser and Evans, 1973).

Although there is no evidence for the presence of barriers in the beds of Facies 7b₁ (Fig. 6.21) it is possible that such sedimentation took place on the submerged tops and along the landward edges of the previously existing ?barrier; the presence of coalified wood fragments with preserved cellular structures (Plate 6.31) indicates the relative proximity of vegetated land areas. The pelletoidal limestones with coarse skeletal debris (samples Am 117, 119, 120) thus represent channels which swept across the bioclastic bodies (forming on top of the submerged ?barriers). These may have been transported by onshore moving currents which considerably broke down and

mixed the shell material with micritic pellets derived from previously formed limestones or locally eroded mud banks. The skeletal grains are not coated and it is evident that they were not extensively winnowed. Sample Am 121 represents the development of mixed calcarenite-bioclastic-oolitic bodies.

In Applecross, a very shallow, poorly winnowed environment developed. Probably the bioclastic channel deposits of Ob Lusa were swept into the quiet, possibly lagoonal environments around Applecross. Such processes are operative today in the Persian Gulf area where, as sediments are swept across the Great Pearl Bank barrier they spill into the Khor Al Bazm lagoon forming steep accretion slopes or small deltas at the end of channels. In Applecross the pelbiosparites of Apc 13 contain a high proportion of large rounded, micritic pellets and reworked, transported ooliths. These are overlain by poorly sorted biopelmicrites with rounded grains and bioclasts mostly composed of echinoderm plates and high spired gastropods. Bivalves including some ?inoceramid fragments are entirely coated by micritic mud with ?algal linings present; most grains show dark, opaque ?organic coatings and patches (Apc 14). It is possible that the bivalves and gastropods were reworked in the wholly muddy environment of a nearshore lagoon; thick coatings of ?algal mud developed around them while echinodermal debris was washed into the environment, barely having the chance to acquire a substantial mud coating.

As a result of the availability of siliciclastic erosion products in the environment, thin beds of very fine (0.08 to 0.1 mm. in diameter) calcareous sands are seen (Apc 17); the presence of pockets of ooliths indicates the possible reworking of these beds. Coalified ?wood fragments associated with other scattered, dark particles points to possible proximal terrigenous sources while abraded echinodermal debris indicate an open-marine connection. The mud-coated bivalves become frequent towards the top of the beds and micrite pellets are abundant (Apc 18). These beds are possibly of mixed origin, the quartz being transported into nearshore, subaqueous environments possibly due to wind action which brought them into the lagoon where clay

deposition was already under way. The presence of cross laminated siltstones (0.06 mm.) in Loch Eishort (Am 51, b 32) at this level, may be analogous to these beds in Applecross; the cross laminated, well washed nature of these beds in the former area points to strong bipolar currents (Figs. 6.18a and 6.19a). As previously mentioned, there is no evidence pointing to transportation under aeolian conditions (although ?reactivation surfaces may be seen in the polished hand specimen of Am b 32); the sediments were transported by vigorous aqueous action probably induced by wind. Possibly the main source of the silts was the development of patches of wind-transported material; beach berms and dunes in the present day are mostly typical of arid regions and they seldom develop under humid conditions similar to those prevailing during the Liassic times in northwest Scotland. Towards the top of the beds of Facies 7 (a_2 , b_2 , c_2), shallower yet less agitated conditions of deposition are encountered in the northern area of study. The development of extensive beds of faintly cross laminated sediments with ?rippled tops in Loch Eishort containing grains of superficial oolites and mud-coated particles indicates the lack of prolonged agitation (samples Am 52, 53); rounded micritic grains indicate the presence of areas with exposed limestone beds or the reworking and redeposition of lagoonal muds which existed in the Applecross area (Fig. 6.21 & 6.22). Poorly reworked bodies of quartz sand are present in northern Strath (Am 122) and comprise poorly sorted subangular, fine sandstones (60% of the grains are 0.2 mm. in diameter while 40% are 0.8 mm. in size). The grains are mostly polycrystalline with sutured crystal contacts typical of a metamorphic origin and the beds are also faintly cross laminated. The strong bimodal grain size of these sandstones points to the probable formation of the finer fraction as a result of strong impact among the larger grains. Although less information is available on the external morphology of these beds, it is possible that they acted as some form of buffer beyond which silty oosparites and oobiopelsparites (Am 123, 124) were being deposited. The high percentage of superficial and

asymmetric ooliths signifies deposition under moderate to low turbulence and admixtures of undulose, polycrystalline quartz sand grains points to the possible association of the sands and ooliths. The equivalent beds in Applecross comprise oobiomicrorites at their base (Apc 20) which were probably deposited in agitated lagoonal channels; some ooliths consist of a bioclastic nucleus, rounded micritic pellets are present and show ?algal laminations. The biomicrosparite of Apc 21 may represent skeletal carbonate muds which developed in lagoonal channels in Applecross where abundant gastropod, bivalve and ostracod fragments and shells were available; some washed-in echinodermal material can also be seen.

Very rich ostracod assemblages occur in the nearshore muds of the Persian Gulf (Hughes, Clark and Keij , 1973). In eastern Abu Dhabi where the Great Pearl lagoonal system is discontinuous, carbonate muds rich in imperforate foraminifera and gastropods occur in the lagoonal ends of the channels and in intertidal and subtidal areas of the lee of the barrier islands. Imperforate forams are absent from the Broadford Beds Arenaceous Group but the laminated biopelmicrorites of Apc 21 possibly formed in partially enclosed, lagoonal channels such as those seen in the present day Great Pearl Bank. Here, marked concentrations of carbonate mud with bivalves are seen immediately nearer shore, partly due to the progressive shallowing and fragmentation of the lagoon and partly as the result of the protection afforded by the barriers. Although this situation can be envisaged for Facies 7 (a_2 , b_2 , c_2), the presence of some echinodermal and crinoidal debris points to a marine connection. The occurrence of thin layers of pelletal, calcareous siltstones (Apc 23) is possibly due to wind action and the erosion of previously formed limestones. High spired gastropods together with bivalves and some echinodermal material are present but are seldom coated with carbonate mud. It is possible that these formed as a result of the erosion of previously deposited relief-forming limestones around which

bioclastic sediments together with calcareous sands developed as accretionary islands of carbonate material becoming intermixed with land-derived quartzitic material.

The absence of thick mud coatings signifies a well agitated environment where the mud was winnowed away. The increasing occurrence of shell debris with thick coatings of algal mud (Apc 24) up the succession, while the quartz particle content diminishes, probably signifies the progressive proximity of broad, flat-lying mud-fringes of the lagoon.

Compound grains are common and the coatings are mostly seen on bivalve and gastropod shells whereas the sparse echinodermal fragments do not usually show coatings. As seen the mud pellets show very faint laminations and some crenulations.

Higher up in the succession, beds entirely composed of large (2-3 mm.) mud-coated shell fragments and pellets are seen in pelletoidal micritic limestones; the coated grains mostly show a two generation origin and the presence of black rinds containing limonitic and some coalified material significantly points to a nearshore environment of deposition. The absence of steep dips or coral fragments excludes the possibility of their being reef-flank detritus produced by the activity of mud producing coralline algae. Although many workers have studied the development of algae in carbonate depositing environments (e.g. Logan et al., 1964; Black, 1933a; Monty, 1967), the work on English and Scottish Jurassic rocks has only revealed the presence of calcareous algae in the Purbeck Beds (Upper Jurassic) of Dorset (Pugh, 1968; Brown, 1963b; 1964; West, 1965; Shearman, 1966) and the Great Estuarine Series (Middle Jurassic) of Skye (Anderson, 1948; Hudson, 1962, 1970); Sellwood et al. (1970) also reported such occurrences from the Junction Beds (Lower Toarcian) of west Dorset and Hamilton (1961) reported algal growths from the Rhaetic, Cotham Marble of southern England. Stromatolites were found in the White Limestone Formation (Middle Jurassic) of central England (Palmer and Jenkyns, 1975) and Solenopora occurs in the Cotswolds Great Oolite

(T.J. Palmer, pers. comm. 1978).

Patches of calcareous algae on the shallow water banks and lagoons of modern seas produce grains that accumulate as carbonate sediment (Blatt et al., 1972). They also disintegrate into 0 ϕ size flakes (size of skeletal elements) and into 10 ϕ carbonate mud, i.e. the size of the aragonite crystals secreted by the algae (Folk and Robles, 1964) which contribute to the formation of the micritic limestone beds (Kendal and Skipwith, 1969a; Lloyd, 1971; Cloud, 1969); Black (1933a) demonstrated the formation of algal laminated sediment in the Bahamas. However, the works of Monty (1965, 1967) and Gebelein (1969) shows that algal sediment can form under subtidal conditions. The consideration of environmental features, also evidence from the underlying and overlying strata cannot sufficiently refute the biogenic origin of the algae-like features seen in the laminated, mud-coated pellets seen in the beds of the Milton Formation but the presence of vadose pisolites (Dunham, 1969b), oncolite-like calcrete deposits (Read, 1976), reworked speleothems (Thrailkill, 1976) or geyserites (Walter, 1976) can confidently be ruled out.

6.13 Breakish Coral Member

This member is represented by the beds of facies 8 which comprise ~ 3 m. of lenticular (1.30 m. long - 25 to 30 cm. thick) limestones in Applecross and northern Strath. These show wavy lower and upper surfaces and are interbedded with calcareous shales (Plate 6.39a, b) which contain up to 25% smectites. Each limestone lens is a mound-like structure, mainly formed by the branching coral Thecosmilia martini, in a microsparite cement. These lenses contain hair-thin seams of clayey material within which quartz fragments are seen. Some indication of syndepositional compaction is seen but there is no evidence of erosion on a microscopic scale. Thecosmilia colonies seem to abut against the layers in thin section and also in the field. As seen, most of the

in situ corals seem to be inclined with respect to the bedding surface. In addition to the coral Thecosmilia, other fossils reported from this bed (Hallam, 1959) are Calcirhynchia calcaria, Gryphaea arcuata, Isocrinus sp., Plagiostoma gigantea, Ptychomphalus expansus, Zeilleria perforata. In the Ob Lusa section the shale interbeddings which are so well developed in Applecross, are totally baked due to the intrusion of a doleritic sill, the base of which lies on top of the uneven surface of the coral bed.

This facies is not fully represented by corals in Loch Eishort. As in the coral Facies 6, the coral structure is only well preserved within the outer few millimetres of the beds, with the corals appearing as masses of sparry calcite to the naked eye. Silt and sand grade detrital quartz material is less common. The shale interbeddings contain echinodermal, bivalve and brachiopod fragments, the bivalves include calcitic shells such as pectinids, oysters and limids. Recrystallised aragonitic shells with micritic envelopes coating irregularly developed crystals of sparry calcite are common in the limestone beds.

6.14 Interpretation of the Breakish Coral Member

Like those of Facies 6, the beds of Facies 8 are laterally impersistent and show no talus deposits, therefore they probably did not stand much above the surrounding deposits.

The micritic, peloidal matrix, the relative scarcity and dendroid growth habit of the coral and the presence of the widely adaptable Thecosmilia (which commonly formed in fore-reef and back-reef environments of the Northern Calcareous Alps) argue strongly against any wave resistant coral-reef-rim interpretation for the beds of Facies 8. Bivalves are very common with some gastropods. Brachiopods, bryozoans and foraminifera are not seen and crinoid debris is found as small patchy developments which together with the presence of cidarid echinodermal remains probably represent accumulations of an open

marine flank of the "banks". The constituents may be termed "lime-mud skeletal limestones" (Dunham, 1962; Heckel, 1974). The shape of the body representing this facies is not readily recognisable, nevertheless the outcrop is linear and appears to have separated different facies on either side, but lack of positive evidence precludes the usage of the term barrier. There is no evidence for the existence of shells or colonies which were permanently attached to or encrusted on other objects (except for some ^{or foraminiferal} algal/encrustations on the coral walls, shown in Plate 6.40). Most of the shells are abraded and the corals show no direct evidence of attachment or binding (although they are near their original growth position). The beds of this facies may be regarded as representing an abraded, "mixed-diverse skeletal-lime-mud build-up". The elimination of signs of original attachment may be attributed to diagenetic changes and the lack of proper exposure; the autochthonous, branching Thecosmilia probably provided a rigid organic framework around which, in place and derived skeletal material and other organically induced (and/or derived) carbonate material accumulated. The corals originally grew as patches and therefore induced locally non contiguous buildups probably due to the limited areal extent of favourable conditions for larval settling and luxuriant adult growth. This facies is composed of an average of 1 to 3 beds of similar composition in the various outcrops. Advantages gained from such "buildups" probably included the following (Heckel, 1974):

- i. Providing suitable habitats for organisms unable to live elsewhere.
- ii. Inducing better water circulation patterns for nutrient and oxygen replenishment and waste removal.
- iii. Improving or maintaining a position in the euphotic zone.

The first and third points are of great importance here. The continual upward growth of the corals which is intermittently interrupted by 3 to 5 cm. thick shale beds together with their known sunlight requirements indicate a gradual rise in sea level or subsidence of the land areas.

Several modes of origin for carbonate buildups have been discussed by Heckel (1974). It is possible that certain organisms found a more favourable environment on an initial buildup, somewhat like those of the "intermediate homocline highs" (Purser, 1973) of the Persian Gulf, and then preferentially colonised it. The position of this buildup may have been further stabilised by the secretion of carbonate material; sediment trapping and binding by the rooted, attached corals, further assisted the resistance of the buildup against movement or destruction by waves or changes in current directions. The profuse growth of erect organisms (e.g. Thecosmilia) on the buildup may have impeded water movement over it and caused more sediment to settle-out (Heckel, 1974). The stabilisation of sediment and abraded shell fragments may have been further enhanced by the action of subtidal gelatinous organic mats which have been described from the carbonate sand areas of the Great Bahama Banks (Bathurst, 1967c; Scoffin, 1970; Neumann et al., 1970). Although no trace of such organisms is left after burial, such gelatinous mats would stabilise a buildup until it was buried. Pusey (1964) found that similar organic films augmented the naturally greater cohesiveness of carbonate muds and further stabilised the buildups in northern British Honduras, rendering them even more resistant to current erosion. Carbonate mud buildups in Florida Bay were found to be little affected by the powerful hydrodynamic forces generated by hurricane Donna (Ball et al., 1967). There is no direct evidence for algal presence in the beds of Facies 8, but the features exhibited by the pellets and the rounded, micritic grains present in the beds of Facies 7c₂ strongly suggest their presence. In places the coral bed consists of a biopelmicrite wackestone containing a varied fauna ranging from recrystallised disarticulated ostracod shells to crinoidal and echinodermal debris which are coated by algal mud in many places ; these possibly represent "interbank" areas which received abundant mud pellets and shell debris from the "bank" areas. Although direct evidence cannot be

obtained, it is possible that the coral banks also developed first, on piles of bivalves and crinoids and grew by the development of Thecosmilia heads in calcareous mud. Thecosmilia was very adaptable and is also commonly found in a micritic matrix. The micritic peloidal beds with coral and other shells represent rim deposits probably formed below the effective wave base. Also the existence of ostracods in the beds of Facies 8 may indicate the occurrence of fresh to brackish waters nearby and would strengthen the notion of the activity of ?blue-green algae. A blue-green algal origin for Jurassic buildups of the Navajo country was also suggested on the same grounds (Harshbarger et al., 1957). Thus stabilised, the buildup was probably augmented by a "regenerative feedback" process (Ball et al., 1967).

It is possible that the minor buildups of Facies 8 resulted from a process involving both organic and hydrodynamic factors continually reinforcing each other to promote localised accumulation of sediments. The corals of Facies 8 are termed hermatypic due to the general possession of unicellular algae (zooxanthellae) in their tissues, but it cannot be proved that these algae had evolved by the early Lias and it is impossible to determine when coral stocks became symbiotic with the unpreserved zooxanthellae; this symbiotic relationship probably provided them with an efficient excretory mechanism (Vaughn and Wells, 1943), but which restricted them to warm, sunlit waters. These hexacorals remained major contributors, particularly of framework, to most tropical marine, shallow water, reef complexes of all sizes up to the Holocene; solenoporid red algae, stromatoporids and/or hydrozoans were major binding organisms in these reefs up to the Eocene. The modern restriction of hermatypic hexacorals to warm water suggests that older reefs of this general composition have been tropical or subtropical for some time (Heckel, 1974). As evidenced by the associated fauna, Thecosmilia corals were stenohaline therefore an environment of deposition with normal marine salinities may be envisaged for Facies 8.

As indicated in Chapter 5 a shallow marine origin is confirmed by the relative low faunal diversity of the buildup assemblage.

By studying the general models for the formation of organic buildups proposed by Heckel (1974), it is evident that Facies 8 represents buildups formed on topographic highs. These may have been organic buildups on the sea floor left from a previous sedimentary regime, "hydrodynamic" or "aerodynamic" accumulations such as dunes and lime mud or sand bars, drowned erosional residues, differentially less compacted sediment or tectonic features. Preferential organic proliferation involves:

- i. Sunlight for the growth of calcareous algae and algae-dependant invertebrates.
- ii. Better circulation, hence better nutrient provision and waste removal which promotes prolific organic growth.
- iii. Position above deeper, stagnant, saline or anoxic water that is inimical to life.
- iv. Decreased likelihood of being overwhelmed by terrigenous sediments.

The buildup probably grew upwards as a counterbalance for bottom subsidence (or sea level rise) and/or grew laterally to compensate for sediment production in excess of that needed to compensate for subsidence. It seems that these beds were at times unable to match the rate of subsidence and were periodically inundated by muddy waters followed subsequently by a steady growth after a temporary stabilisation of the environmental conditions. factors
Such/limiting coral growth are also observed in the Upper Oxfordian of England (Ali, 1977). Buildups of Facies 8 probably formed on uneven floors with irregularities similar to the "intermediate homocline highs" (Purser, 1973) of the Persian Gulf, where the living "reefs" develop on all flanks of these features except their leeward sides (see Fig. 6.23).

6.15 The Upper Sand Member

In northern Strath and Applecross, the coral Facies 8 is overlain by a succession of interbedded fine sandstones and marly limestones which show wavy contacts and are cross-bedded in places (Facies 9). At all localities this facies may be divided into a lower unit (9a) of limestones and an upper unit (9b) of alternating limestone/sandstone beds.

i. Limestone Facies 9a

In Ob Lusa a 1.10 m. thick sill immediately overlies the coral Facies 8; this has a variable dip and although it lies parallel to the bedding on the coralliferous limestones (as seen along the shore) it penetrates the strata obliquely in the outer reefs. It is therefore considered that bed 10 of Hallam (1959) is part of the limestones of Facies 7c₂ which apparently overlie the sill. The lowermost part of facies 9a is poorly exposed in southern Strath. A 4.10 m. gap is seen above the sill at this stratigraphic level, but at low tide a 0.50 m. thick limestone bed can be seen underlying the sandstones which constitute the outer ledges here; limestones which are regarded as the basal beds of Facies 9a are thin bedded (10-15 cm.) blue calcilutites. Petrographically these are matrix supported biopelsparites; sand size quartz fragments are seen and are mainly polycrystalline. Recrystallised bivalve and ostracod shells are abundant but echinodermal fragments are not common. Coral fragments are seen which are coated with layers of finely broken coalified debris and pyrite; small scale, round cellular structures are also present (?algae) and rounded micrite pellets are very common (very fine sand size) (Plate 6.41).

ii. Sandstone/limestone Facies 9b

The beds of this facies are composed of a basal, thin (3-4 cm.), parallel bedded creamish weathering unit; these are well sorted, washed calcareous

siltstones (4 ϕ).

The feldspar content of these beds ($\sim 5\%$) is mainly microcline. Polycrystalline quartz fragments with > 3 crystals per grain occur and no coating is formed on the grains, mica is rare and rounded limestone pellets are found. Towards the top of this unit (7.8 m.), the thin parallel-bedded sandstones become wavy and their thickness varies within short distances; they are very fine to fine grained, thin (1.5 - 3 cm.), undulating, sandstone sheets (Plate 6.42) interbedded with broad lenses of limestone up to 1 m. in length.

The sheets are laterally persistent, cross bedded and show hummocky surfaces. The cross bedding data is presented in Fig. 6.24 and shows a predominant inclination of the sets towards the northwest; the tabular beds show nearly horizontal stratification. At the base of this unit the sandy beds are very thin bedded and are transitionally interbedded with silt beds. At the interfaces of the sheets of sand, coalified debris 2 - 3 mm. are seen and the average thickness of the sandstone beds decreases towards the top of the unit (Plate 6.42) with the sands becoming finely laminated.

In the lower parts, the contacts of the sandstone sheets and the limestone lenses are well defined and somewhat erosional, lateral equivalents of the higher parts of these show transitional, extensively burrowed contacts.

The sandstone beds of this facies show the characteristics of hummocky cross stratification (Harms et al., 1975) which are as follows:

- i. Lower bounding surfaces of sets are erosional and commonly slope at angles less than 10° , though dips can reach 15° .
- ii. Laminae above these erosional set boundaries are parallel to the lower bounding surfaces, or nearly so.
- iii. Laminae can systematically thicken laterally in a set, so that their traces on a vertical surface are fan-like and dip diminishes laterally.
- iv. The dip directions of erosional set boundaries and of the overlying laminae are scattered.

The limestone lenses are mostly poorly sorted, matrix-supported silty biopelsparites; superficial ooliths are very common and reworked ooliths occur in the limestone lenses together with micritic limestone pellets (rounded). The overall quartz content of this facies increases towards the top; the thickness of the sandy sheets increases and they become poorly sorted; however the median grain size remains constant (3 ϕ) and the feldspars are predominantly microcline. These beds are represented along the road leading from Broadford to Heast (NG 652182), where the topmost beds of this facies are exposed.

Here the sandstone beds are only faintly cross laminated and a gradational, bioturbated contact is seen between the limestone and sandstone beds. The sandstones have become much thicker and are laterally persistent.

Towards the western side of Beinn na Chairn (NG 635787), beds of this facies are seen to overlies the Breakish Coral Member. Matrix supported, sandy biopelsparites with up to 20% quartz alternate with sandy lenticular (shaly) partings (Plate 6.43). Here the sand-shale alternations are faintly cross laminated together with the well indurated limestone lenses (data presented in Fig. 6.25). Subangular quartz pebbles occur at the base of these sandy beds with diameters of 2-3 cm.; the limestones are sandy biopelsparites (grain supported) with 20% quartz, which are poorly sorted, subrounded, fine to medium sand size (2-3 ϕ).

The quartz fragments are mostly polycrystalline (> 3 crystals per grain), reworked; rounded limestone pellets (0.1-0.2 mm.) comprise 15-20% of the calcareous beds. The bioclasts are mainly echinodermal and bivalve fragments together with totally recrystallised shell material, of which only a thin "micrite envelope" (Bathurst, 1971) remains (30%). Ooliths are not common. The sandy lenticular partings vary in thickness from 3-5 cm. to a few millimetres. Although the quartz content of the sandstones increases towards the top of this facies, the sorting is much poorer and the limestone beds thicken at the expense of the sandstones.

Although the Breakish Coral Member is not seen in Loch Eishort, the

beds overlying the Lower Sand Member (Facies 7a₂) overlie a 2.50 m. thick unit of limestone lenses (10-15 cm.) interbedded with sandstone sheets (4-8 cm.). At the base of this succession the matrix of the limestones contains abundant large micritic pellets and some coral fragments.

Towards the top of this succession the limestone/sandstone alternations are similar to the same development in northern Strath, but the sandstone sheets are much thinner and do not show cross lamination.

In Applecross the coral Facies 8 is overlain by a development of interbedded sandstone and silty micaceous shales. The sandstone beds are bioturbated and show wavy contacts with the shales; they are 20-30 cm. thick and the occurrence of benthic fauna (ammonites) in the shale beds was reported by Hallam (1959). The shales are dark and calcareous at the base of this facies with a sand content of 8-10%; towards the top of this unit the sand content rises to 35% while the CaCO₃ diminishes proportionally, the total thickness here is 9 m. The undersides of the sandy limestones show very well developed Thalassinoides burrows. At the base of this facies the sandy limestones are matrix-supported silty biopelmicrosparites, which contain up to 30% of moderately sorted, silt size (4ø) quartz grains.

Towards the top the sandy limestones contain up to 35% quartz grains (most of which are polycrystalline) with 3% microcline feldspars. The sandy oobiomicrosparites are poorly sorted with sizes ranging from 2ø to 3ø; coated grains are mostly of pseudo-oolitic origin; some superficial ooliths are also present. The superficial oolitic rims commonly form around the pseudo-ooliths. Multiple ooliths are found together with rounded micritic limestone pellets.

6.16 Interpretation of the Upper Sand Member

Facies 9a and 9b

The bedded biopelmicrite wackestones with rounded ?algal mud clasts and a reworked assemblage of bivalve and echinodermal shells indicate and probably represent the fringing rims of coral bank developments. The presence of ?ostracod shells indicates deposition in a marine environment. The succeeding lensoid limestones and sandstones of Facies 9b are taken to represent very shallow marine, nearshore depositional conditions either due to a sudden influx of sediments or as the result of a general lowering of sea-level (regression). The samples Am 54 to 58 taken from Loch Eishort represent this facies in southern Strath. The sorting of the quartz grains within the sample improves strikingly from Loch Eishort to Applecross and the fine sand (0.2 mm.) quartz grains of southern Strath are represented by very fine sands (0.08 mm.) in Applecross; the echinodermal content also diminishes towards Applecross.

In the Ob Lusa area, very fine sandstones with hummocky cross stratification (Harms et al., 1975) are seen. The characteristics of this feature have been described in previous sections.

The described characteristics suggest that during deposition, the beds were scoured into low "hummocks" and shallow "swales" which were not well organised in their orientation (Harms et al., 1975). The resulting topography was then mantled by laminae of material swept over these features. The hummocks are 65-75 cm. high and are spaced up to 1.40 m. apart. These structures have not been produced in flumes or wave tanks and their formation has not been observed in present day environments, therefore their interpretation is rather speculative. These features are only seen in the very fine sands of Ob Lusa which are interlayered with lenses of pelletoidal limestones with a mixed, shallow marine fauna. The echinoderm debris content of these beds diminishes from southern Skye to Applecross and more or less

uniform palaeocurrent flow directions are seen.

Although wave ripples were not observed, the lack of well organised directional trends and the mantling relationships of laminae to lower set boundaries with sharp basal contacts suggest formation by strong surges of varied direction, generated by relatively large storm waves.

"Large storm waves in shallow offshore environments have been observed to cause bottom surges with a few metres of displacement under each passing wave crest and trough. With such surge velocity begins at nearly zero, and the bed develops embryo ripples that quickly disappear and then flattens with increasing velocity. At peak velocities that may be on the order one metre per second or more, fine material moves as a suspended layer above flow reverses. Stormy seas, dominated by locally generated waves, have complex surface patterns, and the bottom wave surges and current drift can be varied in their direction". (Harms et al., 1975).

Hummocky cross stratification is attributed to strong wave action with surges of greater displacement and velocity than those required to form wave ripples and a bed of hummocky cross strata may be the product of a storm event (Harms et al., 1975).

The mentioned features are confined to the Ob Lusa area in Skye. It is possible that very local topographic differences were also influential in the formation of these beds. The succession in Applecross is burrowed and Thalassinoides markings are seen at the underside of beds. The abundance of large, coalified material indicates vegetated, adjoining land areas in Applecross, whereas at Ob Lusa a black material lining the thin beds is tar-like and appear to exude out of the bed interfaces, showing no evidence of reworking.

6.17 The Breugh Pebble Member

The Breugh Pebble Member is subdivided into three facies as follows:

i. Bedded sandstone Facies 10

ii. Pebbly Facies 11

iii. Rippled sand Facies 12

i. Bedded sandstone Facies 10

This generally termed facies includes deposits ranging from very fine to very coarse sand. The coarser beds are mostly pebbly and sorting is variable. The grain size varies and may be medium to very coarse sand.

Pebbles are present in the form of bands in these beds which become interbedded with calcareous siltstones towards the top of the unit; they show

faint cross lamination together with wavy contacts.

In places the

calcareous siltstones contain coalified plant material.

these
Laterally /

beds grade into very thin bedded sandstone sheets, groups of which have a fan-like appearance.

These are well sorted friable, fine sandstones with coalified plant material scattered along the bedding planes. Their thickness varies from 1-2 cm. and they are very faintly cross-laminated; the total thickness of this unit is 2.5 - 4 m. In northern Strath, towards the western side of Braigh nan Skulamus, this facies is probably represented by the faintly cross laminated sandstones which contain thin quartzitic pebble beds. The pebbles are well rounded and up to 3 cm. in diameter (Plate 6.44).

ii. Pebbly Facies 11

This comprises pebble gravels in which poorly defined horizontal stratification is sometimes apparent. The pebble beds are 20-60 cm. thick, well rounded, sorted, clast-supported (Plate 6.45) and show grading. They are interbedded with laterally persistent (and in places impersistent) lenses of siltstone or cross laminated sandstones. The clasts are not imbricated

but lateral clast size variation is present, e.g. the same facies is represented in northern Skye by rounded coarse-sand pebble grains; however the coarsest grains are, in places, considerably larger (up to 8 cm. diameter) than those found in Applecross (Plate 6.46). In places these pebbles occupy broad-shallow channels typically 20-30 cm. deep and 1-2 m. wide (Plate 6.49). The channels sometimes cut into each other both laterally and vertically, all showing an erosional base. As seen in Plate 6.45, the pebble beds grade upwards into finer, better sorted sandstones which show evidence of rippling on their top surfaces.

iii. Rippled sand Facies 12

The estimated total thickness of this facies is 2 m. (Plate 6.48). The topmost beds of Facies 11 grade upwards into 50 cm. of lenticular sheet sands containing abundant rounded quartz pebbles. These are poorly sorted, well rounded and contain pink quartz fragments (the composition of these beds is shown in Fig. 6.26). This lower lenticular pebbly sand unit is followed by 75 cm. of well sorted, friable sandstones showing linsen and flaser-type structures. The topmost unit shown in Plate 6.48 b consists mainly of ripple cross laminated sandstones. The ripples are asymmetric with amplitudes of < 5 cm. The grain size of the sands ranges from coarse to very fine and the beds contain rounded quartz pebbles. The ripples may be classified as climbing ripples due to their appearance as Kappa and Lambda-cross stratification (Allen, 1963).

The undifferentiated, cross bedding orientation data are portrayed in Fig. 6.27; these were mostly (80%) obtained from beds of Facies 11.

The various facies of the Breugh Pebble Member were not found in the Strath district, however a distinct but laterally impersistent pebble bed overlies the beds of Facies 9 in Ob Lusa (Plate 6.46), which reappears as isolated NE-SW trending elongate bodies, as one traces it from Ob Lusa to Ob Breakish.

In northern Strath the beds immediately overlying the Pebble beds between Ob Lusa and Ob Breakish comprise calcareous siltstones with impersistent coal beds up to 2.5 cm. thick; these are in places overlain by extensively bioturbated siltstones which are covered locally by 1 m. of very fine sand showing rootlet markings.

6.18 Interpretation of the Breugh Pebble Member

It is hard, if not impossible, to make a coherent, detailed interpretation of the strata representing Facies 10 to 12 in the northern area of study. A consideration of the sedimentary features will help in deducing the physical processes which gave rise to the three facies of the Breugh Pebble Member. It is important to note the generally unidirectional palaeocurrent pattern together with the erosive nature of the basal contacts of the pebble and gravel beds in Facies 11. Fining upward sequences of calcareous gravel, sand and silt are attributed to waning current velocities as previously incised channels were being infilled.

A partially, if not wholly terrigenous (fluvial) depositional environment may be inferred for the above mentioned facies based on a combination of the following evidence:

- i. Abundance of coalified fragments and the occurrence of very thin (2 cm.) coal beds capped by a rootletted horizon.
- ii. Transportation of pebbles and gravels by scouring and channelling.
- iii. Generally unidirectional palaeocurrent data.

The sparse occurrence of crushed bivalve shell fragments and oolites in the various beds indicate marine influence and winnowing effects, as the mentioned facies conformably succeed and are also overlain by beds of undoubted marine origin. Two possible genetic relationships should be considered in detail; firstly the braided stream model is examined which is then complemented by an example of a coarse grained prograding shoreline.

It is believed that the Breugh Pebble Member was deposited under a combination of the processes operative in the two considered depositional environments, the influence of the latter being of greater importance.

i. The braided stream model

The braided stream model of the basic fluvial systems outlined by Fisher and Brown (1972) is considered as a possible framework for the genetic interpretation of the features examined in the Breugh Pebble Member; the three facies comprising this Member are shown in Fig. 6.28.

In braided streams the flow is split by mid-channel bars on a small scale in relation to the channel width (Collinson et al., 1977).

a. Setting

Braided streams are best developed as upstream, very high bed-load facies where slope is relatively high. The increased slope of laboratory channels in erodible sediments induces a progressive change in plan view from straight through meandering to braided channels for a given discharge (Leopold and Wolman, 1957; Lane, 1957; Schumm and Kahn, 1972).

This facies is normally deposited under low flash-discharge conditions and develops locally downstream, where it traverses noncohesive fill or along a zone of tectonic or physiographically induced slope increase. Due to the sedimentary features and characteristic outcrops of the Breugh Pebble Member it is not possible to comment on their mode of development.

b. Geometry of channel units

The channel units are essentially flat bedded, discontinuous, lenticular and tabular; longitudinal bars are commonly about 1.20 m. wide and up to 120 m. long (in well exposed areas) with a thickness of up to 0.6 m. While the axes of longitudinal bars are parallel to flow, transverse bar axes are normal to flow.

Leopold and Wolman (1957) showed that experimentally, mid-channel

longitudinal bars develop by the segregation of patches of coarser grains in sandy channels of mixed grain sizes, the patches then grow both vertically and laterally eventually splitting the flow by the down-cutting of the deeper areas to either side. These bars show a complex pattern of growth and development (Krigstrom, 1962); on the top surface of the bar (pavement) sediment accretion takes place by the addition of material whereas at the downstream face it advances by sediment avalanching.

Internally the bars show horizontally bedded gravels which indicate the vertical growth of the bar top pavement; these pass downstream into a cross bedded unit reflecting the downstream advance of the avalanche face (Collinson et al., 1977). The clasts in the Breugh Pebble Member are not sufficiently flat to show any imbrication and the cross bedding and laminations do not show indications of water stage fluctuation by the presence of discontinuities ("reactivation surfaces" of Collinson, 1970).

Large structures such as channels show the lowest directional variability in the braided stream environment (Allen, 1966). Pebbly cross bedding shows a low variability which will only develop at relatively high water stages; this contrasts the directional structures found in the sands which are also active during falling stages when flow is highly complex as it finds its way around bars (Collinson, 1971; Bluck, 1974). It is possible that the asymmetrical growth of longitudinal bars by deposition on only one of the two oblique downstream facing margins could lead to an overall bimodal pattern of palaeocurrents derived from cross beddings, with the modes distributed symmetrically about the downstream direction (Collinson et al., 1977). This situation may be seen to exist in the palaeocurrent data obtained from the beds of this Member (Fig. 6.27). It can be seen that two modes ($200 - 190^{\circ}$ and $230 - 240^{\circ}$) are almost symmetrically distributed about the probable downstream direction ($210 - 220^{\circ}$).

In general as the rapid shifting of bars and channels in a pebbly braided stream is likely to produce a rather complex and random assemblage, the

internal organisation of the sediment will not fall into any well ordered pattern.

As noted earlier, the important elements in the braided stream environment are bars. Numerous terms have been used to describe them (Smith, 1971a); works by Hein and Walker (1977) have indicated that certain bar types are members of evolutionary sequences and that the morphology of any particular type may be very ephemeral.

The presence of the different types of bars (Miall, 1977) in the Breugh Pebble Member is very questionable and may only tentatively be inferred to exist in the beds of Facies 10 and the upper parts of Facies 11, where fining upward pebble beds are seen together with horizontally bedded units; these show erosive bases and are no more than 50 cm. thick with a lateral extent of 2 to 3 m.

No comment can be made on the possible morphology and other channel characteristics. However, as mentioned earlier, ?channel deposits resting on scour surfaces containing basal lag deposits are seen; these have been reported from ancient braided stream deposits by Miall (1970a), Smith (1970), Cant and Walker (1976). An important observation by Cant (1976) should be mentioned here. He stated that in a braided river, the lower parts of the sedimentary structures created by large bedforms will tend to be preserved because smaller bedforms cannot cause deep erosion. The deposits of infrequent, large floods should therefore be preserved selectively (Miall, 1977). This process would explain the occurrence of incompletely fining upward cycles in the beds of the Breugh Pebble Member.

Miall (1977) interpreted the principal depositional facies of braided streams in terms of five main processes, none of which can be related to the development of the beds under consideration with any certainty.

c. Vertical profile

The identification of a vertical profile and lateral lithological

variability are the most important factors for the positive interpretation of facies assemblages and depositional environments. Several types of repetitive vertical sequences, or cycles, were described for the braided river environment by Miall (1977), who has constructed some tree diagrams from published descriptions or illustrations of vertical sequences in various modern braided stream deposits; these together with published Markov chain analyses of measured ancient stratigraphic sections (see Miall, 1977), are used here to establish a meaningful comparison with the Breugh Pebble Member. The crude fining upward cycles shown by the beds of Facies 11 may indicate lateral accretion at palaeochannel bends such as those observed by Williams and Rust (1969) in the Donjek river. The Lower Silurian Shawangunk conglomerate of the north-central Appalachian Mountains (Smith, 1970) also shows a development of facies similar to those of the Breugh Pebble Member.

The Donjek River (Rust, 1972) shows varied facies with fining upward cycles of different scales. The thicker cycles are regarded as reflecting sedimentation at different topographic levels within the channel system; successive events of vertical aggradation followed by channel switching can also be invoked (see Miall, 1977).

The principal facies assemblages in the Breugh Pebble Member are roughly similar to those described from the modern examples of the Donjek River (Doeglas, 1962). Similar examples from the ancient record are also described by Kelling (1968), Steel (1974), Corner (1975) and Cant and Walker (1976).

It is interesting to note that in some of the fine silty and calcareous beds of the Breugh Pebble Member (Facies 10), indications of marine organisms and mud-coated quartz grains are seen which denote a background of marine depositional conditions. Braided rivers which flow directly into the sea are not common at the present; most pass through a deltaic setting (Van de Graaf, 1972) and some evolve downstream into meandering, suspension-load rivers as they approach base-level (Miall, 1977); deltas with braided distributary channels were probably abundant before the appearance of extensive land

vegetation (Schumm, 1968a), these have been termed fan deltas (McGowen, 1970; McGowen and Scott, 1974), being identified and described in the stratigraphic record by many workers (Miall, 1976b; Young, 1974; Young and Jefferson, 1975; Flores, 1975). It is hardly convincing to envisage the ?braided stream of the Breugh Pebble Member debouching directly into the sea due to unusually sparse land vegetation and sudden increased sediment supply from the hinterland; the interbedded fine grained units with indications of marine influence, being of probable tidal origin.

Despite the foregone similarities and explanations, the braided stream depositional environment cannot be accepted as responsible for the formation of facies 10, 11 and 12 for the following reasons:

- a. The evidence of marine influence is overwhelming and seen throughout the section.
- b. There is no firm and convincing evidence for the existence of mid-channel bars.
- c. Evidence for subaerial exposure is minimal.
- d. There is no evidence for overbank sedimentation.
- e. There is no firm evidence showing the sinuosity of the (?)stream.

For the above reasons, the depositional facies of prograding shorelines are also considered.

ii. Prograding shoreline deposits

The seaward building of foreshore and shoreface environments (Reineck and Singh, 1973; Harms et al., 1975) as facies layers resting on offshore deposits, occurs when sediment is added to a shoreline profile during stable sea-level conditions (Harms et al., 1975). This definition of progradation was broadly accepted and applied by Sears et al. (1941), Bernard et al. (1962) and Curray et al. (1969). In lateral distribution, the foreshore and then shoreface zones should drop out of the sequence moving toward the seaward limits of the prograding wedge. Although continuous outcrop is lacking and

the contact of the Strath and Milton Formations is faulted in southern Strath, the general lateral NE-SW fining of the beds of the three facies should be noted.

Examples of prograding sandy shoreline and conglomeratic prograding shorelines are described by Harms et al. (1975). These are compared with the beds under consideration and their similarities are emphasised.

a. The prograding sandy shoreline

An example from the Cretaceous Gallup Sandstone, southwestern San Juan Basin, New Mexico (Harms et al., 1975) shows the successive development of four facies; these compare well with the beds under study.

1. Offshore

Shales found beneath the shoreline sandstones contain marine fossils. These laminated siltstones and shales represent the offshore environment and thin beds of hummocky cross stratified sandstones are seen which are burrowed.

Although the sandstone beds under consideration do not succeed thick marine shales such as those described by Young (1955), they overlie a succession which may be termed "transgressive, regressive" and the shale beds overlying the Breakish Coral Member (Facies 9a, b) in Applecross have yielded ammonites (Hallam, 1959; Oates, 1978) which are generally accepted as marine faunas.

2. Lower shoreface

This depositional environment consists of hummocky cross stratified sandstones and laminated siltstones which are burrowed.

The characteristic type of cross bedding mentioned above is seen in the beds of Facies 9 in northern Strath, together with the almost horizontally stratified beds of fine sandstone signifying a continuous influence of strong surges of various directions, probably induced by a combination of shallower

depths and strong wave action with surges of great displacement and velocity.

3. Upper shoreface

The medium coarse to pebbly sandstones with abundant trough and some tabular cross stratification with dips showing preferred orientations to the west-southwest (possibly offshore in the palaeogeographic reconstruction) are best exposed in the Applecross section. The trough sets lined with pebbles show upward grading of the clasts. These may be interpreted as having been deposited by currents moving dunes or, for the more tabular sets, by sand waves. The directional data suggest that dominantly unidirectional flow systems carried siliciclastic debris from the east or northeast into a depocentre situated somewhere in the western or southwestern areas of northern Strath. The dominance of pink quartz and reworked chert fragments over other transported material suggests a Torridonian (?Applecross Group) source. Rapid transportation under turbulent conditions is indicated by the fining upward, discontinuous nature of the pebble beds. There is no record or evidence of flow reversals. Horizontally stratified beds and tabular sets imply fluctuations of somewhat higher and lower flow strengths.

4. Foreshore and backshore

No evidence of low angle swash cross lamination, characteristic of beach deposits, was found in the topmost parts of Facies 11 or 12. Although some tabular cross laminated sandstone beds are seen in Applecross, the general lack of foreshore, swash cross stratification may also be attributed to removal due to erosion. Although coalified material is common in the uppermost beds of Facies 11 and 12 there is no sign of root markings and/or coal beds in southern Strath; in northern Strath however laterally impersistent calcareous sandstone sheets with root markings and thin coal beds are seen. It is suggested that the thin, discontinuous rootletted horizons overlying very thin coal beds in northern Skye represent periods of

subaerial exposure. The topmost part of Facies 12 is composed of a 1 m. thick, small-ripple bedded pebbly sandstone unit. Small ripple bedding is made up of foreset laminae produced from the migration of small current ripples (Reineck, 1961, 1963a, 1967a). Such current ripples may form in any environment with non cohesive sediment; most commonly they form in the inter-tidal flat, tidal channel and inlet environments (Reineck and Singh, 1973), but as there is no criterion suggestive of tide-dominated depositional conditions (e.g. herringbone cross-lamination and other data for reversible currents), it is most probable that they represent backshore and foreshore deposits.

It is seen that although some features described from the vertical profile of Facies 10, 11 and 12 fit very well into the general regressive sandy shoreline models significant differences exist. Recent examples of such sandy shorelines have been described from Galveston Island (Bernard and others, 1962), Sapelo Island (Howard and Reineck, 1972b) and the Gulf of Gaeta (Reineck and Singh, 1971).

The vertical sequence developed in the examples just given, differs from that of the beds of the Breugh Pebble Member in that the typical foreshore and backshore deposits with swash cross stratification are absent from the Liassic beds and rootleted horizons are rare.

The upper shoreface units are pebbly and thicker than those of modern beaches. Examples and sets of steeper cross strata of significant thickness are rare in all of these modern examples, being restricted to the upper metre or two of the shoreface (Harms et al., 1975). In Sapelo Island (Howard and Reineck, 1972b) cross stratification is well developed and in the modern examples, textural breaks such as abrupt increase in size are only reported by Bernard et al. (1962). The maximum grain size is reported within the foreshore for modern examples whereas the coarsest sediment in the Breugh Pebble Member occurs in the upper shoreface zone. In general the beds of Facies 10, 11 and 12 are coarser grained and contain substantial amounts of

pebbly material which are conglomeratic in places.

Conglomeratic prograding shoreline deposits are not well known and the studies of Clifton (1973, as reviewed in Harms et al., 1975) and Clifton et al. (1971) form the basis of a comparative interpretation here.

b. The prograding conglomeratic shoreline.

The typical sedimentary structures and stratification sequences for a conglomeratic prograding shoreline deposit given by Clifton (1973) are as follows:

1. Offshore

Unbedded siltstones and fine grained sandstones with abundant biotite; burrowing is common disturbing the bedding and molluscs are present. The facies commonly merges downwards into thin widespread conglomerates which rest on an erosion surface truncating other sequences.

The fine grained facies implies sedimentation under non turbulent conditions whereas the conglomerate is thought to be a transgressive lag deposit.

Although burrowing is common in the Breugh Pebble Member there is absolutely no sign of a transgressive lag deposit forming a basal thin conglomerate for the beds of Facies 9a, b or 10, but the lowermost beds may be regarded as offshore deposits.

2. Lower shoreface

This environment is characterised by bedded, fine-grained sandstones with nearly horizontal lamination or tabular cross stratification with locally abundant burrows. Between beds of fine grained sandstones which are commonly less than a metre thick (near the top of the unit) are cross laminated, lenticular beds of well sorted coarse sand or granules. The rhythmically lenticular shapes were formed by large ripple forms. The tabular cross strata may have been deposited by sand waves which are normally

regarded as formed under lower flow energy than those needed for the generation of trough sets. The beds with nearly horizontal stratification were thought to resemble hummocky cross stratification (Harms et al., 1975), formed under strong wave surges, and the ripples of well sorted coarse sand granules support the notion that material of "uncommon coarseness" was occasionally introduced into this facies possibly by storm events and then formed into large landward migrating ripples by wave surge.

It can be seen that most of the features described above and interpreted as representing the lower shoreface depositional environment for an ancient prograding conglomeratic shoreline, are present in the Breugh Pebble Member as facies 10 which also succeeds Facies 9 in Applecross. The above interpretation is also accepted for the beds of Facies 10 without further qualification.

3. Upper shoreface

Trough cross stratified conglomeratic, pebbly, coarse grained sandstones with rare burrows and shell fragments are deposited in the upper shoreface depositional environment (Clifton, 1973); the sandstones are poorly sorted and pebbles are common, which can scatter through the sand or sharply segregate into thin persistent beds which rest on gently inclined erosion surfaces. Pebble imbrication is poor in these beds and stratification within this zone is neither high angle, trough cross stratified, horizontal nor nearly horizontal.

The sets of trough cross strata range in thickness from 15 cm. to 2 m. and dips are predominantly seaward, although a less conspicuous secondary mode directed parallel to the shoreline exists. The base of this unit is an erosional contact and channels are fairly large in places.

The above abridged description of the "second facies" of the Miocene Branch Canyon sandstone (Clifton, 1973) was given to emphasise its striking resemblance to the pebbly beds of Facies 11 of the Breugh Pebble Member which are therefore regarded as representing an upper shoreface depositional

environment. The interpretation of the various features of this facies are somewhat open to question. The sets of cross strata may indicate the prevalence of currents, moving pebbly sands as dune forms; most sets show offshore flow whereas some indicate longshore flow. The basal scour surfaces of Facies 11 indicate channelised eroding currents. The interpretation of poorly defined, horizontal stratification is not quite clear. Harms et al. (1975) maintained that because of their association with trough sets and their pebble content, such horizontal beds are related to high flow energies, but they also mentioned that the lamination could be related to wave surges which were unsteady in velocity and varied in direction.

4. Foreshore and backshore

In the Branch Canyon sandstone example, this environment was characterised by low angle swash cross stratified medium grained sandstones which showed rare burrows and was structureless at the top (Harms et al., 1975). Sandstones and conglomerates with vertebrate remains or oyster shells suggesting flood-plain or shallow lagoon deposits are found together with an overlying root zone.

Except for the isolated examples found in northern Strath, no evidence of rootleted horizons or coal beds was seen in the top beds of Facies 11 and 12. Swash cross lamination and low angle beach stratification are also entirely lacking, but these were most probably very susceptible to erosion.

By the interpretations of Clifton (1973), the three Facies (10, 11, 12) of the Breugh Pebble Member, together with the beds of Facies 9 exposed in Applecross may be related to processes on a non-barred, high-energy shoreline. Although some characteristic features (e.g. hummocky cross stratification in Facies 9) were described in the section related to prograding sandy shorelines, it is evident that the increasing turbulence of the environment through time and the coarse, poorly sorted clastic supply from the hinterland substantially modified the depositional features through time.

Deposits representing the beach and swash zones are lacking in the Breugh

Pebble Member, the only evidence for probable backshore subaerial emergence was found immediately below the Strath Formation along the Ardnish peninsula.

The contact of the Milton and Strath Formations is faulted in southern Strath and Raasay, which may explain the absence of the backshore, shoreline deposits from these areas. Although the topmost beds of the Milton Formation are only seen in Applecross they entirely represent upper shoreface, shoreface small ripple bedding with abundant pebbles. There is no evidence suggesting the post depositional erosion of the topmost beds of Facies 12, therefore it is concluded that the shoreline and backshore belt areas may have existed further towards the north and northeast where no Lower Liassic exposure is known; probably the foreshore belt areas extended from Applecross to Skye and into Arndamurchan.

The coarse grained, cross bedded sandstones of Facies 11 were probably deposited by a variety of depositional processes but within the area of wave buildup and surf, where rip currents and longshore currents continually rework and sort the sediment.

The fine to medium grained, bedded sandstones of Facies 10 were probably laid down in the lower shoreface where wave action is dominant and reworks the sands commonly brought up by rip currents or other (?storm) processes; this facies merges transitionally into the underlying shales, stilstones and limestones which represent offshore conditions where finer sediment existed and organisms reworked material faster than the rate of deposition.

In common with the trough cross bedded, conglomeratic beds of the Branch Canyon Sandstone (Clifton, 1973), the coarse grained, cross bedded sandstones of Facies 11 are most difficult to explain in terms of a barred high-energy model. Harms et al. (1975) explain that although cross bedding is developed beyond the breaker line of the modern shoreline, the substantial thickness and orientation of cross strata in the ancient examples are difficult to interpret. Clifton (1973) maintained that the offshore transport

directions suggest rip current activity. If this is true, the rip currents extend to significant depths below the beach level and substantially expand the interval influenced by breaker and surf zone processes.

A rip channel at depths of 4-6 m., with seaward migrating sand waves was reported by Ingle (1966), but this channel is the deepest of its kind observed so far.

Reasons for the preferential preservation of thicker beds of the high angle cross stratified facies resting on scoured surfaces as compared to the beds of other facies are given by Harms et al. (1975) and briefly stated here:

- a. It is possible that the unusual event of the deepest scours and the beds distributed farthest offshore by the severest storms are preferentially preserved.
- b. Average wave activity and tidal range were greater in the past. As a greater tidal range would produce a thicker foreshore facies it is possible to conclude that tidal range was probably negligible during the time of deposition of the Breugh Pebble Member. Although larger waves may be responsible for a thicker shoreface facies it is not easy to envisage wave activity in the shallow epeiric seas of the past, with their limited fetch, to be greater than those which exist now on the open coasts of the seas and oceans.

iii. Tidal channels

Although tidal and river channels significantly shape variations in prograding sequences, their influence in generating the bedding characteristics of Facies 11 is doubted. There is no significant evidence of bipolar flows to confirm the existence of tidal currents or channels in the Breugh Pebble Member, nor is there any evidence of significant lateral migration of channels, and backshore lagoon deposits signifying extensive tidal influx are absent from the beds. The development of facies in the Breugh Pebble Member is shown in Fig. 6.29.

6.19 Strath Formation

The Strath Formation marks the commencement of substantial argillaceous sedimentation in the northern area of study during *Semicostatum* Zone times. The overall pattern of sedimentation is that of argillaceous beds alternating with thinner silty limestone beds. These different lithologies are marked by distinctly different body and trace faunal assemblages together with different sedimentological attributes. The major facies units with their subdivisions are considered in this section; these will be related together later. The sequence is especially well exposed along the shores of Loch Eishort in southern Strath.

6.20 The Lower Teampull Chaon Member

A total of 23.50 m. of alternating micaceous shales and calcarenite beds overlie those of the Milton Formation in southern Strath, their contact is not seen and a gap (?faulted) separates them.

The lower 8.80 m. of the calcareous argillaceous Facies 13 consists of silty (<15%) micaceous shales with four Gryphaea and subsidiary calcilutite (biomicrosparite) beds. The first Gryphaea bed is found at the base of the shales while the others occur at levels 3.30, 4.10 and 4.30 m. above the base.

The second Gryphaea bed occurs within a calcilutite bed but the others form more or less calcareous beds crowded with Gryphaea. The calcilutite beds are 10-12 cm. thick and show wavy contacts with the surrounding shales. Clay mineral analysis of the shales shows the principal constituent to be illite with some kaolinite; in places illite is the only clay mineral present (see Chapter 4); smectites are present in trace quantities (1-2%) and chlorite forms up to 8% of the shales. The carbonate content of the biosparite beds is as high as 85% and the immediately subjacent shales contain 35-45%.

Quartz is virtually absent in the biomicrosparites, but clusters of framboidal pyrite and faecal pellets are common. The matrix contains

carbonaceous matter in the form of brownish films and coatings dispersed throughout the matrix cement which contains abundant echinodermal and brachiopod shell material. Towards the top of the more prominent calcarenite beds, they contain up to 3% very fine silt size, normal quartz fragments. Relatively little amounts of mica are present in the limestone beds.

The lower unit is overlain by a series of coarsening upward cycles with a total thickness of 4.50 m. These cycles consist of a lower portion of laminated silty micaceous shales which are overlain by lenticular calcareous siltstones. The thickness of each cycle is 35-40 cm. with the shale beds normally 25-30 cm. thick. The finer lowermost portion of these occasionally laminated repetitions are calcareous shales with up to 35% CaCO_3 ; the coarsest parts of the cycles are mainly of two types:

i. Broad shallow lenticular features which range in diameter from 0.50 - 1 m. and in depth up to 10 cm. (Plate 6.49) consisting of well sorted, fine sandstones (75% quartz) which are cemented by calcite and show no internal structures. Crinoidal debris is common at the base of the sand bodies. The sand lenses show very faint ?cross laminations.

Biogenic structures include vertically retrusive Rhizocorallium (Plate 6.50) which normally occur at the bottom of the sandy lenses, and Diplocraterion, which normally occur in the topmost portion of the underlying shales.

ii. Silty micaceous biomicrosparite beds with up to 25% quartz (normal, 4 ϕ). Rounded micritic limestone pellets are seen (4%) and opaque, brownish ?organic and pyritic material is present. Contacts with the underlying shales are gradational and no sedimentary structures are seen. Vertically retrusive Rhizocorallium are scarce but Diplocraterion is present (the burrows show reddish iron oxide stains) together with horizontal feeding marks with surface scratches.

Above this unit is a 9.90 m. thick sequence of alternating, thin bedded,

silty shales and calcareous sandstones is seen (Plate 6.51). The silty shales are up to 10-15 cm. thick and are densely packed with Gryphaea shells. Overlying these is a shaly calcareous siltstone (8-10 cm.). Although no sign of burrowing activity is seen, the sediment is well reworked and un laminated. The silty shale beds are crowded with broken and reworked Pinna hartmani fragments towards the top (Hallam's bed 11) where the beds tend to become generally finer grained and more shaly (Plate 6.52).

On the top of these beds, very large sometimes partially replaced ammonite casts are seen together with the trace fossil Kulindrichnus langi (Hallam, 1960a), which show an average diameter of 1-1.5 cm. and a depth of 3-4 cm. They possess a comparatively thick wall (3-4 mm.) which contains shell infillings and is entirely replaced by calcium phosphates (see Chapter 9). Most of the relief forming trace fossils are preferentially replaced by phosphatic material.

The cement of the hardened calcareous siltstone is ferroan calcite as revealed by staining (Dickson, 1966). The fauna reported from these beds (Hallam, 1959) includes Piarorhynchia juvenis towards the top of the succession, and Chlamys sp. together with Pinna hartmani; ammonites are also common.

succession

They are represented in Raasay where an analogous, sandier/develops (Pl.6.53) Here calcareous micaceous siltstones (4 ϕ) up to 40 cm. thick are interbedded with silty shale beds. They show wavy contacts and the siltstone beds contain sand size rounded fragments of quartz. The CaCO_3 content of the silty limestones is mostly due to the presence of echinodermal debris (Plate 6.54) which is abundant at the base of the beds and decreases considerably towards the higher parts (Plate 6.55). The beds are extensively laminated and pyrite is abundant in the form of clusters, patches and laminae. The surfaces of the sandy, coarser beds show abundant solitary concretionary spheres and elongate bodies, replaced by phosphatic matter. These spheres are up to 1.5 cm. in diameter and are locally replaced by

orange-weathering goethite. The sandstone/calcareous shale contacts are wavy and gradations are seen within the coarser beds (crinoidal limestone to micaceous calcareous siltstone) and the top surfaces of these coarser beds are entirely populated by vertical and horizontal burrowers which are frequently replaced by goethite; coalified fragments are also present. The shales contain Gryphaea and Piarorhynchia was also reported by Hallam (1959). Kulindrichnus langi is also present and its dimensions correspond to those found at the same stratigraphic level in southern Strath. Thalassinoides is more common on the top surfaces of the beds here , than in southern Strath and they do not show signs of replacement. The siltstones contain up to 65% quartz silt grains and are well sorted, their muscovite content is however very high (20%) and towards the top of this unit, the siltstone/shale cycles become thinner bedded (10-12 cm.), with the iron oxide-weathering nodules becoming very abundant. This facies is also present in Ob Breakish (northern Strath) where calcareous siltstones are interbedded with shales crowded with Gryphaea shells. Here the siltstone/calcareous shale sequences are extensively reworked and vertically retrusive Rhizocorallium is seen to occur at the base of the coarser beds.

The beds of this facies (beds 17, 18 and 19 of Hallam) consist of silty micaceous shales alternating with calcareous siltstone lenses of up to 15 m. length and 10 cm. depth. Petrographically the lenses are mainly matrix-supported, containing shell fragments and micritic limestone pellets in a sparry calcite cement; they are moderately sorted with a size range of approximately very fine sand. The quartz grains (up to 22%) are mostly polycrystalline (grains with >3 crystals and undulose extinction); opaque finely powdered coalified material is common as coatings on shells.

At their base, these lenses contain less quartz (11%) and are composed of crinoidal debris together with brachiopod shell fragments. The quartz particles are very poorly sorted and the size varies from silt to fine sand. Epidote and zircon occur (up to 0.05%) and are very well rounded.

Bioturbation is very common, horizontal and vertical cylindrical burrows (?siphonites) can be seen; non replaced Thalassinoides burrows are also common on the top surfaces of the coarser beds. Coalified plant material is scattered on top of the cycles together with patches of broken and finely crushed (?digested) shell material, the centres of which show evidence for the commencement of ferruginisation. On the top surfaces Kulindrichnus langi (smaller variety) appear to be replaced by dark phosphatic matter together with very few belemnite remains which occur both in vertical and horizontal position. The vertically oriented belemnites penetrate both sandy and shaly layers. Towards the top of this facies the fine, sandy calcareous beds become less silty (12%) and contain greenish "chamosite" pellets and a few reworked oolites (up to 5%). The beds are poorly sorted and the quartz ranges up to 2 ϕ in diameter.

At Loch Slapin (southern Strath) the limestone/shale repetitive sequences characteristic of the other major localities is not present. Here the Liassic dark siltstones and orthoconglomerates (Steel et al., 1975), together with the Cambro-Ordovician Durness Limestones are overlain by at least 10 m. of alternating limestone lenses and laminated shales of Semicostatum Zone age, at the base of which, derived fragments of Durness limestone and chert are present. The lower 1.65 m. is composed of an extensively bioturbated alternation of indurated shales and biomicrosparite limestones consisting mostly of crinoidal fragments (Plate 6.56); the sequence is presented in Fig. 6.30. Overlying the described beds is a 9 m. thickness of lenticular limestones and shales, the lower 6.75 m. of which contains beds crowded with Pinna sp. and Gryphaea arcuata. Higher up the sequence very small Pinna are seen in life position, penetrating both limestone and shale layers, the limestones are laminated and wholly composed of crinoidal debris; they contain up to 5% pyritic material. The silt content is minimal but increases towards the top of the sequence. At the 6.50 m. level above the base of these beds, the specimen obtained

(Am 187b) is a poorly sorted, grain supported laminated silty micaceous biopelsparite with 15% quartz (3 ϕ diameter). The matrix is ferroan calcite as revealed by staining and green, "chamosite" rounded pellets are seen. Gastropod and echinodermal debris which have been replaced by ?siderite are common and all shells show evidence of extensive abrasion. The zircon content of the lowermost calcareous and shale beds of this sequence is approximately 0.01%,; they are well rounded and show no coatings.

6.21 Interpretation of the Lower Teampull Chaon Shale Member

Facies 13

The subject of cyclic sedimentation has remained a centre of debate and controversy for decades. A large body of the geological literature is dedicated to the explanation and interpretation of this phenomenon (Duff et al., 1967; Hallam, 1975). Therefore it is not proposed to review the topic here and only those features directly related to the beds of the Broadford Beds Arenaceous and Argillaceous Groups are considered.

The rhythmic units seen in Facies 13 (and 15) are reminiscent of sedimentary cycles described from the Jurassic of England (Brinkmann, 1929; Arkell, 1933). While Arkell (1933) subdivided the Lower Jurassic successions into cycles of clay, sandstone and limestone representing a progressive shallowing sequence attributed to epirogenic oscillations, Hallam (1960b, 1964, 1967c) attempted to solve the long-standing controversy which stood over the origin of "limestone/shale rhythms" in the Blue Lias (Hettangian-Lower Sinemurian) of southern England. Sedimentary and faunal evidence led him to conclude that their origin was composite, being partly primary and partly secondary.

Although the "limestone/shale" type of cycles are the most obvious types, other cycles of more clastic facies are present in the Raricostatum and Jamesoni Zones of England and Scotland. Sellwood (1970a) recognised three types of minor sedimentary and faunal cycles developed in such clastic and

calcareous sequences. In each case an increase in environmental energy towards the tops of the cycle was recognised and related to a shallowing of the sea. A combination of eustatic rise of sea level and tectonic/compactional subsidence could lead to a relatively rapid deepening. Small scale effects are accentuated in shallow extensive epeiric seas and cycles would definitely develop as a result of the superposition of intermittent subsidence and small eustatic sea level changes. Sellwood and Jenkyns (1975) interpreted the clay-sand-limestone sequences in the Pliensbachian to the Bajocian strata of southern England as shallowing upward sequences, in which sedimentation was controlled by faulting in the basement which was related to the rifting and spreading of the Atlantic and western Tethys, not specifically as a result of sea level changes.

i. Facies 13

The features seen in the shale/calcareous siltstones of Facies 13 are almost identical to the description given by Sellwood (1970a) for the type 1 cycles of *Raricostatum* Zone age at Robin Hoods Bay (Yorkshire) and part of the Jamesoni Zone at Carsaig Bay (Mull).

The clays found at the base of the cycles represent a slowly accumulating deposit below the influence of wave action. General variables including the density of suspended matter, depth of water, and tidal current influence, control the deposition of muds (as opposed to sands) on a shelf sea (McCave, 1971).

In the modern Celtic sea, muddy deposits occur at depths below 80-90 m. and sand is present in shallower zones, thus wave activity is assumed to be the major factor controlling the deposition of mud which is laid down only below the level of wave effectiveness.

Also in the modern Tyrrhenian Sea, the lower limit of wave action lies at shallow depths of 6 m. and marks the transition from sands to silty muds (Reineck and Singh, 1971); the transition of clay to sand and silts in the

Scottish Lias is thus interpreted as a shallowing sequence similar to that described from the Pliensbachian-Bajocian facies of England and Scotland by Sellwood and Jenkyns (1975). The vertical and horizontal tubular burrows represent the tunnels of infaunal deposit feeders (which are very abundant in the *Semicostatum* Zone deposits) and are mostly concentrated in the more muddy sediments. The nukulids were similarly infaunal (Table 5.2).

The Rhizocorallium are mostly associated with the sandier sediment. This was interpreted by Sellwood (1970a) as being the result of either of the following:

- a. They were not present during the accumulation of the more muddy deposits.
- b. They were present but remained undetectable.
- c. They were present but acquired a different feeding habit which did not require the construction of the characteristic burrow form.

The sandy beds representing the tops of the cycles show retrusive Rhizocorallium burrows at their bases which indicate rapid sedimentation which induced the typical burrow form. The well sorted nature of the sandy beds is interpreted as being due to strong wave and current action. The traces of Diplocraterion which become quite abundant towards the top of the cycles (Diplocraterion burrows in the sandier beds are often infilled with mud whereas those occurring in the muddy beds show pyrite infillings) are essentially suspension feeding types, thus the final phase in the cycles represents very shallow and turbulent water conditions (Seilacher, 1967; Sellwood, 1970a). To interpret the life habits of the Rhizocorallium animal, although the work of Seilacher (1967) has suggested a deposit feeding habit, taking into account Sellwood's (1970) arguments, it is more likely that Rhizocorallium represents a mode of life similar to some recent callianassid crustaceans which show deposit and suspension feeding tendencies at different stages in their life history signifying adjustments to particular needs at different times. Such callianassids deposit feed whilst they excavate the

system and then suspension-feed when the systems have been completed (McGinitie, 1934).

Thalassinoides burrows are mostly seen on the top surfaces of the coarser members of the cycles where they are wholly replaced by dark phosphatic (dahlite) material. These may have occurred throughout the sandy portion where they were reworked and obliterated by the later burrowers; the Kulindrichnus langi indicates some form of burrow occupied by a sedentary organism, possibly a cerianthid sea anemone (Hallam, 1960a). This trace fossil normally occurs at the top of the coarser beds of the cycles where it usually attains a subvertical cylindrical shape with a phosphatic sheath and a variably phosphatic interior.

As phosphatic limestone surfaces are also found elsewhere in the coarser units replacing Thalassinoides burrows, as spherical and cylindrical aggregates composed of faecal pellets and also as discrete laterally extensive thin nodule beds, the action of seawater on the exposed surface of the burrow cannot be ruled out. Nevertheless the organism itself must have been responsible as well. Careful petrographic examination of the phosphatic walls of the trace fossils, the phosphatic nodules and phosphatic internal moulds of bivalves has failed to trace the presence of any faecal linings and/or clusters in sufficient quantities to contribute to the total phosphatisation of the features present. It is possible that the combined effect of decaying organisms and periodic phosphorus enrichment of the seawater enhanced the process of phosphatisation. The precipitation of carbonate apatite and the mechanism of phosphatisation probably involves the radial migration of the phosphorus-rich material outwards from the burrow and a chemical interaction with and/or surface adsorption on grains in soft calcareous mud; the presence of phosphates cannot be related to the lithology of the rock. Hallam (1960a) considers the existence of crushed shell fragments and minute shells together with traces of organised structure in some specimens as evidence of activity on the part of the animal which

inhabited the burrows; the decay of the organism or its abandonment of the burrows is thus held responsible for the partly collapsed shelly walls and the disorganised structure of most specimens.

The presence of concentrically coated grains and quartz silt fragments infilling the burrows strongly points to their adventitious filling by the action of currents operating close to the sea floor, especially in those cases in which they are packed to the top with shells (Hallam, 1960a).

The change in the iron concentrations within the unit of each cycle is reflected by the fact that iron-replaced burrows and features are abundant in the more shaly units whereas such replacements are less common in the sandier units of the cycle tops where the beds are more calcareous. Curtis and Spears (1968), Hallam (1967a, 1975) and Sellwood (1970a) have interpreted such a situation as reflecting the transition from a terrigenous to a more marine situation.

It is possible that the sediments of Facies 13 were deposited under conditions similar to those existing for Sellwood's (1970a) type 1 cycle. The entire cycle possibly represents a shallowing upward sequence, with quite rapid deepening occurring while the very topmost portions of the cycles were being burrowed. The rapidity of the deepening is inferred from the abrupt change from sand to clay at the cycle tops which gives an extreme asymmetry to the cycles. The presence of burrowers probably caused the removal of some fine grained components in the upper portions of the cycles, by their throwing of fines into suspension and the subsequent winnowing action of currents. The scour features were probably cut and filled rapidly, they normally lack signs of burrowing but in thin uppermost laminae, vertical tubes attributed to escape features are seen. The sandy streak of the topmost part of the coarser beds suggests that the scours resulted from strong wave action which was possibly induced by storms, the scours being produced by eddies which acted around inhomogeneities on the sea bed such as small communities of crinoids or occasionally ammonites (Sellwood, 1970a).

ii. Calcispheres

Calcispheres of various types, hitherto not reported from the Lias of northwest Scotland, are frequently seen in the sandier, more calcareous beds of facies 13 (Plate 6.57a to 6.59).

These minute spheres were first recognised and described by Williamson (1880) from the Lower Carboniferous of Wales and the Devonian of Ohio; he defined the genus Calcisphaera, which referred to a group of micro-organisms of uncertain affinities, and a generic name was recommended as "not involving any premature hypothesis, respecting their nature".

a. Occurrence

Although Stanton (1963) noted their aggregate occurrence, such as those of plant spores or reproductive bodies, they normally occur as disaggregated individuals in the beds of Facies 13.

b. Wall structure

Calcispheres consist of a calcite wall enclosing a spherical chamber. Although the walls are known to have up to seven concentric layers, the calcispheres here display dark, uniform walls which occasionally show very faint evidence of fine layering.

The calcite wall shows an inner and outer concentric surface, the latter of which may bear spines (Plate 6.57); the main wall forms between the two surfaces. The thickness of the main walls may be 3 - 30 μm and the equatorial external diameter of the calcispheres varies from 60 - 225 μm (Bathurst, 1971).

In many cases the wall consists of a single layer of radially arranged translucent, fibrous calcite enclosing the large central body cavity composed of a mosaic of microcrystalline calcite. Hexagonal wall-forms such as those described by Baxter (1960) are not present and none of the forms show evidence for the existence of orifices except in one specimen (Am 148).

A common form of calcisphere in the Strath Formation has alternating layers of thin dark coloured, microcrystalline calcite and translucent, radially oriented calcite (Plate 6.57b); the thin walls are about 2 μm and the radial calcite layers 5 - 6 μm thick. Termed "annular" walls by Baxter (1960), they show a maximum of three to four dark coloured layers with two or three enclosed translucent layers but in most specimens, only remnants of the innermost, microcrystalline layers are preserved.

c. Dimensions, body chamber and spines

Stanton (1963) showed that the overall increase in the diameter of the calcispheres (up to 500 μm) and the cell wall (50 - 170 μm) is largely due to the presence of spines, which appear as syntaxial diagenetic overgrowths but contain radial canals up to 6 μm in diameter. The spines are elongate calcite crystals which extinguish with their vibration directions oriented radially and tangentially (Bathurst, 1971). Some examples found in the beds of the Strath Formation indicate the presence of spines. The spherical to ovoid chambers are enclosed by a non-spinose calcitic wall in most cases, but some examples show a dendritic outer surface reminiscent of ?spines (Plate 6.58 a-b).

The chambers are filled with calcite (spar) cement and internal fillings of micrite are rare; features representing a subsequently removed "erstwhile innermost layer" (Stanton, 1963) were not seen. The central portion of the sphere is mostly occupied by clear calcite, the clear calcite rarely appear in the form of translucent fibrous calcite which is oriented normal to the sphere wall and appear as a cross under polarised light. In most specimens the central cavity is composed of a mosaic of calcite crystals whereas single crystals filling the body chamber are extremely rare.

Origin and occurrence of calcispheres

Calcispheres are invariably regarded as the products of a number of

plant and animal taxa (Williamson, 1880; Cayeux, 1929; Konishi, 1958; Reynolds, 1921; Kaisin, 1926; Derville, 1941; Lombard and Monteyne, 1952). As previously mentioned, Stanton (1963) regarded them as plant reproductive bodies; Rupp (1967) noted the resemblance of non-spinose calcispheres to the reproductive cysts of the modern dasycladacian Acetabularia. Wilson (1967b) regards calcispheres as representing restricted marine or brackish water conditions.

Newell et al. (1953) noted that along with a host of other biological evidence, calcispheres indicate conditions of abnormal and/or fluctuating salinity and water temperature together with very shallow depths. This interpretation cannot be accepted in view of the fact that calcispheres are also common in chalk deposits. Stanton (1967b) found that some calcispheres occur in marine facies.

Calcispheres are regarded by Wilson (1975) as representing restricted circulation on a marine platform. Inorganic structures formed about air bubbles trapped in a calcareous sediment and subsequently infilled with calcite may also form calcispheres (Pia, 1937).

Baxter (1960) found sphere-like objects containing smaller spheres in a central cavity surrounded by a thin wall; he regarded the specimen as "representing some sort of reproductive capsule".

Based on their association with a host of different microfaunal assemblages and lithologies, Mamet (1976) regarded calcispheres as representing a restricted shelf, lagoonal to an intertidal, supratidal environment.

In the present study, no conclusive evidence can be given on the origin of calcispheres. Multispherical structures (Baxter, 1960), which are better examples of possible reproductive tests, were not observed; nevertheless numerous forms with annular walls similar to Calcisphaera polyderma Derville (1941) and a single specimen with an apparent orifice were found.

The beds of Facies 13 represent a period during which a shallow sea

gradually advanced over low lying, vegetated lands and as such the presence of plant reproductive organisms in the calcareous ("shallower") beds is not surprising but expected; it is reasonable to assume that during these "shallower" periods algal spores were more widely distributed due to storms and vigorous current action. Circumstantial and direct evidence for the existence of algae in the younger beds was given in previous sections; nevertheless the possibility remains that the calcispheres showing no internal structures are no more than air bubbles, subsequently infilled with calcite.

It should be mentioned that some micro-organisms were found with spiral-shaped walls which resemble Liassic foraminifera such as Involutina sp., Involutina liassica and Trocholina granosa (e.g. Plate 6.59).

6.22 Breakish Ironstone Member

Ironstone Facies 14

This 40-60 cm. bed was first reported from the Ob Breakish locality by Hallam (1959). It is in most places a matrix supported, sideritic "chamosite" oolite with ooliths and bored, abraded bivalve shell material very commonly present (Plate 6.60 a-g). The matrix is ferroan calcite (68%) with some siderite whereas silt (mainly normal quartz) constitutes 15% of this rock (see Chapter 9).

This bed occurs as laterally impersistent broad lenses which can be seen to thin out locally on the Ardnish peninsula. It is recognised in the field by its dark reddish-brown-black weathering surface and a bluish-green-grey fresh surface appearance. Although surface indications of burrowing is lacking, in thin section, there are extensive signs of such activity. The rock is poorly sorted and because of the bioturbation, the ooliths occur in discrete pockets and clusters. The main constituents are echinodermal and bivalve shell fragments, gastropods and brachiopods are less frequent. The Breakish Ironstone Member is not fully developed in southern Strath and

Raasay, but beds of contemporaneous age may be recognised due to the presence of a high proportion of "chamosite"-siderite matter as shell filling, together with reworked and scattered "chamosite" ooliths. This bed passes laterally into reworked shaly calcareous siltstones, autochthonous fossils are seldom present in this bed and they have mostly a derived origin; this is mainly based on the scarcity of articulated, unabraded shell fragments.

The oolitic "ironstones" of Facies 14 are also present in Ardnamurchan. In the Mull area, ferruginous mudstones also occur which are lateral equivalents of Facies 14 but were deposited under different depositional conditions and are not oolitic.

As both of the oolitic and non oolitic ferruginous beds of this study are genetically related, they should be considered and interpreted together. Details of their petrology and mineralogy are of importance and for this reason the composition and palaeoenvironmental significance of the "ironstones" in the Strath and Leacach Formations together with the associated nodular phosphatic beds, are considered in some detail in Chapter 9. The development of Facies 13 and 14 is schematically represented by Fig. 6.31.

6.23 Upper Teampull Chaon Member

Sandy argillaceous Facies 15

The beds immediately overlying the "ironstone" of Facies 14 in Loch Eishort are fine grained calcareous, shaly siltstones, which are extensively bioturbated. Goethite which is seen replacing the various features, especially vertical and horizontal (?Siphonites) burrows in the topmost parts of Facies 13, is not common at the base of Facies 15, but phosphatic replacement of the various biogenic features is common. At the base, silt and shale beds are well defined, with coarser beds forming thicknesses of up to 25 cm. and the shales 5-10 cm. (Plate 6.6¹). Towards the top of this sequence transition interfaces are less well defined; silty beds become thinner and the thicker shales (18-28 cm.) show a gradational transition into

the overlying coarser strata (Plate 6.62 a). They are all extensively bioturbated and the thickness of sandier beds is minimal at higher horizons. The quartz content of the coarser units does not exceed 32% in the lower stratigraphic positions; they contain up to 4% polycrystalline (> 3 crystals per grain) quartz, are micaceous and moderately sorted. The size of the quartz grains is in the region of fine to very fine sand (3ϕ) and the beds show no internal sedimentary structures being extensively burrowed (Plate 6.62 b,c) mostly by annelid worms. The cement is ferroan calcite and shows negligible amounts of chloritic ("chamositic") flakes. Replaced and coated clasts are present, being less than those found in the higher horizons. The annelid worm tubes (Plate 6.63 a,b) found in a specimen taken from lower stratigraphic levels, resemble those of Serpula (Cycloserpula) socialis Goldfuss which is common in the British Lower Lias and was also reported from the Lias of Baden and Wurtemberg by Schloz (1972); they are replaced and staining suggests that the cement found in the tubes is mainly ferroan calcite. At stratigraphically higher positions, vertical and horizontal tubes are seen which are infilled by goethite.

The top cycle, coarser beds tend to become better sorted and shaly at progressively higher stratigraphic positions. These are highly micaceous (8%) and contain 11% quartz in places, 2% of which is polycrystalline (> 3 crystals per grain) and strained showing wavy extinction; the feldspars are mainly plagioclase (3%). The cement is composed of calcite with pyrite coating the grains together with yellowish-green (limonite) ferruginous patches. The top surfaces of these siltier beds show isolated Thalassinoides burrows and concretionary spheres which are mostly replaced by phosphatic material. Kulindrichnus langi is abundant, showing a thicker wall and larger diameter than those found at lower stratigraphic levels; smaller varieties are also present (Plate 6.64); these form on the top surfaces of the coarser beds and are shown together with Pinna shells which occur in life position as well (Plate 6.64). Chondrites penetrate from the silty, coarser units into

the topmost parts of the more shaly ones, these are infilled by the overlying silty beds towards the top of each cycle. The Rhizocorallium found within the cycles are simple U-tubes and occur in the silty beds and the uppermost parts of the shaly beds where they are filled with very fine silt and display definite "spreite".

The shales contain Fe_2O_3 -replaced vertical and horizontal (?Siphonites) burrows and the top surfaces of the sandier parts also show non distinct features which consist of central composite features with two or three interfering spheres, each 0.5 - 0.8 cm. in diameter with 2-3 slender semiconical arms (3-4 mm. diameter and up to 3 cm. length), radiating outwards. They consist of very finely crushed shell material and are diagenetically replaced by dark carbonate apatites (Plates 6.65 , 6.66 , 6.67).

The total carbonate content of the matrix here is 28% most of which is in the form of ferroan calcite. The clay minerals predominantly consist of illite with some kaolinite and chlorites (see Chapter 4).

The sandy argillaceous Facies 15 in northern Strath occurs above the "ironstones" of Facies 14; the lower beds contain a high proportion of "chamosite" siderite-replaced and/or infilled particles, deformed ooliths are common ("spastoliths" of Wilson, 1966) and there is abundant evidence for abrasion and reworking (broken shell fragments and mottling) at the top surfaces of the coarser beds of each cycle. The beds become more compact towards the top where ooliths tend to diminish at the expense of ferruginous pellets and replaced, reworked shell fragments. The constituents are packed in a grain supported biopelsparite.

The cycles here are 20-30 cm. thick at their bases; this thickness increases towards the top where the shale thickness increases to 90 cm. The top surfaces of the silty beds show abundant Thalassinoides and Kulindrichnus features with Pinna shells in life position. The shales are wavy, micaceous

and show Fe_2O_3 -replaced Siphonites burrows. It should be noted that the top cycle beds immediately overlying the "ironstones" are very faintly cross laminated. The topmost 6 metres of this unit is composed of silty micaceous shales (Hallam's beds 25, 26), in which biogenic structures other than siphonites are hard to see due to the exposure conditions. These shales are mainly illitic (see Chapter 4) and the CaCO_3 content is up to 35%. Whereas Gryphaea beds are frequent at the base of this facies crowding the shale beds, they are less frequent at higher stratigraphic levels.

This facies is represented in Raasay by Hallam's beds 19, 20 and 21. Although the ironstone beds of Facies 14 are not well developed in Raasay, they are represented here by calcarenites, containing pellets and ferruginous brown-weathering concretions. The sandy limestone/shale pattern seen in southern and northern Strath is repeated here, the sandy beds are up to 40 cm. thick and in places are in the form of lenses with evidence of loading at their base (Plate 6.⁶⁸) the sandstones become thinner towards the top of this facies and the sandstone/shale contacts become less well defined and more gradual. The trace fossil assemblages reported from Skye are present here; the top surfaces of coarser beds show some evidence of bioturbation mostly by Rhizocorallium and Thalassinoides, whereas in the shale beds, bioturbation by siphonites is more common, the latter being mostly replaced by iron-oxide material.

At the base, the sandier beds contain up to 34% quartz and are mostly sideritic sandy micaceous biomicrosparites with abundant mica (5%); the quartz fragments are in the size range of very fine sand. The rocks are matrix supported and poorly sorted, most grains being subangular. The other bioclasts are mostly bivalve shells. The top cycle beds at the upper parts of this facies contain 25% quartz with 2% mica; they are sandy biosparites and contain up to 7% echinodermal and brachiopod shells.

At the topmost stratigraphic level in this facies, dark phosphatic rounded nodules are seen which form discrete, thin nodular

beds (Plate 6.69) locally within the shales.

6.24 Interpretation of the Upper Teampull Chaon Shale Member

It can be seen that Facies 15 displays many of the characteristics described by Sellwood (1970a) from the sedimentary cycles of the English, Scottish and Welsh Lias; more specifically some striking resemblances can be found between features found in the beds of Facies 15 and characteristics of Sellwood's (1970a) type III cycles. On the basis of sedimentary features and some faunal evidence, these beds may be interpreted as representing "small-scale" shallowing upward cycles.

Although specific faunal attributes such as pectinid-nuculid assemblages at the base and Pinna-myid assemblages near the cycle tops cannot be readily seen, the above statement bears some truth in many respects. The general clastic nature of the beds with cross laminated and increased trace fossil activity towards the cycle tops may indicate deposition in waters with periodic gentle wave and/or current activity.

The faunal assemblage characterising the base of the unit and interpreted by Sellwood (1970a) as representing quieter waters is not obvious in the field but many features of these lower strata certainly bear-out this suggestion. Rare Pinna in life position are found immediately below the top of the beds, exhumed but articulated Pinna are commonly seen lying on the top surfaces but there is no indication of a preferred orientation.

Although much evidence is found in the sediment for the existence of serpulids, they are not seen encrusting the Pinna shells. This evidence fits remarkably well into a general shallowing upward framework. The increase of suspension feeding fauna suggests an increase in environmental turbulence with the possible lowering of wave base towards the cycle tops. The increased turbulence led to a short time-interval during which the infauna, especially Pinna, was exhumed and also scouring took place. Sellwood (1970a) recognised

erosion surfaces which cut the cycle tops. Features exhibiting such surfaces were not found in the cycles of Facies 15 and although phosphatic pebbles are absent the diagenetic replacement of most of the fossils and bioturbation features found at the cycle tops may indicate extremely shallow conditions. Evidence in the form of crinoid, echinoderm and other bioclastic debris accumulations at the base of faintly cross laminated lenticular beds suggests possible scours filled due to subsequent winnowing. The fact that coarser beds at the top of the shallowing upward cycles are abruptly overlain by less silty shales with no indication of wave or current induced laminations suggests a rapid reversion to less agitated environments. No cases were found showing the Pinna in life position to continue above the sandier beds surface which would indicate that the change in energy regime probably occurred at a slower rate (Sellwood, 1970a). The development of Facies 15 is schematically shown in Fig. 6.32.

6.25 The Dun Borerraig Sandstone Member

Facies 16a, b; 17a, b; 18a, b

This Member is stratigraphically the highest part of the Strath Formation and overlies the Upper Teampull Chaon shale Member in Raasay, northern and southern Strath. The contact between these two Members is gradational and best seen in Raasay (Rubha nan Leac). The base is faulted on top of the Semicostatum shales in southern Strath. Two isolated outcrops occur in Loch Slapin and Rubha nan Eirreanach. Marine fauna are present in the basal beds of the different facies whereas they diminish towards the top of the beds.

i. Ferruginous sandstone Facies 16a

The total thickness is 17 m. and it can be divided into lower, middle and upper units.

In its basal parts, the "lower" unit is represented in southern Strath

by 7 m. of parallel bedded greenish coloured siltstones. These are very fine grained (40), moderately sorted and micaceous (3% muscovite); the cement is mostly ferruginous and organic matter is present in the form of red-brown opaque, irregular patches in bedding parallel laminations. The feldspar content (4%) is mostly plagioclase and "chamositic" flakes also occur.

Towards the top of the unit, the sand grains (40%) become well sorted, well rounded, well packed with ferroan-calcite sparry cement. Organic matter is common (25%) and occurs in irregular patches and blebs within the matrix; microcline feldspars (2%) are also present. Reworked glauconitic pellets and "chamositic" oolites occur and the quartz (>3 crystals per grain) content is 3%. The organic matter forms grain coatings (9%) in the beds which immediately underlie these beds (Plate 6. 70a-b). Their top surfaces show faint Thalassinoides burrows together with Chondrites which penetrate the bedding surfaces (Plate 6. 71a-c). The sandstones are horizontally laminated and locally contain shale strings; the beds are 20-30 cm. thick towards the upper (5-7 m.) intervals and show faint cross lamination (Plate 6. 71c).

A massive, trough forming body of green fine grained sandstones with thin impersistent marl beds erosively overlies the lowermost strata. The sandstones contain a high proportion of pyritic and other opaque substances; vertical and horizontal siphonite burrows are very common in this middle unit, well sorted, rounded massive sandstone (65% quartz).

The "upper" unit is composed of thin bedded (10-15 cm.) purplish sandstones which show uneven lower contacts; frequently the beds are seen to wedge into one another laterally, containing sporadic subangular pebbles. The upper parts tend to become thicker bedded (20-25 cm.) and show very faint tabular and trough cross laminations; bioturbation is frequent and though Thalassinoides markings are not common, vertical and horizontal burrows together with siphonite tubes filled with a ferruginous matter are frequent. At the topmost 2 m. interval, a planar cross laminated sandstone is seen.

These sands contain small pea-size spherules with a phosphatic (?collophane) outer cast and orange-weathering (?iron compound) nucleus. The sandstones are well rounded, poorly sorted with a moderate muscovite content; quartz fragments are in the size range of very fine sand and glauconite is also seen. Very fine, undeformed, colourless, finely laminated ooliths are present (only the extinction cross is seen under cross nicols) and may be entirely composed of kaolinite flakes. The data for the "lower" thin bedded unit and that of the "upper" medium bedded unit is presented in Fig. 2.16 and the cross bedding data obtained is presented in Figs. 6.3.3 and 6.3.4.

ii. Sandy shale Facies 16b

This is the lateral equivalent of Facies 16a of southern Strath and is exposed in Raasay. It is represented by up to 20 m. of silty micaceous shales with laterally impersistent, hard silty ferruginous limestones at the base. Thin beds of nodular phosphatic matter are also found here. The carbonate content of the lower 5-8 m. is around 40% whereas at higher levels, it decreases to 25%; here bioturbated micaceous calcareous siltstones develop beds up to 1 m. thick. These show gradational contacts with the upper and lower calcareous shale beds. No cross laminae are present.

iii. Purple sand Facies 17a

In southern Strath, this facies is represented by 8.5 m. of purplish, lens shaped, well indurated sandstones. The individual lenses (Plate 6. 72) are 1-4 m. long, 30-50 cm. thick, and show mainly convex upward tops and flat bases. Although this is the prevalent morphology, smaller (1 - 1.5 m. long, 30-40 cm. thick) sand bodies with flat tops and convex downward bases occur between the larger ones. These are very faintly cross laminated and occasionally contain broken bivalve and echinodermal shells. The quartz fragments are well sorted, subrounded, fine sands and constitute up to 70% of these beds. The quartz grains are mostly coated by opaque material which

is thought to serve as an additional binding agent together with the ferroan calcite cement containing 4% muscovite. Glauconite with micritic rounded limestone pellets are seen and the quartz fragments are mostly strained, showing wavy extinction; feldspars are also seen. Bioclasts are mostly coated by thick films of opaque material and at the topmost 1.5 m. of this sequence, the beds show distinct planar cross laminations; a 90 cm. thick bed composed of poorly sorted, very large abraded silicified bioclastic material (mainly bivalves) is seen containing large rounded quartz and phosphatic pebbles which show banding and fibre textures (Plate 6. 73). Some intraclasts are rounded pebbles formed of glauconite-cemented fine sands with ooliths (Plate 6. 74). Ochrous rounded pebbles of ?haematite are also present. The cross lamination data taken from this facies are seen in Figs. 6.35-6.39. Thin beds (10-15 cm.) of tabular cross bedded, purplish calcareous sandstones containing abundant phosphatic and mud pellets together with rounded quartz pebbles are seen in Rubha nan Eireannach (NG 645249).

iv. Purple sand Facies 17b

The total thickness of this facies found in Raasay is 4 m. It is represented by brownish-yellow weathering very fine grained (3ø) sandstone lenses (20-30 cm. thick, 2 m. long). These show flat bases and convex upward tops. Smaller scale lenses occur between the larger ones showing flat tops and concave bases (Plate 6.75). These sandstone lenses show orange-coloured spherical concretionary objects up to 1.5 cm. in diameter; the lenses are bioturbated and coalified driftwood is present at their bases. The sandstones are cemented by ferroan calcite (25%) and in most places the fresh unweathered rock is greenish; no cross lamination is seen. At the basal 1 m. of Facies 17b lies a fine grained sandstone with very thin shale laminations, which is bioturbated by vertical and horizontal burrowers; no evidence of sedimentary structures is present.

v. Cross bedded sandstone Facies 18a

It is represented by up to 6 m. of yellowish-orange weathering, fine to very fine sandstones (Plate 6.76 a-b). There are up to six coarsening upward cycles which consist of a lower 30-50 cm. thick mottled, very fine grained micaceous friable, marly sandstone with abundant coalified plant fragments and mud pebbles (mud drapes are also seen). Small scale lenticular and wavy bedding (Reineck and Wunderlich, 1968b) occur in the lower portions. Burrowing activity has reworked these sediments, leaving behind vertical and horizontal, iron oxide-replaced tubes; the topmost parts are parallel laminated.

A cross laminated, well sorted fine sandstone unit 1 - 1.5 m. thick overlies the basal mottled beds. These consist of up to 75% quartz grains of fine sand size. Strained quartz is very common and rounded micritic and phosphatic pellets together with abraded, rounded bivalve and echinodermal debris are seen. The cement is ferroan calcite and in places shows opaque strings within the cement. These beds consist of large lenses with erosional basal contacts (Plate 6.76a). The lenses show low angle festoon shaped cross laminations which cut into one another and in rare cases show convex upward laminae. Herringbone cross lamination is also seen; the beds do not show evidence of burrowing activity, however their top surfaces show stem-like features which may also have formed due to weathering effects. The cross bedding data is presented in Fig. 6.40. Impersistent well rounded quartz pebble beds are also seen.

vi. Shale-sand Facies 18b

In Raasay the equivalent of Facies 18a is represented by an 18 m. thickness of more than five coarsening upward cycles which may be seen in the cliff, forming the Hallaig waterfall (NG 595385). The upper portion of the cliff is inaccessible and the coarsening upward cycles are hard to distinguish at the coastal outcrops north of the waterfall. The cycles

consist of lower fine grained silty micaceous shales containing abundant marine fauna including ammonites; spherical brown weathering, sideritic concretions (up to 2 cm. in diameter) are abundant and shells are mostly coated by rims of yellowish layers. They are up to 1.20 m. thick and are transitionally overlain by 2 m. of micaceous calcareous sandstones.

The sandstones are bioturbated by vertical and horizontal burrowers (Plate 6.77) and contain bivalve and echinodermal shell debris./ Coalified plant material is also seen, with fragments as large as 10 cm. in diameter, which show a thick (?jarosite) yellow coating. The sandstones show normal Rhizocorallium and Thalassinoides burrows towards their tops; the top surfaces of which are normally marked by a 2-3 cm. thick, laterally persistent, straight bed of yellow-orange, iron-oxide matter (Plate 6.78). The cycles tend to become less prominent and conspicuous towards the top of the facies where they form a 4.50 m. thick sequence of medium-thin bedded shaly calcareous sandstones. Towards the top of this facies an 80 cm. thick, sandy limestone bed was seen. It shows ochrous upper and lower weathered surfaces and is in the form of thin sheets in places. The top surface is burrowed by a meandering, relief-forming double line with internal spreite, resembling a lengthened normal Rhizocorallium (Plate 6.79). This bed is well burrowed internally and shows no appreciable difference in terms of clast composition and density in the inner and outermost portions.

It seems that the ferruginous crust is a product of erosion. Staining results suggest that the calcite cement is ferroan; no cross bedding or other directional feature was seen.

6.26 Other outcrops

The Dun Boreraig sandstones are also found in Loch Slapin, Rubha nan Eireannach and immediately overlying the Upper Teampull Chaon Shale Member in northern Strath. The definite stratigraphic and/or facies position cannot

be readily distinguished due to their scattered and non distinct appearance. Their main characteristic is that they are ochrous and contain abundant haematitic pebbles together with banded, concentric structures (Plate 6.73). They are tabular cross laminated and thin; laterally impersistent lenses of fine sandstone are seen with burrowed top surfaces. Rhizocorallium and horizontal siphonites are also seen.

In Loch Slapin this sequence is overlain by a development of distinctly lensoid, clean sands which have no other correlatable representative elsewhere. They are tentatively correlatable with the topmost parts of facies 16a or the lower parts of Facies 17a and the palaeocurrent data obtained from these is presented in Figs 6.36 to 6.39.

Isolated outcrops of the Semicostatum Zone strata occur immediately overlying the Milton Formation in Suisnish, southern Raasay where a three metre thick development of dark blue limestones alternating with metamorphosed silty shales crowded with shells of Gryphaea is seen (Plate 6.80)

, this is also seen in northern Strath at Loch Sligachan. The Gryphaea bearing strata are always represented by silty dark blue limestones. The Milton Formation here is also silt free and is mainly represented by limestones with undulating shale laminations.

6.27 Interpretation of the Dun Boreraig Sandstone Member

The grain size variation of the beds of this Member indicate a general coarsening upward character; consideration of sedimentary structure assemblages together with the trace fossils present suggests deposition under shallow marine, steadily "regressive" conditions. Knowledge of modern shallow marine sand bodies is steadily growing with the publication of vast amounts of oceanographic data on modern shelf and nearshore sediments (e.g. Stanley, 1969b; Stanley and Swift, 1974; Davis and Etherington, 1976). The application of modern data to the interpretation of ancient shelf and nearshore sediments is greatly impeded due to the lack of

vertical control in the former and three dimensional outcrops in the latter case. While the morphological forms of modern equilibrium shelf sand bodies are being discovered, their internal structures are unknown. In order to interpret the depositional environment of the beds of Facies 16, 17 and 18, they are compared with offshore shallow marine sands studied by Spearing (1974), and which were thoroughly discussed by Harms et al. (1975).

Spearing combined a basic understanding of hydraulics with the careful observation of sedimentary structures and sequences to arrive at an interpretation without recourse to a well established model.

The Upper Cretaceous Shannon Sandstone, Wyoming is an offshore shallow marine sand body that was transported consistently but intermittently, in a longshore direction by a combination of tidal or oceanic currents and storm waves (Spearing, 1974). Somewhat similar depositional conditions may be envisaged for the prevailing environment while the Dun Boreraig Sandstone Member was being deposited.

i. Facies 16

Although the basal contact of the beds of this facies together with that of the topmost beds of Facies 15, is faulted and obscured in southern Strath, it is well exposed and gradational in Rubha nan Leac, Raasay.

It is evident that while coarser sediment came to be deposited under somewhat turbulent environmental conditions, the intermittent shallowing process characteristic of Facies 15 was still operative. Characteristic trace fauna and faunal assemblages are absent due to the overwhelming supply of clastic material. Thalassinoides and Chondrites occur at the top surfaces of the coarser, sandy bed in southern Skye whereas the equivalent beds in Raasay are shaly and show evidence of reworking by vertical burrowers. No sign of reworked phosphatic pebbles or coalified wood fragments were found, thus although shallower more turbulent depositional conditions prevailed in southern Skye, this may not have been directly connected with the proximity

of a land area. The occurrence of nodular phosphatic beds in Raasay may represent periods of non deposition prior to the influx of similar material into the basin of deposition.

The silts and sands of the lower unit were brought into the environment under less turbulent conditions and deposited as thin, parallel laminated sheets with the organic matter forming thin layers parallel to the bedding planes. In places the lower unit in southern Strath is planar cross laminated and indicates a bipolar transportation direction (Fig. 6.33). This may have been caused by weak multidirectional underwater currents which moved the sands intermittently as evidenced by the thin ?rippled, fine sand beds found occasionally separated by suspension clay laminae.

Between the times of the transport of fine sands, flow velocities were so slow that suspension clays were deposited over the rippled sand sheet to form the intervening clay laminae. The beds of this facies dip 14° and show a strike of 350° ; while it is evident that a deeper, muddier, basin of deposition developed in eastern Raasay during the formation of Facies 16, it is seen that northeast-southwest currents were responsible for its formation. Figure 6.33 shows a northeast flowing current component to be the dominant mode, indicating southwesterly currents.

The topmost massive trough-filling beds were probably formed due to a sudden influx of quartz sand into the environment and its subsequent reworking by burrowing benthic organisms which destroyed the sedimentary structures.

The lowermost beds of Facies 16a in southern Strath show low angle planar sets. These vary gradually up the sequence to form higher angle planar sets of tabular form with tangential bases; trough-filling sands showing faint trough-shaped sets are seen at the top. Although Fig. 6.34 shows the existence of diametrically opposed palaeocurrent directions, no herringbone type cross lamination is seen.

An upward shoaling with increased current and wave velocities is indicated by the upper beds; the bedforms may have shifted from sand waves to

dunes. The lack of typical beach cross stratification and soil zones or root traces indicates that no emergence took place. Fig. 6.34 shows that there was no preferred current direction present.

The sand layers which make-up this facies may indicate thin sheets of moving sands. It is not possible to identify wave ripples, but some beds originated as such ripples. The sand sheets may possibly have formed similar to the "sand ribbons" in the present North Sea (Stanley and Swift, 1974). These sand sheets built up intermittently and layer upon layer. During slack periods between sand transport, suspension clays were laid down in the Skye area forming thin shale beds between the sand sheets. The quartz clastic sediments were being deposited while shale deposition provided a suitable background.

ii. Facies 17

The process under which beds of Facies 17 were deposited are somewhat problematic. The discrete elongate, lensoid sand bodies which make up Facies 17 consist of sands that are coarser and better sorted than those of the underlying beds. They are feldspathic and show distinctly deformed and altered kaolinitic oolites (the concentric laminae are entirely replaced with silica). The upward change of bedforms from thin sheets to "lensoid" features possibly representing sand waves and dunes which formed in response to higher shoaling energies. Suspension clay laminae in the beds of this facies in Raasay indicate sand and silt deposition against a shaly background sedimentation. Waterlogged driftwood supplied from land areas also found its way into the depositional area and were laid down in the basal parts of the sand bodies.

As the body built vertically the environmental turbulence consistently increased. This is evident towards the top of Facies 17 where a conspicuous bed of coquinooid sandstone is seen which forms channels in an ideal profile.

The sediment which "paves" or fills the channels consists of thick and thin shelled bivalves, large rounded quartz fragments up to 2 cm. in diameter and abundant rounded phosphatic, glauconitic and mud pellets.

They show no evidence of surface borings and/or encrusting by various organisms therefore although a terrigenous source may be inferred it is very unlikely that they were subaerially exposed for long periods of time.

The large sizes of the various constituents indicate transportation under flow conditions much greater than those which were needed for the transportation and deposition of the sand size particles. The infrequent nature, localised appearance, erosive bases and upward grading together with coarse sediments with a maximum grain size much higher than the underlying sediments, suggest short-lived high-energy conditions in this facies. Such high energy conditions may be considered as being storm-induced (Brenner and Davies, 1974; Kumar and Sanders, 1976).

The products of such conditions (i.e. the channels) represent surge channels which were cut through the sediments, while storms passed over the sand bodies under very shallow conditions. The topmost "storm lag" sand overlying the beds of Facies 17 forms a 90 cm. thick bed which shows a straight base and truncates the convex upward tops of the sandstone lenses. This probably formed due to the "beveling" effect of storm activity which allowed storm-lags to form horizontal beds at the sediment/water interface (Brenner and Davies, 1974). The palaeocurrent data obtained from the coquinoid sand beds (Fig. 6.35) indicate a transport direction which remained remarkably constant, flowing in southwest and eastward directions. This contrasts with the palaeocurrent data obtained from the "lensoid" beds of this facies in Rubha nan Eireannaich which do not show a constant direction of flow.

The beds representing the lateral equivalents of Facies 17a in northern Strath are mostly planar cross laminated, rippled sand sheets, containing abundant phosphatic and mud pellets together with other reworked elements. These may indicate a lower shoaling energy which created current-modified

sand bodies. The palaeocurrent data obtained (Fig. 6.38, 6.39) does not show a preferred direction of transport.

iii. Facies 18

The beds of Facies 18 indicate persistently recurring periods when deposition under calm, deeper marine conditions progressively gave way to deposition in shallower more turbulent environments. The coarsening upward "couplets" represent shallow, lower shoreface conditions. The lowermost units in southern Skye which show lenticular and wavy bedding, represent deposition under conditions of current or wave action depositing sand alternating with slack water conditions when mud was deposited (Reineck, 1960a, b; Reineck and Singh, 1973).

In southern Skye, the sandy layers and lenses are made up of foreset laminae of wave ripples. It is usually difficult to distinguish between asymmetric wave ripples and small, current-formed ripples with straight crests (Reineck and Singh, 1973; Boersma, 1970; Tanner, 1967). The small size (0.75 cm. x 1.5 cm.) of the lenticular bedding impedes the proper identification of their internal features. The investigations carried out by Tanner (1967) and Reineck and Wunderlich (1968a) in Recent sediments, cannot be used here for distinguishing among asymmetric wave ripples and small current ripples since both papers deal with the outer form and shape of the ripples. The irregular lower bounding surfaces and existence of inclined foreset laminae which generally pass the trough and peak-up again on the other flank of the adjacent ripple (Tanner, 1960) indicates the existence of wave generated lenticular bedding.

The wave generated lenticular bedding formed where a change most likely took place between slack water and turbulent water conditions; subtidal conditions for their formation are favoured by (Reineck, 1963a; Reineck et al., 1968).

Mud pebbles and coalified wood fragments seen at the bases of the

coarsening upward cycles in southern Strath indicate very turbulent conditions and possibly the relative proximity of land areas (mud pebbles show no sign of burrowing activity).

The basal shaly beds of Facies 18a were deposited under fully marine conditions as evidenced by the presence of abundant marine benthonic fauna; the lack of directional sedimentary structures suggest that deposition took place beyond the reach of normal shoaling waves and the absence of bioturbation features may indicate poorly oxygenated bottom conditions unfavourable for the development of actively burrowing organisms. The gradual upward coarsening of the beds of Facies 18a and 18b is of great interest. The beds comprising the top (coarser) unit of each couplet in southern Skye shows trough and festoon cross laminated medium sandstones indicating a dominant transport direction towards the southwest (Fig. 6.40). The sandy beds of Facies 18a were formed in a very shallow marine depositional environment. The distribution and formation of festoon cross bedding is a subject of disagreement among various workers. McKee (1966b) showed that these forms are rare in dunes, also there is no evidence to suggest that they can develop without the presence of confined unidirectional flow (Visser, 1972) and thus they most probably developed in tidal, distributary and fluvial channels. Deeper marine depositional environments may be envisaged for the beds of Facies 18b.

The characteristic features of the coarser beds of Facies 18b indicate environments favourable for the development of vertical and horizontal burrowing organisms and as marine benthonic organisms are found throughout the beds here, it could be concluded that they formed in considerably off-shore areas where they were beyond the reach of normal wave and current activity. Alternatively the action of marine burrowing organisms has been responsible for the destruction of any record of sedimentary structures. Within the beds of Facies 18b shallow, wide but laterally impersistent troughs with faintly cross laminated red ferruginous beds which also show some

evidence of reworking and bioturbation by horizontal Rhizocorallium. The sharp, even top surfaces of the coarser beds of Facies 18b are seen as discrete iron-rich (?goethitic), laterally persistent surfaces which are abruptly succeeded by the shaly beds of the overlying couplet.

Attempts were made to compare the stratification sequence and sedimentary features displayed by the Dun Boreraig Sandstone Member with those of reported offshore, shallow marine sand bodies and those of shoreline deposits (e.g. shoreline, barrier island and offshore marine bar sands). In every case similarities existed but there were significant differences as well.

Although the beds of Facies 18 intertongue with marine shales towards Raasay (?seaward), they are overlain and underlain by marine shales. No evidence is present for any associated fluvial, floodplain or lagoonal facies and there are no consistent and recognisable landward, seaward facies changes as one would expect in a shoreline sequence. For a shoreline sequence, capping remnants of beach deposits with very low angle, swash-generated beds which slope gently seaward are needed. No beach stratification is seen in the Dun Boreraig Sandstone Member outcrops. Evidence of emergence i.e. root-traces or soil zones was not found.

Shoreline progradational units usually have sharp bases and they overlie shallower marine or non marine facies. The lenticular and wavy bedding, the non-bored phosphatic and clayey pellets at the base of Facies 18a indicates slow sand migration over a seabottom and not shoreline progradation.

The stratification seen in the Dun Boreraig Sandstone Member is closely similar to the lower parts of a shoreline sequence (Campbell, 1971) and in this respect also resemble offshore marine sand bodies both of which are shoaling sequences.

Suspension clay laminae occur throughout the beds of the Dun Boreraig Sandstone Member. This indicates that alternating traction and suspension deposition persisted throughout the period of their deposition. Suspended clays and clay laminae cannot remain in the turbulent wave breaker zone of

a shoreline for long periods; they are found in the foreshore and upper shoreface facies of most present day shorelines. Clay clasts are not common in shoreline deposits. When they occur in such environments they are always crowded with holes and are bored by various organisms (Harms et al., 1975). As mentioned earlier, bored clay clasts are very rare to absent from the beds of Facies 16, 17 and 18.

Shoreline transportation is predominantly unidirectional and in some cases bidirectional and mainly oriented parallel to the shore; some offshore transport components are preserved. There are no persistently recurring transport directions in any of the three facies under consideration except the storm-deposited coquinoid sandstones. Glauconite is abundantly present throughout the beds under discussion. Soft clay grains such as glauconite are removed from the shoreline setting by attrition.

The beds under consideration cannot be regarded as representing parts of marine bar sands or barrier islands for the following reasons; the facies represented by the Dun Borerraig Sandstone Member represents a regressive sequence and although typical barriers and marine bar sands are locally regressive sequences, various dissimilarities can be pointed out. Barrier islands (Hails and Hoyt, 1968; Howard and Reineck, 1972a; Howard et al., 1972; Oertel and Howard, 1972; Oertel, 1972, 1975) are coastal constructional features that are emergent at high tides. Not only is there no evidence of subaerial emergence of the Liassic beds under consideration, there is no evidence of tidal activity. The vertical sequence of sedimentary structures and grain size in complete barrier sequences has been presented by Davies et al. (1971) and this was compared with "marine sand bars" (Brenner and Davies, 1973, 1974). It is seen that the development of preserved low angle beach cross lamination and overlying root zones characteristic of complete barrier sequences are completely lacking in the Dun Borerraig Sandstone Member. Of course it is possible that the straight, eroded top surfaces of the festoon cross laminated beds of Facies 18a represent emergent parts of the sequence

which were eroded prior to the deposition of the overlying sediments. In some cases the lateral migration of tidal inlets in barrier systems may rework the upper level of barrier deposits and leave behind modified channel sediments (Harms et al., 1975), but the vertical sequence of textures and sedimentary structures in Facies 18a does not resemble vertical features in channels (see Brenner and Davies, 1974).

The beds of the Dun Boreraig Sandstone Member overlie marine silty shales and consist of rippled and trough cross laminated fine to medium sandstones which become coarser and better sorted towards their tops, with bioturbational features and storm surge channels cutting through them in a manner somewhat similar to the marine bar sands described by Brenner and Davies (1974) but the actual "bar" morphology is lacking and consequently, muddy "interbar" areas do not exist and it is doubtful that the sand bodies formed discrete bottom morphologies. However, the coarsening upward vertical sequence and the vertical arrangement of sedimentary structures indicate an upward increase in energies attributable to shoaling. Waves were most effective in the transport or at least final deposition of the sand beds of Facies 17a, but currents become more important over the shallower areas (Facies 18a) cutting channels and forming the trough cross strata. The extremely variable flow directions recorded from the various facies may have formed due to a number of shallow marine processes involved in distributing the sand in the "shoal areas" of a shallow shelf, these may have been storm waves, storm driven currents, tidal currents, and even circulating offshore currents. The lack of detailed further information regarding the beds of Facies 16, 17 and 18 precludes the possibility of distinguishing among the various effects of the above mentioned processes.

It is possible to envisage an environment in which initial bottom topography somewhat controlled the nature of sedimentation. On the Persian Gulf coast of the Abu Dhabi complex, Bathurst (1971) observed that in the inner shelf environment, from the 6 m. contour a limestone platform sloped

gently seaward; this platform was more or less covered by lime sand and sparse faunal elements. The distribution of grain sizes was critically affected by changes in bottom topography as slight as 1 m. The coarser sands lie on topographic highs and muds collect in hollows. It is possible that during the deposition of the beds of Facies 18 the siliciclastic sediment which was spread and transported onto the sea floor, preferentially formed shales and dark clays in the various pockets and hollows (topographic lows) while the coarser siliciclastic material occupied the topographic highs (the development of the Dun Borerraig Sandstone Member is shown in Fig. 6.41).

CHAPTER 7

FACIES DETERMINATION IN ARDNAMURCHAN

In Ardnamurchan, strata of the two Broadford Beds Groups were deposited under conditions intermediate between those operative in the two major depositional basins and they represent facies referred to as 'transitional' hereon.

The transitional facies mainly represents the Milton and Loch Aline Formations; representatives of the Strath and Leacach Formations are present at Mingary Castle and Garbh Rubha Bay but the various facies identified in these Formations in the northern and southern regions are not altogether recognizable in this region due to poor exposure and lack of index fauna. Interpretations presented for the various facies of the Skye area may be equally applied to the facies identified in Ardnamurchan.

7.1 Mingay Pier

No index fossils of definite stratigraphic value were found here; lithologically these beds are blue micritic limestones with undulose calcareous shaley interbedding (Plate 7.1) and they directly overlie fossiliferous, very calcareous sandstones with uneven bedding. The basal sandstone overlies Triassic pebbly conglomerates and may be of ?Rhaetic age in the absence of fossils. The 2.60m. thickness of blue biomicrite contains Ostrea sp. and Cypricardia porrecta. It is hard to relate these beds to any of the facies represented by the Broadford Beds Arenaceous or Argillaceous Groups. the alternating sandy shales and limestones together suggest that they are facies equivalents of calcareous sand Facies 3 and shaly Facies 4 of the Broadford Beds Arenaceous Group, however, this does not presume their age equivalence.

7.2 Mingary Castle

A 2.30m. succession of blue biomicrites with sandy shale partings can be seen here, which overlies 3.60m. of massive calcareous sandstones, with a white sandstone bed at their base. This succession is exposed at the southern end of Rudha a Mhile and was first compiled by Richy etal (1930) who found Modiola halliana, Cardinia sp. and Pleuromya sp. in the shales and limestones.

The white sandstone is regarded as being of Rhaetic age and the calcilutite/shale alternations may represent Facies 7 of the Milton Formation, the subfacies cannot be determined.

Beds overlying this succession are faulted and cannot be placed within the stratigraphic column, further inland blue limestones with shaly partings are seen to overlie a 2.10 m. thickness of pebbly, crosslaminated calcareous sand. The limestones consist of dark calcilutites with shale partings and minute bivalve shell fragments; the calcareous sandstones are cross-laminated in places and show contorted bedding features together with somewhat unidirectional cross lamination (Plate 7.2, Fig.7.1), they are graded at the base and erosively overlie the limestone beds. At the contact surface, short, vertical narrow protrusions into the underlying limestone are seen which are infilled by the sandier sediment. The contact surface is wavy Plate 7.3 and these beds become lenticular towards the top of the succession; the pebbles are well rounded and consist mainly of white quartz fragments (1-2 cm. diameter), also very large, rounded gneiss pebbles (up to 8cm. diameter) are seen, they are poorly sorted and occur in the form of lenses at the base of the succession; these pebbly sands become thin (5-8cm) and wavy bedded towards the top and are interbedded with limestone lenses (10-12cm), their tops are faulted and covered by peat.

This facies is represented by the Breugh Member of the Milton Formation in the northern area; the strata in Ardnamurchan may be regarded as representing the Bedded Sandstone Facies 10 and Pebbly Facies 11. Beds of the same lithology and facies occur in northern Ardnamurchan at Garbh Rubha, these are regarded as representing part of the Conybeari Subzone on palaeontological evidence. It is reasonable to assume a roughly similar date for these beds on account of their relative proximity.

At the base of Mingary Castle a 9m. continuous succession can be measured (Fig.7.2), the succession consists of silty biosparites at the basal 2m.; they show undulose surfaces, are 9-11 cm. thick, show very thin (1-2cm) thin sandy shale partings and also contain very small mound-like features (1cm.high) which contain U-shaped (?shell) forms.

The basal portion is overlain by 3.40m. of blue limestones with calcareous shale partings full of Gryphaea shells, these show worm-tube like surface features and have yielded specimens of Arnioceras sp.

A thinly laminated echinodermal silty shale (50cm. thick) succeeds these beds (Plate 7.4), the crinoidal debris is well rounded and the specimen is grain supported with silty carbonaceous, calcareous shreds forming the matrix, it appears to be oolitic at first sight in the field. The cement is microsparitic and the carbonaceous silty shreds contain a cryptocrystalline aggregate cement with high birefringence (Chamosite). Opaque patches (pyrite) are present and sometimes form shell coatings on brachiopods and gastropods. This bed is overlain by 2m. of very faintly cross laminated silty bio-microsparite (9-11cm) alternating with thin sheets of calcarenites (2-3 cm.thick), these contain some scattered quartz pebbles at their topmost surfaces.

The presence of Arnioceras sp. suggests a Semicostatum Zone age; however, in the absence of further evidence it can only tentatively be suggested that they were deposited during this time.

Samples obtained from the ?Semicostatum succession at the Mingary Castle section show concentrically arranged greenish grey oolitic laminae around aggregate crystals with shell fragments and reniform kaolinite books, within a microsparitic cement; these contain $\approx 10\%$ silt fragments (angular and poorly sorted). In places the cement shows a pseudo isotropic character and is bioturbated, this oolitic horizon was regarded as representing the Sauzeanum Subzone (Oates, 1976) and is taken to represent Facies 21a in this area. (Plate 7.5).

7.3 Swordle (NM554772)

Along the northern shore of Ardnamurchan peninsula, a somewhat fragmentary but complete section of the Broadford Beds Arenaceous Group can be measured, these are best exposed along the northern shore of Swordle, from Ockle Point (NM 550717) to the cave west of Garbh Rubha (NM 33007085). They succeed ?Triassic conglomerates at every locality.

7.4 Ockle Point (550717)

Here a total of 8.50m. of ?Liassic strata are exposed (Fig.7.2); at their base 1m. of medium bedded, pure blue clacilutites are followed by 0.60m. of sandy biomicrosparites, forming beds which wedge out laterally. A 4m. succession of dark calcareous sandy shales alternating with regular undulose, less sandy shale partings is seen above the sandy wedges; these become less sandy towards their tops and form blue micritic limestones with narrowly spaced, irregular shale partings. The succeeding 2m. is composed of

calcareous sandstone wedges and is overlain by pure micritic limestone; they are fossiliferous and in the Geological Survey's report (Richey et al. 1930), the occurrence of Modiolus hillana and Perna infraliassicus has been mentioned; as these beds overlies conglomerates of ?Triassic age and also resemble those of the Lower Lias elsewhere, it may be assumed that they partly represent the Rhaetic or possibly beds of the Liassic 'arenaceous' formation of Skye.

7.5 Port an Eilean Mhor (NM 548710)

The thickness of strata immediately overlying the conglomerates of ?Triassic age in this area varies remarkably over short distances. Along the shore, about 200m. east of where Allt Sordial flows into Port an Eilean Mhor (NM 546707), a 3.60m. sequence of homogeneous blue fossiliferous limestones is seen (Fig.7.2) to overlies pebbly conglomerates; they are interbedded in places with 10-15cm. thick beds of calcareous shale at the basal parts. Towards the top they become calcareous sandstones with undulose sheet-like (?erosive) interbeddings. In the absence of definite fossil evidence they are considered to be of ?Rhaetic-Lower Liassic age, this sequence can be seen to become thinner laterally in the same locality. The strata are well exposed within the bay itself, where the compilation of a section and detailed study is hampered due to the very low angle of dip ($1-2^{\circ}$), dip-slope exposures and corrugated surface weathering. The beds here are composed of light greenish blue calcilutites with abundant broken fragments of thin-shelled bivalves; their estimated thickness is 3.40m. They are succeeded by 1m. of dark blue laminated micritic limestones with faint cross laminae at their base and indurated shale partings at the top; these may be correlated with parts ofacies 3, 4 and 5 of the northern area of study.

7.6 Gah Rubha

The base of the Lias is best seen at the east of the bay (NM 540710), where a 7.60m. thick succession of sandstones and limestones are exposed with ?Liostrea at their base. As shown in Fig.7.2, a fine grained calcareous sandstone is in contact with the underlying ?Triassic conglomerates and red mudstones. This bed contains ?Liostrea sp. and Modiolus shells together with other small, non-distinct bivalves and shows no evidence of sedimentary structures; its contact with the conglomerates is non-erosional.

A 60cm. thick silty biomicroparite containing Pleuromya shells gradationally overlies the previous units; towards the top of these beds up to the base of the first sill (shown on Fig. 7.2), a 2.30m. thickness of uneven bedded blue limestone with irregular sandy partings is seen. Overlying the sill a 90cm. thick succession of sandstone sheets (5cm. thick) with uneven partings overlies a 10cm. calcilutite bed. Blue limestone with baked shale partings succeed the sandstone sheets with an abrupt but non-erosive contact. Three medium-bedded (30cm. thick) sandstone beds and a 20cm. thick bed of fossiliferous blue limestone form the top most part of this section. The whole of which

may represent partly Facies 3, 4, and 5 of the northern (Skye) area. Within the bay (NM 536706) a ca. 35m. thick succession of Lower Liassic strata may be measured (Fig.7.2); the contact with the underlying strata cannot be seen here due to beach-covering. At the base of this succession a blue limestone with indurated, undulose shale partings is seen which is 1m. thick and is interrupted at the 15cm. level from the base by a 10-15cm. bed of calcareous sandstone full of bivalve shells. The blue calcilutites below the sandstone contain many broken oyster shells together with ?Isastrea markings; the surface of the 50cm. limestone above the sandstones is full of Liostrea hisingeri shells and other bivalves. This bed is succeeded

by 8-9m. of dark blue calcilutites which are finely bedded in places and show beds covered with thin shelled bivalves and occasional gastropods.

The succession is shown in Fig.7.2, and this lower part can be regarded as of Hettangian age (Schlotheimia sp. at the base, together with Liostrea sp.). The lower limestones do not contain siliceous clastic material and are entirely composed of calcilutites; they are regarded as representing Facies 6 and 7 of the northern (Skye) area.

A 2.30m. thickness of biomicritic limestone containing definite horizons of Thecosmilia sp. (Plate 7.6) overlies the previous unit; as seen in Plate 7.7, this upper succession can be divided into a lower (30cm.) massive horizon, a middle (1.10m) medium-thin, undulose uneven bedded horizon and an upper medium bedded horizon with undulose bedding contacts. The relative position of these beds in the stratigraphic column and their faunal content suggests a definite correlation (although not in terms of their age) with the Thecosmilia beds of Facies 8; the micritic limestones with Thecosmilia beds are transitionally overlain by 1.50m. of micaceous tabular cross laminated (faint) calcareous, fine sandstone lenses. These contain broken bivalve fragments, show no pebbles and are overlain by 3.20m. of tabular and trough cross bedded muddy sandstone lenses with abundant poorly sorted, rounded pebbles at their bases (Plate 7.8); the pebbles are mostly (90%) quartz with some gneiss and black chert fragments (some 6cm.in diameter), they also tend to line the cross laminae and within each lense, they show upward grading (although incomplete). The lenses are broad (80-90 cm.), shallow (10-15 cm.) and the cross laminae almost invariably dip towards the northwest (Fig.7.3 the morphology of the lenses and also the combination of trough and tabular cross lamination produces a herringbone-like appearance which cannot be regarded as true herringbone cross lamination.

The upper 0.7m. of this unit contains less pebbles and shows no evidence of current activity; it consists of lenses of sandy biomicroparitic wackestones interbedded with uneven, indurated silty shales (Plate 7.9) and together with the underlying beds may represent Facies 10, 11 and 12 in this area, the contact of these beds with the overlying strata cannot be seen. Above the pebbly conglomerates, 8m. of alternating nodular limestones with shale partings which tend to become sandier towards the top of the succession are seen; their lower 4m. consists of occasional Pinna shell fragments (and in life position) lying in beds of shaly limestone with undulose surfaces interbedded with silty micaceous shales. The first Gryphea shells found in the limestones occur at the 20.20m. level together with whole and reworked Pinnashells (Plate 7.10) this bed is followed by muddy, shaly limestones with no obvious beddings, full of crinoid, Pinna and Gryphaea debris, which were regarded by Oates (1976) to represent the Conybeari/Rotiforme subzonal boundary. Towards the top of these beds, Gryphaea shells become very common and Pinna is less frequent (Fig. 7.2); due to faulting and vegetation the higher parts of this section cannot be adequately described or correlated, but a 5.60m. succession of cross laminated pebbly sandstone may be seen with medium bedded (15-20cm.) tabular cross laminated sandy biomicroparites containing long (8-9cm. diameter), well rounded quartzite pebbles, the cross laminated data are shown in Figs 7.4 and 7.5. These are interbedded with thin (2-4cm.) shaly calcareous siltstones and show undulose contacts. The lowermost parts consist of blue limestones with bivalve shell fragments and abundant coalified plant material; the cross laminated sandstones contain laterally thinning beds of silty calcareous shales which become thinner towards the top and show wavy contacts.

The sandstones are very rich in muscovite and also show reddish (ferruginous), yellow weathering rounded pebbles; stratigraphically

significant fossils were not found but their morphology, relative position, high muscovite content, large relatively less abundant pebbles, matrix cement and the presence of reddish (ferruginous) pebbles suggests that they may be correlated with those overlying the Thecosmelia beds in Garbh Rubha, possibly they represent the pebbly cross-laminated strata which underlie shales and limestones of the Conybeari/Rotiforme subzone in Garbh Rubha Bay (NM536706).

7.7 Loch Mudle (NM 553653)

A fragmented ?Lower Liassic succession can be measured along the stream which flows to the north from one of the tributaries of Lochan na Gruagaich, east of Loch Mudle. The total thickness of the succession is 7.80m. and the basal 0.80 cm. directly overlies Moinian schists. As shown on Fig.7.2 white sheets (creamish weathering) and lobes of sandstone (14-20 and 5cm) containing coalified wood and rounded quartz fragments (3mm.maximum diameter) form the lowermost part. The succeeding 40cm. thickness of blue calcilutite abruptly overlies the sandstones and contains fine shelled bivalve fragments together with markings which resemble those of ?Isastrea seen in the northern area of study; a 60cm. thickness of ferruginous-weathering markings overlie the blue calcilutites.

The 1.90m. of marlstones (Fig.7.2) show a light-blue greyish colour and orange-weathering surfaces; the lower $\frac{1}{4}$ comprises thin sheets of limestones and pockets of friable marlstone, towards its top, the friable marls become massive, showing surfaces crowded with marlstone nodules. The marls are overlain by a 35 cm thick bed which shows uneven contacts and quartz fragments scattered on the top surfaces; a yellow-weathering, dark blue silty biomicroparite bed full of small pectinid shells and other indistinct bivalves (100 cm. thick) follows the 1.20m. gap which is shown on Fig.7.2 . The beds over-

lying the fossiliferous limestones are 1.30m. thick and consist of blue limestones with sandy shale partings, quartz pebbles are rare and the uppermost 50cm. forms wedges of ferruginous sandy biopel-sparites, the wedges are 25cm. long and up to 5cm. thick. Richey et al. (1930) maintained that these beds are overlain by 'shales (?Pabba Beds)' which were not seen by the present author, but Gryphaea-bearing shales can be seen to form part of the hill sbpes above the track (NM 551644) leading to Loch Mudle; the strata underlying the Gryphaea bearing shales here may be regarded as equivalents of Facies 2-4 of the skye area.

7.8 Ben Hiant (NM539642)

Along a track (covered and indistinct in most places) a 19m. sequence of ?Lower Liassic strata may be measured towards the north-east of Beinn na h' Urchrach (NM 536642). A bed containing Gryphaea arcuata is seen below the track along the eastern side of the stream on the south side of the fence. Dark shales are also exposed in the more westerly stream due to faulting, the basal contact of these beds with the underlying strata is obscure and mostly covered by vegetation. Sandy limestones up to 30cm. thick which frequently contain strings of sandier material constitute the lower 3m. of this succession, the topmost 50cm. of which is composed of wavy calcareous sandstones with beds up to 10cm. thick. The wavy calcareous sandstones are overlain by 9m. of silty biopelmicrosparites 8-15 cm. thick, with wavy sandstone sheets (5-7mm.); the topmost 2.50m. consist of blue oobiopelmicrosparites with carbonaceous shale shreds and ?Thecosmilia remains, if the coral Thecosmilia is present it may be correlated with the same bed which occurs in Swordle (Facies 8).

A series of even bedded, blue calcilutites with no obvious partings overlies the ?Thecosmilia bed; calcilutites become sandier

towards the top of the succession and are overlain by a bed of pebbly calcareous sandstones 80cm. thick. Undulose partings are common and form lenses which become coarser and poorly sorted at higher horizons, they form the topmost 2m.; the pebble grains are poorly sorted and mainly rounded, sizes vary from 3 to 10cm. and large rounded gneiss fragments are seen together with smaller feldspar fragments.

Quartzite and rounded clasts are very common, some as large as 12cm. in diameter; the pebble beds exposed in the Ben Hiant locality may be regarded as facies equivalents of the pebble beds exposed along the northern shores of Swordle. Although no stratigraphically significant fossils were found in the Ben Hiant Liassic exposures, based on lithology and their distance from the pebble beds exposed in Garbh Rubha Bay, it would be reasonable to suggest a pre-Conybeari/Rotiforme age for them.

CHAPTER 8

FACIES DETERMINATION AND INTERPRETATION

IN MULL AND MORVERN

Facies description and interpretation in the southern area of study (Mull, Morvern) is greatly impeded due to the lack of lateral and vertical control which is a result of limited exposure and extensive metamorphism of the predominantly argillaceous strata. Two distinctly different sedimentary facies may be recognised in the Broadford Beds Argillaceous Group as follows (see also Fig. 2.24):

i. Sandy facies:

The strata are mainly found in eastern Mull and contain a higher percentage of sand-size quartz fragments. These will be referred to as Facies 21b, 22B, 22b.

ii. Shaly facies:

The strata are mainly found in western Mull and in Morvern, they contain a minimal percentage of quartz and mostly comprise alternations of limestones and shales, which are frequently crowded with bivalve shells; these will be referred to as Facies 19, 20, 21a, 22a.

8.1 Wilderness Shale Member

Limestone/Shale Facies 19 (Blue Lias Facies 1 of Oates, 1976)

This represents the lowermost Member of the Loch Aline Formation (Wilderness Member) and is only found at Rubha na h'Iolaire, Ardmeanach in western Mull.

The strata consist of 10.50 m. of alternating shales and limestones (Plate 8.1a, b); the succession is marly at the base showing abundant biogenic features of which vertical and horizontal tubes are the most abundant; these are filled with pyrite (mostly in the form of complete cubes) and in places appear to be branching and interconnected. The limestone beds show no signs of burrowing activity at the basal 2 metres of the sequence and are entirely composed of

dark-blue biomicrites, 30-40 cm. thick.

At the 3 m. level the beds become extensively bioturbated by shallow infaunal burrowers such as normal and vertically retrusive Rhizocorallium (Plate 8.2a,b) coalified plant fragments are very common and occur both in the limestones and shales.

The limestones are silty biomicrosparites and biomicrites; brachiopod and echinodermal debris is scattered while pyrite occurs as shell coating, euhedral crystals or framboidal clusters.

Silt grains are generally <10% and in the size range of 4-5 ϕ ; the limestones contain up to 20% pyrite and the carbonate content is 40-50%; their mica content is minimal whereas rounded micritic and glauconitic pellets (4 ϕ) occur and constitute up to 20% of the rock. The limestone/shale contacts are sharp and show no gradation between the two lithologies. The thickness of limestone beds at the top of the sequence is 20-25 cm. and as seen in Fig. 2.17 their thickness regularly decreases from the basal parts of the succession to the top. Shales are silty (10%) and micaceous with a minimal (2-8%) carbonate content; they are rich in smectites (29%) whereas the kaolinite content is minimal (3-5%); the clay mineralogy of these beds is shown and discussed in Chapter 4.

At the 5 and 4 m. horizons here, limestone beds crowded with numerous ammonites (Psiloceras planorbis) are found; both deep burrowing (Pholadomya) and shallow burrowing (Cardinia) bivalves are present.

Pinna is also seen in horizontal and vertical positions and the bivalves listed by Oates (1976) from this facies are as follows: Plagiostoma sp., Modiolus sp., ?Pteromya sp. and crushed pectinid shells, echinodermal fragments (mostly stems) are present and the ostracod most frequently found is Ogmoconcha sp. (Clark, 1969), coalified plant remains are scattered throughout the section.

It should be mentioned that, whereas the Geological Survey Memoir includes

6 m. of "Red mudstone micaceous calcareous sandstone and greenish grey shales with cream-weathering cementstones" (Lee and Bailey, 1925) in the Liassic succession; the base of the Lias here is taken at the base of the silty marlstones and limestones containing abundant oysters, which overlie the red mudstones.

The uniform succession of the limestones and shales as shown in Fig. 2.17 is disrupted by an irregularity in the form of a down-warped portion in bed no.5, where the limestone bed is attenuated and the shale bed thickens, forming a trough-like feature within the succession (Plate 8.1a).

The shales on either side pass inwards to a sandy glauconitic calcarenite in the centre of the feature (Oates, 1976). Dr. Oates kindly provided the present author with specimens and thin-sections obtained from the sandstone body. The hand specimen is composed of scoop-shaped, thin (4-5 mm.) sheets of glauconitic-calcareous sands stacked one on top of the other forming an elongate body at the centre of which an elongate semicircular (in transverse section), concave upwards, burrow-like feature is found.

In thin section, mica-rich regions containing virtually no calcite cement can be distinguished from areas with high calcite cement content without any mica; it is possible that while an immature, minimally transported sediment was being reworked by tube-forming organisms, the intergranular pore space produced, was infilled by the surrounding carbonate cement. Intergrown subangular quartz grains showing minute fractures are also seen. Displaced feldspar twin lamellae clearly show a post depositional compression and as explained by Oates (1976), the diagenetic migration of CaCO_3 from limestone layers into the more permeable sandstones, resulted in the attenuation of the limestone beds towards the sandy unit. However the interpretation of the sandstone body as an "east-west trending channel" is not acceptable here for the following reasons:

- i. The sediments are not graded.
- ii. The sediments do not show current induced cross lamination.

- iii. No evidence of scouring, channel forming or other primary sedimentary features, which would indicate the movement of currents with sufficient energy to transport this sand in submarine channels is seen.
- iv. Horizontal and vertical burrowers are abundant.

It is possible that while an immature, less transported sediment body was being reworked by tube-forming organisms (which are also abundant in the other beds of this facies), the intergranular pore space produced, was infilled by the surrounding carbonate cement. Despite the above tentative explanations, the origin of the sandstone in this predominantly quartz free marine facies is problematic and may be related to Triassic-Rhaetic "relict" (Swift et al., 1971) sediments.

At Allt nan Teangaidh, Gribun, 95 cm. of laminated grey shales overlie a disturbed development of the limestone/shale succession seen at Wilderness. The fauna is entirely composed of small Modiolus sp. and "non calcitic" Lingula sp. (Oates, 1976) together with abundant coalified driftwood and plant fragments.

There is no faunal or other evidence for the existence of the equivalent of this facies at Craigmure Bay in Mull, where it was either not deposited or faulted-out (the former alternative is preferred here).

8.2 Leacach Nodular Limestone Member

i. Facies 20 (Blue Lias Facies type 2 of Oates, 1976)

This facies is represented by the lowermost 4.40 m. of alternating limestones and shales exposed along the Allt Leacach stream in Loch Aline (NM 693453); it probably represents the Angulata Zone here. The calcareous beds are almost entirely composed of biomicrites, although microsparitic developments and replacement of the various shell types is seen; they show irregular wavy upper and lower contacts with the shales and are mostly <20 cm. thick.

The closely spaced limestones contain relatively small amounts of silt

size quartz fragments (<8%) and are mainly bioclastic, composed of bivalves, gastropods and echinoderms which are mostly diagenetically altered and replaced by sparry calcite. In the lowermost beds, shreds of pyritic material are present and lined with quartz silt fragments; they also contain up to 15% glauconite pellets, their surfaces are irregular and crowded with very small oysters.

The shale beds are up to 25 cm. thick and contain shells of Liostrea hisingeri and other broken bivalves at their base, they are micaceous and coalified plant material is very commonly found in them. Surface markings on the limestones include small blister-like features around which long, wavy, thin tubes of mud are seen. The clay mineral content of the shales shows up to 40% smectites (see Chapter 4) and the mica content is 9%. The succeeding 8.47 m. consists of thin beds (up to 15 cm. thick) of nodular impersistent often closely spaced limestone at the base, containing some gastropods and ?Lima sp.; the shale interbeddings contain up to 35% smectites.

The limestone beds are often composed of biomicrites together with brown, rounded micritic pellets (1-2 mm.); bivalves are seen and are mostly present as uncrushed shells in the shale beds. Pelagic organisms (i.e. ammonites) are found and the marine ostracod Ogmochonca sp. (Clark, 1969) has also been reported. Coalified plant fragments together with wood are scattered throughout the strata; the fauna reported by Oates (1976) are as follows:

<u>Plicatula spinosa</u>	<u>Liostrea hisingeri</u>	<u>Cardinia</u> sp.
<u>Camptonectes punctatissimus</u>	<u>Modiolus laevis</u>	<u>Isocardia arenacea</u>
<u>Chlamys subulata</u>	gastropods	<u>Astarte</u> sp.
<u>Lima</u> sp.	cidarids	<u>Pholadomya</u> sp.
<u>Plagiostoma gigantea</u>		<u>Pleuromya</u> sp.

ii. Facies 21a (Blue Lias type 3 of Oates, 1976) (25 m.).

Beds of this facies, overlying those of Facies 20 along the Allt Leacach stream comprise nodular, widely spaced limestone beds 25-35 cm. thick composed of microsparite with brownish carbonaceous films which give it a somewhat dark appearance; silt-size quartz is less than 5%. The smectite content of the shales is up to 25% and the CaCO_3 content is higher than that of Facies 20a; pyrite is present in the form of framboidal clusters and the glauconite content is minimal; no evidence of current activity is present. Gryphaea frequently crowd the beds and shells of small oysters are scattered within the shales together with other shell debris. The shales are 15-20 cm. thick and contain a higher proportion of Gryphaea and Liostrea sp. fragments; the limestone/shale ratio of this facies is greater than the previous facies and this is evident from the thickness and number of the limestone beds present (Plate 8.3a, b).

The biomicrosparites mostly consist of echinodermal debris together with brachiopods and Gryphaea shells; a few horizons containing small brownish-green "ironshot" limestone pellets with spherical to subspherical form are seen; they do not contain coalified debris. The top surfaces of these beds are marked by a thin ochrous-weathering limestone; ammonites are present and the fauna reported by Oates (1976) are:

<u>Chlamys</u>	<u>Lingula</u>
<u>Plagiostoma</u>	<u>Rhynchonella</u>
<u>Gryphaea</u>	<u>Squamirhynchia</u>
<u>Oxytoma</u>	<u>Pinna</u>
<u>Modiolus</u>	<u>Nuculana</u>
<u>Pleurotomaria</u>	<u>Pleuromya</u>

The top 2.45 cm. is composed of calcareous shales with abundant shell fragments (mostly echinoderm debris); they become less calcareous and more indurated towards the top. The mica content of these shales together with

their silt content is minimal; the topmost 45 cm. consists of alternating biomicrite beds with abundant Gryphaea and echinoderm debris (no quartz) alternating with 10 cm. thick shale beds. Pinna occurs at the lower parts and is mostly disturbed and flat lying; Gryphaea becomes frequent towards the top parts of this facies forming most parts of the calcareous nodular beds. On top of these beds an ochrous-weathering nodular limestone bed with no distinct surface features is seen; it appears to have formed as a crust over the shales; in thin section it is composed of fine micritic lime-mud with opaque pyritic patches and scattered fine siltstone fragments with a few mica plates. The 60 cm. thickness of this Fe-stained bed is composed of thin 5-10 cm., narrowly spaced, nodular layers interbedded with very thin micaceous shales. The lowermost layer appears as a thin veneer over the shales. It shows an uneven gradational lower contact with the interbedded shales. A thin section obtained from this crust shows abundant opaque patches present with a cryptocrystalline aggregate cement which appear to be isotropic at first sight. This cement contains numerous small, rounded, recrystallised objects 0.5 - 1 mm. in diameter which possibly represent remains of reworked micritic pellets and shell material. Very rarely, pockets of rounded, micritic pebbles are found as infills of depressions in the top surface of these beds; Oates (1976) has reported ?pisolitic cementstone nodules and pebbles accumulating on top of the uneven surface of this ferruginous crust in Allt Mor. Plate 8.4a, b were obtained from a thin section of a reworked pebble from Allt Mor (loaned to the author through the courtesy of Dr. M.J. Oates); it shows pockets of poorly sorted rounded recrystallised calcareous pellets with superficial concentric oolitic laminations imbedded in a cryptocrystalline (semi-isotropic) aggregate phosphatic cement. The coating shows a characteristic polarising cross under crossed nicols resulting from the concentric arrangement of radial fibre-like crystals. The fibres are greenish grey with low birefringence and negative

optic sign (chamosite). The occurrence of superficial oolitic rims around these micritic pellets in the Loch Aline vicinity is important and will be discussed later in the section on ironstones.

iii. Facies 21b (Limestone/Sandstone Facies)

Strata representing this facies are best exposed at Craignure Bay (Eastern Mull, NM 731360), where they are mostly metamorphosed beyond recognition; the sequence is also faulted to a great extent. Fig. 8.1 shows the lithostratigraphy as far as it could be determined.

The total thickness of the ^{Craignure} section is uncertain but it is taken to be at least 44 m. from the base (immediately overlying Triassic conglomerates) to the top (underlying green, burrowed, micaceous marls with ironstone concretions and large totally replaced horizontal burrows). This section was not adequately described by the Survey; Lee and Bailey (1925) refer to the lower portion which contains a thin bed of quartz pebbles. Above this the limestones become darker and muddy with Gryphaea increasing in abundance. The contact with the underlying Triassic conglomerates is not clearly visible and in the absence of index fossils for the Angulata Zone, Oates (1976) maintained that the existence of this zone at this locality must remain unconfirmed for the present. At Torosay an extensively baked calcareous siltstone sequence occurs which does not contain a variable faunal assemblage; these beds were tentatively assigned to the Semicostatum Zone (Oates, 1976); they are bioturbated and cannot be studied in detail.

The presence or absence of Liostrea hisingeri is mainly facies controlled (Hallam, 1971c; Hudson and Palmer, 1976); it is common in the Rhaetic Beds of Dorset, Somerset and Gloucestershire and also exists in the "pre-Planorbis Beds" of the Lias of southwest England, Yorkshire and Northern Ireland. The presence of L. hisingeri in the Rhaetic-Liassic beds of northwest Scotland is well known, thus it may confidently be assumed that while it occurs in the

beds immediately overlying the ?Triassic pebble-conglomerates (as elsewhere in Scotland) it represents beds of Rhaetic-Liassic age, deposited under unique facies conditions. The suggestion by Oates (1976) that the Hettangian is

"either absent altogether, faulted down out of sight or is represented in the conglomeratic facies which has been previously considered as of Triassic age"

cannot be accepted here; the Liassic strata conformably overlies the Triassic-?Rhaetic beds of Craignure Bay and the base of the succession consists of very thin sheets (5-15 cm.) of laterally impersistent limestones and siltstones; the siltstones show no sedimentary structures (Plate 8.5), and contain >70% quartz fragments which are mostly normal and well sorted. The limestones are white and crowded with Liostrea hisingeri at the 2.40 m. level; they also contain echinodermal shell fragments in places. The bed containing Liostrea may be the equivalent of the numerous Liostrea hisingeri bearing strata seen in the southern area of study, and which usually occur within the Rhaetic beds or are associated with basal representatives of the Hettangian. Above these beds a 3.50 m. thick alternation of sandstones and limestones is seen; they acquire a nodular appearance and become more indurated. At the 6 m. level above the base of the Lias a bed containing Gryphaea arcuata and Lima gigantea are seen in the limestones, which are metamorphosed biosparites, containing some echinodermal debris; the sandstones are very thin and appear as sheets 1.5 cm. thick.

The succeeding 6 m. consist of a basal 30 cm. bed of quartz-conglomeratic marlstone, above which 1.85 m. of interbedded echinodermal limestone and micaceous argillaceous siltstones are seen. The siltstones contain up to 30% quartz which are rounded and mostly consist of normal quartz; plagioclase feldspars are also seen. Towards the top of this facies (above the sill indicated on Fig. 8.1), the beds become alternating limestone and siltstones; although the beds are thermally altered, their top surfaces show abundant shell remains of Gryphaea and bivalves. The sandstones become very common and thin bedded towards the top

of the sequence; at the topmost parts a very hummocky, almost nodular surface is seen with small conical features (1 cm. high, 3 cm. basal diameter) , which appear as small partly replaced Kulindrichnus langi (Hallam, 1960a) or alternatively, resemble ?annelid or water escape features.

8.3 Leacach Shale Member

i. Facies 22a

The representatives of this facies immediately overlie the ochrous-weathering nodular limestone bed; at the base in Allt Leacach (Loch Aline) an 80 cm. thick development of 5-10 cm. reddish-weathering nodular limestones are seen which are narrowly spaced and show interbeddings of hard, fissile shales. The shales are micaceous (5%) and contain 10% CaCO_3 . They are rich in minute ammonite juveniles; their thickness is 2-3 cm. at the base of this facies, and increases above the lowest development of the basal limestones. Towards the top of the sequence, the thickness of the shales becomes 20-50 cm. at the expense of limestone beds which thin out to 2-3 cm. each; the limestones show a characteristic reddish weathering and are nodular (Plate 8.6).

This facies is distinctly different from that represented by the Loch Aline Formation in terms of clay mineralogy. The smectite content of the shales abruptly diminishes at the base and remains as low as 2-3% throughout the facies (in contrast to $\sim 38\%$ in Facies 21a and below it).

The shales contain <5% silt and remain so until the topmost parts are reached. The sequence is terminated by a 0.50 m. bed of sandy micaceous shale containing abundant limestone pebbles (Plate 8.7). The top 10-15 cm. of which is a matrix supported, glauconitic, micaceous silty biosparite with up to 30% quartz; the bioclastic content is mostly composed of foraminifera, brachiopods, echinodermal fragments together with non distinct, large bivalve shells and smaller finely crushed shells. The quartz content is mostly normal with <2% undulose quartz present, polycrystalline quartz is absent and the grains which are mostly subangular are in the size range of 4ϕ .

A thin section of the "biosparite" beds which overlie the Ardtornish House shale Member (variable age) was provided by Dr. M.J. Oates this specimen is a sandy oobiopelsparite packstone containing ~ 5% fine sand grade quartz particles. The quartz grains are mostly polycrystalline with > 3 crystals per grain and all litho and bioclasts show a thin pseudo-oolitic micritic rim; deformed micrite pellets are present and most constituents show signs of extensive reworking and abrasion. The proper ooliths (Carozzi, 1960) are broken in places and the quartz grains mostly show a superficial oolitic rim. Most grains are rounded and some skeletal fragments show bored surfaces, gastropod tests are mostly infilled with micritic and silt size quartz fragments. The bioclasts are mostly bivalves together with gastropods and some echinodermal debris is also present.

ii. Facies 22b

This facies succeeds the strata of Facies 22B at Craignure Bay (NM 731360) and consists of thin (3-6 cm.), laterally impersistent nodule-like beds of bluish marlstone with Gryphaea; other crushed shells and mud drapes are commonly seen; they are only exposed at low tidal conditions and are found at the western and eastern ends of Port an Seilisdier. The rocks are metamorphosed and nothing much can be seen in hand specimen or even thin sections. Nevertheless it is evident that they were biomicrosparites before transformation. They also show tubular, worm-like ?burrow marking which are replaced by calcite. The limestones contain abundant echinodermal debris together with bivalves. The total thickness of this facies in Craignure Bay is 16 m. where the topmost parts show Thalassinoides markings (Plate 8. 8) and consist of even bedded (10-15 cm.) alternating marlstones and Gryphaea bearing limestones. Their true thickness cannot be determined; nevertheless they are distinctly different from the underlying strata which are mainly ferruginous, greenish, micaceous silty marlstones with large scale

bioturbation features, entirely replaced by iron compounds (red staining oxides).

iii. Facies 22B

(15 m.)

The strata of this facies/underlie beds of Facies 22b and overlie those of Facies 21b at Craignure Bay (NM 731360). Nevertheless the contacts are not readily recognisable in the field due to the intrusion of a sill; at its base limestones are well developed and contain a considerable amount of quartz fragments. They contain sparse, totally metamorphosed bivalve shells and vertically retrusive Rhizocorallium (Sellwood, 1970) are also seen. Towards the higher parts of this sequence (19 m. above the base of the Lias), the limestones become full of bivalve shell material, and Pinna together with gastropods become very abundant; the top surfaces of the beds show hump-like escape features, Zoophycos, Thalassinoides and worm-like meandering tubes with spreite in places (Plate 8.9). The limestones and marlstones are 4-7 cm. thick towards the top and their alternations become 8-14 cm. thick at higher horizons; they are crowded with Gryphaea and Pinna shells (concentrated in the more marly beds). The quartz content of the limestones and marls is minimal (<5%); a sample (Am M48) taken from this facies appears as a well indurated ?metamorphosed marlstone bed in hand specimen; in thin section it appears to consist of a cryptocrystalline (pseudo-isotropic) groundmass in which echinodermal debris was found together with other shell material and some angular quartz fragments < 4 ϕ in diameter. The cement is uncharacteristic of the surrounding rocks (Plate 8.10a-b) and X-ray diffraction studies show them to consist of calcite together with chloritic and collophane mud; marlstones occur, alternating with limestone beds up to the 25 m. level at Craignure Bay becoming abundant towards the top of this facies. They are crowded with Gryphaea shells and show abundant surface markings. The marly beds become darker and show vertical and horizontal

burrows. Mound-like features are seen (Plate 8.11a, b) which are surface expressions of features produced in the underlying less indurated marls; these are extensively burrowed areas, containing broken bivalve and ammonite shell fragments and rounded quartz pebbles. Pinna shells are common, mostly broken or flat-lying, the beds are 10-15 cm. thick and the limestones are mostly composed of recrystallised echinoderm debris; quartz is virtually absent.

8.4 Facies interpretations of the Broadford Beds

Argillaceous Group

The origin of the limestone/shale beds in the Broadford Beds Argillaceous Group is discussed in Chapter 10; it is most likely that they owe their present appearance to the secondary accentuation of primary differences. As the sedimentary structures are scarce and lithological variety is limited in Loch Aline, little interpretation can be derived from the field exposures alone. In addition, thermal metamorphism and faulting has drastically altered and disturbed the succession exposed along the shore of Craignure Bay. The clay mineralogical investigation carried out in the succession of Morvern and western Mull is of some importance to the palaeoenvironmental interpretations (see Chapter 4).

The depositional environment of the beds exposed in the southern area of study was explained in detail by Oates (1976) and while the present author has independently examined and interpreted the Lower Liassic succession, some duplication has inevitably resulted while differences of interpretation still remain.

Generally speaking, the beds of Facies 19, 20, 21a and 22a comprise narrowly spread outcrops found in Loch Aline and Gribun; these two areas constitute a roughly northeast-southwest trending outcrop.

The beds consist of somewhat regular, small scale alternations of microsparitic argillaceous limestones with shales. Such deposits are also seen

in the Hettangian and Lower Sinemurian "Blue Lias" formation of southwest England and south Wales (Hallam, 1960b, 1964). They are rich in benthic fossils and were evidently laid down in quiet, shallow waters. These beds pass laterally to the north and northeast into the shallower water bioclastic and conglomeratic limestones, shales and sandstones (with scour marks) of Ardnamurchan and Skye which have transgressed onto a variety of pre-Liassic rocks.

Towards the east and southeast the limestone/shale beds pass laterally into coarser siliciclastic, ferruginous beds which show evidence of shallower water sedimentation in eastern Mull (Facies 21b, 22B and 22b). Such lateral changes were also reported from the Hettangian-Sinemurian stages of South Wales (Wobber, 1965, 1968).

i. Limestone/shale Facies 19

The limestones and shales of this facies were deposited in a wholly marine depositional environment as indicated by the presence of ammonites.

There is no evidence to suggest wave or current activity, and as already explained, the existence of sandstone lenses suggest the reworking of the underlying formations by burrowing organisms rather than by the activity of submarine channels. The scarcity of detrital silt and other terrigenous material suggests the relative distal position of a shoreline. The somewhat paradoxical occurrence of coalified material in some of the beds together with flat lying and broken Pinna shells suggests the occasional activity of currents which not only crushed and transported plant debris offshore from the land areas situated in the east, they also managed to disturb the shallow bottom conditions and dislodge the Pinna shells. No evidence was found suggesting the removal of sediment or its non-deposition. The abundance of bioturbating organisms as evidenced by the totally reworked beds is evidence for nutrient-rich bottom conditions which were able to support the prolific growth of trace-producing, soft-bodied infauna and the

bivalves. Although sedimentary structures are lacking in the beds seen in Gribun, the depositional environment may be postulated as being of somewhat abnormal salinity, due to the reduced faunal diversity and the presence of Lingula sp. which had a high salinity tolerance. High contents of coalified driftwood and plant debris as reported by Oates (1976) would also indicate nearshore deposition. There is no evidence for the development of lagoons or enclosed basins here (see Fig. 8.2).

ii. Facies 20

Although benthonic fauna are found in these beds, increasing evidence in the form of large coalified wood fragments and "mound-like" escape features which resemble those of Callianassa islagrande, (Hill and Hunter, 1976), suggest nearshore depositional environments.

Mound-like escape features produced by the activity of annelid worms are common in the intertidal flat environment and C. islagrande burrows are the dominant biogenic sedimentary structure throughout the bar-trough system in the lower foreshore environment of northern Padre Island, Texas (Hill and Hunter, 1976). The constant reworking of sediments in the above mentioned environments precludes the preservation of all biogenic and sedimentary features produced on the top surfaces of the beds; furthermore the occurrence of ammonites and echinoderm debris suggests that periods of marine sedimentation were frequent.

The diversity and density of the fauna present does not indicate any salinity reduction or environmental restriction. The beds of Facies 20 were probably deposited in calm, periodically disturbed shallow offshore waters. The lack of siliciclastic fragments may be due to the existence of muddy beaches and shorelines which formed a source for ample plant and wood debris capable of supplying and maintaining the nutrient level required by the abundant burrowing bottom dwelling organisms. The beds of Facies 20 have no lateral representatives in eastern Mull and it is thought that they were either not deposited or faulted out (see Fig. 8.3).

iii. Facies 21a

The abundance of brachiopods, echinoderms and ammonites, suggests deposition in offshore marine environments and the presence of laterally equivalent beds in eastern Mull (Facies 21b) suggest seas which were of greater lateral extent than those in which the beds of Facies 19 and 20 were deposited. The greater lateral extent of sea waters and the consequent increasing remoteness of the beach and shoreline areas may also explain the general lack of coalified plant material in these beds. The formation of framboidal pyrite textures is thought to be a result of crystallisation independent of any pre-existing materials (Elverøi, 1977).

The vertical succession in Loch Aline shows a variety of sedimentary conditions. At their bases, the resemblance of these beds to those of the underlying facies indicates similar shallow water depositional conditions. The succeeding shales contain a greater variety of shell fragments including crinoid and echinodermal fragments, which indicate deposition under fully marine conditions.

The preference of the modern semi-infaunal pinnacean bivalves Pinna carnea and Atrina rigida for sea floors carpeted by growths of Thalassia and eel grass respectively and the abundance of Pinna sp. in the beds of Facies 21a led Oates (1976) to suggest that the sea floor was probably well vegetated during the deposition of some of the beds. The unbroken but uprooted Pinna shells were probably reworked by gentle currents which were not sufficiently strong to break down the shells.

A comparison with the environments in which present day Pinna inhabits and assuming comparable tolerance levels for the Liassic Pinna makes it possible to envisage sunlit, shallow seafloors with muddy bottoms. Evidence for the existence of ?marine plants in the beds of this facies are present.

A reddish-weathering ferruginous, uneven calcareous crust caps the top

of this facies in Loch Aline. Although no surface markings or borings were found on these beds, they were interpreted as signifying a major erosion surface (Oates, 1976; MacLennan, 1953). The mentioned erosion which affected the whole of the Loch Aline area is thought to be the result of a gradual rise in the sea floor until the variable action of waves and wind-induced surface currents managed to remove the topmost beds. Oates (1976) found that erosion was greater in the north of the Morvern area, progressively higher beds being generally preserved at the top of the Leacach Nodular Limestone Member southwards as far as Craignure Bay where a section of Semicostatum Zone beds still remains. The eroded surface of the beds of Facies 21a was subjected to ferric encrustation and pockets of reworked phosphatic pebbles accumulated in the surface depressions. Bernoulli and Jenkyns (1974) interpreted ferric and manganese encrustation and nodule formation as a deeper water process but Oates (1976) regarded the Morvern ferric encrustation as a culminating process which occurred over a long period of time; as the surface underlying the beds of Facies 22a was gradually planed smooth and probably subject to erosion and water current movement, the latter alternative is preferred here. In places, current action during the early part of the Semicostatum Zone time reworked and transported crinoidal and shell debris on the eroded surface of the beds of Facies 21a. Similar deposits elsewhere were interpreted as "sand waves" of shelly fragments resorted by current activity (Bernoulli and Jenkyns, 1974). The initiation of laminated shale deposition over the ochrous erosion surface may have been caused by a rapid subsidence of the sea floor which subsequently accelerated the inundation of the land areas by sea water.

iv. Facies 21b

The beds of this facies represent shallow water equivalents of Facies 21a. The presence of some oysters with normal salinity tolerance indicates

a marine depositional environment while the abundance of sandstone sheets which fine and thin upward, indicates a ?gradual subsidence of the sea floor.

The faulting and metamorphic events to which the beds of this succession were subjected, has obliterated most sedimentary and faunal features; the only recognisable feature seen in these beds was the small, mound-like structures which may have been produced by organisms in the shallow foreshore-shoreface environment (the development of Facies 21a and 21b is schematically explained in Fig. 8.4).

v. Facies 22B

The generally finer grained beds of this facies contrast with the coarser undulating sandstone sheets of the previous facies. The presence of Zoophycos suggests a shallow water, nearshore environment of deposition (Frey and Howard, 1972), Thalassinoides and Callianassa-like features also support this interpretation.

The association of Gryphaea and Pinna shells toward the top of the beds indicate periods of shallow marine sedimentation which is also marked by the presence of abundant reworked echinodermal material

The beds are somewhat ferruginous and contribute to the formation of the ferruginous beds of the facies overlying them (see Chapter 9).

vi. Facies 22a

The presence of brachiopods and echinodermal debris indicates deposition under marine conditions, below effective wave base and far from land areas. The shales were deposited below or very near the limit of oxygen availability as indicated by the lack of normal bottom fauna. Thin laterally extensive nodules of ferruginous limestone may have originated by a periodic incursion of carbonate rich waters of facies 22b. At various horizons however, shell accumulations occur which may represent periods when the benthonic community prospered either due to an increase in water circulation or to a slight

bathymetric change (above the limit of oxygen availability) when more oxygen became available; they may also represent periods of lower sedimentation rate and the development of condensed beds. The pebble beds overlying the beds of this facies are regarded as representing the reworked products of previously formed micritic nodules which were being exposed at the basin margins and were probably being eroded (the development of Facies 22a and 22B is explained in Fig. 8.5).

vii: Facies 22b (fig. 8.6)

The beds of this facies are extensively bioturbated, indicating nutrient-rich sediments which were deposited in relatively calm waters which were not subjected to wave and current action.

Some beds are entirely composed of a phosphatic cement, the origin of which is uncertain. Towards the top of these beds, the replacement of very large Teichichnus burrows by reddish-brown weathering (goethite) ferruginous material occurs and the siliciclastic material is much more common than in the previous facies. This facies was developed in an iron rich mudstone basin which has been discussed and explained in the chapter on ironstones.

CHAPTER 9

FERRUGINOUS BEDS

The ferruginous strata which occur in the lowermost beds of the Lower Lias in northwest Scotland, are confined to the Strath and Læcach Formations of the Broadford Beds Groups. Two types of 'ironstones' are seen as follows:

i. Strata with ferruginous ooliths

ii. Strata with ferruginous muds

i. In the Strath Formation of Skye and Ardnamurchan, beds of calcareous 'chamosite' oolite (Hallam, 1967c), are seen occurring in the basal Reynesi and Sauzeanum Subzones respectively. A limestone bed containing 'chamosite' ooliths is also seen to occur at the base of the Sauzeanum Subzone in Ardnamurchan (Mingary Castle). These oolitic ironstones contain abundant coated echinodermal plates together with abraded bivalve shell debris. The ooliths are of several types showing evidence of pre-lithification compaction in the form of distorted ooliths. Compaction and colloidal shrinkage of the interiors of some ooliths have produced irregular spastoliths (Rastall and Hemingway, 1941). The ooliths consist of a nucleus of quartz around which flakes of 'chamosite' (cronstedtite) form concentric layers; often broken shell fragments form the centre of the ooliths. The ironstone contains up to 15% very fine sand (3-4 ϕ) grade quartz which is unevenly distributed within the sideritic microsparite cement, the bed is laterally noncontinuous and forms broad, laterally impersistent lenses.

The oolitic ironstone bed found at the base of the Sauzeanum Subzone in Ardnamurchan shows less variation among the shapes of the ooliths which are mostly 'perfect' and are found in a microsparitic 'chamosite' siderite cement. In these areas the ironstones do not show evidence of cross-lamination.

ii. In the southern area of study (Craignure Bay), the sediments of the Sauzeanum-Birchi Subzone consist of alternating silty calcareous shales and marlstone beds. The microcrystalline calcite cement is frequently overshadowed by the presence of a cryptocrystalline cement which appears to be (pseudo) isotropic when viewed under crossed nicols and is of very low birefringence (Plate 8.10 a-c). These beds appear as alternating bluish green bioturbated strata in the field. They show distinct horizontal, surface burrows (Plate 9.1) which are in most cases replaced by iron compounds and weather to reddish-brown coloured goethite; the extreme bioturbated nature of these beds and the high iron content of the cement gives them a distinctive surface appearance (Plate 9.2). The bioturbation mostly consists of large non-branching Thalassinoides-type tubes. It should be noted here that this iron-rich unit essentially comprises interbedded siltstone/marlstones in which the cement is composed of microcrystalline chloritic mud. In Loch Aline 'ironshot' limestone occurs at the base of the Reynesi Subzone and also towards the top of the Bucklandi Zone. The mentioned beds consist of small subsphaerical, yellow-brown weathering peloids together with fragments of large bivalves and other shell debris in an argillaceous microsparitic matrix.

9.1 Petrography

As mentioned, two different categories of 'ironstone' are found in the ferruginous beds, these are either oolitic or non-oolitic. The former type occurs in Skye and Ardnamurchan, whereas the latter is characteristic of beds in the Mull area.

i. Oolitic rocks

Oolitic ferruginous beds occur mostly in the Strath Formation in Skye, Raasay and Ardnamurchan; they are seen occurring in the basal Reynesi and Sauzeanum Subzones of the three mentioned areas. The limestone bed containing ferruginous oololiths in Ardnamurchan occurs near Mingary Castle. Only the beds exposed in Ardnamurchan and Ob Breakish (nothern Strath). are truly oolitic; those of Loch Eishort, Loch Slapin and eastern Raasay are mainly pelletal with only traces of oololiths present (Plate 9.3). They are considered together with the 'oolitic' rocks only because of their similar genetic and facies relationships.

Thin section examination of the samples obtained from the beds of the 'Breakish Ironstone Member' in the northern area of study has shown that although each sample is texturally very different from the others, they are composed of the same constituent elements and also show a similar pattern of mineralogical variation.

Five major textural features have been analysed in each section:

a. Oololiths: Bodies showing a degree of tangential chlorite flake orientation, some degree of concentric banding or both, are regarded as oololiths no matter how badly deformed they are. The degree of tangential chlorite flake orientation is usually indicated by a radial extinction brush seen under cross polarized light and occasionally by a pleochroic brush; secondly some degree of concentric banding which may be marked by subtle variations of colour zones, or thin lines of iron oxides, usually limonite or bands of non-chlorite minerals such as collophane or kaolinite (Plate 9.4a-d). The second criterion is used in identifying those oololiths that are completely replaced by non-chlorite minerals. In the majority of the rocks, deformed oololiths are also found. It is difficult to determine whether the shape is an original feature, whether it is due to post depositional compaction or merely a reflection of

tectonic strain.

b. Pellets

These are well defined generally rounded bodies, lacking any tangential flake orientation or banding; they are composed of chlorite and its various oxidation products. The size of the pellets ($\approx 0.2\text{mm}$) is mostly smaller than the ooliths ($0.3\text{--}0.4\text{mm}$).

c. Quartz and feldspar

These are detrital or authigenic; discrete grains and crystals of feldspar or quartz are found either in patches or constitute a main character of the rock.

d. Bioclastic material and calcispheres:

e. Matrix

This consists of calcium carbonate, collophane, chlorite, siderite, iron oxides, very finely crystalline quartz and kaolinite; these form that part of the rock in which all the elements previously mentioned (a to e) are found.

The Breakish Ironstone Member may be termed a sandy silty calcitic sideritic chlorite siderite goethite oolite; it is matrix supported (Wakestone).

a. The ooliths range gradationally in size from 0.2 to 0.3 mm in diameter and their concentration varies in different beds; they comprise up to 10% of the rocks in Ob Breakish while they are virtually absent in the Loch Eishort, Loch Slapin and Rubha nan Leac (Raasay) successions. The oolith concentration is patchily developed with the ooliths frequently clustering together.

The oolith shapes vary from very well rounded to distorted spastoliths (Rastall and Hemingway, 1941). The deformation may be due to many factors one of which could be the result of the compression

produced due to the gentle impact of quartz and feldspar grains on the ooliths.

Two possibilities should be taken into account: if the quartz grains have a detrital origin, then the deformation is possibly due to post depositional compaction and squashing of the grains onto ooliths that were still plastic, i.e. prior to lithification of the rock (Weinberg, 1974; Wilson, 1966). If the grains were authigenic then they must have grown prior to rock lithification. As the grains seldom penetrate the ooliths (Plate 9.5) tectonic compression of the rock after lithification can be ruled out. True spastoliths are rare in these beds.

The causes of spastolith development are not well known; probably they are formed due to differential compaction of the unlithified sediments on a small scale (Weinberg, 1974). The long axes of elongate ooliths are randomly oriented in the specimens, therefore the causes of deformation are questionable and cannot be directly attributed to tectonic stress. Many ooliths contain bands of collophane, these are less affected by deforming stresses.

Some ooliths contain single quartz grains as nuclei. These are angular or rounded and occur irregularly; other nuclei observed include echinoderm plates which are coated or have their pores infilled by chlorite and minute gastropod tests with some shell fragments.

The ooliths more often appear not to have a nucleus; this may be a reflection of the oolith plane cut by the thin sections. Ooliths frequently contain cores that are different from their outer parts in terms of mineralogy and texture. This core may be composed of finely divided quartz grains (sutured or poly-crystalline), collophane or very fine, rounded, mosaic-like masses of vermicular kaolinite crystals (probably replacing feldspars).

Tangential orientation of the chlorite flakes is seen. This is usually well developed towards the oolite margins and sometimes this arrangement is observed to diminish towards the oolite centres. The alteration of chlorite and/or its replacement by minerals such as quartz or rarely by siderite mostly but not frequently commences at the centres of the oolites. This was interpreted by Weinberg (1974) as being due to the concentration of poorly developed, non-oriented chlorite flakes in the centre of the oolites.

In very rare cases, collophane patches are seen together with oolites which contain bands or a core of pale yellow to brown isotropic collophane. As previously mentioned, the concentration of oolites within one sample and even within the area of a single thin section is variable. Compound oolites consisting of at least two subsidiary oolites within a sheath of tangentially oriented chlorite occur but are extremely rare. In many cases extensive alteration of chlorite to goethite has occurred, which may mark or mask the oolite banding. Such altered oolites show no signs of any type of nucleus. Chlorite is more abundant in the oolites than in the matrix; it is also replaced by granular goethite and minute cubes of pyrite. The presence of magnetite cannot be discounted. Goethite is very common in the beds occurring as granules with an appearance that is reminiscent of goethite replaced siderite grains.

b. Pellets occur in the 'ironstone' and other ferruginous members. They are smaller than the smallest size of oolites; it is not known why the oolitic features are lacking in them. Possibly their small size precludes the formation of such features (Weinberg, 1974).

c. Quartz and feldspar

Up to 15% quartz grains of size ranges between 0.06 to 0.08mm. (silt) are found in the 'Breakish' and Loch Salpin ferruginous beds

whereas the sand and silt content of the ferruginous beds in Raasay and southern Strath is up to 40%, with the grains being mostly angular to well rounded. In most cases the quartz grains form discrete bands, show undulose extinction and are undoubtedly of a detrital origin. Some grains show sharp non-undulose extinction which cannot suggest an authigenic origin, and they are well rounded. Most grains show signs of alteration around their margins, however, it cannot be determined whether this is due to the replacement of quartz by chlorite and calcite or vice versa; patches of very finely crystalline quartz are seen. (Plate 9.6a,b)

Feldspars are very rare indeed and large-enough grains were not found to allow a petrographic determination. Grains of definite diagenetic origin are present. Although Weinberg (1974) attempted to explain their presence, it should be said that these grains are present due to a variety of reasons none of which can be affirmed or denied at present.

d. Bioclasts and calcispheres

Gastropods and bivalve fragments are by far the most important bioclastic constituents of the ferruginous beds in Strath. Tests of the former are frequently infilled with iron-rich goethitized muds or form oolith nuclei while the fragments of the latter, 0.01mm. to 2cm. in size, show characteristic shell features which are seldom replaced by ferruginous material. Crinoidal ossicles and echinodermal plates become more abundant (up to 5%) in the Loch Eishort, Rubha nan Leac and Loch Slapin sections while their presence is at a minimum in the beds of Ob Breakish (0.5 to 2%). In Ardnamurchan the bed containing chlorite ooliths contains echinodermal plates (Plate 9.7) and is overlain by a crinoidal packstone containing reworked collophane pebbles and replaced crinoidal ossicles.

Brachipod shell debris is extremely rare to absent in the 'Breakish'

beds and only occasionally occurs in the succession of Ardnamurchan and Raasay.

Inoceramid shell fragments are commonly seen in the beds of Raasay and calcispheres are quite frequently found in the ferruginous beds of Skye, Raasay and Ardnamurchan; they are invariably infilled with ferruginous mud and are mostly of the 'radiolitid' type (Horowitz and Potter, 1971) showing radial divisions in the walls.

e. Matrix

The matrix consists of calcite spar within which patchy developments of granular siderite is seen. The grains found within any area of the rock are uniform in size and there is a gradation between clear siderite and that which has been replaced by goethite. The patchy development of siderite suggests that 'micro-environments' of physico-chemical conditions existed at the time of their formation (Sellwood, 1971; Weinberg, 1974).

Collophane which appears as a submicroscopically crystalline or amorphous phosphate, only forms as a part of the matrix in the Ardnamurchan area. Colloform banding is seen in the oolites in the vicinity of Skye and reworked collophane pebbles are common in the beds found in Raasay and southern Strath. Apart from the occurrences in the form of pebbles, the collophane most probably also formed in situ, this occurs in the form of discrete nodules in the Strath Formation of Skye and Ardnamurchan.

The nodules are widespread and frequently coalesce to form laterally contiguous beds (1 to 5 cm. thick) with flat bottoms and hummocky, wavy tops. Commonly the nodular collophane is not distinctly delineated from the general rock matrix but merges into it. The ubiquitous presence of broken (? digested) shell fragments and structureless, calcite-replaced spheres (? air bubbles) of various sizes (up to 0.5mm) and their absence from the rest of the rock suggests that the association is

related. Although richly phosphatic accumulations formed by pene-contemporaneous subaerial alteration of animal excrement are seen in the modern environment (guano), their presence in the ancient rocks has not been reported (Blatt, et al. 1972); nevertheless such an origin (marine guano) remains a tentative possibility for the phosphatic nodules found in the Strath and Leacach Formations (see later section in this chapter).

The matrix of the 'ironstone' is invariably composed of calcium carbonate and although patches of granular siderite are seen, goethite replacement seems to favour the ooliths rather than the matrix. Pyrite occurs most commonly as minute cubes and irregular clots.

ii. Non-oolitic rocks

The rocks of this class are relatively problematic in terms of mineral composition and no standard term can be adopted to describe their textural constitution.

Non-oolitic ferruginous beds are confined to the outcrops of the Loch Aline Formation found at Craignure Bay in Mull (Facies 22b and 22B). As previously mentioned, these beds are intensively burrowed and become progressively siltier from Facies 22B to 22b. Attempts were made to determine the mineralogy of these beds. X-ray and petrographic studies have produced inconclusive results and although these rocks consist fundamentally of different types of chlorite, further detailed studies are needed.

Two varieties may be distinguished, these are chloritic quartzose mudstones and chloritic siltstones.

a. Chloritic quartzose mudstones

These consist of a very pale brownish, green grey, non-pleochroic, almost isotropic groundmass in which very finely crystalline quartz occur together with echinoderm and bivalve fragments. Samples M55 and

and M48 were taken from the beds of Facies 22B and Plate 8.10 a-c shows the petrographic characteristics of M48,55. It is possible that the near isotropic nature of the matrix reflects the submicroscopic size of the crystals.

X-ray diffraction work on sample M48 showed the possible presence of sepiolite, nevertheless the characteristics of the rock are mostly masked by the very fine grained nature of the matrix. The possibility of isotropic phosphatic material forming the cement was also investigated but characteristic X-ray peaks were not observed on the diffractogram.

It should be noted that both samples M48 and M55 show a degree of resemblance to argillaceous rocks with residual pyroclastic textures but no glass 'relict' textures are seen and X-ray analysis shows minerals of a volcanic suite to be absent.

b. Chloritic siltstones

Very fine silt-size grains occur in these samples (Facies 22B) which may vary in size from $2\mu\text{m}$ to $9\mu\text{m}$ and range from angular to well rounded; it is usually extremely difficult to determine whether they are detrital or authigenic (Plate 9.8a,b). In the overlying beds of Facies 22b the quartz fragments are in the size range of silt to fine sand and the beds exhibit wavy or parallel lamination of the siltier bands which would indicate a detrital origin for them (Plate 9.9)

In some cases the very sharp non-undulose extinction of these quartz grains may suggest an authigenic origin; some of these grains are better rounded and thus the roundness in this case does not indicate their detrital maturing. Limonite and goethite staining is very common and opaque minerals which include pyrite as irregular clots or finely disseminated material and magnetite in the form of minute cubes are also seen; haematite has also developed as the result of recent surface weathering of the rocks. Mica flakes (?muscovite) are very common.

Nodules or pebbles of collophane are absent and this substance is most probably scattered throughout the rock; some of the sub-microscopic material of these rocks may be kaolinite. Finely granular goethite with patchy colour differences in shades of reddish brown, deep red and reddish orange has developed which probably represents different degrees of replacement of the matrix by goethite. This is seen both in the form of nodules and patches, wisps and stringers developed in the weathered rock.

9.2 Mineralogy

Attempts were made to determine the composition and mineral type of the ironstones and phosphatic rocks. Operational details of the methods used are given in appendix 3, 5, and 6. These include X-ray diffractometry of whole rock specimens. X-ray powder photography was carried out on carefully separated ferruginous ooliths of sample Am66-1; major element determination was made by X-ray fluorescence (XRF) and wet chemical methods.

Sample Am66-1 was taken from the Breakish Ironstone Member in northern Strath, Am68 was obtained from a phosphatized internal mould of a brachiopod found within the micaceous shales of the ?Reynesi Subzone of southern Strath, sample Am88-2 and Am88-22 were found in the top-most beds of the micaceous, calcareous silty shales of Facies 16. Samples Am155 and Amb72 were obtained from immediately below the beds of facies 14 (Breakish Ironstone Member) in northern Strath, Am79 was taken from a position at the base of the beds of Facies 16 (in Loch Eishort) above the lateral equivalent beds of the Breakish Ironstone Member. Specimen Ra32 was obtained from the phosphatic beds of the Semicostatum Zone shales of Raasay and AD58 was taken from the shales of Semicostatum Zone age in Ardnamurchan.

i. X-ray diffractometry and powder photography

The results of X-ray diffraction analysis of powdered ($<2\mu\text{m}$) samples are presented in Tables 9.1 and 9.2. Rectangular cavity mounts were prepared from the finely powdered samples, thus preserving random particle orientation to a large extent. The compression employed in mounting the sample produces some degree of orientation of the slaty minerals present but this does not affect the other minerals present and as no attempt was made in the quantification of the relative proportions of minerals present, the use of these 'random orientations' proved to be useful although not strictly conclusive.

Reflections of $\text{CuK}\alpha$ radiation were measured from $2^\circ=2\theta$ up to $40\text{--}50^\circ=2\theta$ at a scanning speed of $1^\circ=2\theta$ per minute.

It should be noted that for the analysed samples, all the peak intensities are given relative to an intensity of 1 for the most prominent peak and position and in decreasing order of prominence thereafter; the ASTM relative intensities are measured against the most prominent line with an intensity of 100.

The superposition of reflections from various different minerals produces a generally crowded and diffuse diffractogram from each sample and the 2θ values cannot be accepted with great confidence. Furthermore, as Differential Thermal Analysis (DTA) was not carried out on these samples, the results should be considered with caution. In an attempt to separate the various textural elements of Am66-1 and in order to minimize the matrix effects and superimposed peaks found in the whole rock analyses of the Breakish Ironstone Member, a sample was gently crushed to small granules and subsequently the dark ferruginous oolites were hand-picked under a microscope; they were then powdered in an agate pestle and mortar and prepared for X-ray powder photography.

The results shown on Table 9.1 and 9.2 show that a considerable proportion of the samples obtained from the nodular phosphatic beds

consists of calcium hydroxy apatite. An accurate distinction between calcium hydroxy apatite and calcium fluorapatite is not possible with the XRD method. The method of Altschuler et al. (1952) could not be used and as it proved to be impossible to determine the fluorine content of the sample by wet chemical methods, it should be concluded that most of the phosphates are of the calcium hydroxy apatite type.

Although the X-ray diffractograms of the iron-rich beds are very inadequate, most samples except Am88-2 and Am88-22 show a closer affinity to cronstedtite than the other chlorites ('chamosites'), indeed the X-ray powder photograph obtained from the individual ooliths (Plate 9.10) shows the three prominent lines of this mineral ($d\text{\AA} = 7.09$, $I = 1$; $d\text{\AA} = 3.54$, $I = 2$; $d\text{\AA} = 2.722$, $I = 3$) and thus its presence here may be inferred with some degree of confidence.

ii. Analytical chemistry

The conventional methods used for sample collection for geochemical purposes were not used here due to isolated exposures of the phosphatic rocks and the lack of properly defined horizon/strike in the ironstone. In the latter case, samples were collected along the Ob Breakish main exposure. These were cleaned of weathering surfaces and crushed in a steel pestle and mortar. To minimise the contamination effects, the steel pestle and mortar was precontaminated by crushing up to 200gr of the samples (discarded) before preparing the material used for the study. The samples thus crushed were thoroughly mixed and crushed in an agate swing mill to a fine powder. Great care was taken in order to minimize contamination during each stage of the procedure. Ideally the separation of 'detrital' and 'non-detrital' fractions from the sediment was sought (Hirst and Nicholls, 1958; Gad and Le Riche, 1966); but this was abandoned due to its obvious impracticality.

Major element determination was carried out on 10 whole-rock

samples using a combination of X-ray fluorescence spectrometry (XRF), atomic absorption spectrophotometry (AAS), colorimetric and gravimetric methods; the samples were exactly the same as those used for X-ray diffraction study. Specimen Am88-2w was taken from the silty micaceous calcareous shales in which Am88-2 was found. The analyses were attempted under the guidance of Dr. A. Wilson who also kindly undertook part of the XRF analyses and performed all the final computations applying his original computer program modifications. The obvious uncertainties concerning 'representative' samples and analyses, undoubtedly remain unsolved particularly in the case of rocks with the heterogeneous textures mentioned above.

The XRF method of Norrish and Hutton (1969) was followed to determine SiO_2 , Al_2O_3 , TiO_2 , Fe_2O_3 (total iron), MnO , MgO , CaO ; the method described by Wilson (1955) was used to determine FeO (see also Shapiro and Brannock, 1962). The difficulties involved in taking the phosphatic rocks into solution together with instrumentation problems called for a wet-chemical procedure which mainly developed through trial and error (see appendix 6). The Na_2O content was determined by the AAS method and the CO_2 content was determined by a gravimetric method following Bauer et al. (1972). No attempt was made to determine the CO_2 content of the apatites, (Gulbrandsen, 1966), similarly the distinction between the CO_2 combined as calcite in the carbonate bearing apatites or present in some form other than calcite (Silverman, et al. 1952) was not made.

The United States Geological Survey Standard rocks were used for the calibration of the analyses and the results obtained were compared with the values recommended by Flanagan (1973). See Tables 9.3 and 9.4.

9.3 Phosphatic rocks

As the phosphatic nodules and thin discontinuous beds rich in

phosphorus found in the Strath and Loch Aline Formations show a relationship with the ferruginous beds, they are briefly considered and tentatively interpreted within the framework of this section.

In the field they may appear as dark, bluish grey, brown rocks and are sometimes carious.

Petrographically these rocks are composed mostly of a light yellow to light brown massive, colloform cement which is somewhat isotropic due to the amorphous nature of the collophane which forms the main constituent of the rock. Collophane (Rogers 1922) is a collective term which is applied to a variety of crystalline phosphate minerals (carbonate apatites, francolites.... etc). Within this cement various types of constituent grains are seen; it should be noted that reworked collophane pellets are commonly found as scattered constituents of the beds equivalent to or overlying the Breakish Ironstone Member.

The phosphatic nodules are well consolidated rocks composed of intact and fragmented, non-replaced shell debris (bivalves and echinoderms) together with microfossils (mainly forams) and calcispheres set in a collophane/micrite matrix. The sample taken from southern Strath is mainly composed of quartz fragments (up to 55%) together with large gastropod tests, echinoderm plates, bivalves and microfossils together with calcispheres; kaolinite flakes are present, forming oolitic rims around some quartz grains. The skeletal material in these rocks is invariably calcite and has not been replaced by phosphate minerals. It appears that the cementing material is an intimate mixture of collophane and micrite with the colour of the cement being dependent on the different proportions of these minerals.

The microfossils and gastropod chambers are invariably infilled with phosphatic cement and although some calcispheres show non-replaced 'radioles', their central sphere is infilled with phosphatic material.

The collophane-rich cement is pale to moderate yellow under the microscope and an increase in its micrite content is responsible for the darker patches; minute brown particles of ?goethite are common throughout the cement which may become brown inside the microfossil chambers (?goethite). Optically it is not possible to distinguish between the collophane, micrite and goethite of the cement. The fossil fragments form grain supported textures in parts of the specimen taken from Ardnamurchan whereas within one thin section slide, areas can be seen which are entirely matrix (cement) supported.

The Ardnamurchan specimen (AD58) contains abundant spherical objects which show no obvious sign of internal banding. They show various stages of formation shown in Plate 9.11 evolving as it were from an apparently 'reniform' feature; very few of these 'ovulate' features show remnants of what may have once been concentric banding (Plate 9.12). Some reworked oolitic fragments are seen and indeed some of the spherical bodies contain previously formed 'ooliths' and they all invariably show a brownish inner section. The various 'pinch and swell' features together with the 'necking' of these ovules suggests an origin due to the rupture of soft sediments in a turbulent environment at the time of deposition or to effects of compaction (Conley, 1977), it is also possible that they were diagenetically formed due to the surface and/or volume changes which took place during the diagenesis of pore waters. It should be noted that some of these 'ovules' found in specimens taken from Loch Aline show a thin coating of chlorite (?kaolinite) flakes. (Plate 8.4a,t

It is probable that the subsequent reworking of previously formed 'ovules' formed a nucleus around which flaky minerals gathered; such processes during the deposition of pre-'Breakish Ironstone' phosphatic nodules may be partly responsible for the formation of those chloritic ooliths in the ironstone which show no signs of a definite nucleus.

9.4 Depositional environment of ironstones

Although economically unimportant, the formation and deposition of iron-rich sediments during Semicostatum and Turneri Zones in northwest Scotland may serve as important palaeoenvironmental indicators. A problem posed by both older and younger iron-rich sedimentary rocks, is that of the source of iron. This question, together with the problem posed by the presence and nature of the ooliths, is also briefly considered here. After considering the various models proposed for the formation of ironstones, an explanation for the formation of the Scottish Lower Liassic 'ironstones' is given. Lower Liassic ironstones reported from northwest Europe are presented in Table 9.5.

i. Source of the iron

Weinberg (1974) showed that, in general, ironstones were enriched in terms of iron by 'a factor of more than seven' as compared with 'average' shales. A number of processes may be responsible for this phenomenon.

a. Diagenetic processes - The works of Porrenga (1965, 1966, 1967) and Rohrllich et al. (1969) in Recent sediments has shown that authigenic iron minerals may form due to the mobilization of iron after deposition. post burial and/or sea floor reactions were held responsible for the release and subsequent selective reprecipitation (and concentration) of iron (Hallimond et al 1951; Goldschmidt, 1954; Borchert, 1960, 1965; Dunham 1960; Strakhov, 1969). The remobilization of iron after deposition was confirmed by Curtis (1967) and Curtis and Spears (1968) who found siderite in many ironstones.

The amount of iron remobilized in the above mentioned works is inadequate to explain the formation of large accumulations of economically exploitable iron-rich beds. Major objections have also been made to Borchert's (1960) work by Hallam (1966, 1975), effectively rejecting his hypothesis for the formation of Jurassic ironstones.

b. Volcanic, hydrothermal and metasomatic origin - Many workers have favoured such origins for the genesis of ironstones (Hise and Leith 1911; Oftedahl 1958, 1959; Harder, 1963; Dzotsenidze, 1972). Castano and Garrels (1950) suggested that the Clinton iron ore deposits were formed by the metasomatic alteration of pre-existing sediments. White et al. (1963) have shown that high concentrations of iron exist in volcanic waters.

Although Mesozoic volcanic activity has been reported from southern and eastern England by Hallam and Sellwood (1968, 1970), Sellwood and Hallam (1974), Bradshaw (1975) and Jeans et al. (1977) there is no direct connection with the Mesozoic ironstones of England, Lorraine and Luxemburg. Permo-Triassic volcanic activity in the Hebrides and north west Europe may be envisaged (Amiri-Garroussi 1977) nevertheless, the establishment of a genetic relationship with the Breckish Ironstone Member is not possible.

c. Terrestrial weathering - The weathering of land areas can produce enough iron, which if transported and deposited under favourable conditions may produce commercially viable iron-rich beds. It is known that weathering under humid tropical and subtropical conditions releases the iron (Gruner, 1922; Moor and Maynard, 1929; Taylor, 1949, 1969; Sakamoto, 1950; Alexandrov 1955; Catt et al. 1971; Bubenicek, 1961; James, 1966; Millot, 1970). The iron derived from such weathering action is transplanted by one or a combination of three possible ways: Firstly, in true solution either as ferrous iron or organo-metallic complexes. The iron concentration of present day surface water is very low (Lewis and Goldberg, 1954; Durum and Haffty, 1963; Blatt et al., 1972); and it is highly unlikely that high concentrations of

iron existed in solution as simple ions in the Liassic sea waters. Organo-metallic complexes (Huang, 1973) are found in the present day surface waters and it is quite possible that significant source areas for organic matter existed on Liassic Lands.

Secondly, ferric hydroxides with iron in colloidal or particulate form may contribute to the formation of the ironstones (Lewis and Goldberg, 1954). The works of Laevestu and Thomson (1958) showed that coastal waters generally contain a higher iron concentration than oceanic waters because of the source area proximity and supply of organic matter.

The shortcoming of the above-mentioned process is that it cannot provide adequate concentrations of iron.

Thirdly, iron may be transported into basins of deposition, adsorbed onto clay micelles or as a constituent of the sheet silicate structure; James (1966) regarded the latter condition quantitatively less significant. The work of Carroll (1958) showed that the iron content of soils is dependent on the proportion of its clay fraction which would be affected during weathering under warm and humid climatic conditions (Mohr and van Baren 1954; van Houten 1961; Tomlinson, 1916). The erosion and transportation of these soils from the surrounding pre-Cambrian and Palaeozoic schists, gneisses and granitic land areas may have contributed to the formation of 'ironstones' in the northern and southern areas of study during the Liassic times.

ii. Formation of ooliths and depositional environment - Results of the present work suggest that the concentration of iron and of ooliths in the Breakish Ironstone Member are somehow related; although Weinberg (1974) showed that this is not necessarily the case with all oolitic ironstone beds, previous researchers working on the Phanerozoic minette-type ironstones assumed that in all cases, an inter dependence

existed among the ooliths and the amount of iron. The majority of hypotheses dealing with the mode of oolith formation in ironstones are based on an analogy with calcareous examples because chloritic ones have not been found forming in present environments.

a. Replacement of original ooliths

The formation of ooliths in ironstones and the problems involved were first considered by Cole and Jennings (1889) who favoured an interpretation based on the replacement of originally calcareous beds through metasomatic alteration. This view was also supported by Fearnside (1907, 1910), but not for long (Fearnside and Davies; 1943 see also Cayeux 1922). Calcareous material is very abundant in the Breakish Ironstone Member in the form of cement and replaced shell fragments; the ferruginous beds of Craignure Bay, Mull also contain a modest amount of CaCO_3 and shell debris is present, albeit rare. It is less possible that the beds in question were originally calcareous.

b. Direct formation of iron-rich ooliths

Curtis and Spears (1968) showed that ferrous minerals cannot be found in the depositional environment. It is therefore highly unlikely that iron-rich ooliths formed by the agitation of nuclei on the sea-floor in waters supersaturated in Fe^{+2} and Al^{+3} and S^{-2} (contrary to the view held by Hays (1915), and Hallimond et al (1951)). The formation of iron minerals requires extreme physico-chemical conditions found in some pore waters. Such conditions would only be found in highly anoxic depositional waters rather than agitated, aerated waters (?interstitial).

The possibility remains that originally ferric minerals formed and the iron in them was subsequently reduced during diagenesis. In such cases (i.e. when the ferric ooliths developed as calcareous varieties)

an upper size limit would be expected of the ooliths (Twenhofel, 1950) due to the limited ability of the agitating currents to move

ooliths of more than a certain size, causing the discontinuation of their growth by mechanical accretion (Carozzi, 1957). It is also possible that growth balance is reached either by accretion of material and abrasion caused by inter-oolith collisions (Weinberg, 1974); such a size limit was not seen in the beds of the Breakish Ironstone Member and although an upper size limit of 1mm. is seen, All sizes less than this are present in the rock. Moreover, bivalve and echinoderm shell fragments up to 1cm. long together with large gastropod tests are very common. As shown in Plate 9.5 some ooliths were originally soft and liable to flattening and plastic deformation; these would have been very susceptible to reworking processes. The amount of deformation seen in the ooliths is not related to their sizes and normal oolith-forming conditions are hard to envisage for these forms.

Knox (1970) suggested that the chlorite flakes carried in suspension by gentle ocean-floor currents may adhere to suitable nuclei and be rolled into an accumulation site as a result of scouring; this removal of newly formed ooliths from the site of formation would prevent the sediment surface area from over-crowding which would result in the cessation of the process.

Although the ooliths formed by this mechanism could be buried by bioturbation, a number of objections have been noted for this process by Weinberg (1974) which are relevant to this study as well. Firstly, the ooliths found in the Breakish Ironstone Member do not necessarily show a central nucleus. Secondly, other than in the present ferrous state, the iron would have originally had to be in a ferric state. Thirdly, in the ironstones under consideration, the chlorite flake orientation starts at the oolith margins and only rarely are they seen to extend to the centres; according to Knox's (1970) work, this phenomenon would be expected to occur throughout the body of the ooliths.

The tangential orientation of chlorite flakes mainly at the oolite margins would be explained by the fact that the oolite margins would start to show the diagenetic reduction of iron to the ferrous state; nevertheless, Weinberg (1974) failed to observe the transition of the ferrous minerals on the margins, to ferric minerals in the central portions of ferruginous oolites.

c. Biogenic origin

An origin based on the biogenic forms such as algal growths or faecal pellets was proposed by various workers (Porrenga, 1967; Rohlrich et. al, 1969). None of the forms seen in the oolitic ironstones of northwest Scotland show the size, geometric irregularities and typical mammilated structures distinctive of algal onkolites.

Many shell fragments show surface borings and some boring marks are also seen on the concentric laminae of a few oolites (Plate 9.5).

Onkolites are sometimes enriched in phosphatic material, nevertheless it cannot be shown that the phosphatic material concentration in the Breakish Ironstone Member shows any relationship to the concentration of oolites.

Although information regarding the phosphatic and carbonate content of originally formed structures is lacking and despite the fact that there is no direct evidence of organisms with sizes large enough to produce coprolites up to 1.5mm. in diameter, a faecal pellet origin cannot be ruled out.

d. Concretionary origin

The slow evaporation of a concentrated solution of ferric chloride by Bucher (1918) produced spherical bodies with concentric growth rings and a radial crystal structure. Pulfrey (1933) suggested that the same process could be responsible for oolite formation

this origin would explain the wide size range of the ooliths observed together with the concentric banding.

Jenkyns (1972) also observed the transformation of the constituent crystals from a radial to a tangential orientation during the recrystallization of aragonite to calcite in the process of oolith formation.

Evidently the concretionary development of iron-rich ooliths requires an original sediment mainly in the form of a colloidal emulsion. The presence of a cement which is mainly composed of calcium carbonate rejects such an origin for the ooliths of the Breakish Ironstone Member. although the possibility that they were transported from their generating environment still remains.

e. 'Mud Balls'

Weinberg (1974) suggested that mud balls (Bell, 1940) of various sizes (diameters range from centimetres to less than a millimetre) described from various present day sedimentary environments (Nordin and Curtis, 1962; Dickas and Lunking, 1968; Stanley, 1969a) may be responsible for the formation of ooliths in the Welsh Ordovician ironstones; he postulated that unconsolidated sediment would be subjected to numerous earthquake shocks. Some would mobilize the sediment which would form turbidity currents of localised distribution. During this erosional process mud balls could form and would then be redeposited from a waning current. Although the iron in the Liassic seas may have been preconcentrated during lateritic weathering of low lying land areas in a manner somewhat analogous to the Welsh Ordovician ironstones (the iron being transported into the basin absorbed onto clay flakes), it is impossible that any locally derived volcanic material was responsible for their formation. Although indirect evidence points to an unstable hinterland and although these beds were most probably deposited in

relatively shallow waters, they are not associated with non-sequences at this level. Furthermore, the beds are not graded and evidence of current activity other than shell fragmentation and very faint cross lamination is lacking. The presence of delicate calcispheres and foraminiferal tests suggests less vigorous current activity. If 'mud balls' were redeposited despite the mentioned constraints, chlorite formation would continue as previously mentioned, the chlorite flakes developing a random orientation. The growth would commence at or near the margin of the 'mud balls', being controlled by the percolation of pore waters (Weinberg, 1974).

The original formation of the Breakish Ironstone Member as 'mud balls' as described above is quite a possibility.

The occurrence of bands of iron oxide, kaolinite and collophane within these 'balls' together with the associated chemical variations were semi-quantitatively investigated with the aid of an electron-probe microanalyser by Weinberg (1974); as the mentioned chemical and mineralogical banding phenomenon is of key importance to the interpretation of the origins of these oolitic features. Several attempts were made by the present author to reinvestigate this phenomenon on a more refined scale. Nevertheless this proved impossible due to more than frequent equipment failures, inaccuracy and the obvious length of time involved (the proposed project was abandoned in April, 1977).

A certain minimum concentration of the components of any mineral should be reached before precipitation will take place. Weinberg (1974) maintained that under the conditions prevailing in the 'mud balls' chlorite formed and growth continued inwards from the oolite margins. As the calcium and phosphorus ions present would not be incorporated in the chlorite structure, they would be pushed towards the oolite centre to become progressively more concentrated by a mechanism similar to the

'industrial progress of zone refining'. At stages when the pore solutions become supersaturated with Ca^{+2} and P^{+5} ions a band of collophane would form and consequently chlorite begins to precipitate as a result of undersaturation with respect to calcium phosphate. When these successive super- and under-saturation effects occur during the developmental stages of the oolite, different bands or a core of collophane would be formed. The relative impoverishment of iron would allow the formation of kaolinite rather than chlorite and after depletion of the excess aluminium and silica in the diagenetic fluids, chlorite formation would resume.

This process of 'zone refining' is preferred to the concept of 'formation in an oxidizing environment' by Weinberg (1974), to explain the bands of goethite which are seen in the oolitic structures. This process would appropriately be responsible for the same phenomenon occurring in the Breakish Ironstone Member since there is no evidence for an emergence causing the rapid change in the Eh and pH values which would contribute to the formation and precipitation of the iron oxide minerals. Furthermore, an overall change in the redox potential would be 'registered' in most to all the iron-bearing grains. This is not the case with the beds under study and only some oolites show this banding phenomenon.

The susceptibility of the chlorite flakes to chemical attack may be a function of the 'closeness of the packing of the flakes' (Weinberg 1974). As the zone refining process is accepted here to be partly if not wholly responsible for the formation of the iron-rich oolites in the considered beds, it is quite possible that the loose packing of chlorite flakes in places would render them prone to chemical attack and the formation of iron oxides in the form of bands, patches and microlenses. This is certainly the case with the beds here. Compound

ooliths may be the result of more than one cycle of deposition, reworking and chlorite growth.

f. Feldspar destruction

In addition to the points mentioned above it should be emphasized that feldspars are very rare to absent in the Breakish Ironstone Member (0 to 0.6%) and the nuclei of most 'ooliths' in these beds are composed of finely vermicular kaolinite, which would hardly be resistant to any form of agitation whatsoever.

Occasionally ooliths are seen which consist of an angular grain of feldspar as nucleus showing very well defined twin planes but strongly altered margins which gradationally merge into a thick coat of vermicular kaolinite. This coating covers the whole of the feldspar grain and is in turn covered by a 0.3mm.thick layer composed of chlorite flakes which are flattened (Plate 9.13a-c).

The inter-related effects of weathering and transportation on feldspars are poorly known. Nevertheless the effective strength of reworked feldspars is less than fresh ones, this may be significant in their destruction during transportation. Structurally unstable surfaces such as twin composition planes, phase boundaries between perthite lamellae and cleavage planes may contribute to this destruction. As mentioned, there is ample evidence in the studied rocks for the alteration of the feldspar grains to kaolinite. This is seen in the form of grains that have all of the geometric attributes of a detrital feldspar fragment but are entirely composed of kaolinite vermicules, a composition which would be highly unlikely to survive as an original detrital grain.

Pettijohn, Potter and Siever (1973) have shown the reaction of the hydrolysis of a potash feldspar and have indicated the relevant chemical variables. Following Garrels and Christ (1965), they predicted that if $\log a_K/a_H$ values fall much below 5 or 6 in the chemical environment

H_4SiO_4 concentrations drop below 10^{-3} moles/lit., then potash feldspars will kaolinize if thermodynamic equivalence is reached.

The in-place kaolinization of feldspars may be due to the existence of groundwaters which are low in dissolved solids, including K^+ and H_4SiO_4 and which have a relatively low pH because of dissolved H_2CO_3 . Ground waters of meteoric origin within one to two hundred metres of the surface fit the general characteristics needed for waters responsible for the formation of kaolinite also described by Millot (1970).

As explained, the development of vermicular kaolinite in sandy beds and sands is a result of the precolation of acid waters through porous sediments. Kaolinite develops in various lateritic profiles. Tropical and equatorial rainfall ensures vigorous lessivage during humid seasons; the ions are released through hydrolysis and the electrolytes removed. When a permanent water supply provides the required silica, alumina organizes with it to form kaolinite. The kaolinite thus formed will develop a ring around the central feldspar nucleus. The inevitable process of equilibration with the diagenetic fluids may promote the formation of chloritic flakes which will be forced into a tangential orientation due to mechanical requisits.

The clay mineral analysis of the shales of the Strath Formation shows them to be devoid of smectitic clays (contrary to the Milton and Loch Aline Formations) whereas chlorite and illitic clay minerals develop (see Chapter 4). Probably the environmental requirements of chlorites (erosion of feldspar) overshadowed those which prevailed during the formation of the smectite-rich underlying beds.

The Strath and Leacach Formations were deposited under similar chemical conditions in two geographically separated areas with different environmental characters. In both areas deposition took

place under moderately deep subaqueous conditions, nevertheless, the northern area received a greater amount of clastic silicates than the southern area. Furthermore, sedimentation took place under agitated more turbulent conditions in the former area as evidenced by the abraded nature of the shells and their poor grading. In the southern area relatively calmer deposition of iron-rich muds prevailed, the iron being transported as adsorbed ions on clay flakes weathered from the surrounding schistose and gneissic land areas, and then transported into the depocentres which stood further away from land. The iron-rich muds deposited in this relatively offshore environment was extensively bioturbated by horizontal and vertical burrowers.

This basin progressively shallowed towards the east, northeast and north. Nearshore shallower situations prevailed in Loch Aline during the deposition of ferruginous muds in Mull. In Loch Aline an adequate 'flushing' system prevented the super saturation of the waters in ferruginous ions, therefore the only form in which they are seen concentrated are small pellets ('ironshot') probably washed-in from the Mull basin due to current generation or wind action. The very thin (2 to 3cm) ferruginous nodular beds of micritic limestones seen may have also developed due to the periodic incursions of iron-rich waters by a similar mechanism. X-ray diffraction analysis of whole-rock specimens show the presence of pyrite in these beds which may be secondary but indicates the existence of appreciable amounts of iron.

Towards Arndamurchan, the offshore areas become slightly shallower and although quartz clastics are present and the cement is highly calcareous; ferruginous muds similar to those of the Mull area are common together with pellets derived from the ferruginous muds. Feldspars with vermiculite coatings and chlorite-kaolinite outer layers are common; foraminiferal and echinodermal remains together with calcispheres suggests

that shallow water conditions were probably not present.

In the northernmost area (Skye) offshore shallow depositional conditions may be envisaged. Ample terrigenous material was supplied into the small basin. Some pellets and 'mud balls' originating in the Mull basin were carried into this northern basin where the transformation of feldspars was rapidly taking place and the waters were increasingly acid. The fauna was still dominated by echinoderms and forams but bivalve shell fragments and gastropods are seen together with calcispheres.

The bivalve shell debris are nowhere seen to be replaced or composed of iron-rich material, therefore they were possibly transported into the environment from less terrigenous conditions whereas the entirely coated echinoderm plates and infilled calcispheres most probably have their origins in the deeper, offshore areas of the basin namely Mull. Some gastropods show geopetal features with their tests partly infilled with ferruginous muds containing a partially transformed feldspar (Plate 6.60). Although the basin of deposition during this time was a continuous one with beds containing oolites and 'ironshot' limestone representing basin margins and near(er) shore conditions; the oolitic beds of Ob Breakish and Ardnamurchan are non-contiguous and the sandy silty ferruginous beds of southern Strath occur between them. Beds of similar age in Raasay and southern Strath contain a very high proportion of quartz sand but only sparse, reworked oolites were observed. A model was proposed by Huber and Garrels (1953) in which topographic depressions on the sea beds near submarine highs attracted most of the clastic sediments while undiluted chemical deposits such as ironstones formed on the 'swells'. Although this model is accepted by Sellwood and Jenkyns (1975) for the Pliensbachian-Bajocian ironstones of England and Scotland, Brookfield (1971) contested this idea; but Knox (1970)

found evidence which suggested that a thin Middle Jurassic ironstone bed in Yorkshire was the site of accumulation and not of formation of the ooliths. Thus this model would be accepted here under the terms explained by Sellwood and Jenkyns (1975). The lack of characteristic condensed horizons (Dachbanke) or hardgrounds within the ironstones and with respect to the laterally correlative clastic deposits suggests that such an interpretation can only be tentatively put forth, furthermore the beds only faintly show very shallow current generation features.

9.5 Depositional environment of phosphatic rocks

The origin of phosphatic rocks presents a long-standing controversy since they were first found during the Challenger Expedition in the 1870's (Tooms et al. 1969; Gulbrandsen, 1969). Two main lines of thought have dominated the theories to explain their occurrence in the marine environment.

Kazakov (1937) and Dietz et al. (1942) believe that phosphatic rocks are the result of direct precipitation from phosphate-saturated solutions. Kazakov (1937) postulated a general relationship between the phosphate-rich beds found on submerged continental platforms and the phosphorous-rich upwelling oceanic waters.

Direct precipitation of apatite out of natural waters other than pore waters is not well documented (Richards et al. 1965; Murphy, 1973) but Kazakov's (1937) proposed model gained wide acceptance.

After the consideration of various pre-requisites necessary for the formation and precipitation of apatite, Burnett (1977) showed that apatite probably precipitates from pore waters of anoxic sediments rather than from the overlying ocean water. The various conditions suitable for the precipitation of apatites as discussed in detail by Burnett (1977) include:

- i. High flux of dissolved inorganic phosphate

- ii. Magnesium depletion during diagenesis
- iii. High pH
- iv. Suitable nucleation sites
- v. Warm conditions

He has also found conclusive evidence for the replacement origin of phosphates in the form of elongated apatite crystals grown in the interstices of biological debris i.e. diatom frustules. The calcispheres seen in the samples taken from Skye and Ardnamurchan, although they contain a phosphatic central body-sphere, show very delicate, non-replaced calcareous radial features.

Some of the ancient phosphatic deposits form on or along the flanks of tectonic basins (Youssef, 1965) in conditions of relative isolation from general oceanic circulations. D'Anglejan (1967) found marine phosphorites forming off Baja, California. The deposits occur on a shallow platform marginal to a trough, restricted by submarine banks. He maintained that the phosphates are the end products of imperfectly known processes. Doyle et al. (1978) detected phosphatic concretions in present day mollusc kidneys and proposed a partly biogenic origin for their formation.

Ames (1959), D'Anglejan (1967), Summerhayes (1970), Parker (1971, 1975), Parker & Siesser (1972), Kennedy and Garrison (1975) and Bruckner (1977) favour the replacement of pre-existing calcite by apatite as a result of exposure to phosphate-rich solutions.

Although the phosphate-rich rocks described herein commonly contain as much as 22% P_2O_5 , which is well within the limit (18%) of P_2O_5 described by Bushinsky (1961) for true phosphorites, the term is not used here due to the thin non bedded nature of the nodules together with their infrequency.

The phosphatic nodular beds discussed are texturally and faunally similar to the surrounding beds, differing only in the mineralogy of their cements.

In the absence of further, more detailed geochemical data it is not possible to postulate a definite depositional environment for these beds; most probably a combination of the mentioned processes was responsible for the formation of the phosphatic nodules and replacements. It was seen that the most common occurrence of these phosphates is in the form of replacements of Thalassinoides, Kulindrichnus langi outer shells and not internal parts of shells. It is also probable that an original compositional difference (e.g. high phosphorus content of adsorbed ions on clay flakes) led to the enhancement of the nodules as a result of percolating diagenetic fluids. The diagenetic processes leading to the replacement of micrite by carbonate apatite and 'collophane' are not adequately understood and detailed geochemical work is needed in order to elucidate the problems. The nature of surface-to-centre (Parker and Seisser, 1972) gradations observed in certain phosphatized limestones, probably indicating the micrite to collophane replacement, should be investigated in greater detail by electron probe microanalysis. The evidence shown by non-phosphatized calcitic skeletal fragments within a phosphatic cement (commonly regarded as evidence for the replacement of porous lime, mud/micrite sediments by collophane, due to its susceptibility to penetration by phosphate rich solutions and subsequent chemical reaction) is also intriguing and needs to be further investigated.

In most of the nodules seen, the phosphate seems to have permeated the original sediment, i.e. the non-phosphatic part of the nodules consist of quartz and calcium carbonate sand and silt grains and some clay flakes which resemble the enclosing rock.

The faunal associations found in most of the phosphate rich beds of this study are mostly reworked and cannot be regarded as the representatives of the prevailing depositional conditions. Nevertheless their association with abundant detrital silt and fine sand fragments in

the Skye area may suggest the proximity of land areas and formation under shallow water conditions. In the Ardnamurchan and Mull areas terrigenous matter is less common in the phosphatic nodules, and the broken shell material show a characteristic 'digested' appearance (probably due to boring organisms) (Plate 9.14). The two criteria just stated may indicate relatively less turbulent depositional conditions in the southern area of study, where the supply of terrigenous material was less, allowing the freer precipitation of phosphatic material (if the deposition of phosphorus is believed to be a primary phenomenon) and the more vigorous action of burrowing organisms. The origin of the calcareous 'ovules' seen in the phosphatic pebbles and nodules found in Mull and Ardnamurchan is not clear.

Carbonate fluorapatite has been recognised as the compositional variety of apatite in marine phosphates since the works of Altschuler et al. (1952) and Silverman et al. (1952).

Authigenic marine phosphorites and phosphatic rocks are composed of carbonate-fluorapatite and the composition of Recent marine phosphorites has been shown to be basically fluorapatite (Gulbrandsen, 1969). The phosphates of the Phosphoria Formation (McKelvey, 1959; Gulbrandsen, 1969), contain up to 3.1% fluorine. The concentration of fluorine in Recent phosphorite rocks is generally quite high i.e. up to 3.3% (Parker, 1971; Parker and Seisser, 1972; Dietz et al. 1942; Burnett, 1977; Gulbrandsen, 1969). The mean concentration of fluorine in the ocean water was found to be 1.35 to 0.03 ml/l (Bewers, 1971) which does not increase with depth (Riley 1965; Kester, 1971). This concentration is greatly reduced in rain water. Most of the fluorine concentration of such waters originates from the sea as small droplets of foam are caught up with the wind (Correns, 1956) and may be carried into continental areas. Estimates of such concentrations vary

from zero to 0.089 mg/lit. (Sugwara, 1967), for additional reference see Allmann and Koriting (1972).

The main constituent of the phosphatic rocks in the Strath and Loch Aline areas is hydroxy apatite, therefore it would be reasonable to infer that the environment in which they formed was depleted in fluorine content and its concentration did not reach levels suitable for its incorporation within the apatite structure. Shishkina (1972) reported that fluorine contents of pore waters of oceanic 'productive' areas reached up to three times the level attained by pelagic ooze-waters, but he concluded that the fluorine was controlled by the solubility of fluoride rather than apatite.

The environment in which the Lower Liassic phosphates formed may have been far from normal oceanic upwelling' points and the phosphorus may have precipitated out of nearshore (? restricted from normal seawater), alternatively, post-depositional diagenesis in the presence of meteoric waters (depleted in terms of fluorine content) was responsible for their formation.

Most of the specimens studied show no signs of replacement phenomena such as pseudomorphism embayed contacts and automorphic crystals transecting earlier structures are either completely lacking or very subordinated in the samples (Pettijohn, 1957); although pseudomorphism of microfossils (Parker, 1971) may be considered as conclusive evidence for replacement, they were few and not of much quantitative significance.

Diffuse calcite/apatite crystal boundaries and partly replaced shell fragments and calcareous pellets are present and considered strong evidence supporting a diagenetic replacement.

CHAPTER 10

LIMESTONE/SHALE RHYTHMS IN THE LOWER LIASSIC ROCKS OF NORTHWEST SCOTLAND

Minor rhythms and cycles of limestones and shale occur in the Strath Formation of the Broadford Beds Arenaceous Group, they also constitute the strata of the Broadford Beds Argillaceous Group. The resemblance of these rocks to the Blue Lias of England and Wales suggests that similar processes were responsible for their formation. The origin of the regular cyclic alternations of fine-grained microsparitic limestones and shales characteristic of most of the Hettangian-Lower Sinemurian rocks of England and Wales has been discussed by many geologists (see Hallam, 1960a, for references). The main element of debate lies with the question of their origin; were they caused by original differences in sedimentation (primary) or by a form of diagenetic segregation of calcium carbonate within homogeneous shales and/or marls (secondary), or by a combination of the two.

In the following account, the various lines of evidence given by Hallam (1964, 1975) favouring each of the mentioned processes operating for the Liassic strata of England are explained in terms of their occurrence in the Scottish Lias.

i. Evidence suggesting a primary (sedimentary) origin.

a. Trace Fossils

As in Dorset, the rocks of the two Broadford Beds Groups are mottled due to the burrowing activities of organisms which produced Chondrites, vertically retrusive Rhizocorallium and Rhizocorallium. Kulindrichnus langi (Hallam, 1960b) which is so widely found in the Scottish Lower Lias and is also a common trace fossil in the Blue Lias, represents ?Cerianthid burrows which were accentuated and preserved in beds showing original contrasts with the underlying and overlying ones. In very few cases (notably Chondrites), sediments of the overlying bed infills the depressions

and burrow-tubes formed in the underlying beds.

b. Certain limestone bands (esp. in the Leacach Nodular Limestone and Leacach Shale Members) are very regular; however, this cannot be accepted as an unquestionable line of evidence since such regularities are generally regarded as secondary.

ii. Evidence suggesting a secondary (diagenetic) origin.

The Hebridean Lias consists (in parts of the areas under study) of continuous limestones with irregular hummocky surfaces which are not thought to be representative of the original configuration of the sea bottom (Hallam, 1957); the thickness of some adjacent limestones may vary antipathetically, and in some cases fossils were found in limestones which have been fractured and forced apart by dilation of the matrix during diagenetic recrystallisation. The above points are also cited as criteria suggesting that the limestone bands seen in the Lias of southern England are the result of some solution, migration and recrystallisation in the sediment, after deposition (Hallam, 1964).

Other criteria cited by Hallam (1971c, 1975) as evidence for the diagenetic origin of the limestone-marl cyclic sequences which may also be applied to the Hebridean Liassic Limestone/Shale alternations, are as follows:

a. No correlation can be established between the limestone shale rhythms and any other feature of the rock (sedimentary and/or faunal). Shell beds are distributed randomly with respect to the rhythms. This is the case with beds of the Broadford Beds Argillaceous Group and persists towards the northern area, being also seen in the Broadford Beds Arenaceous Group of Skye.

b. Although it is hard to trace regular limestone beds to see if they pass into beds of accepted secondary origin, it can be seen that the micro-

textures of the limestones give no indication of primary deposition but are consistent with interstitial crystallization during diagenesis.

c. The thickness of the total succession varies considerably from northern and southern Skye to Mull, but in different localities, the thickness of the limestone beds remains constant on the average and the number of such bands declines in approximate proportion to the thinning of the total sequence. Fig. 10.1 shows that this decline in the number of limestone beds is proportional to the total thickness of the succession; however, some successions (Planorbis Zone, Wilderness) do not conform to this generalisation; this is probably due to the dual origin of these rhythms.

It should be noted that due to the pronounced facies changes from the northern (Skye) to the southern (Mull) areas, the rhythmic limestone/shale beds seen throughout the Broadford Beds Argillaceous Group, are only represented by the Upper and Lower Teampull Chaon Shale Members of the Strath Formation. The Loch Aline Formation shows fundamental contrasts to the Milton Formation.

The thin (3 cm) laterally persistent nodular limestone beds of the Leacach Shale Member can only be compared with the beds of the Dun Bore-raig Sandstone Member if they are considered to be of primary origin.

The modal thickness and range of thickness for the Loch Aline and Leacach Formations (limestones) are given in Table 10.1 and are compared with those of the Lias of Portugal, Dorset and Glamorgan; they are very similar and suggest that the rate of sedimentation was irrelevant to the formation of the rhythms.

It can also be seen that the modal thickness of the limestone beds does not coincide with those recorded by Hallam (1964, 1971c) elsewhere.

d. From the above-mentioned criteria it is evident that no simple origin can be suggested for the limestone/shale alternations found in the Lower Lias of the Scottish Hebrides. While trace-fossil mottling is present in

the limestone beds of the Upper and Lower Teampull Chaon Shale Members, they are less abundant in the beds of the Wilderness Shale Member, Leacach Nodular Limestone Member and are totally absent from the Leacach Shale Member; some beds of the Wilderness Member show no sign of bioturbation.

The regularity observed in the occurrence of limestone beds of the Broadford Beds Argillaceous Group is less apparent in those of the Broadford Beds Arenaceous Group, although the beds forming the Strath Formation show a recurring rhythmicity.

A composite explanation is needed, taking into account both primary and diagenetic changes. Probably as Hallam (1964) has explained, more CaCO_3 was precipitated in shallower water sediments while in the deeper parts, the muds were secondarily enriched in CaCO_3 ; after deposition, the solution and reprecipitation of parts of the limestone had begun, CaCO_3 segregated into thin beds of limestone within the shales, "accentuating certain primary rhythms and creating others". Only this combination of sedimentary and diagenetic processes could produce groups of distinctive limestones in the Broadford Beds Groups.

	Number	Thickness		
		Mode	Mean	Range
Loch Aline and Wilderness	158	4.5	12.64	1 - 45cm
Glamorgan (Hallam, 1964)	233	7.5	12.50	2.5 - 30
Dorset (Hallam, 1964)	77	7.5	11.25	2.5 - 30
Somerset (Donovan, 1956)	75	7.5	9.5	2.5 - 30
Midlands (Dudley, 1942; Hallam, 1959)	61	10	16	3.75 - 50

Table 10.1 Data for limestone beds of the lower four Zones of the Lias(see also Fig. 10.2)

It should be noted that in the Skye area, porosity of the sediment was likely to have influenced the limestone thickness. Evidence for this is provided by the fact that the limestone beds are much sandier than the

shales. Although the distribution of the limestone beds and nodules is not controlled by the presence of shell material, individual limestones thicken around some shells.

The possible diagenetic origin of the limestones has been investigated petrographically.

10.1 Petrography

The possible diagenetic history of the limestone beds of the Strath and Loch Aline Formations can be discussed in terms of cementation, crystal enlargement, stylolization and silica diagenesis.

i. Cementation

In simple terms this includes the sum total of physical and chemical processes involved in transforming loose carbonates "which would tickle us between our toes" (Bathurst, 1971) in the Liassic seas, into an indurated limestone bed. The term "cement" includes all passively precipitated, space-filling carbonate crystals which grow attached to a free surface (Bathurst, 1971).

ii. Crystal enlargement

Crystal enlargement or "neomorphism" was defined by Folk (1965) as:

"all transformations between one mineral and itself or a polymorph ... whether the new crystals are larger or smaller or simply differ in shape from previous ones. It does not include older pore-space fillings; older crystals must have gradually been consumed and their space simultaneously occupied by new crystals of the same mineral or polymorph".

iii. Stylolites

These are pressure-solution surfaces which represent the interface between two rock masses. The contact surfaces are recognised by their

irregular, roughly columnar shapes which fit together.

Stylolites transect the whole rock rather than isolated grains; in biomicrites and biosparites, they are continuous surfaces cutting across both grains and micrites or grains and cement or across cements only (Bathurst, 1971); these occur in early and late diagenetic phases with a maximum development in the latter phases.

10.2 Diagenetic changes in the Broadford Beds Groups

To identify diagenetic changes in carbonate sediments, a knowledge of their original properties is needed to form a basis for comparison. Adhering to general uniformitarian principles, the broad pattern of Recent and sub-Recent shallow water carbonate depositional environments may be applied to the original Liassic sediments.

For the purpose of identifying diagenesis in present-day carbonate rocks, the following points should be noted:

- i. Aragonite is more common than calcite (Mg calcite)
- ii. Crystal sizes are commonly less than 10 μm .
- iii. Porosity is 60 - 90%.
- iv. The Sr^{+2} content of aragonite muds and muddy pellet-sands is very high.

The diagenesis of carbonate sediments is produced by pene- and post-depositional processes (Purdy, 1968). Pene(syn)-depositional changes occur below and near the sediment/water interface; these changes affect the sedimentary structures (e.g. effects of vagile infauna) present in a lime mud. The textural changes result from the recrystallization of various grain types, in this process the pre-existing crystal fabric of the affected grains are obliterated; these changes involve the replacement of aragonite and calcite by calcite. Post-depositional effects resulting

from the subaerial exposure of unconsolidated carbonate deposits result in carbonate solution, replacement and precipitation. Calcite and aragonite are preferentially dissolved by meteoric waters, gradually being replaced by low Mg calcite. These precipitations and/or replacements transform the unconsolidated carbonate deposit into a limestone (Purdy, 1968).

The following stages may be distinguished petrographically (Germann, 1968):

i. Early diagenesis - 1

a. Aggregation

This stage refers to the organic chemical and physical changes which are preserved in the sediment; these result in the formation of ooliths and pellets by aggregation and their break-up by biological and physical agents.

b. Cementation by ferroan calcite

On the basis of some fabric criteria for the recognition of cements (Bathurst 1971), the two generations of cement recognised by Bathurst have been identified.

c. Fabric Criteria

1. The spar is usually interstitial (interparticle) with well sorted and abraded particles which are in depositional contact with each other. Micrite from which the spars might have evolved by "aggrading neomorphism" is therefore unlikely to have been present in the original sediment.
2. There are no relict structures such as are seen in neomorphic spar.
3. Particles composed of micrite (e.g. peloids) are not altered to spar.
4. Micrite coats on particles are not altered to neomorphic spar.
5. Mechanically deposited micrite is present but unaltered.

6. Contacts between the spar and particles is sharp.
7. The margin of sparry mosaic coincides with surfaces that were once free (e.g. skeletal particles and ooids).
8. Sparry mosaic is seen to occupy the upper parts of a cavity in geopetal features.
9. Some masses of sparry mosaic have the form expected of a pore filling (see Plates 6.17 and 10.1).
10. The intercrystalline boundaries in the mosaics are made up of plane interfaces.
11. The size of the crystals increases away from the initial substrate of the sparry mosaic.
12. Some mosaics are characterised by enfacial junctions among "triple" junctions.

The above 12 criteria among Bathurst's (1971) 16, were observed in the Liassic limestone beds of northwest Scotland.

d. Cement type

Two different types of cement were also distinguished.

1. Granular cement -

A fine-grained cement which is formed in pores of a detrital sediment (Bathurst 1975). "Drusy" mosaic which has a similar fabric, infilling cavities (other than pore spaces) of a detrital sediment is also seen.

2. Rim cement -

Is a syntaxial extension of a crystal which grows (in the limestones of

the Broadford Beds Argillaceous and Arenaceous Groups) as calcite on crinoid fragments and solution on quartz grains.

Staining (Dickson, 1965, 1966; Evamy, 1969) shows that in the limestone beds, granular cement occurs as both ferroan and non ferroan in composition. In all cases studied, the dividing line between the ferroan and non ferroan calcite cement was found to coincide with the crystal faces. Although zoned crystals are present, these are the result of crystal enlargement (recrystallization). The formation of ferroan calcite cement is by the incorporation of ferroan iron into the calcium carbonate during the growth of the cement from solution (Dickson 1966).

e. Cement source

Some of the general clues to the source of cement for the Lower Limestone Shales Group given by Whitcombe (1970) are broadly applicable to the Broadford Beds Argillaceous and Arenaceous Groups. Evidence for the presence of sources other than the CaCO_3 in solution is also invoked for the Lower Liassic Limestones of northwest Scotland as follows:

1. An excess of CaCO_3 (8.48% by volume) is produced by the inversion of aragonite to calcite. This will eventually act as interparticle cement (Oldershaw and Scoffin 1967).
2. Calcium Carbonate is released into the solution by pressure solution along grain contacts, which is common during early diagenesis.

In the limestone bands of the Loch Aline and Strath Formations, glauconite pellets are widespread and iron oxidization products also occur. This would indicate that the environment of deposition contained high proportions of iron in solution. The initial stage of ferroan calcite cement is thought to have formed during or immediately after the deposition of the grains; this was followed by a meteoric stage during which CaCO_3 from pressure solution and aragonite inversion was released to form the cement.

As no evidence for the presence of ferroan calcite has been found in the pellets, ooliths or skeletal fragments, it is proposed that the inversion of aragonite and pressure solution gave rise to iron-free calcite which formed the stage of nonferruginous cement. If this was the case, some of the first stages of ferroan calcite cement would also be released by pressure solution, but this iron would be in too small a quantity to be detected by the staining methods used (staining detects >1% ferroan carbonates in solid solution). This stage of cementation was followed by the final ferroan-calcite stage formed in a similar environment to the initial stage and at the same time as the formation of glauconite and iron-oxide cements in other areas.

f. Transformation -

The major process involved at this stage is the transformation of aragonite into calcite and cementation by non-ferroan calcite. Aragonite is absent from the limestones of the Strath, Leacach and Loch Aline Formations and low Mg calcite is the dominant mineral constituent. The various mechanisms involved in mineral transformation (i.e. aragonite $\rightarrow$ calcite) are well presented and discussed by Bathurst (1971). Detailed petrographic and geochemical examination of Hettangian-Sinemurian limestones in north-west Scotland shows that during early diagenesis, the most common form of transformation diagenesis was one involving a 'wet' system.

Microcavities produced in the limestone as a result of leaching, were filled by a fine-grained mosaic of calcite, the boundaries between the shelly material and surrounding matrix are well defined and generally marked by a micritic envelope. Although fine textures are preserved in the carbonate beds and their constituents, the Sr^{+2} content of these beds never exceeds values of 1000 ppm and it is well known that the Sr^{+2} concentration of recent natural aragonites is 8000-10,000 ppm. The significance of Sr^{+2} distribution in the Lower Liassic limestones and shales of north-west Scotland is further discussed in Chapter 11.

It would be instructive to study the Sr^{+2} distribution in diagenetically altered limestones with the aid of electron probe microanalyzers, to determine their spatial arrangement and concentrations (work in progress by the author).

The preservation of fine textures in ancient limestones is probably achieved by a dissolution-precipitation process that does not involve an intermediate stage of appreciable void development. Migration of films (Bathurst, 1971; Tan and Hudson, 1974) of solution may carry out such replacement at grain boundaries and may produce changes in isotope ratios and trace element concentrations without leaving discernible textural evidence of replacement (Kinsman, 1969).

ii. Early diagenesis - 2

In this stage the completion of cementation by ferroan calcite occurs; this results in a complete lithification of the rock under conditions of low burial.

iii. Late diagenesis - 1

The most important process occurring under conditions of deep burial is neomorphic crystal growth (enlargement of calcite crystals). The enlargement of crystals with no obvious nuclei is common and large scale crystal mosaics of calcite ($>10 \mu\text{m}$) have formed. The crystal boundaries are seen to be sutured with twin lamellae and cleavage surfaces bent (Plate 10.2); these properties are typical of strained calcite (Demirmen and Harbaugh, 1965; Bathurst, 1971). This crystal enlargement effectively destroys some or all of the original sedimentary textures. The extent of crystal enlargement in the western Scottish Lower Liassic Limestones is widespread and can be seen in the majority of the thin sections.

The possibility that sparry calcite is a cement cavity filling or a rim cement deposit of detrital crystals is ruled out for the British Jurassic by Bathurst (1971). Evidence for this is an irregular distribution of

crystal sizes and scarcity of plane intercrystalline boundaries together with a low percentage of enfacial junctions among the triple junctions. The evidence for in situ replacement is amply reviewed by Bathurst (1971, 1975).

iv. Late diagenesis - 2

This process involves crystal enlargement of quartz (associated with the sediments) and the formation of stylolites, joints and their infilling with calcite to form veins. This may have taken place during conditions of deepest burial associated with tectonic disturbances.

a. Quartz grains in the limestones occur as either the nuclei of ooliths or as free detrital grains. Either of these can be affected by silica diagenesis, the result is the enlargement of grains.

b. Veins are very common, and are generally vertical or subvertical, although some are almost horizontal. They vary from micro-sized to 2-3 cm. in thickness and are composed of ferroan and non ferroan calcite. In places the composition appears not to be a reflection of the ferroan and non ferroan calcite content of the rock in which they occur.

c. Stylolites are common and occur in limestones in most cases; both horizontal and vertical varieties are seen and in amplitude they range from the simple to the seismograph-type (Park and Schot, 1968); no rectangular types are seen.

The magnitude of the stylolites ranges up to 1 mm. and they are thought to have occurred during late stages of diagenesis (due to their size and persistence), when pressure from the overlying sediments and from tectonic movement acted on the lithified rock. This process of pressure removal resulted in the redistribution of material and led to the formation of mineral components such as euhedral quartz (common near the stylolite seams).

10.3 Petrographic conclusions -

1. In the limestones of the Strath and Loch Aline Formations, the stable diagenetic carbonate phase is low Mg calcite; no aragonite is present.
- ii. The transformation of aragonite to calcite probably occurred early in the process and involved a "wet, dissolution-reprecipitation" process with the "migration of solution films" (see Chapter 11).
- iii. Consolidation and compaction was completed early and resulted in the loss of pore space in the rock; cement of ferroan and non ferroan composition is present.
- iv. During late stages of deep burial crystal enlargement occurred, which has in places completely destroyed the primary sedimentary textures and produced a coarse crystalline rock.
- v. Tectonic stress and the weight of the overlying sediments during deepest burial formed the majority of the stylolites, euhedral quartz and veins.

CHAPTER 11

STRONTIUM DISTRIBUTION IN THE

LOWER LIASSIC ROCKS OF NORTHWEST SCOTLAND

11.1 Introduction

In the search for suitable geochemical elements as diagenetic and sedimentary facies indicators Sr^{+2} has been among the most investigated elements by research workers. A complete review of the vast amount of data presented in the literature on the behaviour of Sr^{+2} in various situations is beyond the scope of this work.

The general subject of Sr^{+2} distribution in carbonate rocks was amply reviewed by Graf (1960), Wolf et al. (1967), Flügel and Wedepohl (1967), Kinsman (1969), Bathurst (1971, 1975), Milliman (1974) and Füchtbauer (1975); valuable information on the distribution of Sr^{+2} in European sedimentary carbonate rocks is also found in the book on European carbonate sedimentology by Muller and Friedman (1968).

Flügel and Wedepohl (1967), Wedepohl (1969), Kinsman (1969), Kinsman and Holland (1969) and Katz et al. (1972) have discussed the theoretical aspects of Sr^{+2} behaviour and its partition during the precipitation of carbonates. The basic features of Sr^{+2} distribution in Quaternary carbonate rocks is well known (cf. Kinsman, 1969; Dodd, 1967) but the uncertainty in the knowledge of Sr^{+2} behaviour during diagenesis is considerable.

Various works by Lowenstam (1961), Turekian (1964) indicate that during the Phanerozoic the Sr/Ca ratio of seawater was relatively constant.

Rubey (1951, 1955) demonstrated that the atmosphere and hydrosphere are not primitive but must have developed through geological time (see also Cloud, 1968). Ronov (1959) also investigated the post Cambrian geochemical history of the atmosphere and hydrosphere and later (Ronov, 1968) (see also Vinogradov and Ronov, 1956) postulated probable

changes in sea water composition during the course of geological time. Arguments based on residence times of dissolved constituents (Garrels and MacKenzie, 1971), the sedimentary and palaeontological records, palaeosalinity indicators (Reynolds, 1965; see also Perry, 1972), carbon isotope ratios (Becker and Clayton, 1970) and other chemical considerations (Garrels and MacKenzie, 1974) suggest that little change has taken place in the chemistry of sea water over extended periods of geological time. The isotopic composition of sea water throughout Phanerozoic time was studied by Peterman et al. (1967); Veizer and Compston (1974) studied the variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ composition of sea water during the Phanerozoic and found low $^{87}\text{Sr}/^{86}\text{Sr}$ values in the Cretaceous and Late Jurassic but could not support a further structure beyond a general trend through the Phanerozoic, which they tentatively correlated with the rate of continental denudation; this was further supported by Spooner (1976) who suggested that variations in land area may have been partly a consequence of variations in global mean sea floor spreading rate. Veizer and Compston (1976) have also attempted to use Sr^{+2} isotope ratios in Precambrian carbonates as an indicator of crustal evolution.

The works of Odum (1957), Lowenstam (1961, 1963) and Turekian (1964) indicate that the Sr/Ca ratio of sea water was relatively constant during the Phanerozoic. Veizer (1977b) indicated that there is no age trend in the Sr/Ca ratio of pre-Tertiary limestones (see also Kulp et al., 1952). Lowenstam (1961) based his arguments on the relatively similar Sr/Ca ratio of fossil (Mississippian) and Recent marine brachiopods. This also implies that the biochemical factors influencing Sr^{+2} partition were not vastly different from the ones operating at present.

This, together with the variations in Sr^{+2} concentration throughout diagenetic processes affecting Quaternary limestones (see Bathurst, 1975), leaves no doubt that the observed differences are the results of post sedimentary partition. Veizer and Demovic (1974) pointed out that the

diagenetic histories of particular rock types are not clear and not easy to decipher; our assessment of the role of different diagenetic solutions (fresh and marine water) on the lithification of carbonate sediments is in a transitional stage, being constantly revised and added to. The mentioned uncertainties are also reflected in the evaluation of the use of Sr^{+2} as a possible facies indicator and viewpoints differ markedly (Kinsman, 1969; Schroeder, 1969). The adsorption and ion-exchange properties of clay minerals makes them important receptors, carriers and donors of different cations especially Sr^{+2} ; this also adds to the ambiguity of the use of Sr^{+2} as a palaeoenvironmental indicator (see also Bausch, 1968; Kinsman, 1969).

The behaviour and migration of strontium, during a complicated sequence of diagenetic changes, was studied by Shearman and Shirmohammadi (1969), and Al-Hashimi (1976) also discusses the significance of Sr distribution in some carbonate rocks. Recent studies by Gieskes et al. (1975) and Sayles and Manheim (1975) shows that the Sr^{+2} concentration in Recent deep sea sediments is related to the CaCO_3 recrystallisation processes. The Sr^{+2} content of altered and unaltered Recent and fossil calcareous organisms and their relationship with the enclosing limestones and/or other medium has been studied by many workers (Kulp et al., 1952; Turekian and Armstrong, 1961; Hallam and Price, 1968; Veizer and Wendt, 1976). Such studies are of potential value in the identification of sedimentary environments and the response of organisms to their different chemistries through time. The subject is somewhat complex and as unaltered, calcareous parts of organisms were not analysed from the Lower Liassic beds of northwest Scotland, no further consideration will be given to it.

11.2 Strontium in carbonate rocks and minerals

The minerals which form calcareous rocks are few; the great variation in appearance and properties of different limestones arises from the almost

endless variety of organisms and other structures by which the crystals of these minerals may be aggregated.

The carbonate minerals form over a wide range of environmental conditions and their composition is controlled largely by their mode of genesis.

Sedimentary carbonate rocks which are composed of minerals that have the CO_3 group in common are termed "limestone". Approximately 60 of the above mentioned minerals occur in nature but hexagonal calcite and its orthorhombic polymorph aragonite are the most common in the sediments of modern environments. In ancient rocks calcite and dolomite are by far the most common. Calcite is the stable form of calcium carbonate at ordinary pressure-temperature conditions and may be regarded as the principal mineral of limestones. Aragonite is the form which calcium carbonate normally adopts when inorganically precipitated from sea water. Conditions favouring its precipitation in preference to calcite are warm water, high alkalinity supersaturation and abundance of SO_4^{--} in solution. These conditions are usually found in warm current areas of seas and oceans. Waters rich in SO_4^{--} and trapped in sediments may help to preserve aragonite mud longer than normal. Aragonite is a metastable mineral and under normal T-P conditions may be converted into the more stable calcite.

The older limestones normally contain no aragonite and any shells which originally consisted of this mineral are found to be represented by open moulds or by coarsely crystalline calcite.

Calcite and aragonite are never compositionally ideal, other divalent cations commonly substitute for Ca^{+2} in varying amounts in the crystal structure, which produces a whole series of materials with different solubilities and properties. Those elements with smaller ionic radius than Ca^{+2} ($0.99\text{\AA}$), such as Mg^{+2} ($0.66\text{\AA}$), Fe^{+2} ($0.74\text{\AA}$) and Mn^{+2} ($0.8\text{\AA}$) are more readily accepted into the hexagonal calcite structure, while those ions whose radii are greater than Ca^{+2} such as Ba^{+2} ($1.32\text{\AA}$) and Sr^{+2} ($1.12\text{\AA}$) are more readily accepted by the aragonite structure. In order to understand the

minor element composition of natural calcite or aragonite it must be ascertained whether the mineral was the product of biochemical processes within the tissues of organisms and whether it was determined by the metabolic processes of the organism or if the minor element composition was controlled by physico chemical processes taking place in the solution without direct biochemical influence. Any crystal growing in contact with the surrounding water rather than within organic tissues should reflect the conditions existing in the water and not within the organism.

As previously mentioned, the literature on organically induced and influenced distribution of minor elements in carbonate skeletons is voluminous; therefore for the purpose of this study, only the inorganic factors and processes related to elemental composition of carbonates will be emphasised. The minor and trace elements present in calcitic and aragonitic rocks as given by Deer et al. (1966) are as follows:

Calcite - Mg, Mn, Fe, Sr, Ba, Co, Zn.

Aragonite - Sr, Pb, Ba, Mg(?), Mn(?).

11.3 Non carbonate components of carbonate rocks

Non carbonate components in calcareous sediments are inorganic or organic in composition. The sum total of organic matter from diverse sources have a considerable influence on the cation composition of the sediments.

The non-calcareous material may be of primary or diagenetic origin. When attempting to separate the non-carbonate from carbonate fraction, it is significant to consider that not all non-carbonate fractions are insoluble (Wolf et al., 1967). Graf (1960) gave a list of minerals that have been reported from carbonate sediments; those which are found in various concentrations in the studied rocks include: clay minerals, phosphates, feldspars, goethite, micas, quartz, glauconite, chlorite and pyrite. In general one of the most significant and widespread contaminants of sedimentary

carbonates is the clay fraction.

The effect of cation adsorption, membrane filtration (Kharka and Smalley, 1976) and diagenetic changes (Perry and Turekian, 1974) in carbonate rocks with considerable argillaceous components is considered by some (Bausch, 1968; Kinsman, 1969) to be an important control over their Sr^{+2} content.

Aragonite is stable as long as it is in contact with sea water (Friedman, 1964); carbonates are readily cemented and the pore space thus diminishes; therefore aragonite will experience transformation in lithified and covered sediments. When part of the aragonite is present in the partly cemented carbonates, while pore volume is small, the Sr^{+2} concentration will increase in the pore spaces. It is possible that the clay mineral particles present in the system can adsorb more Sr^{+2} from this increased availability than they could have adsorbed directly from sea water. Wahlberg et al. (1965) found that the distribution coefficient of Sr^{+2} between solution and the clay minerals (Na^+ or Ca^{+2} illites) is constant up to concentrations of 10^{-3} , Sr^{+2} in solution.

Because sea water is a 10^{-4} molar Sr^{+2} solution, clay minerals are able to further increase their Sr^{+2} content; if new calcite crystallises in the pore solution, it will also pick up more Sr^{+2} than when in equilibrium with sea water, due to the increase of Sr^{+2} in the pore solution (see Bausch, 1968). Although Knoblauch (1963) found no relationship between Sr^{+2} content and the proportion of clay in a sequence of Upper Jurassic limestones from the Swabian Jura, Bausch (1968) maintained after various considerations that an increase in the Sr content in relation to the clay mineral content is a generally recurring phenomenon.

The differences found in the Sr^{+2} content of fossil shallow water and deep marine limestones, has been explained by the differences in their clay mineral content (Bausch, 1968). The low (100-300 ppm) Sr^{+2} , shallow water (reef) "pure" limestones contain negligible amounts of clay whereas a 5-15%

clay content in limestones causes a corresponding increase of the Sr^{+2} content (400-700 ppm). Carbonates with more than 20% clay content contain 1000 ppm strontium. The above mentioned differences are greatly affected by diagenetic processes; diagenetically old limestones show no relationship between the Sr^{+2} and clay content.

11.4 Strontium content of carbonates through geological time

The study of Vinogradov and Ronov (1956) shows an increase in Sr^{+2} content with a decrease in the geological age of the carbonate rocks of the Russian Platform. The strontium content of all limestones older than Cretaceous varies within known average values of 400-700 ppm (Bausch, 1968) whereas present day carbonates show concentrations of 7000 to 9000 ppm (Kinsman, 1969). This phenomenon was related to the aragonite-calcite transformation in older rocks (Kahl, 1965).

11.5 The use of strontium data for the study of diagenesis in limestones

Modern shallow marine, warm water carbonate sediments are composed predominantly of the unstable carbonate minerals aragonite and high Mg calcite. They contain only minor amounts of low-Mg calcite. Carbonate rocks are composed essentially of two stable minerals, calcite and dolomite. If one can assume that Pleistocene and older limestones were chiefly metastable carbonate minerals originally, virtually all limestones have undergone recrystallisation, which may be a solid state inversion process or may be achieved through the solution of metastable carbonates (Kinsman, 1969; Bathurst, 1975).

The works of Linck (1909), Johnston and Williamson (1916) and many others shows that a temperature of about 400°C is needed for aragonite to invert rapidly to calcite through a dry, solid-state reaction. This dry process is called "polymorphic inversion" (Spry, 1969). This process is

regarded as valid by some researchers insofar as it accounts for the textural perfection of fine carbonate rocks which now consist of calcite but which were most probably aragonitic in their earlier history (Taft, 1967). The inversion temperature needed for such a solid state recrystallisation depends on the history of the rock (e.g. storage of energy in a deformed lattice) and on its crystal size; below 400°C the inversion still goes on but at rates that can be slow even judged against the geological time scale. Fyfe and Bischoff (1965) summarised the relevant data, which shows that below 100°C the time required for the completion of the dry reaction is in the order of 10^7 years; the preservation of ancient aragonite (Hallam and O'Hara, 1962; Hallam and Price, 1968) indicated that times of at least 3×10^8 years are required for inversion in the normal range of diagenetic temperatures and pressures.

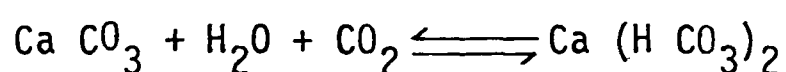
On a more refined scale Kinsman (1969) suggested that if such a process is responsible for the transformation of aragonite into calcite (i.e. absence of a fluid phase) then the $m\text{Sr}^{+2}/m\text{Ca}^{+2}$ ratios of the crystals are expected to be somewhat comparable, this is clearly not the case. It has been explained that the aragonite-calcite recrystallisation involves the reduction of lattice positions which could be taken up by Sr^{+2} . A dry state inversion cannot provide the system with an adequately efficient "flushing" mechanism whereby the excess of Sr^{+2} produced as a result of such diagenetic changes could be withdrawn. Such a situation would probably lead to a haphazard distribution of Sr^{+2} in the crystals. This phenomenon can readily be investigated by electron microprobe studies on examples of both dry and wet recrystallisation processes.

Although the delicacy of the textures preserved in some ancient marine oolites has been used as evidence for solid state inversion, no calcitic oolite older than the Pleistocene age shows Sr^{+2} concentrations higher than 1000 ppm. Kinsman (1969) suggested that the preservation of such fine textures is possibly achieved by a dissolution, reprecipitation process that does not involve an intermediate stage of void development. He maintained

that migrating films of solution may carry out such replacements at grain boundaries and may produce changes in isotope ratios and trace element concentrations without leaving textural evidence of replacement.

Recent summaries of the process of diagenesis (Bathurst, 1971, 1975) have shown that the transformation of metastable carbonate assemblages into a lithified calcitic (low-Mg calcite) limestone is a solution-precipitation process during which aragonite is replaced by calcite in situ as selected crystals grew at the expense of their neighbours; Ca^{++} and CO_3^{--} or (HCO_3^{--}) diffuse from one lattice to the other through the action of "solution films" with a consequent conservation of the original lamellar structure of any shell walls present.

The process of solution-precipitation is summarised as follows:



Each solution-precipitation step involves the repartition of oxygen and carbon isotopes in CO_3^{-2} as well as the partition of elements substituting for Ca^{+2} , such as Mg^{+2} , Sr^{+2} , Fe^{+2} and Mn^{+2} . Bausch (1968) showed that this process may be inhibited by a 2% clay mineral content of limestones. The importance of detailed scanning electron microscopy, electron microprobe and stable isotope studies of selected samples, in order to substantiate or disprove the solution-precipitation and other diagenetic processes is obvious. The study of the degree of textural perfection in altered limestones with the aid of a scanning electron microscope may reveal otherwise undetectable imperfections especially in the lamellar structure of altered shell fragments.

The study of the distributions of minor elements (esp. Sr^{+2}) in the vicinity of such imperfections by the electron microprobe may disclose the nature and control of trace element distribution (hence the diagenetic fluids), e.g. it would be possible to establish whether "solution films" aid diffusion from one lattice to the other and if so, whether they act as

"channels" through which the excess of trace elements created due to lattice changes and element substitution were transported, to be ultimately withdrawn from the system (depending on the open or closed nature of the particular diagenetic process). This wet diagenetic process in ancient carbonate rocks would be manifested by the elemental zoning of crystals and the ultimate elemental enrichment of crystal boundaries; depending on the nature of the "solution film" action, element-enriched "relict" areas or fronts may also be distinguished. In particular, the importance of cation adsorption properties of phyllo-silicates and other "insoluble residues" during the process of diagenesis may be investigated in full. The diagenetic stabilisation achieved through the solution-precipitation process (aragonite low-Mg calcite) is believed to be accomplished mostly in waters of meteoric derivation, which are characterised by lower Sr/Ca ratios and lighter $\delta^{18}\text{O}$ than that of sea water (e.g. Kinsman, 1969; Gross, 1964). Increasing post-depositional equilibration with such waters will lead to a progressively lower Sr content and lighter $\delta^{18}\text{O}$ of calcite, a positive correlation of Sr with $\delta^{18}\text{O}$ would thus confirm the role of meteoric waters in the diagenetically altered parts of the rocks under study, whereas other explanations may have to be sought after if the above is not the case for samples with $> \times 10^7$ years old (see Hudson, 1977; Veizer, 1977b).

A project proposed (1975) to investigate the effects of diagenesis on the redistribution of Ca^{+2} , Mg^{+2} , Mn^{+2} , Fe^{+2} and Sr^{+2} in the northwestern Scottish Lower Liassic limestones and ironstones with the aid of a CAMB. 90 electron microprobe was abandoned in April 1977 due to various instrumental and other inconsistencies and the large amount of time involved.

11.6 Use of Sr^{+2} as palaeoenvironmental indicator

It is evident from the foregoing that the Sr^{+2} concentration of limestones is greatly affected through different stages of diagenesis. Knoblauch (1963) examined a sequence of Upper Jurassic limestones and marls

in the Swabian Jura but found no relationship between the Sr^{+2} content of the beds and their clay mineral content and depositional environment.

Bausch (1968) found differences in the Sr^{+2} content of ancient reef and basin limestones which he related to their clay mineral content rather than primary enrichment controls. Evaporitic formations are known to contain distinctly high Sr^{+2} contents (Vinogradov and Ronov, 1956; Smykatz-Kloss, 1966) as compared with those originating from normal marine conditions.

A series of studies on the carbonate rocks of High Tatra Mantle Series Region and the Slovak Karst Plateau by Veizer et al. (1971), Veizer and Demovic (1973) showed a bimodal distribution of Sr^{+2} in the limestones, which was related to the presence of different facies types. It was found that the high Sr^{+2} groups represented the "hypersaline", dark coloured (oxygen deficient) and deep sea facies, whereas the reef detrital, neritic and shallow pelagic limestones contained lower Sr^{+2} concentrations.

Veizer and Demovic (1974) found that the distribution of Sr^{+2} in Mesozoic carbonate rocks of the Central Western Carpathians is facies controlled; the distribution pattern was found to be similar to that of the High Tatra Mantle Series Region and the Slovak Karst Plateau. The causes of the observed bimodality were not known and it was assumed to have been inherited from original sediments and enhanced during diagenetic alteration. The high Sr^{+2} group probably reflects a predominantly aragonitic and low-Mg calcitic original mineralogy whereas the low Sr^{+2} group is indicative of a high-Mg calcite precursor.

Subsequent work on the Lithographic Limestones and Dark Marls from the Jurassic of southern Germany (Veizer, 1977a) has shown a similar bimodal distribution of Sr in the limestones. In order to assess the relative importance of primary differences in the Sr concentration as opposed to later diagenetic effects, $\delta^{18}\text{O}$ data were also obtained for the limestones. Lighter $\delta^{18}\text{O}$ values would have indicated stronger diagenetic effects by meteoric

waters, but it was shown (Veizer, 1977a) that the $\delta^{18}\text{O}$ results were in conflict with the different trace element concentrations and the bimodal differences in the Sr^{+2} concentrations most probably reflected original depositional differences. Further study on the observed bimodal distribution of Sr^{+2} in ancient calcareous rocks combined with $\delta^{18}\text{O}$ work (Veizer, 1977b) showed that the chemical and isotopic compositions of ancient rocks are facies controlled and such data may be used as supplementary indicators in facies studies. It is stressed that the causes for such differences are still not well known and are open to interpretation; detailed explanations for the different processes involved are given by Veizer (1977a, b), Veizer and Hoefs (1976).

11.7 Methods and results

Seventeen samples were selected from three different Formations of the Lower Liassic rocks in northwest Scotland; the Sr^{+2} , Mg^{+2} , Mn^{+2} , Fe^{+2} and Ca^{+2} content of their calcareous (soluble in 10% Hcl) fraction was determined by atomic absorption spectrophotometry (see Appendix 5 and Tables 11.1, 11.2 & 11.3)

Table 11.1

Sinemurian	Turneri				Strath Formation
	Semicostatum	b40 b39 b38, Am60 b37, Am59	13	Lower Teampull Chaon	
Hettangian	Bucklandi	b58, b36 Am57, b35 Am55 Am54	9b 8	Upper Sand Breakish Coral	Milton Formation
	Angulata				
	Johns.+ Liass.				
	Planorbis	M92 91 89 87 85	19	Wilderness Shale	Loch Aline Formation
Stage	Zone	Sample	Facies	Member	Formation

Table 11.2

Sample	Description and classification	Intraclasts	Matrix/cement	Comments
Am54	Very poorly sorted silty biosparite with micritic pellets.	echinoderms, brachiopods, bivalves, pellets.	sparite	Heterogenous texture cement
Am55	Silty biopelsparite, with coalified material.	ooliths, echinoderms, bivalves, pellets.	microspar	Very heterogenous texture
Am57	Calcareous siltstone.	pellets, echinoderms.	microspar, organic material	-
Am58	Silty biopelsparite.	echinoderms, bivalves, pellets.	microspar	Bivalve shells with micritic lining; large echinoderms.
Am59	Fine calcareous sandstone.	-	microspar	-
Am60	Biomicrosparite.	bivalves, brachiopods, echinoderms.	microspar	Framboidal pyrite and organic matter abundant.
b 35	Calcareous siltstone.	echinoderms.	microspar	No ferroan calcite.
b 36	Microsparite.	-	microspar	Pyrite.
b 37	Sandy oopelbiomicrosparite.	bivalve "ghosts", echinoderms, ooliths.	spar	Some veins with ferroan calcite.
b 38	Sandy oopelbiomicrosparite.	ooliths, bivalve "ghosts"; echinoderms.	microspar	Mostly altered.
b 39	Silty oopelmicrosparite.	ooliths, pellets.	microspar	No organic material.
b 40	Silty microsparite.	-	-	-
M 92	Calcareous siltstone.	?aragonitic bivalves.	-	-
M 91	-	-	-	-
M 89	Silty microsparite.	echinoderms, pellets.	-	Echinoderms abundant, some pyrite.
M 87	Silty microsparite.	echinoderms, pellets.	microspar	Framboidal pyrite.
M 85	Silty microsparite.	pellets, some echinoderms.	microspar	Framboidal pyrite organic matter.

Sample	CaCO ₃ 1	CaCO ₃ 2	CaCO ₃ $\bar{3}$	CaCO ₃ %	Sr 1	Sr 2	Sr $\bar{3}$	Ca 1	Ca $\bar{2}$	Ca $\times 10^{-5}$	Mg 1	Mg $\bar{2}$	Mg	Mn	Fe $\times 10^{-3}$
b35	24.45	30.81	30.83	218.70	978.65	1021.24	999.95	1692.88	919.12	1306	548.59	1041.67	795.13	1797.39	3676.47
b36	46.46	54.26	61.21	53.98	697.52	753.34	725.43	687.46	10022.38	854.92	226.79	591.91	409.35	1345.24	1721.91
b37	51.32	66.00	57.95	58.42	700.10	643.53	671.82	326.65	799.53	563.09	452.83	1569.81	8061.32	2973.87	3705.15
b38	63.13	64.50	69.57	65.73	570.17	506.73	538.45	632.02	783.85	707.97	484.20	1013.46	748.83	2889.94	2992
b39	55.99	37.05	33.60	42.21	881.14	481.92	681.53	1009.27	535.52	772.40	803.30	1088.90	946.10	1561.94	3570.15
b40	27.88	23.82	37.13	29.61	369.89	627.92	498.89	1964.43	717.62	1341.03	1075.27	1363.47	1219.57	2179.76	6817.37
Am54	22.47	27.18	21.3	23.65	855.50	978.21	916.86	1122.67	666.96	894.82	273.82	578.03	425.93	1667.41	3557.14
Am55	66.56	72.96	79.1	72.87	412.10	495.20	453.65	561.73	884.69	723.21	430.84	1077.91	729.38	607.74	1050.42
Am57	19.75	19.26	30.4	23.14	821.40	886.08	853.71	2232.50	759.49	1496	522.93	1215.19	869.06	1708.86	5569.62
Am58	54.50	63.49	54.60	57.53	512.60	559.02	535.81	339.82	449.05	394.44	288.85	1119.65	704.25	1191.35	2414.93
Am59	33.42	44.73	33.30	37.15	602.20	1196.17	899.19	803.57	1674.64	1239.11	814.73	1839.11	1326.92	2197.97	5681.82
Am60	66.34	79.99	65.30	68.88	558	481.71	519.86	1403.02	904.98	1154	345.85	1038.69	692.27	3236.49	2860.15
M85	48.62			48.62		504.32	504.32		936.60	936.60		926.31	926.31	349.94	1543.85
M87	54.94			54.94		372.39	372.39		899.18	899.18		862.85	862.85	236.15	2270.66
M89	53.86			53.86		190.54	190.54		841.35	841.35		748.89	748.89	314.35	2311.39
M91	2.96			2.96		1797.39	1797.39		2516.78	516.78		2517.99	2517.99	7550.34	41946.31
M92	46.53			46.53		472.20	472.20		890.75	890.75		1094.66	1094.66	826.36	2897.62

Table 11.3 Elemental composition (in ppm) of the acid-soluble fraction of selected Lower Liassic rocks of northwest Scotland.

11.8 Discussion

It is seen (Fig. 11.1) that if the element values obtained from each sample are plotted against the insoluble fraction of the rock an apparently random scattering results. Nevertheless a somewhat harmonic variation in elemental composition is seen in the different samples.

The cation content of the samples increases with the decrease in CaCO_3 ; the variation of Ca^{+2} , Mg^{+2} , Mn^{+2} and Fe^{+2} between the different samples is somewhat concordant whereas the Sr^{+2} variation does not follow the general pattern shown by the other elements.

Although detailed analysis was not carried out to determine the exact phyllosilicate content of the insoluble residue fraction of each sample, it is seen that the Sr^{+2} content of the samples shows a steady decline as the soluble fraction of the rocks increases. This may indicate that the insoluble residue (in 10% HCl) fraction of the calcareous rocks, affects the Sr^{+2} concentration to some extent. Obviously the quartz, feldspar and other detrital insoluble matter present may significantly affect the results by shifting the results towards the lower CaCO_3 side without affecting the Sr^{+2} content; thus it should be instructive to plot Sr^{+2} values against the clay mineral and organic matter content of each sample.

In Fig. 11.2 the Sr^{+2} concentration of the rocks are given for the different ages. Each age group represents a different facies (Table 11.1, 11.2) and although some degree of overlap exists, it can be seen that the Sr^{+2} distribution is not random.

The range (Fig. 11.2) covers three distinctly different Liassic facies from two regions of northwest Scotland, nevertheless the results cannot be regarded as representative and comprehensive, since only seventeen individual samples were considered.

The distribution of Sr^{+2} in the limestones (Fig. 11.2) is variable but generally somewhat uniform within the groups (no significant variation can be

seen between the groups). A comparison with Table 11.2 shows that these variations are not dependent on any one of the variables such as insoluble residue, clay fraction, neomorphism, texture or skeletal composition and may result from a combination of Sr^{+2} variations in each population. Detailed electron microprobe studies may reveal the different amounts of Sr^{+2} contributed to the rock by each of the individual constituents.

Although all the mentioned variables may be responsible for the Sr^{+2} variation within and between the three groups, the possibility of a facies controlled variation cannot be ruled out.

While the average Sr^{+2} content of the analysed samples is in the region of 684 ppm and they show a more or less similar diagenetic history, the three rock populations (Planorbis, Angulata and Semicostatum) show distinctly different ranges. The Bucklandi Zone group of samples contain the highest amount of Sr^{+2} while the Planorbis Zone group show distinctly lower values than the other two. Sample M91 which contains unusually high contents of all elements and only 2.95% CaCO_3 was taken from a silty marlstone with ferruginous concentrations and delicate, crushed aragonitic bivalve shells in the Wilderness Member. The unusually high iron content of this sample may be explained in terms of post depositional processes and the enrichment of the ferruginous concretions. The Sr^{+2} content on the other hand may reflect the original concentration of the unaltered aragonitic shell material, this in turn may represent the Sr^{+2} concentration of the waters in which the bivalve thrived.

In plotting the population averages onto a frequency diagram (Fig. 11.3), it becomes evident that one is dealing with two limestone groups with Sr^{+2} modes of 400-600 ppm and 600-800 ppm. Relating the facies type of limestones to this somewhat bimodal distribution, a sequence of decreasing Sr^{+2} content may be seen from the well aerated, semi-enclosed, nearshore environments of facies 8 and 9b to the relatively more "offshore", less aerated, shelf deposits

of facies 13 and 19. Although this distribution and relation to facies type is somewhat speculative and based on scanty evidence (only 17 analyses); it shows some similarity to the Sr^{+2} distribution and its relationship to different facies as discussed by Veizer and Demovic (1973, 1974), who found high Sr^{+2} concentrations in limestones of hypersaline and deep, pelagic origins while lower concentrations characterised bank and shelf-deposited organodetrital limestones; a definite trend of decreasing Sr^{+2} content from a hypersaline, lagoon into an algal bank and also towards the deep sea was observed. In the Liassic samples studied, the overlap in Sr^{+2} content between various environmental groups is expected, since they represent a continuous profile and it is natural for their properties to be transitional.

i. Bimodal distribution of Sr^{+2}

It is not known if this bimodality is statistically true. Veizer and Demovic (1973, 1974) proposed that this might be caused either by two types of diagenetic solutions, which would cause repartition of Sr^{+2} regardless of its original distribution or the bimodality is inherited from the original sediments and preserved in spite of the diagenetic repartition. Their first assumption was questioned on geological grounds. The $\delta^{18}\text{O}$ data (cf. Veizer, 1977a) was also inconsistent with such a conclusion. As explained earlier it is generally believed that the Sr^{+2} content of a carbonate rock decreases with increasing degree of alteration by waters of a continental parentage. Thus the decrease in Sr^{+2} should be accompanied by an increasing isotopic equilibration of limestones to such waters and therefore the $\delta^{18}\text{O}$ should be progressively lighter (cf. Degens, 1967; Veizer, 1977a). It also follows that if the bimodality is caused by two types of diagenetic solutions only, the two solutions should be characterised by different $\delta^{18}\text{O}$ values and therefore the distribution of oxygen isotopes would be also bimodal; the studies of Veizer (1977a), Veizer and Demovic (1974) indicated that neither a bimodal distribution or other correlation with the Sr/Ca ratios existed. However the

lighter $\delta^{18}\text{O}$ values found in the samples taken from hypersaline environments than their open sea counterparts indicated that the O-Sr relations were a more or less "primary" feature.

The Liassic samples were not analysed for $\delta^{18}\text{O}$ content, but it is hard to see how the diagenetic solutions could follow the different facies-type boundaries in the rocks.

The distribution of Sr^{+2} in the original sediment may have been of a bimodal type. Veizer and Demovic (1974) showed that the Sr^{+2} distribution in Recent carbonate sediments is of a bimodal type probably due to the proportions of different skeletal material in the system. They also pointed out the relative importance of high-Mg calcite in open sea neritic and shallow bathyal environments and some reefoid and reef-detrital facies whereas aragonite is more typical of slightly hypersaline and some littoral and sub-littoral environments. Although the above generalisations may be broadly true, the reports are somewhat conflicting and both minerals are quite variable and their occurrence probably depends on local factors (Bathurst, 1975; Kinsman, 1969; Ginsburg *et al.*, 1971; Friedman, 1964; Land and Goreau, 1970). Bathurst (1975) and Alexanderson (1972) maintained that the mineralogy of the framework and matrix in unconsolidated carbonate sediments favours nucleation of the same mineral as cement.

The observations of Land and Goreau (1970) showed that the preferred nucleation may not work well in littoral and shallow neritic organodetrital sediments.

All the above mentioned factors may have been responsible for the bimodality of Sr^{+2} concentration in the original sediments with aragonite being abundant in the hypersaline, lagoonal facies and high-Mg calcite in algal banks and neritic and shallow, pelagic organodetrital varieties. Oxygen isotope and electron microprobe studies are needed to indicate whether continental waters were involved in the diagenetic alteration of the Liassic deposits. The actual transformation was very likely a solution-precipitation

process; such a process could lead to a bimodal distribution of Sr^{+2} in the diagenetically stabilised limestones.

If the original sediments were composed predominantly of stable low-Mg calcite (deep sea sediments), their Sr^{+2} content (~ 1000 ppm) would be less altered than in the case of the metastable minerals and therefore the diagenetically stabilised sediments will plot into the high Sr^{+2} group. If the above discussion of Veizer and Demovic (1974) is correct then it may explain both the bimodal distribution of Sr^{+2} as well as the increasing trend from offshore limestones to nearer shore deposits of the Liassic strata under study; the increase in Sr^{+2} content towards the nearshore (semi-enclosed) facies could be related to the increasing proportion of high Sr^{+2} , aragonite in the original sediments whereas the relatively slight decrease towards the offshore areas could be related to the increasing content of the stable low-Mg calcite.

The Sr^{+2} data obtained for the Liassic samples of northwest Scotland is comparable to the Sr^{+2} concentrations obtained by Veizer (1977a) for the Posidonienschiefer and Opalinuston of Germany; these values are similar to (or in excess of) equilibrium values for marine calcite. This would require their precipitation from diagenetic solutions with Sr/Ca ratios equal to or higher than that of sea water.

Although $\delta^{18}\text{O}$ data were not obtained from the Liassic samples, staining results suggest that some non-ferroan calcite is present together with ferroan calcite cement, furthermore most bivalve and other shell fragments are composed of non ferroan calcite. The above may indicate that the rocks were at least partly restabilised by solutions of meteoric derivation in which case a bimodal distribution of Sr^{+2} with different values for diagenetic solutions of meteoric and marine origin is expected. Figure 11.3 shows a possible tendency towards bimodality. The dissolution of marine calcite (or low-Sr aragonite with ~ 1000 -25000 ppm Sr^{+2}) in meteoric waters can lead to such high ratios only in an entirely "closed" diagenetic system.

In this case the precipitation of diagenetic calcite with ~ 200 ppm Sr^{+2} leads to an enrichment of the residual solution (e.g. Kinsman, 1969) and thus a higher Sr^{+2} content of the later-precipitated calcites. As seen in Figure 11.1 and Table 11.3 the absence of samples with such low Sr^{+2} concentrations argues against (except M89) this alternative. The dissolution of an original marine aragonite (~ 9000 ppm Sr) in a semi-enclosed diagenetic system may easily account for the required Sr/Ca ratios of the diagenetic solutions (Veizer, 1977a). This seems to support the suggestion that the original sediment or early cement of the studied Liassic beds was possibly high- Sr^{+2} aragonite.

From the above discussion it is clear that as our present state of knowledge concerning the details of diagenetic change and elemental distribution in carbonates is somewhat uncertain, Sr^{+2} values cannot be used independently in the determination of palaeoenvironmental conditions. Combined with sedimentological and palaeontological work it may be used as complementary evidence; Sr^{+2} may be used successfully in deciphering the timing of sequential diagenetic events in ancient carbonate rocks. Combined electron microprobe, $\delta^{18}\text{O}$ and SEM studies together with simple staining and petrography are needed to study individually, the various constituents of carbonate rocks; wet chemical analysis of Sr^{+2} concentrations of whole-rock samples of ancient limestones can be misleading.

"We must take our limestones to pieces, they are not equilibrium systems: they consist of different generations of allochems, cements, veins, vug fillings, etc., all formed under different diagenetic conditions" (Hudson, 1977).

11.9 Iron and manganese

The samples which were taken from three different suites also differ from one another in Mn^{+2} and Fe^{+2} contents.

Accepting 17.4 as a partition coefficient for Mn^{+2} between sea water and diagenetic solution (Bodine et al., 1965), (1.9 and 41100 ppb of Mn and Ca; Turekian, 1972) ; calcite in equilibrium with sea water should contain approximately 30 ppm of Mn. If the estimated partition coefficient for Fe suggested by Veizer (1974) is applied, calcite in equilibrium with sea water (3.4 ppb of Fe; Turekian, 1972) should contain 70 ppm of Fe. These estimates are not within the range measured for the Liassic rocks. On the other hand the iron concentration of oxygen deficient sea water should be 3-30 ppm according to Holland (1973) and the concentrations of Mn are probably < 1 order of magnitude lower (Veizer, 1977a). Taking the lower concentration levels and applying the above partition coefficients, calcite in equilibrium with such waters should have ~8000 ppm of Mn and ~ 60000 ppm of Fe. Although the quoted, theoretical values are rough estimates it is seen that the Mn^{+2} and Fe^{+2} concentrations in the Lower Liassic samples are much lower than the predicted values (Fig. 11.4). The geochemical data is somewhat compatible with the palaeontological and sedimentological data of the Semicostatum Zone beds in the northern area which indicate an oxygen deficient sedimentary environment. The lower concentrations however, may be a direct reflection of the non permanence of the stagnant depositional conditions or "open system" diagenesis where elements were freely exchanged with different diagenetic fluids originating from different environments.

It should also be noted that the concentrations of Fe and Mn are most probably not only related to the calcite lattice but are mainly related to the presence of iron minerals. Both iron and manganese have been introduced into the sedimentary basin in oxidised form and precipitated as coatings of iron compounds (see Chapter 9). It is likely that due to the abundant organic matter and reducing conditions below the sediment/water interface, most of this Fe and Mn was diagenetically redistributed into siderite, pyrite, "chamosite" and cronstedtite. This would be consistent with the general palaeogeography and humid climate during the Lower Liassic times. Detailed

wet chemical and electron microprobe analysis of separated shell fragments and the enclosing rocks would determine if the Sr^{+2} , Fe^{+2} and Mn^{+2} concentrations of the shell calcites were derived by a late diagenetic "secretion" from the enclosing rock.

CHAPTER 12

PALAEOGEOGRAPHY

An understanding of the environmental conditions operating during the deposition of the Broadford Beds Arenaceous and Argillaceous Groups is greatly facilitated if the different facies already described are considered within the framework of seven sequences. It should be noted that these sequential arrangements do not imply depositional synchronicity within the basin of deposition.

The transition from the Rhaetic (?) through the Hettangian and into the Sinemurian marks a worldwide rise in sea level (Hallam, 1969b, 1978). During this time a shallow sea spread over much of northwest Europe. This is the only part of the world away from the Pacific margins or parts of the Tethyan zone where a marine Rhaeto-Hettangian sequence is known (Hallam and Sellwood, 1976).

In general the western Scottish Lower Lias developed in two broadly defined environments, namely in a primarily shaly facies to the south (Mull, Morvern) and in sandier facies to the north (Skye). Strong evidence for the development of Liassic northeast-southwest running shorelines exists in the northern area and it is evident that the northern and southern basins evolved separately.

During the Hettangian stage, the areal extent of the seas was minimal and very shallow water sediments (?continental in places) were much more widespread than their deeper water equivalents.

The distribution of the marine deposits is generally similar to the ?Rhaetian strata on which they rest conformably; in places they locally overlap the ?Rhaetic beds to overlie Triassic conglomerates and/or the thrust complex of Cambro-Ordovician Durness Limestones together with late Precambrian Torridonian sandstones (Hallam, 1975), indicating slight phases of transgression. During the Sinemurian stage a widespread transgression

affected the Inner Hebrides of Scotland with the sea spreading for the first time in the Jurassic (Hallam, 1978). Although no notable regression with respect to the Hettangian is known, a minor regressive "pulse" is recorded during the Turneri Zone in the Skye and Mull areas. Walther's "Law of facies" (1893-4) may be successfully applied here. The model proposed for the successive phases of deposition is that of a shallow marine basin with a marginal, shallow, quartz-clastic/carbonate shelf (somewhat similar to the present day Persian Gulf), which gradually sloped into only slightly deeper offshore waters where finer grained argillaceous sediments accumulated to the south. A multicomponent system controlled sediment distribution and the development of various facies (Fig. 12.1).

In the Lower Liassic strata of northwest Scotland, apart from the beds of the Breakish Coral, Breugh Pebble and Breakish Ironstone Members, individual beds cannot be traced over considerable distances. This and other overlapping characteristics suggest that although sedimentation took place in widespread shallow epicontinental areas with reduced tidal and other circulation (Shaw, 1964) local topographic and/or tectonic control was influential in the development of various facies patterns. Epicontinental seas are regarded as subtle indicators of facies shifts and sedimentation changes because their small scale volumetric changes tend to affect the ocean as a whole (Shaw, 1964; Irwin, 1965). Evidence from the northwest European Mesozoic basins shows that considerable subsidence occurred throughout the era (Sellwood, 1970) which was coupled with tectonic movements (Sellwood and Jenkyns, 1975; Hallam and Sellwood, 1976; Jenkyns and Senior, 1977).

The interaction between the two above mentioned factors, eustasy and epeirogenesis, was most probably responsible for the development of the Lower Liassic "cycles" in northwest Scotland.

As shown in Fig. 12.2 the Hettangian facies of the studied area were developed asymmetrically, with gradual fining upward sequences being abruptly followed by rapidly transported and deposited coarser sediments; in contrast

the Lower Sinemurian marks the development of coarsening upward sequences which are abruptly terminated at their tops. The concomitant faunal changes which occur suggest relative shallowing and deepening as suggested by Duff et al. (1967). The development of such facies differences on a regional scale requires some overriding tectonic (Sellwood and Jenkyns, 1975; Hallam and Sellwood, 1976; Jenkyns and Senior, 1977), eustatic (Hallam, 1969b, 1978) or local tectonic combined with eustatic changes (Sellwood, 1970).

Numerous local basins have been recognised in the Lias of England, Scotland and Wales (Arkell, 1933; Hallam, 1958; Sellwood and Jenkyns, 1975), and although it is known that tectonic subsidence did not occur regularly over wide areas in this region, widespread synchronous cycles of deepening and shallowing water, independent of local tectonic and facies development have been reported by Hallam (1978) in the Jurassic of northwest Europe. Although the development of various facies in the Lower Lias of northwest Scotland can be related to the widespread changes which have been reported to affect most of northwestern Europe during this time, it is seen that a local tectonic control was of considerable importance as well.

The history of the development of the sedimentary basins in the Sea of the Hebrides and the Minch has been recently investigated by Binns et al. (1975) and McQuillan and Binns (1975) based on published onshore and substantial off-shore data. Three principal north easterly faults control the western margins of asymmetric sedimentary basins in this area. The Sea of the Hebrides basin is bounded on the west by the Minch Fault. To the north the North Minch Basin is controlled by the Minch Fault; to the east the Inner Hebrides Basin is controlled by the Camasunary-Skerryvore Fault and displacement on the Great Glen Fault affects the area south of Mull. The local Triassic-Jurassic features which obviously affected the sediment distribution pattern in the Inner Hebridean Basin are as follows:

i. Central Strath High

Palaeogeographic studies by Steel (1971a; 1977b) showed that the central Skye area was a structurally high drainage basin which supplied sediment to Raasay and Scalpay (during the Trias).

During the early Lias this area (Loch Slapin and Torrin in southwestern Strath) remained an area of erosion rather than deposition, but unlike during the Trias, sediment supply from this land area to the depocentres was negligible and confined to very local areas. This is probably a reflection of the more humid climatic conditions which dissolved the Durness limestone constituents. During Middle Liassic times the Central Strath High area was completely covered by the Liassic Sea.

Steel et al. (1975) provided valuable information concerning the latest Triassic and/or earliest Jurassic upland areas in central Strath. An alluvial fan system drained local Palaeozoic and Torridonian areas between Beinn an Dubhaich and Ben Suardal. It dispersed eastwards and a floodplain system drained a metamorphic region northeast of Broadford dispersing towards the southwest. The age relationship of the two sedimentary systems is not clear but as discussed in Chapter 6 of this study, the succession is regarded as representing the youngest and most easterly Triassic sedimentation in the Minch region.

An eastward enlargement (younging) of the Minch Basin complex was suggested in the Triassic period as evidenced by regional thickness changes. This overall change in thickness is interrupted by local thinning and thickening of the strata signifying deposition on an initial topography (e.g. the succession of Triassic conglomerates of 50 m. in the Heast district (NG 654178), together with the Lower Liassic strata of the Milton Formation, thins to zero in the area south of Torrin (NG 575201)).

ii. Camasunary Fault

Although contemporaneous movement on the Camasunary fault has been suggested for the deposition of Permo-Triassic sediments (Steel, 1971a, 1974), there is no direct evidence for such movements during the Lias (Hallam, 1959; Howarth, 1956). Examination of the Lower Liassic outcrops of Torrin (NG 580207), along Allt Slapin (NG 580213) and Allt an t-Statha Bhig (NG 575225) situated east of the Camasunary fault revealed the presence of calcareous micaceous shales (Strath Formation) overlying cross bedded calcareous sandstones (Milton Formation). The calcareous sandstones resemble those found along Abhainn nan Leac (NG 526199) and on the southern flank of Ben Bla Bheinn (NG 528206) (west of the Camasunary fault). The exposures in both areas are very poor, but the character and composition of the beds do not suggest penecontemporary movement along the fault line.

It should be noted that the topography which was initiated in the Trias probably controlled the deposition of the very shallow water, quartz-free carbonates in southern Raasay (Suisnish) and north Skye (Sligachan).

iii. Coll-Tiree

Steel (1971a) showed that the two northern (Skye) and southern (Mull) depositional basins were separated during Triassic times. There is ample evidence that upland areas of basement rock persisted in Jurassic times, and some evidence (Tor Mor Member) has been given for the persistence of upland areas in southern Strath during earliest Liassic times. It is probable that the northeast/southwest upland barrier through Coll and Tiree persistently isolated the two Hebridean basins in Liassic times but with the lack of direct evidence this suggestion can only be very tentative.

iv. The Great Glen Fault System

The Great Glen Fault forms an important structural feature in the southern area of study. Kennedy (1946) believed that it shows a pre Late Devonian -

?Early Carboniferous sinistral displacement; Turner et al. (1976) found no evidence suggesting post Devonian sinistral movement. The continuation of this fault offshore in the Moray Firth is less linear and it has been suggested (Bacon and Chesher, 1975; Bott and Browitt, 1975) that this fault shows a normal vertical displacement during the Mesozoic. This view has been contested by Flinn (1975, 1977). Lee and Bailey (1925) suggested that a possible line of the fault continued southwards to Loch Buie on the south coast of Mull but Binns et al. (1975) explained that the distortion of the area due to Palaeogene igneous activity is such that the location of the fault with respect to the Permo-Triassic and Jurassic rocks of northeast Mull cannot be determined with confidence.

The thickness of Permo-Triassic sediments lying to the west of the fault (Rast et al., 1968) was interpreted as fault-controlled by Steel (1974) but the relationship with the Great Glen Fault is not clear. Holgate (1969) reported a 29 km dextral transcurrent movement along this fault since the Upper Jurassic.

Pronounced breaks in sedimentation occur in the Loch Aline and Mull areas during Lower Liassic times (Oates, 1978). Such breaks together with significant changes in the lithological character of the beds from Loch Aline to Mull suggest intra Liassic erosion and significant topographic differences. Although no direct evidence suggesting Liassic penecontemporaneous movement along the Great Glen Fault was found, such movements cannot be ruled out.

v. The Minch Fault

The Minch Fault is a northeast-southwest trending fault line which controls the western margin of the Sea of the Hebrides Basin and the Minch Basin.

Along this fault, Mesozoic sediments are thrown down to the east against

Precambrian (Lewisian) gneisses and basal Mesozoic sediments.

Information is available for Mesozoic sedimentation along the western edge of the North Minch Basin. Along the Minch Fault (Steel and Wilson, 1975), it is suggested that the Stornoway Formation (4 km. thick) is of ?Permo-Triassic age and the different sedimentation phases within it were tectonically controlled. This formation was deposited as the sedimentary fill within the deep, western margin of an asymmetrical North Minch (Permo-Triassic) Basin (cf. extreme thickness of Inner Hebridean Trias). Further considerations suggested that the western margin of the Basin shifted westward as the locus of faulting and fault generated sedimentation migrated from the Minch Fault.

12.1 Lower Liassic facies relations in the Hebridean Basin

Lower Liassic facies occurrences and their general relationships in the area of study are presented in Figures 12.2 to 12.8. The various facies identified in the Broadford Beds Groups represent 7 sequences in the northern and southern areas of study and it can be seen that there is some similarity in their development. The sequences are represented by lateral equivalents of the "northern" facies in the Mull area.

Facies (Skye)				Facies (Mull)	Sequence No.
1	2	3		↑ 19 ↓	1
4	5	6			2
7a ₁	7a ₂	7 _D	8	↑ 20 ↓	}
7b ₁	7b ₂	7 _D	8		
7c ₁	7c ₂	7 _D	8		
9a	9b				4
10	11	12		21a - 21b	
13	14				5
15				22B	6
16a	17a	18a		↑ 22a - 22b ↓	}
16b	17b	18b			

The general lateral, vertical and compositional relationships among the various facies are as follows:

i. Sequence 1

Facies 1 abruptly overlies the underlying Triassic-?Rhaetic and in places Precambrian rocks in the Skye area; it is represented by conglomerates in Skye, white sandstones in Ardnamurchan and alternating shales and limestones (Facies 19) in Mull.

This facies thins out towards the north and northwest in the Skye area.

Facies 2 transitionally overlies Facies 1; it contains similar clasts as Facies 1 and can be seen in Applecross where it shows erosive contacts with the underlying strata. The graded pebble beds show abrupt and in places erosional contacts in southern Skye and Applecross.

Facies 3 transitionally succeeds Facies 2 and can be seen in southern Strath and along the western Liassic outcrops of Beinn nan Cairn from Heast to Skulamus.

ii. Sequence 2

Facies 4 is best seen in Applecross where it abruptly overlies the beds of Facies 3; in northern Strath Facies 4 is represented by marlstones with limestone pebbles and it is absent in southern Strath.

Facies 5 abruptly follows Facies 4 and is well represented in Applecross, northern and southern Strath; a decrease in the total thickness of this facies is seen from Applecross to southern Strath.

Facies 6 (coral) transitionally overlies Facies 5 in Applecross, northern Strath and central Strath. Indications for its presence in Ardnamurchan are also presented; no appreciable regional thickness change is seen. Facies 6 is absent from the Raasay succession and southern Strath.

iii. Sequence 3

Facies 7 is well represented in the northern and southern areas of study;

there is no apparent break in sedimentation and the beds transitionally overlie those of Facies 6. The lowermost beds of Facies 21b in Mull and parts of Facies 20 in Loch Aline are equivalents of this facies of the Skye area. In the northern area of study this facies shows considerable local variations. The equivalent of Facies 7 in Ardnamurchan comprise calcilutites, biomicrites and biomicrosparites.

Facies 8 (coral) transitionally overlies Facies 7 in northern Strath. Its lower contact is not exposed in the Applecross succession (?faulted). It is developed inland in the Strath district, but is absent from the Loch Eishort succession.

Although the beds of this facies are lenticular, the regional overall thickness does not differ appreciably from that of the type area.

iv. Sequence 4

In Facies 9 no signs of an erosive lower contact with Facies 8 can be seen; it is present in Applecross, northern and southern Strath but its existence in Raasay cannot be established with confidence.

This facies is also present in Ardnamurchan; parts of Facies 21a and 21b represent basinal aspects of this facies.

Facies 10 shows a transitional base in all its major outcrops of the Skye and Ardnamurchan areas. Facies 11 and 12 when present and distinguishable, show erosive, abrupt contacts.

v. Sequence 5

Facies 13 consists of coarsening upward shale/siltstone beds and although it is not well exposed in Ardnamurchan and Loch Aline, it is well represented in northern Strath. Facies 14 represents the shallowest (and ?emergent in the Loch Aline succession) depositional conditions within sequence 5. This sequence is absent from Applecross and its contact with the underlying beds is obscured and poorly understood in all the sections of the northern (Skye) area.

Facies 14 is locally well developed in northern Strath, but only faint indications of its presence may be found in the Raasay, Loch Slapin and Loch Eishort sections.

vi. Sequence 6

Facies 15 is present in all the major sections of the northern area of study where it is represented by coarsening upward repetitions. It is not well represented in Ardnamurchan where an oolitic ironstone was found within the Sauzeanum beds of this facies. A major stratigraphic break eliminates most of the beds from the Loch Aline succession; it is laterally represented by beds of Facies 22B in Mull.

vii. Sequence 7

Beds of Facies 16, 17, 18 of sequence 7 gradationally overlie the strata of sequence 6 in Raasay; this contact is faulted in southern Strath and the sequence is not recognisable in Ardnamurchan.

In Loch Aline this sequence is represented by a thin development of facies 22a. At Craignure Bay (Mull), no break in sedimentation is seen and a complete basinal equivalent of Facies 16, 17, 18 is observed.

12.2 Palaeogeography of the Hebridean Basin

The palaeogeographic evolution of the Hebridean Basin of northwest Scotland during the Lower Lias is summarised in Fig. 12.2 which is based on the seven broadly defined sedimentary sequences already described. The explanations are supplemented by Figs. 12.3 to 12. 8.

i. Triassic

Steel (1971a, 1974, 1977b) has described the development of alluvial fans, playa lakes, floodplain, braided river and tidal flat depositional environments from the Triassic (Storetveit and Steel, 1977) of the Hebridean Basin. He interpreted the deposition of alluvial fan conglomerates as marking the

rejuvenation of the landscape due to activity along northeast-southwest trending faults (see also Steel and Wilson, 1975).

Although the overall sedimentation pattern indicates source areas to the southeast and northwest along the margins of a large graben structure, local source areas as "upthrown blocks" in central Strath and a ridge extending southwest through Coll and Tiree strongly influenced the depositional characteristics in their immediate vicinity.

The northern end of the Inner Hebrides Basin (Binns et al., 1975) showed a complex structure during the Trias that probably closed not far to the north. Sedimentary filling took place down a southwesterly palaeoslope while lateral filling by alluvial fans took place from northeast-southwest ridges (Steel et al., 1975). It is very important to note that throughout the Trias, the central Strath area was a structurally high drainage basin which supplied sediments to Raasay and Scalpay and therefore was itself an area of erosion and not sedimentation at that time.

The possibility that the Inner Hebrides Trough (McQuillin and Binns, 1975) is younger than the trough further northwest, or that it migrated northwards with time cannot be ruled out; in this way the Skye sediments would exemplify the youngest part of the filling (Steel, 1971a).

An important finding of Steel et al. (1975) is that the base of the New Red Sandstone is diachronous, younging to the east, i.e. it is possible that the tectonic basin complex which controlled the extent of sedimentation, migrated eastward with time.

The maturing Hebridean landscape was inundated by the sea in Rhaetic times; estuarine and tidal flat sediments of the Passage Beds (Steel, 1974) were deposited and at localities around Heast (NG 648173), northern and western Strath very thin sediments occur which have been lithologically compared with the Rhaetic strata of Mull (Lowe, 1965). The abundance of cornstone and caliche deposits in the Triassic succession suggests somewhat arid to semi-arid climatic conditions; towards the top of the New Red

Sandstone succession the climate became progressively more humid (Steel, 1974). The clay minerals of the red siltstone and mudstones (floodplain deposits) consist predominantly of illite with some chlorite and kaolinite also present whereas the red mudstones immediately underlying the Rhaetic-Liassic beds of western Mull are dominantly composed of kaolinite with illite and smectites present as well.

ii. Planorbis Zone

During this time open marine deposition occurred in western Mull (NR 404287). This area together with the island of Arran is the only one where fully marine deposits are known in northwest Scotland. Strata of this age are absent from Loch Aline and Craignure, suggesting that sedimentation took place in areas with variable local topography with the land sloping gently towards the southwest. Deposits of this age were either not deposited in Loch Aline or were overlapped by the succeeding Liassic beds. Facies relations suggest a strong topographic gradient between areas of offshore sedimentation in central and western Mull and areas of very shallow (and/or non) deposition in eastern and northeastern parts. The absence of marine Planorbis Zone deposits from Craignure Bay (NM 730360), Glenbyre Farm (NR 578227), Port Ohirni (NR 634201), the entirely clastic nature of the sediments of Craignure Bay and Port Ohirni together with the occurrence of limestones with "deeper" water aspects in Glenbyre Farm and the progressive increase of dip from southeast to northeast (Port Ohirni 15° NE; Glenbyre Farm 80° NE) indicates the presence of a possible northeast-southwest trending upland area with gentle slopes towards the northwest but which was separated from the central Mull basin by somewhat steeper scarps.

No marine deposits of this age are known from the Ardnamurchan and Skye areas. It is possible that continental deposition continued on these gently west-southwest sloping upland areas from Triassic times.

The deposition of marine facies in western Mull is a direct consequence

of the transgression which marks a widespread facies change from the underlying strata in northwest Europe (Hallam, 1978). Deposition of open marine beds in Wilderness during this Zone indicates that while this transgression effectively infilled the deeper local basins in the south and west, it could not affect the northern (Ardnamurchan-Skye) and eastern (Craignure) areas (Fig. 12.3; 12.3a).

The clay mineral suite of the Planorbis marine beds is different from that of the Triassic "floodplain" deposits in that smectites and illites dominate over kaolinite and no chlorite is present. An intermediate suite is recognised for the Upper Triassic - ?Rhaetic beds with moderate amounts of smectites and illite. The consideration of other physical, chemical and palaeontological criteria suggests that these differences may have been influenced by a progressive change of climate from semi-arid to a warm humid one in the early Lias.

iii. Liasicus Zone

Continued sea level rise and/or land subsidence affected the southern area of study where a gently southwest sloping basin was fringed by muddy shorelines and was continually infilled by argillaceous sediments eroded from nearby land areas under humid climatic conditions. The basins which had developed in Loch Aline and Wilderness were still relatively deeper than those in the Craignure area to the southeast and the Ardnamurchan and Skye areas to the north. In the Craignure area, deposition of coarse, very shallow marine (?estuarine) sands continued, which is regarded as signifying the existence of a topographically higher region as compared to Loch Aline. The same situation is true for Ardnamurchan where only white pebbly sandstones and some limestones developed. Although no stratigraphically significant fossils were found in the northern area, it is possible that the beds of the TorrMor Member were being deposited during this time.

Gentle west and southwest sloping land areas may have been present in

this area and probable pulsar movement along ? northwest-southeast trending lines developed steeper slopes toward the north and northwest with the consequent development of local drainage basins off northwest Raasay. No appreciable changes can be recognised in the climatic conditions and while the sediments were mostly derived from the southeast and east, the presence of rounded epidote grains together with zircons suggest possible derivation from the Epidote Grits of the Torridonian, Applecross Group of Strath which show abundant outcrops in the northeast and eastern parts of Strath.

Hudson (1964) attributed the existence of an epidote, garnet, apatite and sphene suite of minerals in the Middle Jurassic sandstones of northwest Scotland to the existence of source areas of basic orthogneisses (Lewisian) in the Outer Hebrides. The association of rounded epidotes with polycrystalline quartz grains of metamorphic origin, together with other direct evidence (Steel and Wilson, 1975) suggests that the outer Hebridean Lewisian basement complex actively contributed to sediments during the Triassic and early Jurassic times. In the Lower Liassic beds under study, epidotes are not associated with sphene, apatite and garnet but with rounded zircons (very scarce), which would indeed suggest the possible influence of Torridonian arkoses; the picture is further complicated by the presence of subrounded chert fragments of ?Durness affinity. It is obvious that various local source areas were contributing sediments to the depocentres while a gradual sea-level rise or land subsidence was affecting the northern area.

iv. Angulata Zone

Continued sea level rise affected both the southern and northern area during this time (Fig. 12.4). The progressive inundation of land areas with variable topography has been reported for both areas. This is indicated in the southern area by the variable thickness of these beds within short distances and their overlapping relation with Carboniferous strata in Loch Aline.

Marine deposition occurred for the first time on the upland areas of Craginure, Ardnamurchan and Skye evidencing the progressive inundation of the land areas. This is manifested by the Liassic beds with Liostrea hisingeri which overlap the previously deposited strata to overlies directly onto Moine Schists in Ardnamurchan.

Horizons of Liostrea-bearing beds are widespread in both the northern and southern areas and signify a widespread rise in sea level which managed to inundate the structurally high areas of Craginure, Ardnamurchan and Skye for the first time in the Lias. The two Coral Facies 6 and 8 in the northern areas signify periods during which relatively more open marine, warm, clear-water depositional conditions prevailed allowing their survival in the photic zone. Such conditions were never reached in the southern area probably because of the generally turbid nature of the waters with the lack of suitably shallow and firm topography. It is ideally desirable to know the exact subzonal position of the coral beds in both regions in order to verify the exact chronological order of the progressive northward inundation which took place. Corals cannot be used as fine stratigraphic markers and although ammonites have been identified from above and below the beds of Facies 8, its true age can only be determined at the zonal level which seems not to vary in either of the two areas (Skye and Ardnamurchan). The thickness of the overlying strata (to the top of the Breugh Pebble Member) cannot be used as a reliable index for the depositional time intervals involved due to the variable tectonic and sedimentary processes which had influential effects on the thickness and distribution of the sediments. It has been shown that the base of the Milton Formation is diachronous (younging towards the north). Therefore it is probable that the coral beds in this formation also followed a northward advance of the sea.

As shown in Fig. 12.2, sequences 2 and 3 consist of two fining upward successions which are interpreted as indicating two, slow, progressive sea level rises/land subsidence which followed intervals of rapid sea level fall.

The periods marking a lower sea level are characterised by cross laminated sandy and oolitic beds with pebbles, the composition of which was shown to vary according to their proximity to local source areas (viz. Central Strath High; Applecross and eastern Strath).

The thickness variation of the Bucklandi Zone successions in the northern and southern areas seems to contradict the notion of a progressive northward shallowing of the basin in northwest Scotland (the maximum thickness of the Angulata Zone beds in Applecross is 35 m. whereas this zone is represented by only 12.58 m. of shales and limestones in Loch Aline). Although it is probable that the lowermost 10 m. of the succession in Applecross does not represent the Angulata Zone, this still cannot effectively resolve the apparent thickness contradiction.

Although no evidence of contemporary faulting and/or movement is found in the southern area during this time, the largely condensed sequence may indicate a progressive syndepositional uplift of the area; this was not reflected in the composition of the sediments due to the distal location of the land areas (illite is the dominant clay mineral in these beds and benthic organism occurrences are greater than in the northern area where quartz clastic sedimentation was predominant). The erosion of metamorphic terrain (mostly Dalradian schists) in the southern area, under warm, moderately humid climatic conditions produced abundant fine grained argillaceous sediments which were transported into the depositional basin off southwest of Loch Aline. In the northern area, quartz clastic material was abundantly eroded off the Precambrian and Palaeozoic outcrops of the Torridonian and Cambrian strata, together with subordinate metamorphic terrains. Thus minor changes in the sediment supply and/or transportation would be reflected in the sediments of the northern area as pebble beds and/or strongly cross bedded strata but this would not be the case for the chemically vulnerable rocks being eroded in the southern area. It is probable that simultaneous deposition and subsidence occurred in the northern area while a gradual syn-sedimentary uplift was

occurring in the southern area. The two phases of rapid shallowing coupled by gradual deepening of the sea in the Skye area probably reflect the instability of the depositional basins which may account for the thickness of material deposited therein. The underlying mechanism of this apparent shallowing in the southern area, while deepening of the northern basin occurred, is of great interest and may signify a progressive tilting of the basins northward.

v. Bucklandi Zone

During this Zone important events took place both in the southern and northern areas while the Liassic sea continued to inundate land areas in both regions (Fig. 12.5).

The continued local sea-level fall/land uplift which was responsible for the development of condensed beds in the southern area during Angulata Zone times continued into the Bucklandi Zone and reached its culmination at the end of this time, as evidenced by the development of condensed beds capped by (?) subaerially exposed beds of the Loch Aline Ochrous Member. Faunal evidence also points to a drop in species diversity in both areas, from the Angulata to the Bucklandi Zone time. This evidence together with the development of prograding shoreface-shoreline deposits of the Breugh Pebble Member (Facies 10, 11 and 12) in the northern area indicates an overall regression although it is impossible to determine directly whether this was due to fall in sea level, rise of the hinterland or a combination of both.

The condensed sequence in the southern area together with the evidence of subaerial(?) exposure in Loch Aline points out a gradual uplift of the land areas here, while the prolonged uninterrupted deposition of albeit very shallow water deposits in the Craignure area (immediately to the southeast) indicates a continuation of land subsidence there and its structural independence from the northern Loch Aline area.

In Skye sequence 4 represents a slow regression which becomes more rapid towards its top due to possible acceleration of erosional processes in the hinterland. The pebble beds of Facies 11 overlying sandy oolitic limestones and marine shales, is accepted here as representing a minor regressive pulse which may have been caused by the general uplift of the whole region (including Loch Aline). Throughout the deposition of the Milton Formation, although the fauna and sedimentary evidence point to shallow unstable, shelf depositional environments, the thickness of these beds is considerable and cannot be accounted for by simple rise in sea level or the development of basin axis etc. It is possible that while the Loch Aline area was undergoing uplift and erosion, the northern area underwent progressive subsidence and the hinterlands were periodically subject to uplift and erosion, providing the immediate basins with an ample sediment supply.

Although the relative change of sea level at this time cannot be commented on, it is possible that a "depositional regression" (Curry, 1964) occurred and which corresponds to "facies change A" of Hallam (1978) indicating regressive/shallowing events. Based on recent ammonite discoveries (Oates, 1978) it is now established that the beds of Facies 10, 11 and 12 occur below the Rotiforme^{-Bucklandi} Subzone in Applecross whereas equivalent pebbly sandstones in Ardnamurchan occur below beds containing faunal assemblages reminiscent of those found at the Conybeari/Rotiforme Subzonal boundary in Morvern. This indicates that the age of the pebbly beds possibly youngs toward the north and the thickness of strata between the beds of the Breakish Coral Member and the top of the Breugh Pebble Member (3-5 m. in Ardnamurchan but up to 15 m. in Applecross) apparently supports the notion of a northward-moving pulse of basin uplift which started in the Loch Aline area during Angulata/Bucklandi Zone times but did not reach Ardnamurchan until before the Conybeari/Rotiforme Subzone, and succeeded to affect the Applecross succession before Rotiforme Subzone time; the obvious time involved also produced to some extent, the differences in sediment thickness from Loch Aline to

Applecross.

vi. Semicostatum Zone

Faunal evidence suggests a strong diversity increase from the Bucklandi to the Semicostatum Zone which may in turn be attributed to a widespread transgression which affected the northern area (Fig. 12.6 and 12.7). This zone marks important changes in the early Liassic depositional environments of northwest Scotland. In the Loch Aline area, (?)emergence or non deposition continued in most parts but the existence of some beds of Reynesi Subzone age indicate a short-lived period of inundation possibly due to the composite effect of variable topography and sea level rise which promoted the development of thin, patchy successions. General subsidence in the Loch Aline area would have produced a laterally continuous, more complete succession.

Deposition of silty calcareous shales occurred in the Mull area and thick oyster beds with some resemblance to those found in the Skye area were laid down.

In the northern area the prograding sequence of the Breugh Pebble Member develops into a retrograding sequence (Reineck and Singh, 1973) towards its top and the lack of a complete transgressive sequence and the absence of transition zone deposits signify a rapid inundation of a low lying topography with gentle slopes (depositional transgression of Curray, 1964). The vertical profile resembles sequences A and B of Hallam (1978).

The Semicostatum Zone silty calcareous shales are characterised by a deeper water fauna than the beds of the underlying strata. In the northern area they are seen to overlap the underlying beds to lie directly upon older Palaeozoic formations. This not only signifies local topography but a rise in sea level.

A possible change in climatic conditions, together with the possible complete erosion or the covering of some sediment source areas is reflected in the clay mineral composition of the shales overlying the Breugh Pebble and

Leacach Ochrous Members. Probably the change of climate (with increased rainfall) from a warm humid one (which somewhat inhibited the formation of $7\text{\AA}$ clay minerals, but facilitated the formation of smectite clays from the erosion of basic igneous rocks under alkaline pH and conditions of poor drainage and/or low rainfall), to one in which the in-place kaolinisation of feldspars rapidly took place (due to the existence of high rainfall, good drainage conditions causing acid leaching in waters of low pH) was instrumental in the creation of such depositional patterns.

The Craignure area was structurally independent from the areas further north. Continued uplift or non deposition continued in the Loch Aline area and in the northern area a rapid rise in sea level was followed by several minor phases of sea level fall which were recorded in the sediments as coarsening upward cycles at the culmination of which "ironstones" developed. Local "basin and swell" topography in the northern area enhanced the formation of oolitic ferruginous beds on areas with local "swell" characteristics centred somewhere near northern Strath whereas pelletal and iron-rich beds formed in areas of relatively local "basin" character such as those in eastern Raasay, Loch Eishort and Loch Slapin. Despite local topography, the "ironstone" beds developed on an extensive, wide platform which extended to the southern parts of Ardnamurchan with gentle slopes towards the west and possibly the north (if some structural unity of the Skye and Mull area is assumed). The lack of any evidence of soft sediment slumping and the non existence of turbidites suggests that although topographic irregularities did exist on the floor of the epeiric sea, they were subdued enough to prohibit the development of such features (Sellwood and Jenkyns, 1975). The sequence of strata signify a regression somewhat similar to that of Hallam's "regressive sequence B" (1978).

The ironstones are again abruptly overlain by a development of deep water ammonite bearing shales (Hallam's transgressive sequence B) which indicate another rapid subsidence which is followed by a slow (somewhat pulsar) uplift

in the northern area of study while ferruginous marls and shales continued to develop in the Craignure area; the Loch Aline locality remained an area of erosion or non deposition throughout this time. It is interesting to note that the ferruginous oolitic beds found in Ardnamurchan are dated as ?Sauzeanum Subzone indicating the possible development of very shallow, turbulent depositional conditions in this area during the late Semicostatum Zone. Continued uplift or sea level fall is seen in the Skye and Ardnamurchan areas towards the end of this Zone.

vii. Turner's Zone

A widespread regression affected the depositional areas of northwest Scotland during this time (Fig. 12.8). In the northern parts the general profile resembles that of Hallam's "regressive sequence B" (1978) with the deeper marine shales of the Strath Formation passing up through sandy shales into shallow, shoreface sandstones; locally ferruginous beds are very common.

This general regression was most probably due to continued uplift or lowering of sea level in the Skye area while subsidence or sea level rise took place in the Loch Aline area; this is reflected by the locally transgressive character (transgression type G and F of Hallam (1978)) of the Turner's silty shales. The regressive sandstones of the Boreraig Member were deposited in the northern area throughout the Brooki Subzone, above which a disconformity marks the Birchi and Obtusum Subzones. The Leacach Shale Member was deposited in the southern area during the Birchi Subzone times, the top of which is marked by the Leacach Pebble Member.

The sedimentary evidence given shows that the Hettangian of northwest Scotland is dominated by episodes of rapid sea level rise/land subsidence which were followed by a slow sea level fall/land uplift. This corresponds to Hallam's "eustatic model D" (op.cit.).

The data so far presented support Hallam's (op.cit., fig. 4) notion that

the general deepening which occurred during the Hettangian-Sinemurian resulted in the formation of transgressive facies pattern types A and B whereas the Turnerian shallowing events prompted the formation of the regressive facies pattern B.

12.3 Lower Liassic sedimentation control in northwest Scotland

The facies distribution and composition of the Lower Liassic strata described in this chapter may be explained by considering sedimentation in "rift zone" areas where tilted fault blocks are the basic tectonic unit (Sykes, 1975; later sections in this chapter). It is readily seen that the subsidence styles of three broadly defined (albeit minor) blocks have controlled sedimentation and the resultant facies distribution to a great extent within the Inner Hebrides Basin from Planorbis to Turnerian Zone times. The blocks are defined as follows (Fig. 12.9):

- i. Skye-Ardnamurchan Block.
- ii. Mull-Morvern Block.
- iii. Eastern Mull Block.

The subsidence of each independent block may at any given time follow one of the following patterns:

- i. Epeirogenic subsidence (e.g. Planorbis-Late Angulata in the three blocks; ?Semicostatum in western Skye and Ardnamurchan).
- ii. Differential tilting (e.g. gradual subsidence in the north of Skye-Ardnamurchan Block during Angulata to Bucklandi Zone while uplift occurred in the Mull-Morvern Block; a period of stabilisation is seen at the base of the Semicostatum when prograding shoreface facies formed in the north while a disconformity developed in Loch Aline. It is possible that some uplift took place in both areas resulting in a regression.
- iii. Reversed tilting and subsidence. This continues in the Semicostatum Zone after the general transgression, while the Loch Aline side of the Mull

Block is a "high" area. The subsided Skye area is uplifted in pulses while Loch Aline remains stationary and is eventually covered by Turneri basinal silty shales coming from the north where shoreface-shoreline environments can be distinguished.

The development of the above patterns depends on several variables, e.g. scarp/dip source, sediment loading at the landward margins, inclination of the basement and the presence of local structural highs (Sykes, 1975).

Sellwood and Jenkyns (1975) explained the importance of structural highs in controlling sedimentation in the English Jurassic. Sedimentation on the Skye-Ardnamurchan tilted block is greatly modified by the existence of a Central Strath High. The Milton Formation is completely absent in this region but the beds of the Strath Formation show a general westward thickening and their equivalent beds (*Semicostatum*) in the Loch Aline area are missing.

It has been shown that the thickness changes in the areas of study are directly dependent on the amount of clastic supply and the nature of tilting in any of the three (minor) blocks (see Figs. 12.10 and 12.11).

In the Skye area, proximity to a fault zone, abundant clastic supply and relative tilting towards the northeast are significantly reflected in the sedimentation style. Relatively offshore shallow depositional environments on a steadily shallowing basin (condensed sequence) with very little clastic supply are recognised in the Mull area. Slopes were gentle in both areas.

The depositional characteristics of the Scottish Lower Lias are intermediate between those recorded from areas proximal to "rift zone" sedimentation (e.g. East Greenland; Viking Graben; Moray Firth Trough (Sykes, 1975; later sections in this chapter) and the "platform" areas of the English Jurassic (Hallam, 1975; Sellwood and Jenkyns, 1975).

Marked fault activity, numerous unconformities, steep slopes, large-scale cross bedding, slumping, the generation of turbidites and the development of thick continental sedimentary facies such as those reported from areas close to "rift zones" during Rhaeto-Jurassic times are not seen in the Hebridean

Liassic succession; but evidence of a persistently stable shelf, with shallow marine conditions, reduced clastic input and the development of conditions of "platform" areas is also lacking. The situation is that of an environment transitional between the two mentioned, where the activity of independent tilted fault blocks against a general rise in sea level produced numerous local shallower and deeper water conditions. The sedimentary facies which developed was also greatly influenced by the sediment availability and input stimulated by fault activity. Despite the more obvious contrasts some similarity to what was described by Sykes (1975) can be seen in the tectonic style of the Lias in the Inner Hebrides, a rather thick succession of marginal marine facies accumulated in the subsiding areas of the tilted fault blocks whereas thinner more offshore sequences were developed on the flanks of the blocks. This differs from sedimentation in the "platform" areas where "basins" and "swells" were the basic controlling features which also produced expanded and condensed sequences (Sellwood and Jenkyns, 1975).

12.4 Regional geology and palaeogeography

The recent interests in hydrocarbon exploration in the northwest European continental shelf areas has meant that vast amounts of hitherto unknown details of the subsurface geology have become available. The consideration of such available data for northwest Europe, Greenland, eastern North America and northwest Africa will aid the development of a tectono-sedimentary depositional framework for the Lias of northwest Scotland.

i. East Greenland

The Jurassic of eastern Greenland has been recently studied in some detail (Surlyk et al., 1973; Birkelund and Perch-Nielsen, 1976). Rhaetic-Liassic coal-bearing fluviatile deposits (Kap Stewart Formation) are quite common in Jameson Land and Scoresby Land but they have not been reported with certainty north of Kong Oscars Fjord (Birkelund, 1975). The disappearance

of red beds from the lower boundary of the Kap Stewart Formation (cf. Wilderness in northwest Scotland) probably indicates a large-scale climatic change which is also known from other places in northern Europe, while an arid depositional climate continued into the Jurassic in other parts of the northern Atlantic (cf. the Nova Scotia Shelf; Jansa and Wade, 1975). This change in climate may be related to the establishment of a North Atlantic seaway during the Rhaetic but may also be caused by latitude shifts (Birkelund, 1975).

Investigations in the Kap Stewart Formation of southern Jameson Land by Sykes (1974a) has shown that the lower conglomeratic sandstones and arkoses represent alluvial fan deposits. The well developed fining upward sequences that overlie the conglomerates show coalified plant material, rootlets, coal beds and other primary sedimentary structures, indicating deposition in a low sinuosity non-braided river environment in close proximity to a south-southeast source area. Towards the north, the sediments become finer grained and suggest brackish-marine intercalations, the main depocentre was probably situated in Scoresby Land.

Clemmensen (1976) showed that the facies of the Kap Stewart Formation in Scoresby Land are quite different from those encountered in the type area in southern Jameson Land and reflect a change towards more marine conditions in the northern parts of the basin.

In contrast to the unimodal northward current pattern in southern Jameson Land (Sykes, 1974a) the palaeocurrent data from the Kap Stewart Formation in the northern region show a bipolar current pattern with the strongest mode towards the south; this pattern was suggested by Clemmensen (1976) to indicate tidal activity with a strong southward mode (it should be pointed out that in the absence of reliable ammonite data for establishing the synchronicity of the successive beds from which the bimodal palaeocurrent data was obtained, the reliability of the conclusion that they are of tidal

origin is questionable). The dominant mode points to a northerly source area and the existence of transverse fault lines in the Kong Oscar Fjord which may have been reactivated in early Jurassic times (Surlyk et al., 1973). Surlyk (1977) proposed a general model for Jurassic sedimentation and tectonism in east Greenland, a summary of which follows. The initial basin was formed by faulting along north-south trending lines accompanied by slight westward rotation of the fault blocks; thus a number of north-south elongated depositional troughs were formed. This pattern had started in the early Carboniferous and continued in the late Palaeozoic and Mesozoic (Vischer, 1943; Haller, 1971). The trough areas were oriented in a north-south fashion and were located over the western down-tilted parts of the blocks, whereas the eastern block margins formed mountain ridges, elongated islands, peninsulas or submarine shoals, depending on the degree of submergence. Subsidence took place mainly by gradual movements along the faults, only in some periods interrupted by strong fault activity (e.g. at the Jurassic/Cretaceous boundary). The general north-south faulting pattern was cut by a number of northwest-southeast trending faults which downthrow to the south.

In contrast to the north-south fault system, this system seems to have been activated in a series of violent movements reflected by the pronounced stepwise reductions in thickness and disappearance of formations when passing the fault zones from south to north.

It is interesting to note that Triassic dredge samples obtained from 64°N, 35°W contain volcanic debris (Johnson et al., 1975). The age of dredge samples can at best be regarded as tentative but the occurrence of such volcanism may be explained in view of the early Mesozoic evolution of northern North Atlantic continental margins and may also provide some clues as to the origin of smectites in the early Liassic shales of northwest Scotland (Amiri-Garroussi, 1977).

ii. Spitsbergen, the Svalbard Archipelago and the Barents Sea

Harland et al. (1974) have not reported the occurrence of any Hettangian-Sinemurian deposits in Spitsbergen.

Extensive recent work has been carried out on the geology of this region and detailed reviews may be found in publications by Ronnevik et al. (1975), Smith et al. (1975), Smith (1975), Smith et al. (1976), Talwani and Eldholm (1977).

A probably continuous succession from Karnian or Norian into the earliest Jurassic has been established for Hopen and Svalbard.

The Rhaeto-Hettangian Lyngefjellet Sandstone Formation (Smith et al., 1975) of Hopen Island consists of interbedded fine to medium coarse sandstones which are in places pebbly, showing cross bedding and ripple marks; they contain plant fragments and deposition under fluvial conditions was suggested.

The existence of mid-Triassic to late Jurassic relatively flat lying marine and non marine sediments of Wilhelmøya and Hellwaldfjellet in eastern Svalbard has been described by Smith (1975). It was discovered that Norian-?Rhaetian beds (De Geerdalen Formation) thicken from 190 m. at the type section in Spitsbergen to at least 370 m. in the more eastern Edgeøya. A thickening of the equivalent strata towards the north (up to 503 m. in Wilhelmøya and 506 m. in Hellwaldfjellet) and south (460 m. in Hopen) was also seen.

These thicker deposits probably include a more complete representation of this poorly known interval in Svalbard between the marine faunas of the mid-Triassic and those of the late Toarcian beds. Although a substantial thickness of Norian deposits was proved for Wilhelmøya and probably for Hellwaldfjellet, breaks in the sequence above this level and the lack of any faunal or floral evidence suggests the non representation of the Rhaetic (and probably Hettangian and Sinemurian) deposits in this area.

The substantial work by Smith et al. (1976) has provided a greater understanding of the geology of Kong Karls Land of eastern Spitsbergen.

The Svenskøya Formation was the name given to dominantly sandy continental beds at the base of the succession in north and south Svenskøya and also in parts of Kongsøya. It is similar in the development of various facies to parts of the De Geerdalen Formation of Spitsbergen. From recent palynological investigations made by the above mentioned authors, the lower part of the Mohnhøgda Sandstone Member (Svenskøya, Kong Karls Land) and Sjøgrenfjellet Sandstone Member (western Kongsøya, Kong Karls Land) is Rhaetian and the main part in each case is probably Hettangian to Sinemurian. The sandstones of the Members are cross bedded and ripple marked in places, containing pieces of petrified wood; thin pebble horizons are also seen. The top of the Mohnhøgda Sandstone Member (?Sinemurian) is marked in places by a thin bed of gravel conglomerate with an apparently eroded top. Stringers and interbeds of clay and shales are seen and rare coal and ironstones are also present in the Mohnhøgda Member. Thin coal seams, plant fragments, wood and lignite are seen in the Sjøgrenfjellet Member together with some slump structures. There is little doubt that the Rhaetic-Hettangian-Sinemurian sediments represent predominantly continental depositional environments; the loose porous sandy facies also suggest no great depth of burial which is also confirmed by the state of preservation of the palynomorphs (Smith et al., 1976; Hughes et al., 1976).

The geology of the Barents Sea off the northern Norwegian continental shelf has been investigated by Ronnevik et al. (1975), Sundvor (1975), Talwani and Eldholm (1977).

Talwani and Eldholm (1977) show that the sedimentary sequence in the Barents Sea consists mainly of Mesozoic and late Palaeozoic rocks. The Barents Sea was divided into two main areas, i.e. the sedimentary basins and the Svalbard Platform. The basins consist of the North Cape Basin, trending east between Norway and Bear Island and the Novaya Zemlya Basin, trending northeast along the west coast of Novaya Zemlya. These basins merge in the southeastern Barents Sea with the Pechora Basin extending from mainland

Russia and they are believed to contain a thick succession of sediments of Mesozoic and Palaeozoic age. The regional sedimentary basins are composed of several sub-basins and ridges and the sediments are considerably thinner beneath the Svalbard Platform where the main portion of the Mesozoic-Cenozoic is probably missing. The above mentioned authors also found that except for north-trending lineaments near the western shelf edge and partly on the Svalbard Platform, the main structural fabric trends northeast. During pre-Tertiary times the area of the Barents Sea developed as a shallow sea in which deposition kept pace with subsidence. Regionally it was part of an epicontinental sea that extended from the North Sea between Norway and Greenland and onto the Russian Platform. The areas of maximum subsidence seem to have been stationary in time (Talwani and Eldholm, 1977).

Hinz and Schluter (1978) studied the western Barents Sea and found that in the north-northeast trending Transø Basin, Jurassic sediments were deformed due to halokinetic movements. Triassic-Middle Jurassic sediments are also found overlapping Palaeozoic sediments in a basin between the western Transø and Bear Island basins, the Spitsbergen Platform in the north and the North Cape Basin in the east.

iii. Norwegian Continental Shelf

The Norwegian continental shelf was subdivided into three major areas by Ronnevik et al. (1975).

<u>Major geological segment</u>	<u>pre-Tertiary sub-basin</u>	<u>Highs</u>
The North Sea	East Shetland basin	Vestland Ridge (Arch)
	Viking Graben (Trough)	
	Central Graben (Trough)	Ringkøbing-Fyn High
	Danish-Norwegian Basin (Embayment)	
Møre-Lofoten	Møre Basin	Nordland Ridge
	Vøring Basin	
	Helgeland Basin	

Barents Sea

Hammerfest Basin

Tromsø Basin

Senja Ridge

Harstad Basin

The Permian-Jurassic structure and evolution of the northern North Sea was dominated by tensional tectonic stresses (P.A. Ziegler, 1975a; Kent, 1975b) which culminated in the creation of a central rift zone.

The sedimentological development in the Norwegian North Sea during the Lower Jurassic is related to a major transgression which gave rise to a shaly facies in most of the Lower Jurassic strata overlying the Triassic continental red bed sequences (Ronnevik et al., 1975; see also Michelsen, 1977).

The Møre-Lofoten segment also shows the development of onlapping sedimentary sequences (Ronnevik et al., 1975) characteristic of epeirogenically subsiding basins (this follows a period during which sedimentation occurred in block-faulted basins and major rift systems were established). Among the pre-Tertiary sub-basins, the Møre basin can be traced into the Viking Graben; one branch may also continue into the West Shetland Basin (Ronnevik et al., 1975). The Møre and Vøring Basins are interpreted as resulting from the same rifting that created the Viking Graben and the West Shetland Basin. This rifting is thought to have started in the first half of the Jurassic as in the North Sea (P.A. Ziegler, 1975a; W.H. Ziegler, 1975).

The Barents Sea segment comprises the main geological basins situated in the area between the Svalbard Bank and the Norwegian mainland. The data of Ronnevik et al. (1975) suggests that in the Barents Sea, the Senja Ridge is related to a north-south fault system together with the Tromsø Basin (east of Senja Ridge). The Hammerfest Basin is related to a north-east-south-west fracture system but remained an epeirogenic basin through most of its history. With the exception of the Hammerfest Basin, the tectonic setting is quite similar to that found by Sokolov (1965) on Svalbard; for this reason Ronnevik et al. (1975) suggested that the proposed model could be extended to

include the whole of the Barents Sea. It was also suggested that in the predrift situation, probably the Old Devonian north-south fault system (Harland, 1969) became activated and relaxed the regional tension.

Important Cimmerian tectonic movements occurred in the Norwegian parts of the North Sea during the whole of the ?Rhaetic-Jurassic. At least three tectonic phases have been reported by W.H. and P.A. Ziegler (1975) as follows:

1. (i. Keuper/Rhaetic boundary
 - (ii. Early/Middle Jurassic
 - (iii. Middle Jurassic
2. (iv. Middle/Late Jurassic boundary
 - (v. Late Jurassic boundary
3. (vi. Jurassic/Cretaceous boundary

It is not known whether the Cimmerian tectonics only occurred as rejuvenation of movements along pre-existing fault systems or were in fact related to new systems (Al-Kasim and Ronnevik, 1975).

The early Jurassic of offshore Norway is marked by a major transgressive phase. At the end of the Keuper, the Norwegian part of the North Sea was a continental sedimentary basin with some intrabasin source areas (W.H. Ziegler, 1975). The early Cimmerian phase at the beginning of the Rhaetic resulted in the creation of sub-basins that were invaded by early Jurassic marine transgressive deposits (Al-Kasim and Ronnevik, 1975) and in addition to the Norwegian mainland, other source areas for the sediments were in existence as structural highs on the Vestland Arch and the Ringkøbing-Fyn High and the east part of the Tampen Spur (Ronnevik et al., 1975).

Sediments of early Jurassic age have been observed in the central Trough, the northern part of the Norwegian - Danish Basin and the East Shetland Basin. They are missing on the structural highs making up the Vestland Arch.

The sediments in general consist of marine sandstones, shales and

siltstones together with conglomerates and ?chamosite-siderite oolites (Young et al., 1975) with minor amounts of limestones which are disturbed by post early Jurassic halokinetic movements in the southern part of the Norwegian Central Trough (Al-Kasim and Ronnevik, 1975).

In the East Shetland Basin sediments of early Jurassic age are found in the Viking Trough and the Stratfjord area; the observed thicknesses vary from 50 to 350 m. The lithostratigraphy of the Lower Jurassic consists of two units, the lower of which is known as the Stratfjord Sand Formation (Bowen, 1975; Deegan and Scull, 1977a, b). This is a fluvial sand sequence ranging in age from late Rhaetian to early-mid-Sinemurian. It averages about 200 m. in the Stratfjord Field and consists mainly of white poorly sorted fine to very coarse and conglomeratic, kaolinitic sand; a lowermost braided stream environment was succeeded by sheets of coastal barrier sands (Bowen, 1975; see also Jones et al., 1975).

In the Danish North Sea, Triassic red beds were deposited in the Danish Embayment together with the Central Graben. Seismic reflection data suggest that the Horn Graben was developed mainly during the Triassic (Childs and Reed, 1975). By early Jurassic time, fully marine conditions were established throughout the basinal areas where Lower Jurassic sediments are represented by dark grey to greyish brown, shallow marine shales, with some very fine sandstone and siltstones.

iv. Northeast Scotland and the Moray Firth

The lowermost Jurassic in northeast Scotland is exposed in the Golspie/Helmsdale area where the Jurassic rocks of the Brora outlier represent the exposed western margin of a large sedimentary basin which extends eastwards beneath the Moray Firth.

The Lower Liassic succession is best exposed on the foreshore southeast of Dunrobin Castle. The Dunrobin Bay Formation (Neves and Selley, 1975) consists of three distinct units the lower of which comprises carbonaceous

siltstones and clays (Hettangian-Sinemurian) signifying essentially fresh-water conditions although some palynological evidence shows the middle of the unit to be marine (Lam and Porter, 1977). The white sandstones succeeding the carbonaceous shales are barren but simple microplankton taxa indicate some open sea influence in what is possibly an estuarine situation (Neves and Selley, 1975). The Lower Liassic succession is clearly a shallow water one and a product of the progressive inundation of the basin margin which occurred from Hettangian to the lowest Pliensbachian.

Although the Geological Survey borehole near Lossiemouth was drilled on the south side of the Moray Firth and very near the Golspie/Brora succession, ammonites in the upper parts of the beds indicate a *Raricostatum* Zone (Berridge and Ivimey-Cook, 1967) and equivalents of the Hettangian-Sinemurian beds have not been determined with certainty.

It is important to note that while carbonaceous sandstones and siltstones occur above the ammonite horizons in the Lossiemouth borehole indicating a return to freshwater conditions, this contrasts with the dark grey siltstones and thin argillaceous limestones found in Dunrobin Bay. A regression due to elevation on the southern margin of the Moray Firth basin was proposed as a probable explanation by Neves and Selley (1975).

The general similarities of the Liassic succession in Sutherland and their development in southern Sweden has been pointed out by Judd (1873), Hallam (1965) and Sellwood (1972). The coal-bearing strata of the lowermost carbonaceous siltstones and clay unit may be compared with the thicker (200 m.) Helsingborg Formation (Lower Hogönas Series). The White Sandstones unit (barren) and Doshult Formation are both marine and of a Sinemurian age. Upper Sinemurian deposits are not present in Scania due to uplift and regression (Neves and Selley, 1975).

The study of Brooks and Chesher (1975) and Bacon and Chesher (1975) concluded that during the Lower Jurassic, the continental conditions that prevailed in the northern North Sea in Triassic times gave way in the Moray

Firth Basin to a littoral marine environment. Sediment deposition varied between marine and non marine and it was suggested that the Moray Firth Basin formed an isolated basin during the Lower Jurassic with marine access to the west coast of Scotland along the line of the Great Glen Fault, between the Grampian and Northwest Highland massifs. More continental conditions prevailed at this time in the central parts of the Viking Graben with the continuation of red shales and sandstone deposition from the Trias. More argillaceous sediments representing a shallow marine environment were deposited on top of these beds during the end of early Jurassic times with a probable marine access from the north.

v. Eastern, Central and Southern England

The nature and mode of development of the Triassic-Rhaetic-early Jurassic sediments in Yorkshire is broadly similar to the other regions already discussed in that the Keuper Tea Green Marls were deposited under coastal plain depositional conditions and are underlain by Triassic fluvial sediments and overlain by Hettangian marine strata representing open marine environments.

In central and southern England evidence has shown that the older parts of the New Red Sandstone was deposited in a series of grabens (Audley-Charles, 1970; Whittaker, 1975), although tectonism decreased towards the end of Triassic times several phases of renewed subsidence are apparent. Marginal marine depositional environments persisted during the Rhaetic and were succeeded by the establishment of fully marine conditions in the Hettangian (Blue Lias).

In general deposition occurred in a series of transgressive, regressive shelf-shoreface sequences. At the base transgressive shallow marine shelf deposits were overlain by shales and limestones (e.g. Blue Lias) indicating deeper marine depositional conditions.

Tectonically the area was relatively quiet and fault-controlled subsidence

may have occurred in grabens near the Hercynian Front (Whittaker, 1975) and other "fractures" (Jenkyns and Senior, 1977). Disconformities developed only in the proximity of swells, i.e. persistent structural highs with only condensed consequences which separate thicker basinal successions (Sellwood and Jenkyns, 1975).

vi. Irish Channel and Continental Shelf

The marine geology of the Slyne Ridge (Bailey et al., 1977), Irish continental margin (Bailey, 1975), the Irish Sea (Wright, 1975), the Celtic Sea and southwestern Approaches (Blundell, 1975) and the North Irish continental shelf (Gerard and Boillot, 1977) have been investigated recently. Geophysical work shows that in the Seabight Trough 5 km. or more Mesozoic to Recent sediments rest on thinned, "quasi-oceanic" crust and the trough is flanked from three sides by drowned continental crust. The Slyne Ridge has sustained some Mesozoic rifting but is essentially a drowned east-west ridge of ancient Caledonian crystalline basement rock, bounded on the south by an old line of structural weakness. In the Bristol Channel up to 1500 m. of Jurassic clays with thin cementstone beds, forming the core of an east-west trending syncline, overlies Permo-Triassic sandstones and marls. The Slyne Ridge is regarded as an extension of a crustal swell in which the Precambrian metamorphic rocks are exposed and on its marine perimeters Caledonoid lines of crustal weakness became the loci of Mesozoic movements. The Slyne Trough is infilled by Triassic-Jurassic sediments. Although Middle Jurassic volcanism has been reported from the Forties and Piper oil fields in the North Sea (Howitt et al., 1975; the age is disputed by Faereth et al., 1976), no volcanic rocks have been reported from the Irish or Scottish Hebridean basins. Upper Jurassic (U. Callovian-L. Kimmeridgian) volcanism has been reported by Knox (1977) in northwest Scotland.

The presence of ?Permo-Triassic to Jurassic formations in the northern Irish Sea is inferred due to the continuation of the strata with the outcrops

of the Cheshire and Carlisle basins. The Bunter Group mainly consists of sandstones with pebbly horizons at some localities whereas the Keuper Group are mainly sandstones which are succeeded by gypsiferous silty mudstones with thick beds of halite. Little is known about the offshore extent of the Permo-Triassic beds, and the thickness variation of these beds together with those beds of Jurassic age which are present. However, some information (albeit inconclusive) is given by Wright (1975).

In the southern Irish Sea knowledge of the offshore Mesozoic bedrock geology is concentrated on the basins of Kish Bank, Tremadoc Bay, Cardigan Bay and St. Georges Channel.

Kish Bank Basin

The existence of this Basin is mainly inferred from the data of Bott and Young (1971) and Dobson et al. (1973) and although a lower Mesozoic base is suggested, no rock samples have been obtained so far.

Tremadoc Bay Basin

This Basin is bound on its eastern margin by the Mochras fault which drops Tertiary and Mesozoic rocks at least 4500 m. (Woodland, 1971) to the west against Cambrian rocks on the Welsh coastline. The Mochras borehole (Lewis, 1972) proved a thickness of 1305 m. of Liassic and 32 m. of Triassic rocks. The Liassic sequence comprises uniform mudstones and siltstones with uniform limestones in the lower part. These indicate uninterrupted uniform marine deposition with no contemporaneous movement on the adjacent Mochras fault. The Upper Triassic sediments comprise calcareous sandstones and siltstones with impure limestones and dolomites which are of ?shallow water origin and are different from the rocks of the same age elsewhere in England (Wright, 1975).

Cardigan Bay

The basin fill consists of Permo-Triassic and Jurassic strata with no Tertiary cover, the thickness of which is not known. The northern side of the

Jurassic outcrops shows grey mudstones of early Liassic (Sinemurian) age; at the southeastern margin of the basin grey mudstones of Upper Sinemurian age were found. These are overlain by grey argillaceous shelly limestones of Middle Jurassic (Bathonian) age and Dobson et al. (1973) suggested that the Middle Jurassic rested unconformably on the Lias cutting down to the Lower Lias in some places. Evidence suggests a uniform shallow marine depositional environment through the Permo-Triassic and Jurassic.

St. Georges Channel Basin

Although no direct evidence is given, Dobson et al. (1973) and Wright (1975) have suggested that in this basin the Tertiary may be underlain by Mesozoic formations such as those known from the Bristol Channel and Celtic Sea areas to the south.

12.5 Tectonic considerations

The tectonic framework proposed for the evolution of the North Sea area and northwestern Europe suggests the development of a Permo-Triassic "intra-cratonic" stage followed by a Jurassic-Cretaceous taphrogenic rifting stage (Ziegler, 1975a, b; Kent, 1975a, b, 1977b; see also Pegrum and Mounteney, 1978; and models proposed for the development of rifted continental margins by Flavey, 1974; Veevers and Cottrill, 1978). In many instances basin subsidence was dominantly fault controlled, but it was also intermittent and not universal within a single region so that some areas (e.g. the broad southern North Sea basin) at the same time were undergoing simple downwarps (Kent, 1976).

Ziegler (1975) explained that during the Permo-Triassic intracratonic stage, following the Variscan Orogeny, large parts of the North Sea were occupied by the rapidly subsiding intracratonic northern and southern Permian basins in which thick clastics and evaporites accumulated. The development of new graben systems in the Trias which effectively modified the Permian

structural pattern, is thought to be related to early rifting movements along the Arctic-North Atlantic rift zone (Haller, 1971; Hallam, 1971a).

Rapid differential subsidence occurred during the Triassic in the Viking Central Graben system of the North Sea. Other similar subsiding Triassic grabens are mentioned by the above named authors which include those developed along the Atlantic seaboard of Scotland and Ireland as well as those in the Celtic Sea and Western Approaches. Such "intracratonic" development of Triassic basins in the Scottish Hebrides has been explained in great detail by Steel (1971a), Steel and Wilson (1975) and Steel et al. (1975). The development of the North Sea rift system started during the Triassic and dominated the palaeogeographic setting of the area during the Jurassic and Cretaceous (taphrogenic stage). The evolution of the North Sea rift was related to the development of the Arctic-North Atlantic rift zone (Ziegler, 1975; Kent, 1975a, b). The latter reached the stage of crustal separation in the early Tertiary (Ziegler, 1975).

Hallam and Sellwood (1976) concluded that regional tensional tectonic stresses were important in the development of the various sedimentation patterns in the Triassic and Jurassic of Britain. These were created by either vertical faulting which occurred in the Hercynian basement or by more extensive upwarping and graben collapse of continental crust associated with stretching (this may or may not have been followed by sea floor spreading).

Stewart (1971) and Kellog (1975) explained that graben tectonics generally occur in regions of transextension. They may also be characteristic of normal extensional stresses in an area. As transform faults were not a part of the structural style in the British-Greenland Mesozoic, it has been assumed that regional extension governed the main rift structure (Sykes, 1975). Although this fault-controlled subsidence is typical of the initial phases of continental collapse as seen in many miogeosynclines, it is not possible to recognise the three characteristic sedimentation phases (i.e.

orthoquartzite-carbonate, flysch and molasse suites) of Mitchell and Reading (1972) in the Liassic sediments of northwest Scotland, northeast Atlantic and the Greenland-Norway areas. Such a resemblance may be suggested for overall Jurassic sedimentation in the mentioned areas (Sykes, 1975).

The large-scale Mesozoic vertical movements in the North Sea, the Hebridean areas and northwest Europe in general, are indicative of extensional tectonic activity which developed on continental crusts and was associated with relatively sparse volcanism. The above features were emphasised by Sykes (1975) who compared the Mesozoic tectonic development of northwest Europe with the features of an aulacogen-type trough (Hoffman et al., 1974). These are transverse linear troughs showing dominantly vertical tectonic movements in which rifting was insufficient to produce much oceanic crust and little or no subduction occurred.

The major Mesozoic sequences which developed in southern Jameson Land were compared with those reported from the Athapuscow aulacogen in the Precambrian of northwest Canada (Sykes, 1975) and it was shown that they were deposited in an aulacogen-type trough or "rift area" (Salop and Sheinman, 1969), which revealed the characteristic evolutionary stages of triple junction development (Burke and Dewey, 1973; Burke and Whiteman, 1973).

Initial extensional stresses in the late Permian and Triassic marked the commencement of the tectonic sequence and grabens were formed over a wide area of Greenland, Britain, eastern U.S.A., northeast Canada, Labrador Sea, northwest Africa (Hallam and Sellwood, 1976; Kent, 1976, 1977a; Naylor and Mounteney, 1975; Bott, 1976; Ballard and Uchupi, 1975; Rodgers, 1970; Burke, 1977; Sheridan, 1976; Jansa and Wade, 1975; Strong and Harris, 1974; Strong, 1977; papers in Yorath et al., 1975; van Houten, 1977; van Houten and Brown, 1977). Contemporaneous faulting resulted in periodic rejuvenation of the drainage systems which was in turn reflected by coarsening and fining upwards fluvial megasequences. This occurred in areas which were near the

rift zones where faults remained active and abundant clastic debris was carried into the basins. For example the graben stage of southern Jameson Land (Triassic-Jurassic) as described by Sykes (1975) shows that deposition took place within the basin in fluvial, coastal plain and marginal marine environments throughout the Trias. During the Rhaetic-Hettangian rejuvenation affected the area with alluvial fan and floodplain deposits forming. A widespread disconformity is also recorded in the Sinemurian. This pattern of deposition is not seen in areas furthest from the faults where a somewhat more uniform marine sequence developed (e.g. northwest Scottish Lias).

The development of aulacogens is closely linked with the hypothesis of "trilete rift arm" system proposed for the North Sea area by Whiteman et al. (1975), although it was suggested that this trough system originated in Hercynian times (ibid.). If so there seems little doubt that it was being reactivated during Triassic-Jurassic times (Ziegler, 1975a).

Talwani and Eldholm (1977) found that in the northeast Atlantic, the limit of Mesozoic rocks appears to be symmetrical to the boundaries of the Caledonian Front; the Mesozoic basin continues north from the North Sea and continues through the Greenland Shelf north of Scoresby Sund. They also noted that the time of initial opening lies well within the Mesozoic basin. The time of formation of the Mesozoic grabens however is not seen to coincide with the time of opening but predates it by a large interval. There is a similarly large interval between the time of formation of the Triassic grabens along the east coast of north America and the subsequent initiation of sea floor spreading that separated north America and Africa.

It is also interesting to note that the western Barents Sea was already divided into basins and ridges in Palaeozoic times. The Svalbard Platform acted as a more stable area and has subsided less since Palaeozoic time than the areas in the south where subsidence and sedimentation varied in space and time (Hinz and Schulter, 1978).

The Permo-Triassic horst-graben topography seen in the northeast Atlantic

was probably caused by the initial tectonic stresses which were precursors to the establishment of the now extinct triple junction south of Greenland (Kristoferson and Talwani, 1977).

Based on a general consideration of Laurasian global tectonics at the time of volcanism in the North Sea, a model was proposed by Gibb and Kanaris-Sotiriou (1976) according to which the volcanic activity signified developments along a line of incipient subcrustal rifting during the early opening of the Atlantic. They explained that before crustal spreading took place the spreading axis probably migrated westward and was responsible for the volcanism which developed in the Northwest Tertiary Igneous Province. Although the speculative nature of the model was mentioned by the above authors, it should be stressed in addition that the age relationships of such a westward migration is somewhat inconsistent with the development of extensive Permo-Triassic basins east and south east of Greenland. Moreover recent seismic work has provided convincing evidence for sedimentary troughs of Mesozoic age on the Hebridean continental margin (Jones, 1978).

APPENDIX 1

PETROGRAPHY

A total of 342 thin sections were prepared from the samples. Standard petrographic techniques were followed for the study of compositional and textural variations in the different lithologies.

Quantitative analyses were made using the Glagolev-Chayes method for studying thin sections (Galehouse, 1971). Point counting was carried out using an automatic point counter manufactured by James Swift and Son Ltd., and the grain sizes were estimated using a micrometer eyepiece.

The data obtained by point-counting 150 selected samples is presented in pages 1-2 to 1-30; it should be pointed out that this data was not used independently in order to reach a conclusion and was only considered together with the field data.

Slide no.	Apc 1		Apc 2		Apc 3		Apc 4		Apc 5	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	12	1.78	9	1.16	-	-	-	-
Spar	125	19.53	465	68.99	-	-	44	8.16	1	0.19
Recryst. cal.	2.38	37.19	13	1.93	431	55.33	429	79.59	273	50.19
Ech. & crin.	1	0.16	4	0.59	2	0.26	2	0.35	1	0.19
Bivalves	36	5.63	15	2.23	13	1.67	46	8.53	66	35.82
Gastropods	-	-	2	0.29	4	0.51	1	0.19	-	-
Brachiopods	-	-	-	-	-	-	3	0.56	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	2	0.31	35	5.19	5	0.64	2	0.35	1	0.19
Ooliths	219	34.22	108	16.02	58	7.45	9	1.69	192	12.30
Pellets			1	0.15						
Algae	-	-	✓	✓	18	2.31	-	-	-	-
Qtz. (und.)	14	2.19	4	0.59	208	26.70	2	0.35	-	-
Qtz. (polyx.)	-	-	-	-	-	-	-	-	-	-
Qtz.	-	-	-	-	-	-	-	-	-	-
Felds. P=Plag. O=Or.	2	0.31	-	-	8	1.03	-	-	-	-
Mica	-	-	-	-	-	-	-	-	1	0.19
Opagues	3	0.47	15	2.23	20	2.57	34	6.31	1	0.19
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.3-0.4		0.3		0.16- 0.74		-		0.32	
Total	640	100	674		779		572		536	
Name	Bio oosparite wackestone		Oobiospary micrite		Sandy Biopel-micro sparite		Biosparite wackestone		Bio oosparite wackestone	

Slide no.	Apc 6		Apc 8		Apc 12		Apc 13		Apc 14	
	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	28	8.54			450	73.05	1	0.18	-	-
Spar	144	43.90			6	0.67	244	44.6	4	0.74
Recryst. cal.	36	10.98			40	6.49	168	30.71	328	60.97
Ech. & crin.	-	-			1	0.16	2	.37	11	2.04
Bivalves	14	4.27			22	3.57	104	19.01	81	15.06
Gastropods	-	-	-	-	4	0.65	4	0.73	5	0.93
Brachiopods	1	.30	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	2	0.37	-	-	2	0.37
Ooliths	97	29.58	-	-	47	7.63	✓	✓	11	2.04
Pellets	-	-	-	-	13	2.11	-	-	62	11.52
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	-	-	-	-	-	-	-	-	-	-
Qtz.	2	0.61	-	-	8	1.30	10	1.83	-	-
Felds. P=Plag. O=Or.	-	-	-	-	-	-	3	0.55	-	-
Mica	-	-	-	-	-	-	-	-	-	-
Opagues	6	1.83	-	-	23	3.73	12	2.19	34	6.34
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.					0.3-0.4, ool 0.1-0.2, qtz		0.16- 0.2		0.1-0.3	
Total	328				616		547		538	
Name	Oobio microspar- ite wackestone		Biopel sparite + dismicrite		Sandy oobio- micrite wackestone		Sandy bio oo pel microspar- ite wackestone		Bio- pelsparite	

Slide no.	Apc 17		Apc 18		Apc 20		Apc 21		Apc 23	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	355	48.83	319	48.41	637	77.86	162	75	343	58.14
Ech. & crin.	5	0.69	5	0.76	-	-	4	1.85	3	0.51
Bivalves	1	0.14	14	2.12	32	3.62	17	7.87	14	2.37
Gastropods	-	-	-	-	3	.34	5	2.31	-	-
Brachiopods	1	0.14	1	0.15	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	21	9.72	1	0.17
Ooliths	2	0.28	3	0.46	163	18.43	-	-	1	0.17
Pellets	-	-	81	21.29	-	-	-	-	52	8.81
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	-	-	-	-	-	-	-	-	-	-
Qtz.	293	40.30	205	31.11	-	-	-	-	160	27.12
Felds. P=Plag. O=Or.	9	1.24	7	1.06	-	-	-	-	4	0.68
Mica	6	0.83	2	0.30	-	-	-	-	4	0.68
Opauques	6	0.83	4	0.61	-	-	7	3.24	4	0.68
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.08-0.1		0.04-0.08		0.5		-		0.04-0.14	
Total	727		659		885		216		596	
Name	calc. siltstone		Silty bio oipel sparite		Oomicro sparite wackestone		Bio micro sparite		Silty bio oipel sparite	

Slide no.	Apc 24		Apc 25		Apc 26		Apc 28		Apc 30	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	1	0.18			-	-
Spar	9	1.70	1	0.25	-	-			-	-
Recryst. cal.	304	57.51	270	67.67	189	33.16			349	58.45
Ech. & crin.	14	0.265	8	2.01	-	-			5	0.83
Bivalves	38	0.71	37	9.27	-	-			10	1.67
Gastropods	-	-	1	0.25	-	-			-	-
Brachiopods	2	0.37	2	0.5	-	-			-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	8	1.52	-	-	-	-			-	-
Coliths	58	10.98	44	11.03	-	-			15	2.51
Pellets			15	3.76	-	-			-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-			-	-
Qtz. (polyx.)	-	-	-	-	-	-			-	-
Qtz.	90	17.04	9	2	347	60.88			194	32.42
Felds. P=Plag. O=Or.	4	0.75	-	-	18	3.16			15	2.51
Mica	-	-	-	-	-	-			4	0.67
Opagues	1	0.18	4	1	15	2.63			5	0.83
Epidote	-	-	-	-	-	-			-	-
Zircon	-	-	-	-	-	-			-	-
Max. gr. size mm.	0.04		0.8-1						0.08	
Total	528		399		570				599	
Name	Silty biopel- sparite		Bio oopel- sparite		Calc siltstone		Silty biomicro sparite		Silty biopel microsparite	

Slide no.	Apc 33		Apc 35		Apc 36		Apc 37		Apc 41	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-			27	5.103	-	-
Spar	2	0.35	1	0.22					-	-
Recryst. cal.	306	54.06	231	51.67			263	49.716	283	47.40
Ech. & crin.	2	0.35	5	1.11			4	0.756	-	-
Bivalves	-	-	4	0.89			19	3.591	8	1.34
Gastropods	-	-	-	-			-	-	-	-
Brachiopods	-	-	-	-			1	0.189	2	0.34
Corals	-	-	-	-			-	-	-	-
Foram. & ost.			5	1.11			1	0.189	-	-
Ooliths	2	0.35	26	5.81			4	0.756	-	-
Pellets	1	0.17	19	4.25			34	6.427	2	0.34
Algae	-	-	-	-			-	-	-	-
Qtz. (und.)	-	-	✓	✓			-	-	-	-
Qtz. (polyx.)	-	-	✓	✓			-	-	-	-
Qtz.	241	42.57	137	30.64			170	32.136	283	47
Felds. P=Plag. O=Or.	6	1.06	15	3.35			-	-	17	2.85
Mica	2	0.35	2	0.44			5	0.945	-	-
Opakes	4	0.706	2	0.44			-	-	2	0.34
Epidote	-	-	-				-	-	-	-
Zircon	-	-					-	-	-	-
Max. gr. size mm.	0.08		0.12				0.08		0.04-1	
Total	566		447				529		597	
Name	Sandy biomicro sparite		Sandy bio oo micro sparite		Dismicrite		Sandy biopelspar- ite		Silty bio pelsparite	

Slide no.	Ra 27		Ra 33		Ra 34A		Ra 34B		Ra 34C	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	374	57.63	441	73.01	122	44.36	94	39.5	100	58.48
Ech. & crin.	2	0.30	-	-	-	-	27	11.34	33	19.30
Bivalves	17	2.69	3	0.50	-	-	-	-	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	5	0.77	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	-	-	-	-	-	-	-	-	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	11	1.69	-	-	-	-	-	-	-	-
Qtz. (polyx.)	25	3.85	-	-	-	-	-	-	-	-
Qtz.	186	28.66	132	21.85	139	50.55	96	40.34	30	17.54
Felds. P=Plag. O=Or.	20	3.08	4	0.66	-	-	2	0.84	-	-
Mica	-	-	3	0.50	-	-	1	0.42	1	0.58
Opagues	9	1.39	21	3.48	14	5.09	18	7.56	7	4.00
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.2		0.04-0.08		0.08		0.06		0.06	
Total	649		604		275		238		171	
Name	Calcareous fine sand		Micaceous calcareous siltstone		Calc. siltstone		Calc. siltstone		Silty biosparite	

Slide no.	Ra 1 - 2		Ra 2		Ra 3		Ra 4		Ra 5	
	no.	%	no.	%	no.	%	no.	%	no.	%
Constituents										
Micrite	481	83.22	-	-			-	-	-	-
Spar	37	6.46	-	-			-	-	-	-
Recryst. cal.	-	-	428	55.37			211	31.87	753	86.25
Ech. & crin.	13	2.25	-	-			-	-	18	2.06
Bivalves	18	3.11	-	-	✓		-	-	45	5.15
Gastropods	4	0.69	-	-			-	-	-	-
Brachiopods	12	7.08	-	-			-	-	-	-
Corals	-	-	-	-			-	-	-	-
Foram. & ost.	2	0.35	-	-	✓		-	-	2	0.23
Ooliths	-	-	-	-			-	-	4	.46
Pellets	2	0.35	-	-			-	-		
Algae	-	-	-	-			-	-	-	-
Qtz. (und.)	-	-	20	2.62			18	2.72	-	-
Qtz. (polyx.)	-	-	119	15.58			27	4.08	-	-
Qtz.	-	-	162	21.20			354	53.47	-	-
Felds. P=Plag. O=Or.	-	-	36	4.71			27	4.08	-	-
Mica	-	-	1	0.13			9	1.36	-	-
Opakes	9	1.56	3	0.39			8	1.21	51	5.84
Epidote	-	-	-	-			-	-	-	-
Zircon	-	-	-	-			-	-	-	-
Max. gr. size mm.	2 - 5		0.2		-		.08			
Total	578		764		-		662		873	
Name	Biomicrite		Calcareous sandstone		Biomicrite		Calcareous siltstone		Bio pelsparite	

Slide no.	Ra 8		Ra 9		Ra 10		Ra 18		Ra 20	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	12	0.14	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	446	74.96	585	66.10	285	5.23	199	54.22	290	44.14
Ech. & crin.	37	56.64	-	-	-	-	68	18.53	-	-
Bivalves	54	9.08	-	-	10	1.94	11	3.00	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	4	0.67	-	-	-	-	8	4.02	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	180	20.43	134	46.91	-	-	-	-
Pellets	18	3.03	108	12.20	87	16.86	-	-	60	9.13
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	2	0.54	8	1.22
Qtz. (polyx.)	20	3.36	-	-	-	-	4	1.08	-	-
Qtz.	-	-	-	-	-	-	58	15.80	217	33.03
Felds. P=Plag. O=Or.	-	-	-	-	-	-	2	0.54	7	1.07
Mica	-	-	-	-	-	-	-	-	39	5.94
Opaquea	16	2.69	-	-	✓		15	4.09	36	5.48.
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	-		0.5 - 0.8		0.1 - 0.2		0.1 - 0.2		0.08	
Total	595		885		516		367		657	
Name	Silty bio pelsparite		Oopelspar- ite		Oobiopel sparite		Sandy biosparite		Sideritic silty mic. biosparite	

Slide no.	Ra 22		Ra 23		Ra 23		Ra 25		Ra 26	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-			-	-				
Spar	-	-			-	-				
Recryst. cal.	317	46.96			269	33.37				
Ech. & crin.	27	4.00			-	-				
Bivalves	42	6.22			-	-				
Gastropods	2	0.30			-	-				
Brachiopods	7	1.04			-	-				
Corals	-	-			-	-				
Foram. & ost.	-	-			-	-				
Ooliths					-	-				
Pellets					-	-				
Algae	-	-			-	-				
Qtz. (und.)	-	-			10	1.24				
Qtz. (polyx.)	6	0.89			28	3.47				
Qtz.	177	26.22		5	467	57.94				
Felds. P=Plag. O=Or.	-	-			13	1.61				
Mica	10	1.48			7	0.09				
Opakes	87	12.89		✓	12	1.49				
Epidote	-	-								
Zircon	-	-								
Max. gr. size mm.	0.1		0.08		0.1		0.1			
Total	675				806					
Name	Sandy biosparite		Silty-sandy biosparite		Calc. ferrug. sandstone		Calc. ferrug. feldspathic sandstone			

Slide no.	Ra 27		Ra 33		Ra 34A		Ra 34B		Ra 34C	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	374	57.63	441	73.01	122	44.36	94	39.5	100	58.48
Ech. & crin.	2	0.30	-	-	-	-	27	11.34	33	19.30
Bivalves	17	2.69	3	0.50	-	-	-	-	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	5	0.77	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	-	-	-	-	-	-	-	-	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	11	1.69	-	-	-	-	-	-	-	-
Qtz. (polyx.)	25	3.85	-	-	-	-	-	-	-	-
Qtz.	186	28.66	132	21.85	139	50.55	96	40.34	30	17.54
Felds. P=Plag. O=Or.	20	3.08	4	0.66	-	-	2	0.84	-	-
Mica	-	-	3	0.50	-	-	1	0.42	1	0.58
Opakes	9	1.39	21	3.48	14	5.09	18	7.56	7	4.00
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.2		0.04-0.08		0.08		0.06		0.06	
Total	649		604		275		238		171	
Name	Calcareous fine sand		Micaceous calcareous siltstone		Calc. siltstone		Calc. siltstone		Silty biosparite	

Slide no.	Am 25		Am 26		Am 40		Am 42		Am 45	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	810	60.99	-	-	163	26.68	-	-
Spar	-	-	386	29.07	-	-	162	26.51	-	-
Recryst. cal.	246	44.17	-	-	273	37.09	130	21.27	438	62.93
Ech. & crin.	10	1.80	-	-	-	-	1	0.16	25	3.59
Bivalves	-	-	-	-	-	-	11	1.80	13	1.87
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	37	2.79	-	-	4	0.65	-	-
Ooliths	-	-	-	-	-	-	138	22.58	124	17.91
Pellets	-	-	-	-	-	-	-	-	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	12	2.15	-	-	20	2.72	-	-	-	-
Qtz. (polyx.)	15	2.69	-	-	149	20.24	-	-	40	5.75
Qtz.	237	42.55	84	6.33	251	34.10	2	0.33	52	7.47
Felds. P=Plag. O=Or.	18	3.23	-	-	40	5.43	-	-	1	0.14
Mica	7	1.26	-	-	1	0.14	-	-	-	-
Opagues	12	2.15	11	0.83	2	0.27	-	-	3	0.43
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.08		-		0.2		0.2-0.3 Oo. 0.1-0.16 Q.		0.2 Q	
Total	557		1328		736		611		696	
Name	Micaceous calcarenite		Silty sparry biomicrite		Calcareous sandstone		Oo biopel micrite		Sandy Oobiopel sparite wackestone	

Slide no.	Am 46		Am 47		Am 48		Am 49		Am 50	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	269	64.98	243	50.84	303	43.16	299	47.92	382	55.04
Ecn. & crin.	-	-	11	2.30	-	-	5	0.80	-	-
Bivalves	-	-	4	0.84	-	-	27	4.33	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	34	7.11	-	-	8	1.28	312	44.96
Pellets	-	-	-	-	-	-	119	19.07	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	7	1.69	-	-	9	1.28	-	-	-	-
Qtz. (polyx.)	109	26.33	50	10.46	72	10.26	22	3.53	-	-
Qtz.	6	1.45	132	27.62	302	43.02	139	22.28	-	-
Felds. P=Plag. O=Or.	20	4.83	4	0.84	7	1.00	4	0.64	-	-
Mica	3	0.72	-	-	8	1.14	-	-	-	-
Opagues	-	-	-	-	-	-	1	0.16	-	-
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.2		0.2		0.1		0.1		0.2 - 0.3	
Total	503		478		702		624		694	
Name	Calcareous fine sandst.		Sandy oo biopel sparite wackestone		Calcareous fine sandst.		Sandy bio pel sparite wackestone		Oosparite wackestone	

Slide no.	Am 51		Am 52		Am 53		Am 54		Am 55	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	16	3.59	350	53.44	296	46.91	322	49.69	390	70.14
Ech. & crin.	-	-	-	-	3	0.48	8	2.48	15	2.70
Bivalves	-	-	21	3.21	14	2.22	6	0.93	64	11.51
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	-	-	-	-	4	0.62	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	165	25.19	15	2.38	-	-	1	0.18
Pellets	-	-	-	-	121	19.18	-	-	28	5.04
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	2	0.45	-	-	-	-	11	1.70	-	-
Qtz. (polyx.)	18	4.04	16	2.44	19	3.01	37	5.71	7	1.26
Qtz.	393	88.12	90	13.74	-	-	213	32.87	15	2.70
Felds. P=Plag. O=Or.	12	2.69	5	0.76	1	0.16	20	3.09	-	-
Mica	7	1.57	8	1.22	1	0.16	18	2.78	-	-
Opagues	abdt.	-	-	-	-	-	9	1.39	36	6.47
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.24-0.3		0.06-0.08		0.02-0.08 0.2-0.4 Q		0.16-0.2		0.2-0.4	
Total	446		655		631		648		556	
Name	Fine to medium gr. qtz.sandst.		Silty oolio sparite wackestone		Sandy peloolio sparite		Sandy biosparite		Sandy biopel sparite	

Slide no.	Am 56		Am 57		Am 58		Am 59		Am 61	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	643	69.66	278	41.90	345	51.65	184	35.73	507	86.82
Ech. & crin.	50	5.42	2	0.29	20	2.99	-	-	-	-
Bivalves	68	7.37	-	-	20	2.99	-	-	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	3	0.33	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	2	0.22	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	26	2.82	3	0.44	98	14.65	-	-	6	1.03
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	25	3.65	14	2.11	7	1.36	-	-
Qtz. (polyx.)	-	-	71	10.37	6	0.90	40	7.77	-	-
Qtz.	76	8.23	253	36.93	148	22.12	263	51.07	20	3.42
Felds. P=Plag. O=Or.	-	-	43	6.28	12	1.79	13	2.52	-	-
Mica	5	0.54	10	1.46	1	0.15	7	1.36	-	-
Opagues	50	5.42	-	-	5	0.75	1	0.19	51	8.73
Epidote	-	-	-	-	-	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.04		0.04		0.06		0.1-0.12		0.04-0.06	
Total	923		685		669		515		584	
Name	Silty biopelsparite		Calc. siltstone		Silty biopelsparite		Calcareous fine sandstone		Silty pyritic micro sparite	

Slide no.	Am 65		Am 66		Am 66-1		Am 69		Am 72	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	432	65.95	446	70.13	476	67.42	271	58.61	380	48.04
Ech. & crin.	-	-	23	3.62	9	1.27	4	0.63	2	0.25
Bivalves	10	1.53	15	2.36	9	1.27	28	4.42	4	0.50
Gastropods	-	-	1	0.16	4	0.57	7	1.11	-	-
Brachiopods	-	-	5	0.79	-	-	8	1.28	-	-
Corals	-	-	-	-	-	-	2	0.31	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	61	0.84	-	-	-	-
Pellets	16	2.44	-	-	-	-	49	7.74	10	1.26
Algae	-	-	-	-	✓	-	-	-	-	-
Qtz. (und.)	-	-	6	0.94	-	-	4	0.63	6	0.76
Qtz. (polyx.)	2	0.30	11	1.73	1	0.14	24	3.79	20	2.52
Qtz.	138	21.07	95	14.94	107	15.16	106	16.75	232	29.33
Felds. P=Plag. O=Or.	3	0.46	3	0.47	1	0.14	4	0.63	55	6.95
Mica	11	1.68	5	0.79	7	0.99	4	0.03	37	4.68
Opagues	42	6.41	20	3.14	31	4.39	16	2.53	8	1
Epidote	-	-	-	-	-	-	clay 6	0.95	clay 37	4.68
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.04-0.06		0.08-0.18		0.01-0.1		0.16-0.2		0.08-0.1	
Total	655		636		706		633		791	
Name	Silty microspa- rite		Sandy biomicro sparite		Silty sandy calc. cham sid. oolite wackestone		Chloritic Sandy bio microspar- ite		Sandy micaceous chloritic biomicro- sparite wackestone	

Slide no.	Am 73-1		Am 73-2		Am 75		Am 77		Am 78	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	333	54.06	352	51.37	288	40.17	395	71.43	382	49.16
Ech. & crin.	15	2.44	8	1.17	8	1.12	1	0.18	10	1.29
Bivalves	29	4.71	47	6.86	13	1.81	-	-	42	5.41
Gastropods	-	-	3	0.44	1	0.14	-	-	3	0.39
Brachiopods	4	0.65	-	-	1	0.14	-	-	7	0.90
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	*clays 11	1.61	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	15	2.44	8	1.17	96	13.39	-	-	2	0.26
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	2	0.32	-	-	3	0.42	-	-	7	0.90
Qtz. (polyx.)	6	.98	5	0.73	29	4.05	-	-	44	7.66
Qtz.	150	24.35	172	25.11	196	27.34	109	19.71	216	27.80
Felds. P=Plag. O=Or.	5	0.81	P*2	0.29	12	1.67	3	0.54	27	3.22
Mica	32	5.19	46	6.72	37	5.16	10	1.80	21	2.76
Opakes	12	1.95	17	2.48	21	2.93	27	4.88	17	2.19
Epidote	*clay 13	2.11	2	1.02	-	-	-	-	*clay 1	0.13
Zircon	-	-	7	0.29	-	-	-	-	-	-
Max. gr. size mm.	0.06-0.1		0.1		0.08-0.12		0.04		0.12-0.24	
Total	616		685		717		553		777	
Name	Fine sandy calc. chloritic biopelsparite		Sandy micaceous biopel sparite		F.sandy micaceous chloritic biisparite		Silty mic. biopel sparite		Sandy micaceous chloritic bio pelsparite	

Slide no.	Am 79		Am 80		Am 81		Am 82		Am 84	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	401	56.32	361	49.86	151	28.60	171	52.78	233	34.93
Ech. & crin.	2	0.28	-	-	-	-	1	0.31	5	0.75
Bivalves	62	8.71	38	5.25	-	-	1	0.31	1	0.15
Gastropods	7	0.98	-	-	-	-	-	-	-	-
Brachiopods	14	1.97	1	0.14	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	2	0.28	2	0.28	18	3.41	4	1.23	12	1.80
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	6	0.84	4	0.55	-	-	7	2.16	13	0.95
Qtz. (polyx.)	11	1.54	36	4.97	23	4.35	44	13.58	19	2.85
Qtz.	149	20.93	238	32.87	274	51.89	181	55.86	316	47.38
Felds. P=Plag. Q=Or.	5	0.70	12	1.66	13	2.46	4	1.23	20	3.00
Mica	20	2.81	18	2.49	27	5.11	-	-	15	2.25
Opauques	27	3.79	8	1.10	12	2.27	11	3.40	8	1.20
Epidote	*clay 6	0.84	*clay 6	0.83	*clay 10	1.89	-	-	*clay 25	3.75
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.3		0.1-0.12		.06-0.1		0.2		0.08-0.1	
Total	712		724		528		324		667	
Name	Sandy micaceous biopelspar- ite		Fine sandy micaceous biopelspa- rite		Chloritic micaceous fine sandstone		Calcareous glauconitic fine sandstone		Calcareous micaceous glauconitic fine sandstone	

Slide no.	Am 85		Am 86		Am 87		Am 89		Am 90	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	185	32.34	104	17.39	315	40.59	122	28.91	171	34.34
Ech. & crin.	10	1.75	1	0.17	4	0.52	-	-	1	0.20
Bivalves	20	3.50	9	1.51	4	0.52	-	-	6	1.20
Gastropods	8	1.40	40	6.69	2	0.26	-	-	-	-
Brachiopods	1	0.17	2	0.33	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	5	0.87	-	-	-	-	-	-	-	-
Pellets	10	1.75	11	1.84	4	0.52	-	-	5	1
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	5	0.87	6	1	6	0.77	5	1.18	5	1
Qtz. (polyx.)	22	3.85	19	3.18	38	4.90	34	8.06	34	6.83
Qtz.	257	44.93	237	39.63	363	46.78	227	53.79	265	53.54
Felds. P=Plag. O=Or.	4	0.70	11	1.84	11	1.42	17	4.03	11	2.21
Mica	-	-	3	0.50	4	.52	1	0.24	1	0.20
Opagues	45	7.87	150	25.08	21	2.71	10	2.37	8	1.61
Epidote	-	-	*chl. 3	0.50	*clay 4	0.52	-	-	-	-
Zircon	-	-	-	-	-	-	-	-	✓	-
Max. gr. size mm.	0.24-0.3		0.2-0.3		0.14-0.2		0.16-0.2		0.1-0.12	
Total	572		598		776		422		998	
Name	Calcareous ferruginous bioclastic sandstone		Calcareous ferruginous sandstone		Calcareous glauconitic fine sandstone		Ferruginous ? sideritic sandstone		Calc. very fine sandstone	

Slide no.	Am 90		Am 91-1		Am 91-2		Am 93		Am 94	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	171	34.34	96	15.56	310	49.28	32	5.89	44	11.20
Ech. & crin.	1	0.20	2	0.32	-	-	-	-	1	0.25
Bivalves	6	1.20	1	0.16	35	5.56	-	-	13	3.31
Gastropods	-	-	9	1.46	1	0.16	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	2	0.32	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	*clays 2	0.32	-	-	-	-
Ooliths	-	-	16	2.59	3	0.48	-	-	-	-
Pellets	5	1	11	1.78	43	6.83	17	3.13	7	1.78
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	5	1	3	0.48	2	0.32	7	1.29	9	2.29
Qtz. (polyx.)	34	6.83	49	7.98	21	3.34	27	4.98	41	10.42
Qtz.	265	53.54	273	44.25	162	25.76	369	67.96	211	53.66
Felds. P=Plag. O=Cr.	11	2.21	6-0 15-P	0.97 2.43	1	0.16	3	0.55	9=0 2=P	2.29 0.51
Mica	1	0.20	1	0.16	2	0.32	7	1.29	10	2.54
Opakes	8	1.61	111	17.99	34	5.41	68	12.52	31	7.89
Epidote	-	-	✓		2		4	(0.74)	11	(2.80)
Zircon	✓		-	-	6		9	(1.66)	3	(0.76)
Max. gr. size mm.	.08-0.1		0.2		0.06-0.3		0.2-0.3		0.3	
Total	498		617		629		543		393	
Name	Calcareous fine sandstone		Ferruginous (?Hamatitic) sandstone		(Silty) bio pel sparite		Ferruginous sandstone		Calcareous sandstone	

Slide no.	Am 95		Am 96		Am 98		Am 99		Am 100	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite			-	-	-	-	-	-	-	-
Spar			-	-	-	-	-	-	-	-
Recryst. cal.			308	46.67	62	10.32	9	1.49	3	0.59
Ech. & crin.	2	0.30	43	6.25	7	1.16	-	-	-	-
Bivalves	25	3.79	1.7	2.58	4	0.67	-	-	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	12	1.82	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	*clay 2	0.33	-	-	-	-
Ooliths	2	0.30	-	-	-	-	-	-	-	-
Pellets	4	0.60	2	0.30	11	1.83	-	-	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	2	0.30	-	-	19	3.16	1	0.17	134	26.27
Qtz. (polyx.)	28	4.24	10	1.51	21	3.50	34	5.62	9	1.76
Qtz.	259	39.24	239	36.21	215	35.77	370	61.06	270	52.94
Felds.	6=0		2=0	0.30	44=0	7.32	7=P	1.16	0=0	-
P=Plag. O=Or.	1=P	1.06	3=P	0.45	2=P	0.33	7=0	1.16	3=P	0.59
Mica	5	0.76	2	0.30	5	0.83	3	0.50	-	-
Opauques	3	0.45	2	0.30	199	33.11	131	21.62	91	1.65
Epidote	16	(2.42)	10	(1.52)	3	(0.50)	-	-	-	-
Zircon	8	(1.20)	7	(1.06)	7	(1.16)	-	-	-	-
Max. gr. size mm.	0.2-0.3		0.2-0.3		0.2		0.1-0.2		0.02-0.2	
Total	660		660		601		606		570	
Name	Calcareous sandstone		Ferruginous sandy limestones		Hematitic fine sandstones		Sideritic calcareous sandstone		?Sideritic fine sandstone	

Slide no.	Am 101		Am 102		Am 103		Am 106		Am 108	
	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	33	5.33	-	-		
Spar	-	-	-	-	-	-	-	-		
Recryst. cal.	486	56.58	312	46.22	279	45.07	234	31.62	346	72.69
Ecn. & crin.	1	0.12	-	-	1	0.16	-	-	6	1.26
Bivalves	15	1.75	-	-	46	7.43	-	-	12	2.52
Gastropods	-	-	-	-	-	-	-	-	3	0.63
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	16	2.58	-	-	4	0.84
Pellets	36	4.19	-	-	82	13.25	-	-	56	11.76
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	3	0.35	3	0.44	4	0.65	5	0.68	3	0.63
Qtz. (polyx.)	30	3.49	30	4.45	11	1.78	8	1.09	11	0.84
Qtz.	279	32.48	315	46.67	140	22.62	469	63.38	22	4.62
Felds. P=Plag. O=Or.	14=O 0=P	1.63 -	0=O 4=P	- 0.59	=O 3=P	0.48	0=O 7=P	- 0.95	1=O 0=P	0.21 -
Mica	2	0.33	-	-	-	-	3	0.41	-	-
Opaques	1	0.12	2	0.30	3	0.48	8	1.08	5	1.05
Epidote	3	(0.35)	6	(0.89)	1	0.16	3	(0.41)	7	(1.47)
Zircon	-	-	2	(0.30)	-	-	3	(0.41)	-	-
Max. gr. size mm.	0.1-0.2		0.2		0.2		0.2-0.5		0.1	
Total	859		675		619		740		476	
Name	Sandy biopel sparite		Calcareous fine sandstone		Sandy biopel sparite wackestone		Calcareous sandstone		Fine sandy biopel sparite wackestone	

Slide no.	Am 109		Am 110		Am 111		Am 112		Am 113	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	335	75.79	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	-	-	352	35.16	302	48	297	49.25	215	46.14
Ech. & crin.	-	-	-	-	1	0.16	4	0.66	21	4.51
Bivalves	8	1.81	2	0.20	63	10.03	61	10.12	35	7.51
Gastropods	-	-	-	-	-	-	-	-	146	31.33
Brachiopods	-	-	-	-	-	-	2	0.33	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	7	1.59	-	-	139	22.13	169	28.03	2	0.42
Pellets	16	3.62	2	0.2	55	8.76	52	8.62	9	1.92
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	9	0.90	1	0.16	-	-	-	-
Qtz. (polyx.)	40	9.05	10	1.00	1	0.16	3	0.50	2	0.43
Qtz.	17	3.85	597	59.64	49	7.8	8	1.33	20	4.29
Felds. P=Plag. O=Or.	0=0 5=P	- 1.13	6=0 5=P	0.60 0.50	1=0	0.16	-	-	1	0.21
Mica	-	-	3	0.30	-	-	-	-	-	-
Opakes	14	3.17	-	-	4	0.64	5	0.83	7	1.50
Epidote	-	-	12	(1.2)	1	(0.16)	2	(0.33)	1	(0.21)
Zircon	-	-	3	(0.30)	1	(0.16)	-	-	5	(1.07)
Max. gr. size mm.	0.2		0.04-0.1		0.06-0.16		0.16		0.06-0.4	
Total	442		1001		628		603		466	
Name	Sandy biopel micrite		Calcareous siltstone- sandstone.		Silty-sandy oobiopel sparite		Sandy oobiopel sparite packstone		Silty-sandy echinod. pellet limestone	

Slide no.	Am 115		Am 116		Am 117		Am 119		Am 120	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	35	5.56	-	-	-	-	2	0.32	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	364	57.78	393	55.43	547	77.48	450	71.20	564	80.69
Ech. & crin.	5	0.79	5	0.71	4	0.57	15	2.37	8	1.14
Bivalves	10	2.54	30	4.23	21	2.97	27	4.27	26	3.72
Gastropods	1	0.16	-	-	2	0.28	-	-	1	0.14
Brachiopods	-	-	-	-	3	0.42	4	0.63	23	3.29
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Coliths	-	-	1	0.14	-	-	1	0.16	1	0.14
Pellets	40	6.35	257	36.25	43	6.09	77	12.18	54	7.73
Algae	-	-	-	-	-	-	-	-	✓	
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	12	1.90	1	0.14	2	0.28	-	-	-	-
Qtz.	147	23.32	5	0.71	82	11.61	51	8.07	18	2.58
Felds. P=Plag. O=Or.	1=0 0=P	0.16 -	-	-	-	-	-	-	-	-
Mica	-	-	-	-	-	-	1	0.16	-	-
Opagues	6	0.95	17	2.40	2	0.28	4	0.64	2	0.29
Epidote	2	(0.32)	-	-	-	-	-	-	2	(0.29)
Zircon	1	(0.16)	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.06-0.2		0.03-0.4		0.1-0.12		0.04-0.1		0.16	
Total	630		709		706		632		699	
Name	Silty-sandy biopel micro sparite		Biopel sparite		Sandy biopel sparite		Silty-sandy biopel micro sparite		Sandy biopel sparite	

Slide no.	Am 121		Am 122		Am 123		Am 124		Am 125	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	32	4.56	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	828	49.70	328	34.03	309	52.02	336	48	522	79.45
Ech. & crin.	1	0.15	-	-	6	1.01	-	-	8	1.22
Bivalves	21	3.18	-	-	24	4.04	-	-	17	2.59
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ool.	-	-	-	-	-	-	-	-	-	-
Ooliths	78	11.81	-	-	98	16.49	272	38.74	2	0.30
Pellets	143	21.67	-	-	95	15.99	-	-	13	1.98
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	1	0.14	-	-
Qtz. (polyx.)	16	2.42	80	11.94	32	5.39	28	3.99	-	-
Qtz.	66	10	350	52.24	29	4.88	20	2.85	30	4.59
Felds. P=Plag. O=Or.	1	0.15	4=P	0.60	-	-	1	0.14	5=O 4=P	0.76 0.61
Mica	-	-	1	0.15	-	-	-	-	4	0.61
Opakes	4	0.61	3	0.45	1	0.17	1	0.14	19	2.89
Epidote	2	(0.30)	4	(0.60)	-	-	5	(0.71)	32	(4.87)
Zircon	-	-	-	-	-	-	4	(0.71)	3	0.46
Max. gr. size mm.	0.1-0.2		0.08-0.2		0.2		0.24-0.32		0.1	
Total	660		670		594		702		657	
Name	Sandy oobio pelsparite wackestone		Fine sandstone		Sandy oopel biosparite wackestone		Sandy oosparite wackestone		Sandy biopel microspa- rite	

Slide no.	Am 126		Am 127		Am 128		Am 129		Am 130	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	140	30.5	-	-	-	-	-	-	-	-
Recryst. cal.	248	54.03	✓		✓		✓		✓	
Ech. & crin.	5	1.09	-	-	-	-	✓		✓	
Bivalves	13	2.83	-	-	✓		✓		✓	
Gastropods			-	-	-	-	-	-	-	-
Brachiopods	1	0.22	✓		✓		-	-	-	-
Corals			-	-	-	-	-	-	-	-
Foram. & ost.	2	0.44	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	6	1.31	✓		-	-	-	-	-	-
Algae			-	-	-	-	-	-	-	-
Qtz. (und.)			-	-	-	-	-	-	-	-
Qtz. (polyx.)	-	-	-	-	-	-	-	-	-	-
Qtz.	-	-	-	-	-	-	-	-	-	-
Felds. P=Plag. O=Or.	-	-	-	-	-	-	-	-	-	-
Mica			-	-	-	-	-	-	-	-
Opagues	27	5.88	✓		-	-	-	-	-	-
Epidote	14	(3.05)	✓		✓		✓		✓	
Zircon	3	(0.65)	-	-	-	-	-	-	-	-
Max. gr. size mm.			-		0.1-0.3					
Total	4.59		-							
Name	Biopel microspar- ite		Biopel microspa- rite		Bio sparite (micritic in places)		Biosparite (micritic in places)		Biomicro sparite	

Slide no.	Am 131		Am 132		Am 135		Am 136		Am 137	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	350	50.58	196	30.77	203	29.34	328	48.81	207	30.44
Ech. & crin.	3	0.43	1	0.10	-	-	7	1.04		
Bivalves	33	4.77	8	1.26	-	-	33	4.91	-	-
Gastropods	1	0.14	-	-	-	-				
Brachiopods	-	-	-	-	-	-	-	-		
Corals	-	-	-	-	-	-	-	-		
Foram. & ost.	26	3.76	7	1.10	-	-	-	-	-	-
Ooliths	3	0.43	-	-	1	0.14	5	0.74		
Pellets	-	-	*clay 1	0.16	*clay 2	0.29	91	13.54	1	0.15
Algae	-	-	-	-	-	-	*clay 1	0.15		
Qtz. (und.)	-	-	1	0.16	-	-	1	0.15		
Qtz. (polyx.)	176	25.43	-	-	-	-	9	1.34	49	7.20
Qtz.	-	-	355	55.73	451	65.17	185	27.53	403	59.26
Felds. P=Plag. O=Or.	-	-	0=P 16=0	2.51= 0	1=P 15=0	0.14 2.17	0=P 2=0	0.30	2=P 3=0	0.29 0.44
Mica	2	0.28	4	0.63	1	0.14			11	1.62
Opakes	4	0.58	30	4.71	9	1.30	5	0.74	2	0.29
Epidote	16	(2.31)	12	(1.88)	9	(1.30)	4	0.60	2	(0.29)
Zircon	15	(2.17)	6	(.94)	-	-	1	0.15	-	-
Max. gr. size mm.	-		0.04		0.01-0.08		0.06-0.16		0.14	
Total	692		637		692		672		680	
Name	Biopel sparite		Calcareous siltstone		Calcareous siltstone		Silty-sandy biopel sparite wackestone		Calcareous fine sandstone	

Slide no.	Am 138		Am 139		Am 140		Am 141		Am 142	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	378	56.42			304	44.90	330	42.58		
Ech. & crin.	6	0.90	✓		10	1.48	-	-		
Bivalves	14	2.09	✓		14	2.07	-	-		
Gastropods	-	-	-	-	-	-	-	-		
Brachiopods	-	-	-	-	-	-	-	-		
Corals	-	-	-	-	-	-	-	-		
Foram. & ost.	-	-	-	-	-	-	-	-		
Ooliths	20	3.14	-	-	28	4.14	-	-		
Pellets	70	10.4	-	-	76	11.22	-	-		
Algae	-	-	-	-	-	-	-	-		
Qtz. (und.)	1	0.15	-	-	1	0.14	-	-		
Qtz. (polyx.)	4	0.60	-	-	13	1.92	4	.52		
Qtz.	148	22.09	-	-	214	31.61	423	54.58		
Felds.	5=P	0.75			5=P	0.74				
P=Plag. O=Or.	-	-	-	-	2=O	0.30	1	0.13		
Mica	7	1.04	-	-	4	0.59	11	1.42		
Opakes	8	1.19	✓		2	0.30	3	0.39		
Epidote	8	(1.19)	-	-	4	0.59	3	0.39		
Zircon	1	(.15)	-	-	-	-	-	-		
Max. gr. size mm.	0.1-0.2				0.1-0.16		0.16			
Total	670				677		775			
Name	Sandy bio pelsparite		Biosparite		Sandy bio pelsparite wackestone		Calcareous fine grained sandstone		Recrystall- ized coral limestone	

Slide no.	Am 143		Am 145		Am 148		Am 149-9		Am 149	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	247	37.03	384	65.20	538	85.95	353	56.21	-	-
Ech. & crin.	10	1.50	13	2.21	2	0.32	39	6.21	✓	
Bivalves	13	1.95	6	1.02	27	4.31	53	8.44	✓	
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	1	0.15	-	-	2	0.32	24	3.82	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	2	0.30	4	0.68	10	1.60	-	-	-	-
Ooliths	-	-	5	0.85	-	-	2	0.32	-	-
Pellets	-	-	10	1.70	5	0.80	-	-	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	24	3.60	33	5.63	-	-	1	0.16	-	-
Qtz.	368	55.17	98	16.64	32	5.11	71	11.31	-	-
Felds. P=Plag. O=Or.	- 2=P	0.30 =P	7	1.19	-	-	3=P =0	0.48	-	-
Mica	-	-	-	-	1	0.16	6	0.96	-	-
Opagues	-	-	15	3.91	3	0.48	27	4.30	-	-
Epidote	-	-	6	(1.02)	3	(0.48)	36	(5.73)	-	-
Zircon	-	-	8	(1.36)	1	(0.16)	13	2.07	-	-
Max. gr. size mm.	0.14-0.16		0.1-0.2		0.06-0.1		0.04-0.2			
Total	667		589		626		628			
Name	Calcareous fine sandstone		Fine calc. sandstone		V.fine calcareous sandstone		Echinoder- mal limestones		Biosparite packstone	

Slide no.	Am 150		Am 156		Am 158		Am 165		Am 170	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	408	62.39	281	63.29	357	54.09	293	52.51	223	43.73
Ech. & crin.	26	3.98	16	3.60	25	3.79	14	2.51	5	0.98
Bivalves	57	8.72	12	2.70	73	11.06	69	12.37	12	2.35
Gastropods	1	0.15	3	0.68	-	-	4	0.72	1	0.20
Brachiopods	1	0.15	-	-	6	0.91	3	0.54	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	1	0.15	3	0.68	-	-	4	0.72	-	-
Ooliths	4	0.60	5	1.36	-	-	1	0.18	-	-
Pellets	24	3.66	13	2.93	12	1.82	26	4.66	4	0.78
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	2	0.30	1	0.23	3	0.45	4	0.72	16	3.14
Qtz.	73	11.16	29	6.53	111	16.82	99	17.74	229	44.9
Felds. P=Plag. O=Or.	=P =O	- -	78	17.77	-	-	2	0.36	4	0.78
Mica	1	0.15	-	-	7	1.06	19	3.41	6	1.18
Opauques	54	8.26	-	-	62	0.94	19	3.41	8	1.57
Epidote	2	0.30	3	(0.68)	4	(0.61)	1	0.18	-	-
Zircon	-	-	10	(2.25)	-	-	1	0.18	2	0.39
Max. gr. size mm.	0.2		0.2		0.1-0.2		-		0.3	
Total	654		444		660		558		510	
Name	Sandy calc. sid., cham. bio pelsparite		Sandy bio pelsparite wackestone		Sandy bio pelsparite wackestone		Quartzose biopel sparite		Calc. sideritic bioclastic sandstone	

Slide no.	Am 171		Am 173		Am 175 Am 176, 185		Am 187b		Am 188	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	115	21.02
Recryst. cal.	289	55.68	-	-	-	-	386	62.16	302	55.21
Ecn. & crin.	18	3.47	-	-	-	-	27	4.35	-	-
Bivalves	37	7.13	-	-	-	-	59	9.50	-	-
Gastropods	5	0.96	-	-	-	-	2	0.32	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	3	0.48	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	5	0.96	-	-	-	-	5	0.81	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	-	-	-	-
Qtz. (polyx.)	9	1.73	-	-	-	-	3	0.48	-	-
Qtz.	135	26.01	-	-	-	-	94	15.14	62	11.33
Felds. P=Plag. O=Or.	1=P 1=O	0.19 0.19	-	-	-	-	-	-	-	-
Mica	1	0.19	-	-	-	-	23	3.70	42	7.68
Opakes	14	2.70	-	-	-	-	15	2.42	11	2.01
Epidote	-	-	-	-	-	-	4	.64	13	2.38
Zircon	-	-	-	-	-	-	3	0.48	2	0.37
Max. gr. size mm.	0.2-0.3		-				0.1-0.2		0.06	
Total	519		-				621		547	
Name	Sandy biopel sparite		Rextall. sparite		Laminated echinoder- mal packstone		Sandy, mic. biopel sparite packstone		Metamor- phised lamin., silty mic. shale	

Slide no.	Am 190		Am C1		Am C4		M3		M10	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	390	67.53	303	62.22	149	33.94	368	64.56	235	49.36
Ech. & crin.	5	0.79	15	3.08	34	7.74	-	-	-	-
Bivalves	-	-	27	5.54	8	1.82	12	2.11	8	1.68
Gastropods	1	0.10	1	0.21	* ch. 2	0.46	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	* gla. 10	1.75	-	-
Pellets	-	-	20	4.11	-	-	78	13.68	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	1	0.23	-	-	-	-
Qtz. (polyx.)	2	0.32	26	5.34	22	5.01	-	-	47	9.87
Qtz.	131	20.73	77	15.91	198	45.10	71	12.46	137	28.78
Felds. P=Plag. O=Or.	- 2=P	- 0.32	1	0.21	4	0.91	-	-	19=O 3=P	3.99 0.63
Mica	-	-	8	1.64	2	0.46	-	-	7	1.47
Opagues	1	0.16	-	-	1	0.23	27	4.74	7	1.47
Epidote	-	-	6	(1.23)	-	-	-	-	9	1.89
Zircon	-	-	3	(0.62)	2	(0.46)	4	(0.70)	4	0.84
Max. gr. size mm.	0.1-0.2		0.2		0.2-0.3		0.04-0.1		0.1	
Total	632		487		423		570		476	
Name	Metamor- phosed sandy lst.		Sandy bio pelsparite wackestone		Sandy bio pelsparite wackestone		Silty-sandy biopel sparite packstone		Calc. micaceous fine sandstone	

Slide no.	M 17		M 23		M 26		M 32		M 60	
Constituents	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-			-	-	-	-
Spar	-	-	-	-			-	-	-	-
Recryst. cal.	307	62.78	319	43.16	169	30.07	225	42.61	135	22.17
Ech. & crin.	2	0.40	2	0.26	3	0.53	-	-	-	-
Bivalves	21	4.29	18	2.43	3	0.53	-	-	-	-
Gastropods	4	0.80	-	-			-	-	-	-
Brachiopods	-	-	-	-			-	-	-	-
Corals	-	-	-	-			-	-	-	-
Foram. & ost.	3	0.61	-	-			-	-	-	-
Oolites	-	-	-	-			-	-	-	-
Pellets	32	6.54	-	-			-	-	-	-
Algae	-	-	-	-			-	-	-	-
Qtz. (und.)	-	-	-	-			-	-	-	-
Qtz. (polyx.)	-	-	23	3.10	12	2.14	-	-	-	-
Qtz.	68	13.90	347	46.95	354	62.99	293	55.49	286	46.96
Felds. P=Plag. O=Or.	3	0.61	2=P 1=O	0.26 0.13	7=P 1=O	1.25 0.18	7	1.32	-	-
Mica	2	0.40	15	2.02	2	0.36	-	-	31	5.09
Opakes	39	7.97	8	1.08	5	.89	-	-	42	6.90
Epidote	4	0.80	2	0.26	3	0.53	3	0.56	-	-
Zircon	4	0.80	2	0.26	3	0.53	-	-	-	-
Max. gr. size mm.	0.02-0.04		70.2		0.1-0.3		0.04-0.1		0.1	
Total	489		739		562		528		494	
Name	Silty biopel sparite		Ferrugin. calc. Sst.		Calcareous ferrug. Sst.		Calc.silst. with pseudo- isotrop. cem.		Mic., calc., f.Sst. (ps.isotrop. cem.)	

Slide no. Constituents	M 65		M 66		M 77		M 96		M 62	
	no.	%	no.	%	no.	%	no.	%	no.	%
Micrite	-	-	-	-	-	-	-	-	-	-
Spar	-	-	-	-	-	-	-	-	-	-
Recryst. cal.	144	30.37	200	37.38	541	64.33	154	33.26	190	32.31
Ech. & crin.	-	-	-	-	2	0.24	-	-	-	-
Bivalves	-	-	-	-	41	2.88	1	0.22	-	-
Gastropods	-	-	-	-	-	-	-	-	-	-
Brachiopods	-	-	-	-	-	-	-	-	-	-
Corals	-	-	-	-	-	-	-	-	-	-
Foram. & ost.	-	-	-	-	-	-	-	-	-	-
Ooliths	-	-	-	-	-	-	-	-	-	-
Pellets	-	-	-	-	6	0.71	5	1.08	-	-
Algae	-	-	-	-	-	-	-	-	-	-
Qtz. (und.)	-	-	-	-	-	-	2	0.44	-	-
Qtz. (polyx.)	-	-	-	-	1	0.12	3	0.66	3	0.51
Qtz.	258	54.43	275	51.40	167	19.86	254	54.86	328	55.78
Felds. P=Plag. O=Or.	1	0.21	-	-	4=O 2=P	0.48 0.24	5=O 7=P	1.08 1.51	-	-
Mica	57	12.02	47	8.78	7	0.83	24	5.18	36	6.12
Opakes	14	2.95	13	2.42	31	3.68	4	0.88	25	4.25
Epidote	-	-	-	-	3	(0.36)	4	(0.88)	6	(1.02)
Zircon	-	-	-	-	-	-	-	-	-	-
Max. gr. size mm.	0.04-0.08		-		0.06-0.08		0.1		0.2	
Total	474		535		806		463		588	
Name	Mic. silst. in isotrop. cement		Mic. silt and sst in pseudo isotrop cement		Silty biosparite wackestone		Ferrug., calc., mic. fine sst.		Micaceous calcareous sandstone	

Slide no.	M 68		M 70							
	no.	%	no.	%	no.	%	no.	%	no.	%
Constituents										
Micrite	-	-								
Spar	-	-								
Recryst. cal.	219	45.82	164	30.43						
Ech. & crin.	-	-								
Bivalves	-	-								
Gastropods	-	-								
Brachiopods	-	-								
Corals	-	-								
Foram. & ost.	-	-								
Ooliths	-	-								
Pellets	-	-								
Algae	-	-								
Qtz. (und.)	-	-								
Qtz. (polyx.)	-	-								
Qtz.	197	41.21	309	57.33						
Felds. P=Plag. O=Or.	-	-								
Mica	79	16.53	64	11.87						
Opagues	29	6.07	-	-						
Epidote	-	-	2	(0.37)						
Zircon	4	0.84	-	-						
Max. gr. size mm.	0.04-0.06		0.14-0.18							
Total	478		539							
Name	Mic. calc. siltst.		Mic. calc. fine sandst.							

APPENDIX 2PALAEOCURRENT DETERMINATIONS

The directional data were obtained by field measurements along vertical, two-dimensional sections using data sheets recommended by Potter and Pettijohn (1963).

As cross bedding mostly occurred in gently dipping but faulted strata, corrections were applied in order to compensate for tectonic disturbance using a stereonet (Potter and Pettijohn, 1963; Hoyt, 1971; High and Picard, 1971).

The palaeocurrent data is presented in pages 2-2 to 2-11 and grouped in rose-diagrams at 20° intervals.

Torr Mor Fig. 6.4					Torr Mor Fig. 6.5				
		Bedding St. N355 Dip 8° SW					Bedding St. N17° Dip 10° NW		
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N307°	25°	N321°	20°	1	N319°	18°	N351°	11°
2	N298	16	N319	10	2	N276	8	N148	3
3	N259	16	N248	8	3	N254	11	N195	6
4	N263	11	N248	3	4	N288	12	N288	2
5	N272	18	N289	10	5	N296	2	N304	10
6	N291	26	N283	18	6	N283	11	N283	1
7	N332	32	N344	29	7	N262	20	N248	12
8	N269	13	N277	5	8	N341	13	N22	16
					9	N298	21	N307	12
					10	N284	16	N284	6
					11	N286	17	N286	7

Loch Bishort Fig. 6.18a					Loch Bishort Fig. 6.18				
		Bedding St. N358 Dip 16° SW					Bedding St. N358 Dip 16° SW		
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N209°	26°	N172°	24°	1	N261°	20°	N231°	5°
2	N270	10	N90	6	2	N232	20	N180	12
3	N295	14	N18	7	3	N218	14	N148	13
4	N288	22	N321	9	4	N246	18	N180	7
5	N292	30	N311	16					
6	N267	30	N262	14					
7	N273	36	N276	20					
8	N238	27	N206	16					
9	N25	18	N54	29					
10	N294	34	N311	21					

Loch Lonachan Fig. 6.25					Ob Lusa Fig. 6.24				
		St. 68 Bedding Dip 18SE					St. 51° Bedding Dip 7°NW		
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N217°	15°	N285°	16°	1	N337°	13°	N358°	7°
2	N308	12	N327	28	2	N314	6	N203	2
3	N99	6	N358	16	3	N358	4	N105	5
4	N175	16	N250	4	4	N323	38	N324	31
5	N32	8°	N358	24	5	N336	12	N350	5
6	N285	19	N308	33	6	N328	50	N330	42
7	N201	17	N276	12	7	N132	70	N132	77
8	N44	12	N3	25	8	N121	5	N129	12
9	N338	6	N338	24	9	348	6	N94	4
10	N210	8	N308	11	10	N329	12	N343	12
11	N224	14	N291	18	11	N333	22	N339	22
12	N281	22	N231	19	12	N302	18	N311	11
					13	N354	9	N40	5
					14	N295	-	N295	-
					15	N356	-	N356	-
					16	N310	12	N294	5
					17	N277	-	N277	-

Applecross Fig. 6.27 Bedding $\text{St. } 53^{\circ}$ Dip 8°NW									
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N282 $^{\circ}$	37 $^{\circ}$	N265 $^{\circ}$	24 $^{\circ}$	26	N327 $^{\circ}$	11 $^{\circ}$	N303 $^{\circ}$	5 $^{\circ}$
2	N250	26	N214	30	27	N234	11	N202	15
3	N272	22	N252	20	28	218	17	N202	21
4	N274	19	N249	15	29	N262	6	N189	8
5	N280	6	N210	8	30	N258	10	N214	11
6	N222	26	N241	22	31	N333	12	N329	4
7	N215	24	N235	11	32	N206	14	N189	14
8	N214	18	N242	16	33	N110	3	N146	10
9	N205	12	N243	8	34	N306	7	N221	4
10	N231	14	N265	14	35	N297	6	N202	6
11	N216	12	N260	10	36	N290	10	N236	9
12	N215	30	N230	27	37	N145	4	N157	11
13	N215	15	N233	11	38	N36	14	N74	12
14	N232	28	N247	28	39	N339	6	N153	2
15	N20	14	N29	8	40	N338	28	N339	2
16	N225	10	N192	15	41	N269	20	N245	29
17	N220	20	N206	25	42	N307	12	N269	6
18	N75	16	N44	15	43	N320	6	N225	3
19	N173	12	N167	20	44	N320	8	N231	3
20	N222	12	N195	17	45	N304	14	N268	8
21	N214	10	N190	16	46	N270	12	N228	12
22	306	8	N226	5					
23	N66	30	N79	29					
24	329	12	N292	6					
25	N300	4	N186	6					

Ardnamurchan Fig. 7.1 ; 7.3					Ardnamurchan Fig. 7.4				
Bedding St. N352° Dip 15°S					Bedding St. N344° Dip 18°SW				
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N256°	14°	N155°	1°	1	N279°	30°	N306°	16°
2	N270	15	N351	2	2	N282	24	N328	12
3	N298	18	N2°	11	3	N279	21	N335	9
4	N266	33	N269	18	4	N285	16	N11	10
5	N278	22	N310	9	5	N269	24	N303	9
6	N292	1	N42°	8					
7	N254	18	N32	22					
8	N294	16	N358	8					
9	N293	20	N337	10					
10	N289	25	N34	13					

Loch Eishort Fig. 6.19					Loch Slapin Fig. 6.36				
Bedding				St. 335 Dip 30°SW	Bedding				St. 334 Dip 20°SW
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N83°	16°	N72°	44°	1	N186°	8°	N90°	18°
2	N200	20	N110	22	2	N272	14	N22	10
3	N292	19	N26	22	3	N111	14	N85	31
4	N255	14	N60	15	4	N238	9	N73	11
5	N295	13	N39	24	5	N283	12	N28	12
6	N225	10	N80	22	6	N218	12	N99	11
7	N216	24	N141	25					
8	N187	10	N86	26					
9	N249	20	N28	3					
10	N182	14	N96	27					
11	N227	20	N99	13					
12	N254	24	N35	7					
13	N268	20	N32	14					
14	N289	12	N45	22					

Loch Slapin Fig. 6.37					Loch Eishort Fig. 6.34				
Bedding				St. N334 Dip 20SW	Bedding				St. N350 Dip 14SW
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N160°	9°	N90°	22°	1	N301°	15°	N7°	10°
2	N256	18	N4	4	2	N250	16	N198	4
3	N221	12	N98	9	3	N222	26	N193	27
4	N345	22	N20	32	4	N268	14	N350	2
					5	N238	14	N158	5
					6	N264	22	N273	8
					7	N213	22	N175	16
					8	N192	23	N166	21
					9	N262	18	N259	4
					10	N237	18	N190	8
					11	N275	15	N312	1
					12	N254	16	N208	32
					13	N274	15	N341	4
					14	N240	12	N130	5
					15	N242	26	N228	12
					16	N248	11	N114	4
					17	N255	15	N219	1
					18	N297	12	N18	8
					19	N226	14	N151	1
					20	N261	13	N83	1
					21	N260	20	N260	40

Loch Eishort Fig. 6.33					Loch Eishort Fig. 6.35				
		Bedding St. N350° Dip 14°SW					Bedding St. N350° Dip 14°SW		
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N329°	22°	N7°	21°	1	N259°	4°	N79°	10°
2	N210	9	N47	9	2	N45	8	N67	26
3	N322	18	N8	16	3	N204	30	N178	24
4	N331	18	N14	18	4	N206	13	N319	12
5	N286	19	N331	9	5	N96	10	N87	23
6	N247	18	N210	5	6	N145	7	N102	28
7	N234	22	N200	12	7	N251	20	N224	8
8	N242	21	N210	9	8	N250	20	N228	7
9	N326	8	N48	13	9	N252	2	N81	12
10	N329	10	N40	14°	10	N256	10	N87	4
11	N263	13	N58	2	11	N257	10	N86	4
12	N249	9	N103	6					
13	N358	8	N93	17					
14	N54	24	N63	37					
15	N236	26	N214	14					
16	N233	22	N201	12					
17	N272	10	N55	4					
18	N308	14	N1	11					
19	N307	12	N14	10					

Rubh an Eirannaich Fig. 6.38					Rubh an Eireannaich Fig. 6.39				
		Bedding $\text{St. N}83^{\circ}$ Dip 7° NE					Bedding $\text{St. N}83^{\circ}$ Dip 7° NE		
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N232 $^{\circ}$	22 $^{\circ}$	N218 $^{\circ}$	26 $^{\circ}$	1	N63 $^{\circ}$	10 $^{\circ}$	N105 $^{\circ}$	10 $^{\circ}$
2	N201	16	N192	23	2	N358	18	N359	11
3	N260	18	N238	19	3	N353	18	N353	11
4	N353	12	N353	5	4	N31	27	N44	22
5	N1 $^{\circ}$	35	N5	28	5	N262	8	N221	10
6	N231	14	N211	29	6	N244	11	N218	15
7	N244	2	N227	23	7	N296	12	N258	10
8	N256	15	N242	17	8	N97	28	N107	30
9	N178	16	N176	23	9	N196	12	N187	18
10	N253	15	N269	18	10	N348	6	N181	10
11	N219	14	N203	19	11	N304	20	N286	16
12	N358	18	N358	11	12	N340	8	N288	2
13	N358	14	N7 $^{\circ}$	7	13	N162	8	N166	15
					14	N105	28	N116	32
					15	N291	30	N278	27
					16	N295	14	N263	12
					17	N256	24	N241	25
					18	N330	14	N312	8
					19	N266	10	N228	12

Loch Eishort Fig. 6.40									
St. N350 Bedding Dip 14SW									
No.	Cross-bed		Corrected		No.	Cross-bed		Corrected	
	Azimuth	Dip	Azimuth	Dip		Azimuth	Dip	Azimuth	Dip
1	N244°	35°	N235°	22°	26	N262°	16°	N293°	3°
2	N234	14	N159	6	27	N276	22	N301	9
3	N226	30	N205	20	28	N204	29	N177	24
4	N260	22	N260	8	29	N262	22	N268	8
5	N33	13	N264	2	30	N227	20	N186	11
6	N276	14	N10	4	31	N235	26	N211	15
7	N255	22	N246	8	32	N206	9	N120	11
8	N206	7	N243	19	33	N244	18	207	6
9	N288	21	N322	11	34	N245	30	N232	17
10	N3°	16	N38°	23	35	245	18	N204	6
11	N227	4	N215	29	36	N255	20	N242	6
12	N245	21	N222	8	37	N242	20	211	8
13	N258	14	N170	1	38	N260	18	N260	20
14	N252	24	N243	11	39	N264	13	N278	4
15	N291	30	N308	18	40	N217	20	N173	13
16	N296	19	N343	12	41	N252	18	N224	5
17	N209	16	N160	11	42	N262	14	N348	1
18	N256	29	N253	15	43	N258	8	N90°	6
19	N238	32	N224	20	44	N258	29	N255	14
20	N233	40	N222	28	45	N269	19	N287	8
21	N247	30	N235	16	46	N246	38	N240	24
22	N246	23	N228	10	47	N270	30	N280	17
23	N242	26	N226	13	48	N268	35	N271	22
24	N281	23	N309	12	49	N272	23	N294	10
25	N87	6	N83	7	50	N295	21	N303	41

APPENDIX 3

X-RAY DIFFRACTION

i. Operational procedures and technique

A Phillips PW1050, X-ray diffractometer was used with nickel filtered $\text{CuK}\alpha$ radiation at 44 kV/22 mA.

One degree divergence and anti-scatter slits were employed with a 0.1 mm. receiving slit. The goniometer speed was $1^\circ 2\theta$ per minute, the diffracted X-rays being detected by a scintillation counter (J. and P. amplifier, single channel analyser and rate-meter; Telsec type V horizontal bed recorder) and passed through a pulse height analysing unit prior to recording on a strip chart. The samples were scanned between $2^\circ 2\theta$ and $20^\circ 2\theta$ generally.

In an attempt to equalise the beam intensity experienced by the sample over all diffraction angles, Shaw (1971) suggested a scheme of goniometer slit adjustments; he varied sizes of divergence and anti-scatter slits with diffraction angles. This practice was not followed in order to simplify the method and bring it into line with the usual practice of maintaining constant slit width. The operating conditions under which the diffractograms were measured are given below:

J. and P. amplifier, single-channel analyser and rate-meter.

Telsec type V horizontal bed recorder.

Full scale deflection : 10^3 (3×10^3) counts

Time constant : $\times 1$

Goniometer speed : $2^\circ 2\theta \text{ min}^{-1}$.

Divergence and anti-scatter

slits : 1°

Receiving slit : 0.1 mm.

Recorder input : 100 mV

Chart speed : 1 cm. min^{-1} .

ii. X-ray powder photography

X-ray powder photography with $\text{FeK}\alpha$ radiation, was performed on one sample (Am66-1) and $\text{CuK}\alpha$ was used for two others (Am79, 88) using a standard Phillips, large Debye Scherrer Powder Camera (PW1024) with a diameter of 114.60 mm.; this was equipped with a standard entry collimator and beam-exit tube.

The powdered specimen was introduced into capillary tubes with 0.3 mm. diameter (Pentax (EMI) Ltd.) directly (diluent was not used). The specimen was mounted and placed onto a camera bracket (PW1012/10), exposed to X-rays and rotated for 48 hours with a synchronous motor (1 rev. min⁻¹.) (PW1033).

The Straumanis technique of film calibration was followed (see Zussman, 1967).

APPENDIX 4

CLAY MINERALOGY

i. Sample preparation

The clay and shale samples obtained were disaggregated in deionised water by mechanical means. The carbonate content of the shales was removed according to the method proposed by Bodine and Fernald (1973) and a few drops of sodium hexametaphosphate were added to prevent the flocculation of the clay minerals. The $<2\mu\text{m}$ fraction of the assemblage was then separated by centrifuging in an I.E.C. centrifuge machine; time and speed being determined from sedimentation nemographs produced by Tanner and Jackson (1947) which give linear plots representing different centrifuge rotational speeds, and relate the minimum particle size sedimented with the function of time and centrifuge parameters ($t/\ln(R/s)$). Obviously the basic assumption by Tanner and Jackson (1947) using Stokes Law, that the particles are represented by a spherical diameter is a shortcoming for platy minerals. Also, sidewall retardation effects in a small centrifuge tube may be quite significant; however it is considered that a reproducible separation on the order of 90% effectiveness may be obtained after three such treatments for each fraction. Samples of the separated clay fractions in suspension were mounted for clay mineral determination.

ii. Clay sample mounting

Clay minerals have structures consisting of stacked, two dimensional aluminosilicate layers, thus their basal reflections in the X-ray diffraction pattern may be enhanced by the alignment of their 001 planes parallel to the face of the mounted specimen. Methods of mounting clay minerals for XRD analysis have been studied by Gibbs (1965, 1968), Carroll (1970), Stokke and Carson (1973) and Bajwa and Jenkins (1978). Recent work (Towe, 1974) has shown that the most suitable preparation methods are those which eliminate the

segregation of clay minerals according to their particle size and morphology. Such methods yield an inhomogeneous specimen and consequently erroneous XRD analytical results. The technique of sample mounting chosen for this study follows that proposed by Shaw (1972). Glass wool, broken into short lengths and mixed with the clay suspension before mounting (Rich, 1975), helped to minimise the curling of some dried specimens.

iii. Clay mineral identification

Clay mineral identification follows the standard four-stage (air dried, glycolated, heated at 400 and 550°C) procedure of obtaining X-ray patterns for each sample (Biscaye, 1965; Millot, 1970). The following table was used in identifying the various types of minerals present (Molloy and Kerr, 1961; Biscaye, 1964; Millot, 1970; Bath, 1973; Yan-Chen, 1976; Thorez, 1976).

Air dry	Glycolated	400°C	550°C	Boiled in HCl	Mineral
~ 14Å	No change	No change	Increased intensity	All reflections	Chlorite
~ 14Å	Expands to 18Å	Collapses to 10Å	No change, or slight collapse	May dissolve	Smectite
> 10-14Å	No change	No change	No change or slight increase	No change	Regular I-Ch
> 10-14Å	No change	No change	No change	May dissolve	Rand. I-Ch.
> 10-14Å	Expands to higher spacing	Collapses to 10Å	No additional collapse	No change	I-Smect.
~ 10Å	Non-symmetrical peak, some shift	No change	No change slight sharpening	No change	I(musc.)
~ 7Å	No change	No change	Destroyed	No change	Kaolinite

Although the samples were not boiled in 10% HCl, the effects of the various heating treatments on intensities, shapes and positions of diffraction peaks were used as an aid in identifying all or part of the peaks belonging

to specific clay mineral types. The effect of ethylene-glycol treatment on smectites is to displace the interlayer water, replacing it with the organic molecule and shifting the (001) peak to $\sim 17\text{\AA}$; an advantage of the ceramic mounts (Shaw, 1972) used in the mounting of the clays in this study, is that glycol liquid can be applied directly, the excess being absorbed by the ceramic base; for many other mounting techniques, glycollation must be carried out by vapour. The interlayering in mixed layer illite-smectite clays is a complex matter and their estimation involves distinguishing between random and regular-interstratification (Reynolds and Hower, 1970); the nature of any mixed layer clays present has not been investigated in detail here. Heating at 400°C causes the collapse of expandable lattices to their basal spacing by expelling the interlayer water or glycol molecules.

The kaolinite lattice is destroyed at 550°C and the $7\text{\AA}$, kaolinite (001) peak is lost. The effect of heat-treatment on various clay minerals is investigated by Austin and Leininger (1976).

iv. Quantification

Methods used in determining the proportions of clay minerals present in a rock sample are semiquantitative and vary among workers (Schultz, 1964; Moore, 1968; Pierce and Siegal, 1969; Devine et al., 1972; Stokke and Carson, 1973; Levy and Francis, 1975; Weir et al., 1975; Kazi, 1975; Austin and Leininger, 1976).

Johns et al. (1954) applied consideration of structure and the variation in scattering with diffraction angle to the estimation of inter-peak intensity factors.

Schulz (1964), Biscaye (1965), Weaver (1967) and Shaw (1971) determined the amounts of clay minerals present in their samples by estimating the intensity factors listed below:

$18\text{\AA}$ pattern	$10\text{\AA}$ peak area	$7\text{\AA}$ peak area	Reference
1 X	4 X	2 X	Biscaye (1965)
1 X	4 X	1.6 X	Weaver (1967)
1 X	4.5 X	*4.5 X)	Schultz (1964)
1 X	4 X	2.2 X)	
1 X	4 X	1.6 X	Shaw (1971)
1 X	4 X	1.6 X	Bath (1973)
1 X	4 X	2 X	This study

* Value depends on the crystallinity of kaolinite.

As it is seen all five authors have estimated that different clay phases at the same diffraction angle will have the same intensity factor, i.e. collapsed smectite, mixed-layer illite-smectite and illite at $10\text{\AA}$ all have the same reflection efficiency as do kaolinite and chlorite at $7\text{\AA}$. No unified approach as suggested by Pierce and Siegel (1969) has yet been developed on an inter-laboratory basis, therefore the significance of the semiquantitative results in this study should be regarded as the relative compositions of samples within this study only.

The computation method adopted here utilises estimated intensity factors to correct the intensities of diffraction peaks and thus produce values which may be compared quantitatively, although there is some diversity in the actual factor used, this is the most widely applied technique and is rapid.

Weighted integrated diffraction peaks were used in this study, following Bradshaw (1975); the computations of the amount of each mineral present (in %) are as follows:

$$10\text{\AA} (G) = I \%$$

$$7\text{\AA} (G) \times \frac{1}{2} = K \%$$

$$10\text{\AA}(550) - 10\text{\AA}(G) = IM + M = Ex \%$$

$$18\text{\AA} (G) \times \frac{1}{4} = M \%$$

$$Ex \% - M \% = I M \%$$

v. See enclosed reprint.

APPENDIX 5ATOMIC ABSORPTION SPECTROPHOTOMETRY

A Perkin-Elmer 306 Atomic Absorption Spectrophotometer was used in conjunction with the Special Burner Control Option. A Deuterium Arc Background corrector is installed for use with certain elements. Filter slit and range settings used were all as noted in the standard conditions section of the Perkin-Elmer Analytical Methods book which are given at the end of this Appendix. Perkin-Elmer single element "Intensitron" hollow cathode lamps were used throughout, at the recommended source currents. The wavelength settings are especially indicated at the end of this appendix as those given by the instrument controls do not correspond to the true wavelengths.

Standard solutions giving the required range of concentrations were prepared by volumetric dilution of BDH standard solutions for atomic absorption spectrophotometry.

Matrix differences especially of acid concentration, affected the analytical accuracy probably by the alteration of solution viscosity, causing variations of solution uptake rate and consequently of atomic population in the flame. Matrix interferences can often be controlled by diluting the sample solution until the effect of dissolved salts or acids become negligible. Where dilution is not possible, matrix effects can be controlled by matching the concentration of major constituents in a sample and the standard solutions. In this study dilution was used to minimise matrix effects and in addition, a dispersing agent (normally KCl solution with a concentration of 1000 ppm) is used for each sample.

The sample preparation procedure is as follows; it should be mentioned that only the carbonate fraction of each specimen was analysed.

1. Sample "chips" are powdered by mechanical grinding.
2. Select 5 to 10 gr. for washing (collect in separate polythene bag).

3. Select 25 ml. beakers, wash, dry, leave to cool then weigh.
4. Take 2-3 full spatulates of powdered sample, place into appropriate beaker and pour deionised water on each (up to 1/3), stir well and leave to settle.
5. Decant and leave in oven to dry.
6. Remove cake from beakers and place in pre-labelled bags making sure that they are powdered (rub against fingers).
7. Prepare a solution of 25% HCl.
8. Weigh accurately 1 gr. of each sample and place into clean, weighted beakers; cover with polythene paper.
9. Pour 10 cc. of 25% HCl over each sample and stir vigorously.
10. After the reactions have simmered, cover each beaker and leave for 12 hours.
11. Prepare a 1000 ppm concentration of KCl solution.
12. Select 100 ml. container flasks and filter funnels, rinse thoroughly in deionised water, select no.42 filter paper and place in funnels.
13. Place 5 ml. of K solution in each flask, including the blank.
14. Wash out beakers into filter funnels, allow sediment to filter; repeat the washing of beakers as necessary until no residue remains in them.
15. Top up flasks to 100 ml.
16. Transfer filter paper + sediment into its corresponding beaker including the blank and allow to dry in oven.
17. Weigh the beakers plus filter paper, plus sediment in order to calculate the % weight lost.
18. Prepare 2, 4, 6, 8, 10 and 30 ppm standards from BDH solutions for each element (the blanks are prepared in 100 ml. flasks).
19. Run on AA spectrophotometer.

It should be noted that standards used for Mg determination are 0.02, 0.04, 0.06, 0.08 and 0.1 ppm concentrations and the prepared solutions are all

diluted by 50 according to their concentrations.

An alternative method for analysing samples for major elements is as follows:

1. Exactly weigh 0.125 gr. of rock powder into a crucible (5% AuPt) and mix with 1.000 gr. of "specure" lithium metaborate (van Loon and Parissis, 1969; Ingmells, 1970).
2. Heat in oven to 1000°C (1 hr. to 3 days) until reaction is complete.
3. Pour fused melt directly into a glass beaker containing about 100 ml. of 50% v/v HCl placed on magnetic stirrer hot plate at 60°C and allow the sample to dissolve.
4. Cool.
5. Dilute volumetrically to 250 ml. with 50% v/v HCl and store in polypropylene bottles.

No deterioration of the stored solution will occur over long periods up to three months.

If the above procedure is followed, the matrices of prepared standard solutions ^{can} be made identical to those of the samples by the addition of the required quantities of "specure" lithium metaborate and HCl.

In order to remove chemical interferences in complex solutions, lanthanum may also be used (Omang, 1969). A solution containing 10% w/v lanthanum ($26.6 \text{ LaCl}_3 \cdot 7\text{H}_2\text{O}$) should be added to both standard and sample solutions to give a final concentration of lanthanum of between 1% w/v and 5% w/v depending on the element analysed.

One or more rocks of known chemistry should be analysed during the course of the analysis of the unknowns in order to provide an independent method of checking the accuracy of the analyses.

Acetylene is the fuel used for all analyses. The cylinder pressure should never be allowed to fall below 75 PSIG and the cylinders should be renewed when the pressure falls to 100 PSIG; failing to do so results in high

background level readings due to the release of acetone.

Details for the preparation of standards from BDH solutions are as follows:

BDH ul. St. ppm	*Ca	*Mg	Mn	Fe	Sr	K Sol. (1000 ppm)
2	100	100	100	100	100	5 ml.
4	200	200	200	200	200	5 ml.
6	300	300	300	300	300	5 ml.
8	400	400	400	400	400	5 ml.
10	500	500	500	500	500	5 ml.
20	1000	1000	1000	1000	1000	5 ml.
30	1500	1500	1500	1500	1500	5 ml.
Blank	0.00	0.00	0.00	0.00	0.00	5 ml.

* Ca and Mg were highly concentrated in the solutions, therefore they were volumetrically diluted.

Container flasks (50 ml.) with the 5 ml. K solution (1000 ppm) were selected and 1 ml. of each solution was transferred into them; new standards were also prepared with 0.02, 0.04, 0.06, 0.08 and 0.1 ppm concentrations. All analyses were determined by calibration curve. Instrumental settings are as follows:

Setting Element	Wave- length	Region	Operating lamp cur- rent (mA)	Slit width	Oxidant	Flame type	Air/ Na ₂ O flow (lit/ min)	Fuel flow (lit/ min)
Ca	211x2	VIS	10	4	Air	lean blue	55	32
Mg	285	UV	6	4	Air	"	55	32
Mn	279	UV	20	3	Air	"	55	32
Fe	248.3	UV	30	3	Air	"	55	32
Sr	230x2	VIS	20	3	Air		52	44

APPENDIX 6

WET CHEMICAL ANALYSES

The ferrous iron, phosphate (P_2O_5) and carbonate carbon (as CO_2) of the samples were determined as follows:

i. Ferrous iron

The method described by Wilson (1955) was used here. It involves the decomposition of samples by cold HF in the presence of excess ammonium metavanadate, which oxidises the ferrous iron in situ and prevents any atmospheric oxidation. The excess is then back titrated with standardised ferrous ammonium sulphate. This method is unsuitable for carbonaceous rocks and gives erroneous results if any reducing agents other than ferrous iron, such as non carbonate carbon or sulphide is present (Groves, 1951). The sulphur content of the rocks under study is negligible and it is assumed that the non carbonate carbon content does not exceed the maximum mean value of 0.81%; thus the error introduced by the existence of non carbonate carbon is small and although the total ferrous iron does not exceed 19% per weight, the slight inaccuracy of the result is compensated for by the convenience and rapidity of the method used. It is impossible to determine to what extent the ferric iron content of the rock is "original" and how much is due to recent weathering, thus the significance of the ferrous iron values is uncertain.

Weinberg (1973) showed that the grinding of these samples in the agate swing mill does not cause their oxidation.

The ferric iron content is given by the following relationship:

$$Fe_2O_3 = \text{Total Fe (as } Fe_2O_3) - \frac{FeO \times \text{Mol. wt. } Fe_2O_3}{2 \times \text{Mol. wt. } FeO}$$

Each of the analyses were duplicated and gave a maximum deviation of 0.05.

ii. Phosphate

The methods normally used to determine the phosphorus (as P_2O_5) in rocks could not be employed here due to the formation of insoluble residue of calcium and magnesium phosphates and pyrophosphates which considerably affect the end results. The conventional methods involve the decomposition of fine rock samples in $HF/HClO_4$ "bombs" prepared in teflon containers and tightly clamped in bomb vessels. An alternative procedure was followed which involved the fusion of a known amount of sample with Spectroflux 100 ($LiBO_2$), acid dissolution by 50% v/v HCl and the determination of P_2O_5 by colorimetric methods. The procedure is described in detail below:

1. Wash Au-Pt crucibles in HCl , dry in oven and allow to dry for 20 min.
2. Weight crucibles accurately and then add 3 gr. of $LiBO_2$ (Johnson-Mathey Chemicals, Spectroflux type 100A) to each.
3. Place crucibles with Spectroflux 100 in oven at $1000^{\circ}C$ for $1\frac{1}{2}$ hr., remove, place on asbestos plates, cover with tops and allow to cool in dessicator.
4. Weight accurately, crucible and flux.
5. Add 0.3 gr. sample.
6. Place in oven at $1000^{\circ}C$ for 3 hours, checking to stir at equal intervals.
7. Remove from oven and cover, allow to cool (initially on asbestos plates then in dessicator).
8. Weight accurately.
9. Place crucible + sediment + flux in 100 cc. beaker and add 50 cc., 50% HCl and heat (do not boil), allow the acid to dissolve flux and sediment.
10. Remove crucibles.
11. Heat to evaporate all but 5 to 10 ml. of solution.
12. Transfer to Au-Pt dishes.
13. Add 10 ml HF , 5 ml. $HClO_4$.
14. Evaporate to near dryness.

15. Add a few drops of water.
16. Repeat steps 13 and 14 if necessary.
17. Add 1 to 2 drops of HCl.
18. Make up to volume (100 ml).
19. Transfer to plastic beakers.
20. Determine P_2O_5 calorimetrically.

The P_2O_5 content was determined colorimetrically after the formation of a phosphomolybdate complex and its reduction to "Molybdenum blue". Each analysis was triplicated and this produced a maximum standard deviation of 0 to 0.1.

iii. Carbonate carbon

The carbonate carbon (as CO_2) was determined gravimetrically following Bauer et al. (1972).

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Origin of montmorillonite in the early Jurassic shales of NW Scotland

K. AMIRI-GARROUSSI

Summary. Clay mineral analysis of shales in the early Liassic Lower Broadford Beds of the Hebrides reveals the presence of abundant quantities of montmorillonite whereas the Upper Broadford Beds contain illite, kaolinite and subordinate mixed layer minerals together with chlorite. Montmorillonite enrichment of the Lower Broadford Beds as a consequence of recent weathering contamination by doleritic sills can effectively be ruled out. It is argued that the montmorillonite may have been derived from the weathering of basic igneous rocks exposed in the sediment source area during Late Triassic times. By Upper Broadford Beds times, the igneous source was either eroded away or transgressed by the early Liassic sea.

1. Introduction

Clay mineralogical analysis of the Lower Broadford Beds (Upper Planorbis – Lower Bucklandi zones) of NW Scotland (Hallam, 1959) has unexpectedly revealed the presence of a significant proportion of montmorillonite in the argillaceous rocks, together with mixed-layer illite-montmorillonites, illite and kaolinite. Previous reliable work on English and Scottish Liassic shales (Hallam, 1960; Perrin, 1971; Sellwood, D.Phil. thesis 1970, University of Oxford; Lewis, Ph.D. thesis 1972, University of Wales; Cosgrove, 1975) has revealed illite and subordinate kaolinite to be virtually the only constituents.

2. Stratigraphic setting and lithology

The Broadford Beds (Hettangian-Sinemurian) are found as small, widely spaced outcrops in two major areas of NW Scotland, here termed the northern region (Skye, Applecross, Raasay) (Lee, 1920; Hallam, 1959) and the southern region (Mull, Morven, Ardnamurchan (Lee & Bailey, 1925; Richey & Thomas, 1930; MacLennan, 1954); Figure 1.

The Lower Broadford Beds consist of bioclastic, micritic and oolitic limestones alternating with sandstones containing scattered pebbles. They were deposited under shallow water, marginal marine conditions whereas during the deposition of the more argillaceous Upper Broadford Beds (Semicostatum-Turneri zones) fully marine conditions were established.

Successive overlap within a short distance, of the Lower and Upper Broadford Beds onto a variety of pre-Triassic rocks, suggests that variable relief existed in much of the northern region. The earliest shallow marine Mesozoic transgression extended over the southern region in Rhaetian times; but did not reach the northern region until the late Hettangian (?) and early Sinemurian.

Shale beds in the Lower Broadford Beds of the northern region are only found at Applecross on the mainland, and in the Isle of Raasay. In the southern region shales were deposited from Upper Planorbis to Lower Bucklandi zone times; coalified plant material is abundant in the alternating limestones and shales of both regions.

The association of calcilutites and shales on the one hand with oolitic limestones and pebbly sandstones on the other, suggests shoal areas affected by variable current activity and free from terrigenous influx, alternating with periods when abundant land-derived material was transported into the sea under turbulent conditions. It is probable that while sedimentation in a shallow water, partially enclosed, marginal marine basin was taking place the hinterland underwent periodic, short-lived phases of uplift and erosion, thereby providing the basin with thin, imper-sistent beds of coarse, poorly sorted conglomeratic material.

The Liassic and underlying non marine Triassic rocks of the northern region, rest unconform-

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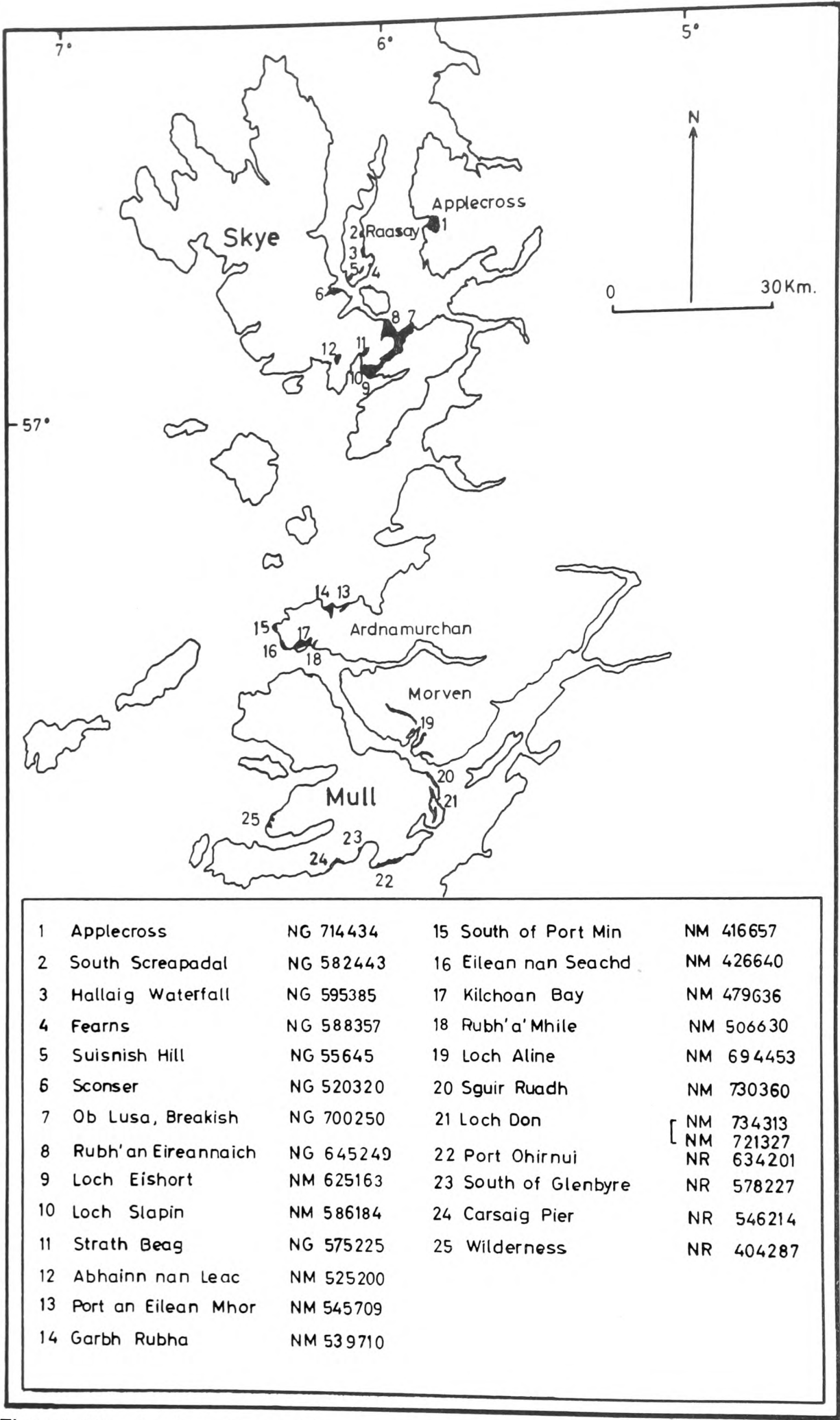


Figure 1. Map showing outcrops of the Broadford Beds of NW Scotland.

ably on a thrust complex of Cambro-Ordovician Durness Limestone and late Precambrian Torridonian Sandstone. Together with the strata of the southern region, they have been intruded by numerous early Tertiary dolerite dykes and sills. A smaller number of granitic intrusions are also present (Stewart, 1965). As a result the shales adjacent to the intrusions are baked but, in general, thermal alteration is very local.

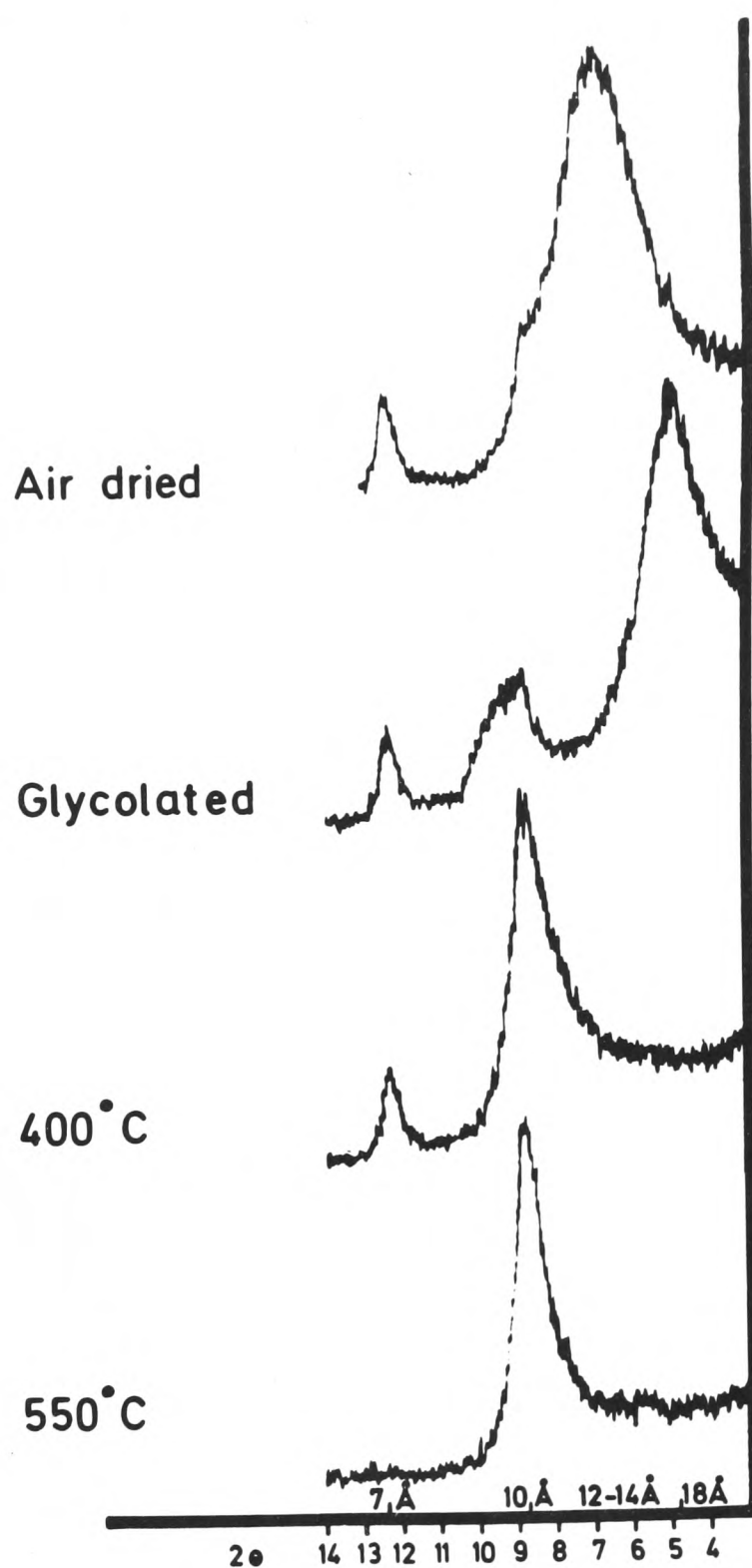


Figure 2. X-ray pattern obtained from a representative sample (M90) from the Upper Planorbis zone of Wilderness (Western Mull). The 14 Å peak shifts to a larger basal spacing after glycolation, indicating the presence of montmorillonite, the other clay minerals, i.e. illite (10 Å) and kaolinite (7 Å) are not affected. Heat (400 °C) destroys the 14 Å peak. At 550 °C the 14 and 7 Å peaks are destroyed, the 10 Å peak is enhanced due to the large proportion of mixed layer illite-montmorillonite (collapsed) present.

3. Laboratory technique

Specimens were obtained at approximately 1 m intervals, in all major sections of both Lower and Upper Broadford Beds, additional samples being taken to determine local variations of lithology.

Sixty-four shale samples from the Broadford Beds were disaggregated in deionized water by mechanical means. After removing the sulphate and carbonate content of the shales (Bodine & Fernald, 1973), a few drops of sodium-hexametaphosphate were added to prevent the flocculation of the clay minerals. The $< 2 \mu\text{m}$ fraction of the assemblage was then separated by centrifuging and subsequently mounted on ceramic tiles following the method proposed by Shaw (1972).

A Phillips X-ray diffractometer was used with nickel-filtered $\text{CuK}\alpha$ radiation at 44 kV/22 mA. Mineral identification follows the standard four stage (air dried, glycolated, heated at 400 and 550 °C) procedure of obtaining X-ray patterns for each sample (Biscaye, 1965; Millot, 1970).

Figure 2 shows a typical X-ray pattern obtained from a sample of the Lower Broadford Beds. Methods used in determining the proportions of clay minerals present in a rock sample are semiquantitative and vary among workers (Schultz, 1964; Pierce & Siegel, 1969; Stokke & Carson, 1973; Weir, Ormerod & El Mansey, 1975; Austin & Leininger, 1976). In this study weighted integrated diffraction peaks were used following Bradshaw (1975).

4. Results

The Upper Broadford Beds contain only illite and kaolinite with some chlorite minerals. In contrast 22 analysed samples from the Lower Broadford Beds contain in addition to illite, kaolinite and mixed-layer illite-montmorillonites, montmorillonite in substantial quantities of up to 30% (Table 1, fig. 3). A slight increase in the overall proportion of montmorillonite is recognizable in the northern region. Depending upon the variations in the chemistry of montmorillonite, its air-dried 001 peak position is obtained at intervals between 12–14 Å (Millot, 1970). This peak shifts to 18 Å after glycolation and entirely collapses to a 10 Å basal spacing after heating up to 400 and 550 °C. Although the Lower Broadford Beds in Raasay are not represented on Figure 3 they have also been found to contain up to 20% montmorillonite. It is significant that samples obtained from red mudstones (Rhaetic) of Wilderness (locality 25, Fig. 1) contain montmorillonite in addition to illite and kaolinite.

5. Discussions

Because of the presence of Tertiary basic igneous intrusions, which might be expected to weather to montmorillonite, the possibility of weathering contamination of the adjacent shales must be considered.

Montmorillonite is known to form at present by the weathering of basic igneous rocks in an alkaline environment in the presence of Mg (Millot, 1970).

A relatively unweathered doleritic sill 1 m thick occurs about 50 m above the base of the Lias at Applecross and a deeply weathered doleritic sill was observed 2.50 m above the Lias/Trias contact at the same locality (NG 725447). Although a characteristic colour zoning (Hourang & Hatton, 1974) was absent, a yellowish-brown weathering friable core surrounded by a concentric layer of grey sandy shale-like material could be recognized at the latter locality. The clay mineralogy of samples taken progressively outwards from the core shows a systematic decrease of montmorillonite and the formation of mixed-layer minerals, kaolinite is absent from the core itself. The montmorillonite content of the core is a product of *in situ* weathering and partial alteration of the dolerite to clay minerals. The montmorillonite enrichment of a thin zone underlying the two sills of Applecross could therefore possibly be related to recent weathering contamination.

Table 1. Proportion of clay minerals in the argillaceous rocks of the Broadford Beds (U = Upper Broadford Beds, L = Lower Broadford Beds)

Sample	Outcrop	Montmorillonite (%)	Chlorite (%)	Illite (%)	Kaolinite (%)	Illite-montmorillonite (%)
Skye						
U177	7	—	—	100	—	—
U168	7	—	—	84	16	—
U167	7	—	—	82	18	—
U166	7	—	7	62	21	10
U162	7	—	—	80	16	4
U161	7	—	—	66	21	13
U159	7	—	—	82	18	—
U155	7	—	—	92	8	—
U154	7	—	—	73	13	14
U153	7	—	—	84	16	—
U146	7	—	—	100	—	—
U144	7	—	—	97	—	3
U125	7	—	—	80	20	—
U114	7	—	—	100	—	—
U107	7	—	—	100	—	—
U104	7	—	—	100	—	—
U79	9	—	—	76	24	—
U77	9	—	—	82	18	—
U76	9	—	—	80	20	—
U75	9	—	29	55	15	1
U74	9	—	—	85	15	—
U71	9	—	9	46	29	16
U70	9	—	8	73	19	—
U68	9	—	—	91	9	—
U67	9	—	—	84	16	—
U65	9	—	—	63	24	—
U64	9	—	—	89	11	—
U63	9	—	—	85	15	—
U61	9	—	—	85	15	—
U59	9	—	—	90	10	—
Ub64	9	—	44	51	3	2
Ub58	9	—	16	84	—	—
Ub56	9	—	—	100	—	—
Ub37	9	—	—	68	17	15
Ub28	9	—	—	83	—	17
Applecross						
L34	1	14	—	19	11	56
L31	1	40	—	33	11	16
L29	1	24	—	40	3	33
L27	1	24	—	39	7	30
L16	1	33	—	36	7	24
L10	1	32	—	47	7	14
L44	1	11	—	31	3	50
L42	1	22	—	37	3	38
L39	1	27	—	28	—	45
Raasay						
U31	3	—	10	45	23	22
U21	3	—	18	48	15	19
L11	3	26	—	53	1	20
L7	3	2	—	40	2	56
Mull						
U76	20	—	—	100	—	—
U63	20	—	11	78	11	—
U19	21	—	10	78	12	—
U16	21	—	18	78	3	1
U14	21	—	24	51	9	6
L90	25	29	—	28	6	37
L89	25	28	—	24	6	42
L88	25	23	—	38	7	32
L87	25	17	—	32	8	43
L86	25	20	—	37	6	37
L84	25	19	—	29	5	47
L83	25	25	—	32	6	37
L4	19	21	—	40	10	29
L2	19	18	—	29	7	46

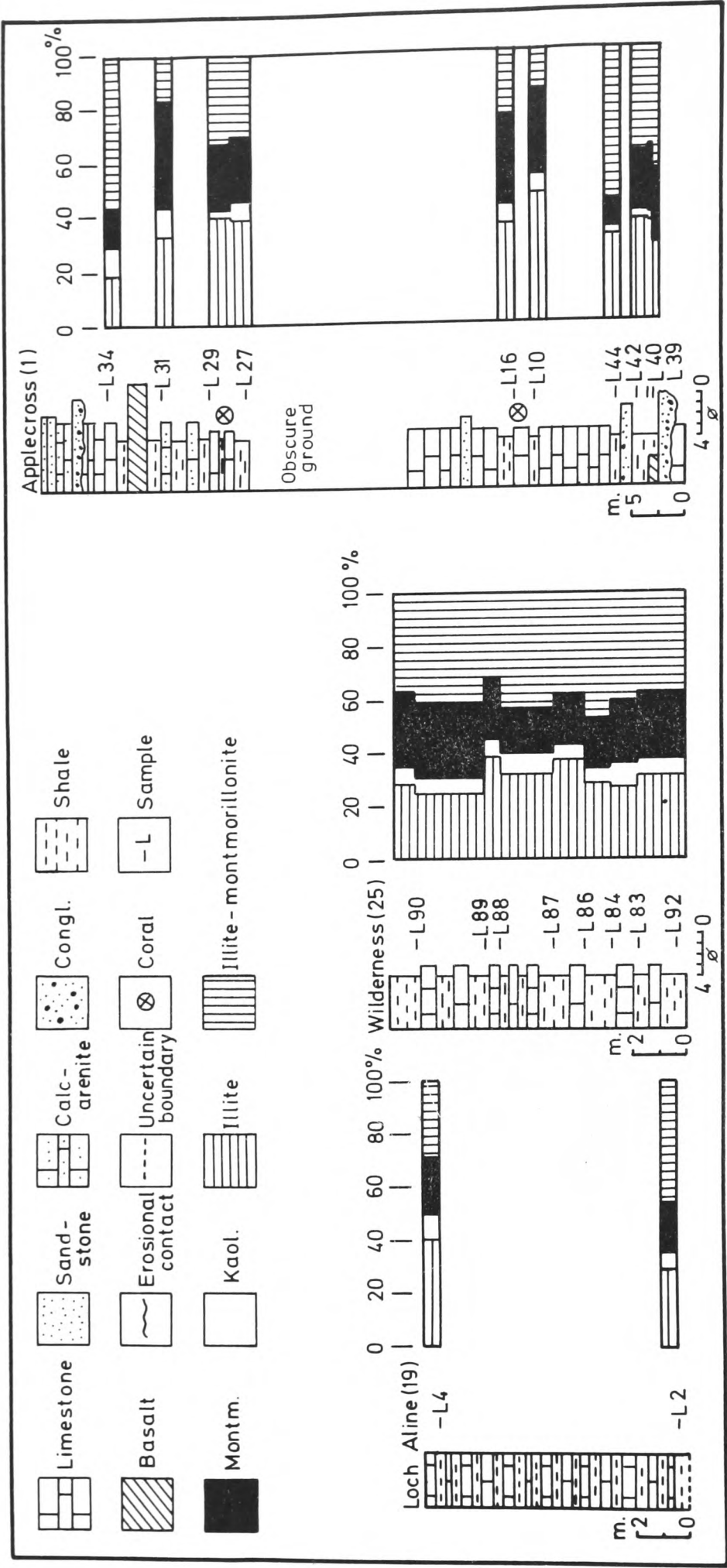


Figure 3. Clay mineral content of the Lower Broadford Beds (Upper Planorbis - Lower Bucklandi zones). Location of sections given in parentheses.

Thermally induced chemical transformations in the immediate vicinity of igneous intrusions into shales are directly related to the composition of the intrusion. An association of quartz+albite+paragonite+muscovite+chlorite is neoformed in montmorillonite bearing shales, under controlled P/T conditions (Althaus & Johannes, 1969). Short-lived high temperature gradients, present at the time of emplacement can create overpressures due to dehydration around the periphery of the sills. The result tends to be fissure emanations and haphazard contamination of sediments above the intrusions.

With regard to the Broadford Beds, a systematic disappearance of kaolinite together with the diminution of mixed-layer illite-montmorillonite near the intrusions (Correia & Maury, 1975; Sauvin, Esquevin & Chenoux, 1975) was not observed. Chlorite and interstratified chlorite-dioctahedral smectites (Blatter, Robertson & Thompson, 1973) are totally absent, and no evidence of metasomatic alteration is seen above and below the sills in Applecross. The overall clay mineral distribution pattern is persistent throughout the stratigraphic sections and there is no systematic enrichment of the sediments in certain minerals with respect to their distance from intrusions. This more or less uniform distribution of montmorillonite in the Lower Broadford Beds, whether intrusions are present or not, together with its total absence in the Upper Broadford Beds, calls for an explanation genetically related to the early Liassic depositional environment.

Montmorillonite may be formed through the dissolution of silica from opaline sponge spicules or flint, removal of alkalic and alkaline earths from glauconite, mica or feldspars and the dissolution of CaCO_3 (Brown, Catt & Weir, 1969). Although feldspars are present in the Lower Broadford Beds together with mica, there is no trace of flint or sponge spicules, and glauconite is rare.

The post-depositional diagenesis of degraded micas in the presence of waters of low ionic strength may also form montmorillonite (Perrin, 1971). This process may possibly be effective in porous sandstones but it has no relevance to the low-porosity shales and limestones of the Lower Broadford Beds.

Montmorillonite may form by the terrestrial and subaqueous alteration of basic volcanic material (Ross & Shannon, 1926; Brindley, 1957; Slaughter & Early, 1965; Hallam & Sellwood, 1968; Biscaye, 1965; Millot, 1970; Perrin, 1971; Bradshaw, 1975). The presence of glass shards is the most direct and convincing evidence for the association of montmorillonite with such igneous activity, and is reported from some recent sediments (Millot, 1970). In the marine environment, volcanic glass undergoes hydrolysis to give rise to montmorillonite and the excess silica produced by such a reaction may appear in the sediment as cristoballite (Brindley, 1957), remain amorphous or combine with Al and alkaline earths to form zeolites. X-ray analysis of whole-rock samples has not revealed the presence of any of the above in the Lower Broadford Beds; only 3.22 and 3.26 Å peaks have been recorded which are generally considered to be characteristic of sanidine. If present, sanidine may be related to volcanic rocks and their corresponding tuffs; however, more evidence is needed to confirm this mineral's presence in the Lower Broadford Beds. No claim is made here for the presence of true bentonites.

Illite-montmorillonite commonly forms from montmorillonite through burial diagenesis (Pollard, 1971), it is readily seen that such a process could provide the Lower Broadford Beds with an original montmorillonite content of up to 70%.

The near-uniform distribution of montmorillonite in all sections, its low degree of cristallinity and the high proportion of mixed-layering together with illite suggests an origin based on the erosion of previously exposed basic igneous rocks, or a previously formed montmorillonite-rich source; the disappearance of montmorillonite in the Upper Broadford Beds would then indicate an elimination of this source. Palaeocurrent data from the calcarenites and pebbly sands,

together with the facies distribution during the Lower Broadford Beds times, indicate a probable primary sediment source in the N or NE.

6. Tectonic considerations

The commencement of continental breakup in the southern parts of the North Atlantic is indicated by subsidence and local marginal faulting associated with graben structures and volcanism in the Late Triassic (Hallam, 1971; Pitman & Talwani, 1972). Major crustal rifting associated with the extrusion and intrusion of basaltic lavas and sills in the Late Triassic is widely reported from the Connecticut Basin and the Newark Group of Eastern United States (Rodgers, 1970), the Gulf of Maine and Georges Bank (Ballard & Uchupi, 1975), the continental margin off Nova Scotia and Newfoundland (Jansa & Wade, 1975) and the Essaovira Basin, Morocco (Ambroggi, 1963); this tensional regime evidently extended further north into the Labrador Strait (Henderson, 1973).

Although less information is available for the British area, recent investigations have provided evidence for a comparable tensional horst and graben regime, initiated in Late Triassic times (Bott, 1970; Audley-Charles, 1970; Steel, 1971, Ph.D. thesis, University of Glasgow; Hallam & Sellwood, 1976).

Triassic basins developed along pre-existing Caledonian and Hercynian trends in the Hebrides of Scotland (Audley-Charles, 1970). It has been shown (Steel, 1971, Ph.D. thesis, University of Glasgow; Steel & Wilson, 1975; Steel, Nicholson & Kalandar, 1975) that contemporaneous vertical movement along these lines of crustal weakness created graben structures which were subsequently infilled longitudinally by floodplain deposits and laterally (marginally) by alluvial fan deposits.

Striking similarities exist between the eastern American and northwestern Scottish Triassic:

(i) Both systems consist of a series of fault-controlled grabens with similar basin-filling processes operative.

(ii) Both partly overlie crystalline basement rocks and partly nonmetamorphosed rocks.

(iii) Faults which bound the blocks are continuations of older lines of crustal weakness.

Despite the above similarities, the widespread basic igneous activity associated with the Triassic of eastern United States and elsewhere has not been documented in NW Scotland, and episodes of volcanic activity in the Mesozoic of the British area have not been reported earlier than Middle Jurassic (Howitt, Aston & Jacqué, 1975). However, K–Ar ages obtained from alkaline dykes in western Norway (Faereth, MacIntyre & Naterstad, 1976) are in the range of 275–165 Ma B.P.; a period of maximum igneous activity at 220 Ma B.P. is also recognized.

7. Conclusions

The Lower Broadford Beds of NW Scotland contain up to 30% montmorillonite. The intrusion of Tertiary doleritic sills has not affected the montmorillonite content of these rocks and its presence is related to the genetic environment of the shales. Older strata with a considerable montmorillonite content appear to be absent from the region. Lack of episodic or pulse-like increase in montmorillonite content, cf. Bradshaw (1975), or more direct evidence of igneous detritus in the sediments eliminate contemporary volcanism as a likely source.

A comparison with conditions operating during the creation of early Mesozoic continental margins, however, renders quite plausible the possibility of igneous activity in northwest Scotland during the late Triassic. It seems quite probable that as a source area with exposed basalts was being eroded, it was progressively transgressed by the Lower Broadford Beds to become completely submerged by Upper Broadford Beds times. Alternatively the source was completely eroded by then. The apparent absence of feeder dykes is worrying, but much of the

relevant area is covered by sea. Further information from offshore investigations in the Hebrides is needed to reveal more about the tectonic history of this region and perhaps throw more light on the origin of montmorillonite in the Lower Broadford Beds.

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Department of Geology and Mineralogy
University of Oxford, Parks Road
Oxford, OX1 3PR, England