Early Devonian stratigraphy, conodont faunas and palaeogeography of the Capertee High

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EARLY DEVONIAN STRATIGRAPHY, CONODONT FAUNAS AND PALAEOGEOGRAPHY OF THE CAPERTEE HIGH

A thesis submitted in fulfilment of the requirements for the award of the degree

DOCTOR OF PHILOSOPHY

from

THE UNIVERSITY OF WOLLONGONG

by

GARY PHILIP COLQUHOUN (BSc)

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1998
"To see a World in a grain of sand,  
And a Heaven in a wildflower,  
To hold Infinity in the palm of your hand,  
And Eternity in an hour."  
- William Blake

"I never saw the moor,  
I never saw the sea,  
Yet I know how the heather looks,  
And what a wave must be."  
- Emily Dickinson

"In nature's infinite book of secrecy,  
A little I can read."  
- William Shakespeare

"It is only by making long excursions in time and space that we come to understand the place in which we live."  
- Henri Poincaré
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View looking south up the Carwell Creek valley from GR 771000 6364900. In the foreground, prominent beds in the Early Devonian Riversdale Volcanics and Carwell Creek Formation can be observed on the limbs of a small-scale anticline; shale of the Early Devonian Yellowmans Creek Formation crops out in the core of the structure. In the middle distance, the farms of "Carwell" and "Lucknow" can be seen on the floodplain of Carwell Creek. The northern end of Kandos No.2 Quarry, exposing the Early Devonian Clandulla Limestone, is visible in the background, behind which occurs the rugged, heavily-wooded terrain of Clandulla State Forest. Prominent in the background is an extensive east-dipping plateau that marks an erosional surface at the base of the Sydney Basin sequence, remnants of which occur as hills and cliffs in the distance.
ABSTRACT

The Capertee High in the Mudgee-Capertee region was a Late Silurian to earliest Middle Devonian shallow marine to subaerial palaeogeographic unit to the east of the Hill End Trough, north-east Lachlan Fold Belt, southeastern Australia. Rocks of the Capertee High occur in structurally complex sequences along with Ordovician basement, thick Late Devonian siliciclastic sequences, and Middle Carboniferous granitic intrusions. Flat-lying Permian-Mesozoic sequences overlap the High in the east and north.

Early Devonian sequences associated with the Capertee High can be broadly subdivided into two categories: (a) Platform sequences, which were deposited on the shallow marine to subaerial platform areas of the Capertee High and have been collectively defined as the approximately 4000 m thick Kandos Group; and (b) Western Platform Margin Sequences, which were deposited in slope and base-of-slope settings along the western margin of the Capertee High platform and have been subdivided into the Queens Pinch and Limekilns Groups. The latter sequences are transitional westwards with strata of the eastern Hill End Trough.

This thesis focuses on: (a) elucidating the stratigraphy, depositional processes and environments, and palaeogeography of the Kandos Group (particularly in the Rylstone-Cudgegong area); (b) establishing a biochronological framework for the Early Devonian succession by obtaining conodont faunas from the carbonates; (c) attempting correlation of the Kandos Group with platform margin units and Hill End Trough sequences; and (d) producing a synthesis of Early Devonian sedimentation and volcanism on the Capertee High and eastern Hill End Trough. Results show that the Kandos Group spans almost the entire Early Devonian and is separated from Late Silurian sequences by a widespread unconformity. The Kandos Group can be accurately correlated with the western platform margin sequences which are also dated by conodonts; however, correlation with the Hill End Trough succession is imprecise and relies on a small number of SHRIMP U/Pb zircon dates (which are inaccurately aligned with conodont zones) and rare macrofaunas.

The Lochkovian portion of the Kandos Group is a transgressive to regressive shallow marine sequence deposited on a west- to southwest-sloping shelf flanked to the east by eroding sedimentary basement and active silicic volcanoes, and passing to the west into deepwater facies along the western margin of the High (basal Mullamuddy Formation) and in the eastern Hill End Trough (upper Crudine Group). Sedimentation in the Kandos Group began in the early Lochkovian *woschmidtii* Zone and comprised (in ascending order): a transgressive sequence of wave-dominated, fan-deltaic clastics (Warrah Conglomerate); a unit of transgressive biostromal to biohermal carbonates (Clandulla Limestone; *woschmidtii* to *eurekaensis* Zones); a transgressive to regressive muddy, storm-influenced shelf deposit (Yellowmans Creek Formation; *eurekaensis* to *delta* Zones); and a thick regressive unit of storm-dominated siliciclastics deposited in middle shelf to nearshore settings (Roxburgh Formation; *delta* to *pesavis* or early *sulcatus* Zones). Facies prograded west during the late Lochkovian regression,
culminating in shelf exposure and some erosion in the east. The early Lochkovian transgression may
reflect a eustatic sea-level rise, whereas the late Lochkovian regression more likely indicates uplift of
nearshore areas ahead of increasingly proximal volcanism.

Thick sequences of dacitic to rhyolitic pyroclastics and volcaniclastics (Riversdale Volcanics
and Huntingdale Volcanics) demonstrate that the Capertee High was dominated by subaerial silicic
pyroclastic volcanism during the early to mid Pragian *sulcatus* and *kindlei* Zones. Medium-scale
calderas, centred southeast of Cudgegong and in the eastern Capertee Valley, were flanked by exten­sive, shallowly-dipping sheets of silicic ignimbrites and epiclastics which probably overlapped, forming
a continuous sheet. Short-lived transgressions on the platform resulted in minor carbonate deposition
during volcanically quiescent periods. Huge volumes of volcanic detritus were transported to the west
and deposited as lowstand fans and aprons along the margin of the Capertee High (Mullamuddy and
Kingsford Formations) and in the Hill End Trough (Merriens Formation).

The Carwell Creek Formation overlies the Riversdale Volcanics and represents a possibly
eustatic transgression in the latest Pragian to earliest Emsian (*pireneae* to *dehiscens* Zones) which
initially reworked then drowned the existing volcanic topography, creating a broad southwest-sloping
shelf with local eroding volcanic islands. Sedimentation in this setting was dominated by a laterally and
temporally extensive regime of storm- and tidally-influenced nearshore to shelf siliciclastic sequences
deposited in barrier island settings, along with local biostratal and biohermal carbonate buildups
deposited in lagoonal and open-marine environments (Carwell Creek Formation, Myrtle Grove Forma­tion and basal Mt Frome Limestone). Transgression cut off the supply of clastic detritus from slope
and basin areas, and condensed mudstone sequences were deposited along the margin of the Capertee
High (Warratra Mudstone and 'Rosedale Shale') and in the Hill End Trough (basal Cunningham
Formation). A transgressive pulse in the *perbonus* Zone was followed by regression, local platform
exposure and erosion of platform areas in the middle Emsian (late *perbonus* to *inversus* Zones), and a
number of carbonate aprons developed along the western margin of the High (basal Sutchers Creek
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A transgression (possibly eustatic) resulted in further deposition of carbonates on the Capertee
High during the late Emsian *serotinus* Zone (middle to upper Mt Frome Limestone and unpreserved
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unknown length of time into the early Eifelian before it was terminated by uplift and regression related
to the mildly manifested Middle Devonian Tabberabberan deformation which effectively ended the
depositional history of the Capertee High.
Modern analogues for the Early Devonian Capertee High-Hill End Trough-Molong High exist in the Sumisu Rift (off the southeast coast of Japan), the southern Havre Trough (north of New Zealand), and the Sulu Sea (between Borneo and the Philippines). The Tertiary Waitemata interarc basin (north of Auckland, New Zealand) displays gross similarities in facies, palaeogeography and scale compared to the Early Devonian Hill End Trough and its flanking highs.
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CHAPTER ONE
INTRODUCTION

1.1 THESIS AIMS AND SCOPE

The area examined is located in the Central Tablelands region of New South Wales, south­
eastern Australia, approximately 200 to 290 km northwest of Sydney by road (Fig. 2.1) and lies
between the towns of Mudgee in the north and Capertee in the south; other major towns in the area are
Rylstone and Kandos in the east. Lake Windamere lies approximately 40 km southeast of Mudgee and
has flooded (since 1984) the location of the former town of Cudgegong (Fig. 2.2).

Complexly deformed rocks of the Capertee Zone of the northeast Lachlan Fold Belt are over­
lapped by flat-lying Permian-Triassic Sydney Basin sequences in the east (Figs 2.1, 2.2). The Capertee
Zone strata comprise Ordovician quartz turbidite fan and mafic volcaniclastic/volcanic deposits; shal­
low marine to subaerial clastic, carbonate and silicic to intermediate volcanic sequences (and their
slope-deposited equivalents) of the Silurian-Devonian Capertee High; thick Late Devonian siliciclastic
sequences; and Carboniferous granite intrusions (see Section 2.2.2 for further description). The geology
of the area is comparatively poorly-known and, prior to this study, there was a particular dearth of
published data. This thesis concentrates on mapping and describing strata of Early Devonian age from
the Mudgee-Capertee region and integrating these with the results of previous studies. Numerous
university (mostly honours) theses, unpublished reports and rare published material have dealt with
aspects of the study area, often producing differing and conflicting interpretations of the geology -
especially of the pre-Carboniferous stratigraphy, structure, and depositional environments. A number
of projects undertaken in recent years have provided more coherent results. The thesis was therefore
undertaken with the aim of producing a synthesis of Early Devonian sedimentation, palaeoenvironments
and faunas of the Capertee High, focussing on the shallow marine-subaerial sequences of the Kandos
Group (see Section 2.2.2) in the Rylstone-Cudgegong district. Specifically, aims of the thesis are to:

(a) produce detailed geological maps of the Early Devonian Kandos Group sequences of
the Capertee High, particularly the area extending from the lower reaches of Carwell
Creek to the upper Carwell Creek valley and around Ilford (Figs 1, 2 and 3);
(b) describe the petrography, facies, fauna, and depositional setting of each unit (Chapter
3);
(c) extract conodonts where possible from the carbonates of the Kandos Group to provide
a biochronological framework for the sequence (Chapter 3);
(d) review previous studies of Early Devonian rocks from the area, integrating them with
the present work, and revising the stratigraphic nomenclature of the Early Devonian
sequences (Chapter 3);
(e) attempt to correlate the Kandos Group sequences with: (i) lesser known platform sequences of the Capertee High and compare the Kandos Group sequences to coeval units on the Molong High; (ii) slope-deposited sequences of the western margin of the Capertee High; and (iii) units of the eastern Hill End Trough (Chapter 4);
(f) develop a synthesis of Early Devonian sedimentation, palaeoenvironments, tectonism and sea-level movements of the Capertee High and eastern Hill End Trough (Chapter 5);
(g) discuss possible palaeogeographic analogues and the tectonic setting of the Capertee High (Chapter 5).

1.2 THESIS PRESENTATION AND ORGANISATION

The text of the thesis, including the figures, tables, and supporting papers, is contained in one volume. The main geological maps (Figs 1, 2, and 3) are located in the back pocket. Detailed geological mapping around the Carwell Creek area is presented as a 1:7 500 scale map (Fig. 1) and mapping in the Carwell Creek south, Ilford and Brogans Creek districts is produced as a composite 1:12 500 and 1:25 000 scale map (Fig. 2). The rest of the mapping was carried out while employed by the Geological Survey of New South Wales and is presented as the Mudgee 1:100 000 Geological Sheet (Fig. 3). The author had responsibility for mapping and compiling all of the eastern half of the sheet with the exception of the non-Ordovician part of the Botobolar 1:25 000 Sheet. Work on the Ordovician sequences east of Mudgee was carried out in conjunction with Chris Fergusson (Section 1.3). Other areas of more detailed mapping are presented as text figures.

During the production of the thesis, substantial portions of the work have been published. These publications have been presented as Supporting Papers 1 to 11 at the end of the thesis text; a summary of the findings is generally reproduced in the text. The reader is referred to the supporting papers for more detailed, or related, information. Grid references are given in standard 13 figure notation of the Australian Metric Grid, giving an accuracy to 1 m, but are rounded off to the nearest 10 m in most cases. Previous geological investigations and synonymy of the various units is covered where appropriate in the text.

1.3 METHODS

Detailed geological field mapping in the Rylstone-Kandos-Ilford district was carried out on film overlays placed over enlargements (to 1:7 500) of the Mudgee Run 5 (June, 1964) 1218/9091 aerial photographs; however, mapping in areas south of Kandos No.1 Quarry relied on 1:12 500 scale enlargements of the Kandos 8832-2-N and Ilford 8832-2-S (1st ed.) 1:25 000 Topographic Sheets. During mapping, geological data from these overlays (e.g. lithology, sample locality, position and accuracy of lithological contacts) were transferred to an overlay on a 1:15 000 scale base aerial photograph.
The final geological maps (Figs 1 and 2) were, however, prepared at the original mapping scale on a corrected base map, and the Australian metric grid was incorporated.

Mapping of the "Lue beds" from Botobolar to Lue was undertaken with Chris Fergusson in August to October, 1994 while employed as a Research Assistant at the Department of Geology, University of Wollongong. Mapping was conducted using 1:12 500 scale enlargements of 1:25 000 sheets and was aided by a 1:250 000 scale radiometric image of the Dubbo Sheet (see below).

Later mapping (January to December 1995) from Brogans Creek in the south to Havilah in the north (Fig. 3) was undertaken while employed at the Geological Survey of New South Wales as part of the remapping of the Dubbo 1:250 000 sheet (SI 55/4). Mapping was carried out on 1:25 000 scale aerial photographs and topographic maps and subsequently compiled onto plastic overlays of the Mudgee 1:50 000 sheet (8832 I & IV) and Rylstone 1:50 000 sheet (8832 II & III). These overlays were digitised at 1:50 000 scale and output as a composite 1:100 000 scale map (the Mudgee 1:100 000 Sheet; Fig 3). Mapping made extensive use of airborne radiometric and magnetic data flown by AGSO over the entire Dubbo 1:250 000 sheet in 1991 at a 400 m line spacing. Radiometrics were used as RGB images, with red as potassium, green as thorium, and blue as uranium; "KTMag" images - which substitute "magnetic activity" for uranium in the blue channel - were also used. These images were commonly produced at 1:50 000 and 1:100 000 scale and were used extensively as an aid to recognising different geological units and inferring geological boundaries. Magnetic data were commonly output as 1:50 000 and 1:100 000 scale TMI (Total Magnetic Intensity) images; 1st and 2nd vertical derivative images were used to sharpen magnetic trends. The magnetic images were used primarily to elucidate the geological structure.

Geological data were mainly obtained from natural outcrops, road cuttings, and quarries, and a number of drill cores from Kandos Quarries were examined. Sections in road cuttings, quarries and short cliff or hill sections were logged in detail using a Jacob's Staff or a 2 m tape measure; other longer sections over less well exposed terrain were paced out and later corrected for dip variations. During the latter part of the field work, a Magellan GPS was used for site location and was found to be very useful for orientation in rough, heavily wooded country. Petrographic and chemical analyses utilised standard techniques; in total, over 1000 hand specimens were collected and 212 thin sections, 18 X-ray diffraction analyses, 4 whole rock geochemical analyses of major, trace, and rare earth elements, 3 polished blocks, and numerous polished slabs were prepared. In addition, several of the carbonate samples were stained with Alizarin Red S and K-ferricyanide, following the techniques of Dickson (1965), to assess the relative proportions of calcite and dolomite.

Palaeontological data were also obtained by standard methods: acetate peels of corals and stromatoporoids were made from polished and etched carbonate slabs; latex casts of fossil moulds were prepared; and silicified faunas were etched in HCl for several days and subsequently picked without sieving. A programme of bulk reconnaissance sampling for conodonts obtained over 200 2 to 15 kg
carbonate samples and several shale samples. These conodont studies were initiated while employed as a Research Assistant at the Department of Geology, University of Wollongong. Approximately 180 carbonate samples were dissolved in buffered acetic acid (following the techniques of Jeppsson et al. 1985) followed by wet sieving and picking; concentration by the sodium polytungstate method and further picking followed where conodonts were encountered. Attempts to disaggregate shales for conodont analysis were unsuccessful but included: boiling in Na$_2$CO$_3$ for 2 hours; repeated freezing and thawing; and soaking in Stoddard solvent (2-4 hours) followed by hot water and detergent.

Structural and palaeocurrent data were largely obtained in the field using a Brunton compass, although a number of oriented samples were extracted and later analysed for sole mark orientation. Structural data were mostly analysed on a stereonet or using a simple public domain DOS program (Quickplot); palaeocurrent data were restored to horizontal using a public domain DOS program written by B.G. Jones.

The possible source and magnitude of any errors in all techniques have been indicated where appropriate, in addition to suggestions for improvements and suitable avenues for future research. Any information used in the thesis which was not derived from this study is acknowledged in the text.
CHAPTER TWO
REGIONAL GEOLOGY AND TECTONIC HISTORY

2.1 INTRODUCTION

This chapter provides a concise summary of previous work on the geology and tectonics of the Capertee Zone in particular, and the northeast Lachlan Fold Belt in general. This information places the Early Devonian geology of the study area in its regional context and serves as a basis for comparison and discussion with results of the present study. Where the results of the thesis have significantly altered the existing information and interpretation, these results are briefly stated and discussed in the relevant section of the thesis. The chapter also summarises briefly some of the work the author has done on non-Early Devonian sequences in the same area. More detailed treatment of the regional stratigraphy and setting is contained in Supporting Paper 2 (Pemberton et al. 1994) and Supporting Paper 8 (Colquhoun et al. 1997). The structural geology of the Mudgee 1:100 000 Sheet is summarised in Supporting Paper 8 (Colquhoun et al. 1997). The stratigraphic code from the Mudgee and Bathurst 1:100 000 Sheets is shown in italics after the relevant stratigraphic unit.

2.2 GENERAL GEOLOGY OF THE NORTHERN CAPERTEE ZONE

2.2.1 Setting

The study area is located in the extreme northeast of the Lachlan Fold Belt (Fig. 2.1) - a large Palaeozoic orogenic belt that covers 300 000 km² in New South Wales and Victoria and constitutes the southeastern part of the Tasman Fold Belt System of eastern Australia (Scheibner 1987). The Lachlan Fold Belt is composed of Cambrian to Early Carboniferous deep marine to terrestrial sedimentary and volcanic rocks, and voluminous granitic intrusions, occurring in variably-deformed sequences of low to moderate metamorphic grade (Gilligan and Scheibner 1978; Cas 1983). The study area (Fig. 2.2) is located on the northern part of the Capertee Zone of Scheibner (1993).

The boundaries of the Capertee Zone coincide approximately with the margins of the former Capertee High (Figs 2.8 and 2.9) - a prominent, north-northwest-trending, palaeogeographic unit which formed in the Middle to Late Silurian (Packham 1960, 1968, 1969; Cas and Jones 1979; Powell 1976; Crook and Powell 1976) and persisted until the earliest Middle Devonian (Packham 1960; Crook and Powell 1976). To the west, the Capertee Zone is separated from the Hill End Zone - the deformed fill of the deepwater Silurian-Devonian Hill End Trough - by the Wiaigdon Thrust (Packham 1960) and the Mudgee Thrust (Colquhoun et al. 1997) to the south and north of the Aarons Pass Granite respectively (Fig. 2.3). The eastern limits of the Capertee Zone are concealed by the Early Permian Rylstone Volcanics (Shaw et al. 1989) and flat-lying Permian-Mesozoic sediments of the Sydney Basin;
Figure 2.1 Location of the study area in the northeast of the Lachlan Fold Belt of southeastern Australia

undeformed Mesozoic sediments of the Great Australian Basin cover the Capertee Zone to the north (Fig. 2.2).

2.2.2 Stratigraphy

Ordovician and Silurian strata are the oldest rocks recognised on the Capertee Zone and have several distinct areas of outcrop (Figs 2.3 and 2.4):

(1) a belt extending along the exposed eastern margin of the Capertee Zone, composed of poorly-known metasediments (Lue beds) and Silurian silicic volcanics, volcanioclastics and carbonates (Dungeree Volcanics Std) (Offenberg et al. 1971) which are unconformably overlain by, and partially faulted against, Early and Late Devonian rocks in the west. Work by Fergusson and Colquhoun (1996) concentrated on these Ordovician-Silurian sequences and considerably increased knowledge of the stratigraphy and structure (Supporting Papers 9 and 10). The Silurian Lue beds of Offenberg et al. (1971) have been recognised as Ordovician (Fergusson and Colquhoun 1996; Supporting Paper 9) and divided into the Early Ordovician Adaminaby Group Oa (quartz-rich turbidites) and the Late Ordovician-
Figure 2.2 Location diagram of the Mudgee-Capertee region with the major structural elements of the northeast Lachlan Fold Belt (after Scheibner 1993). Permian-Mesozoic cover shown in grey; major granites are indicated by crosses.
Early Silurian Coomber Formation *Ocd* (mafic volcaniclastics, lava, intrusives and mudstone). The boundaries and extent of the Dungeree Volcanics, as recognised by Offenberg *et al.* (1971), have been modified considerably and the unit is now placed in the newly defined Tannabutta Group (Colquhoun *et al.* 1996) which includes all the Silurian units of the Capertee High (Figs 2.3 and 2.4). A detailed petrographic and provenance study of the Adaminaby Group and Coomber Formation has also been carried out in conjunction with Chris Fergusson and Stuart Tye (Colquhoun *et al.* in review; Supporting Paper 10).

(2) a western belt extending from Ilford to Mudgee (Fig. 2.3). The section from Cudgegong to Mudgee was studied in detail by Pemberton (1980b, 1989, 1990) and conodonts obtained from the limestones are listed by Pickett (1982b). This belt is composed of Late Ordovician basaltic andesite to basaltic volcanics and volcaniclastics (Sofala Volcanics *Ocs*; formerly the Cudgegong Volcanics of Pemberton 1989) overlain with probable unconformity by the Late Silurian Tannabutta Group. The latter consists of shallow marine to subaerial elastics and carbonates at the base (the Willow Glen Formation) conformably overlain by the Windamere Volcanics (subaerial to shallow marine dacite to rhyolite lava and autoclastics) in the east, and the Toolamanang Formation (volcaniclastic arenite, mudstone and basalt blocks) in the west (Pemberton 1989, 1990; Pemberton *et al.* 1994). The Mills-ville Formation occurs conformably at the top of this sequence and comprises a shallow marine sequence of limestone, dacite, dacitic conglomerate, breccia, and shale (Pemberton 1989, 1990).

A continuation of this belt has recently been mapped by the Geological Survey of New South Wales between Mudgee and Gulgong (Colquhoun *et al.* 1996, 1997; Supporting Papers 7 and 8) where the Late Ordovician Burranah Formation *Och* (mafic volcaniclastics, lava, and intrusives, and mudstone) is overlain in the west by the Dungeree Volcanics; the latter unit possibly passes further west conformably into the Chesleigh Group *Ss* of the basal Hill End Trough sequence (Watkins 1996) (Figs 2.3 and 2.4). Sequences in this area were previously thought to be Early Devonian (Exon 1962; Offenberg *et al.* 1971; Armstrong 1983). A probable continuation of this belt occurs north of Gulgong, between Tallawang and Dunedoo, as a unit previously mapped as the Tucklan beds by Offenberg *et al.* (1971) and also thought to be Early Devonian. Remapping of this unit by the author, Simone Meakin (NSWGS) and Tony Henderson (AGSO) shows it consists of a Late Ordovician mudstone, and mafic volcaniclastic-volcanic sequence (the Tucklan Formation *Oct*) overlain disconformably by the Dunger­ee Volcanics.

(3) Extensive Ordovician and Silurian sequences crop out around Sofala and Palmers Oakey and in the western Capertee Valley and have recently been remapped as part of the Bathurst 1:100 000 Sheet (Watkins and Pogson 1994) (Fig. 2.5). These sequences include Early Ordovician quartz turbidites and chert (Adaminaby Group *Oa*), Late Ordovician basaltic andesite to basaltic volcanics, volcaniclastics (Sofala Volcanics *Ocs*), and an Early to Late Silurian sequence comprising the late Llandover­ery Pipers Flat Formation (Pickett *et al.* 1996) overlain disconformably by the Tanwarra Shale (*Smt*:
Figure 2.3 Simplified geology of the Mudgee 1:100 000 Sheet. See Fig. 2.4 for key to unit names and the text for unit descriptions (from Colquhoun et al. 1997).
shale, limestone and conglomerate), which is conformably overlain by the Bells Creek Volcanics (Sml: silicic volcanics, volcaniclastics) and the Chesleigh Formation Sms (now Ss) (lithic sandstone and slate), the latter forming the basal unit of the Hill End Trough sequence (Packham 1968; Gilfillan 1976; Fergusson 1979; Fell 1984; Pogson and Wyborn 1994; Wyborn et al. 1994).

Correlation of sequences between some of these areas was attempted by Fell (1984), Pemberton (1989, 1990) and Pemberton et al. (1994).

Devonian rocks between Mudgee and Cudgegong occur as two strips, separated from older rocks, by strike faults or unconformable boundaries (Wright 1966, 1969) (Figs 2.3 and 2.4). The western strip at Queens Pinch (16 km south-southeast of Mudgee) consists of 1800 m of intensely folded and faulted Early Devonian conglomerate, lithic sandstone, intermediate to basic volcanics, mudstone and allochthonous and allodapic limestone (Queens Pinch Group Dq) deposited in slope and base-of-slope settings adjacent to the western margin of the Capertee High (Wright 1967; Garratt and Wright 1988; McCracken 1990; Pemberton et al. 1994). Wright (1966) and McCracken (1990) have divided this area into a number of stratigraphic units which are tightly age-constrained (Fig. 2.4); these are summarised by Pemberton et al. (1994; Supporting Paper 2) and Colquhoun et al. (1997; Supporting Paper 8). Recent mapping (Fig. 3) recognised the Queens Pinch Group to the north of Mudgee (Sutchers Creek Formation Dqs) and mapping during this study has identified a new unit in the Group (Kingsford Formation Dqk, Section 3.4.2) to the north of Ilford (Colquhoun et al. 1997).

The eastern strip stretches from northeast of Mudgee to southwest of Kandos and north of Ilford (Fig. 2.3). It is dominated by a thick Late Devonian siliciclastic sequence (the Mt Knowles Group Dp; Colquhoun et al. 1997) occupying the core of the Pine Ridge syncline (Wright 1966) and extending southeast from Mudgee to the west of Rylstone, where it is complexly folded and faulted (Pemberton et al. 1994). The Mt Knowles Group is at least 1700 m thick and consists of the basal Buckaroo Conglomerate Dpc, consisting of quartz-rich purple conglomerate, quartz sandstone and red mudstone representing braided stream and overbank deposits (Wright 1966; Killick 1987). It is conformably overlain by quartz and sublithic sandstone of the Bumberra Formation Dpb deposited in an open, tide- and storm-wave influenced, shallow marine setting. The muddy, storm-influenced shelf deposits of the Lawsons Creek Shale Dpl conformably overlie the Bumberra Formation, and are, in turn, overlain by the regressive, richly fossiliferous quartz sandstones of the Derale Sandstone Dpd, deposited in a storm- and wave-influenced environment.

The Late Devonian sequence is underlain with slight angular unconformity (Powell and Edcombe 1978; Millsteed 1985, 1992; Killick 1987; Colquhoun et al. 1997) by shallow marine elastics, carbonates and silicic volcanics of the Early Devonian Kandos Group Da (Pemberton et al. 1994) (Figs 2.3 and 2.4). Units of the Kandos Group crop out extensively in the Rylstone-Cudgegong district and have formed a particular focus of this study. Parts of this work have been published by Pemberton et al. (1994; Supporting Paper 2), Colquhoun (1995a, 1996; Supporting Papers 4 and 6), Colquhoun et
Figure 2.4 Time-space diagram for the Mudgee 1:100 000 Sheet. Radiometric ages are by U/Pb SHRIMP techniques unless otherwise indicated (from Colquhoun et al. 1997).
The conodont biostratigraphy and correlations of the sequence have been discussed by Colquhoun (1995b; Supporting Paper 5). The Kandos Group progressively onlaps the Late Silurian Dungeree Volcanics near Carwell Creek (GR 772015 6364800; Fig. 1), so that only the uppermost member of the Group (the Carwell Creek Formation) is present north of here, and has been mapped almost continuously to the Havilah area, 30 km to the north, where it is offset by the Havilah Fault and continues north of the Lue Road in the Mt Knowles-Buckaroo-Eurundury district (Fig. 2.3). The group spans almost the entire Early Devonian and comprises an early to middle Lochkovian transgressive sequence of basal fan-deltaic clastics (Warrah Conglomerate \textit{Daw}) overlain by reefal to biostromal carbonates (Clandulla Limestone \textit{Dal}) which in turn is overlain by a muddy storm-influenced shelf sequence (Yellowmans Creek Formation \textit{Day}). The thick regressive siliciclastic sequences of the late Lochkovian storm- and tide-influenced Roxburgh Formation \textit{Dar} overlie the Yellowmans Creek Formation and complete the Lochkovian transgressive-regressive cycle. The Roxburgh Formation is overlain with partial disconformity by the Pragian Riversdale Volcanics \textit{Dav}, a sequence of subaerial to shallow marine, dacitic to rhyolitic pyroclastics, volcaniclastics, epiclastics and rare limestone. The latest Pragian to early Emsian Carwell Creek Formation \textit{Dac} conformably overlies the Riversdale Volcanics and comprises over 1000 m of siliciclastics and carbonates deposited in barrier island and shoreline settings. The stratigraphy of these sequences is discussed in detail in Chapter 3.

Further south, around Brogan’s Creek and in the Capertee Valley (east of Capertee), large areas of dominantly Early Devonian lithic sandstone, limestone and silicic volcanics crop out as inliers surrounded by Permian-Triassic sediments (Bembrick \textit{et al.} 1969) (Fig. 2.2). The sequences here are comparatively poorly-known but a number of workers have noted lithological and faunal similarities with sequences further north (Wright 1966, 1967; Kennedy 1976; Bracken 1977). Conodont dating and some mapping were carried out in these areas in an effort to correlate with the Kandos Group sequences further north (Sections 3.3.7 and 4.4.1 and Supporting Paper 5).

In the Limekilns area (22 km south-southeast of Sofala; Figs 2.2 and 2.5), a thick sequence of Early Devonian fine-grained clastics, allochthonous carbonates and silicic volcanics and volcanioclastics occurs (Limekilns Formation \textit{Dk}; Wyborn \textit{et al.} 1994). The stratigraphy here has been elucidated by Packham (1968, 1969) and modified by Wyborn \textit{et al.} (1994) and age-diagnostic faunas are known from several stratigraphic levels (Wright and Chatterton 1988; Garratt and Wright 1988; Wright and Haas 1990; Mawson and Talent, 1992). Lithological and faunal similarities with the Queens Pinch Group have been noted by Wright (1969) and McCracken (1990). Small areas of Early Devonian mudstone and limestone breccia also occur in the Palmers Oakey district (Fergusson 1979; Bischoff and Fergusson 1982) (Fig. 2.2). Correlation of these sequences is discussed further in Chapter 4.

Younger rocks of the Capertee Zone include several Middle Carboniferous granitic intrusions (Powell 1976; Vickary 1983) including the Aarons Pass Granite \textit{Cag} (328 Ma: Pemberton 1990), the
Figure 2.5 The northeast portion of the Bathurst 1:250 000 Sheet (SI 55-8) Geology (preliminary 2nd edition, October 1994) (Wyborn et al. 1994). See text in Section 2.2.2 for code and descriptions of the main stratigraphic units.
Gulgong Granite (319 Ma: Evendern and Richards 1962, recalculated by Cas et al. 1976), the Havilah Granite \( Chg \), the Pyangle Pass Granite \( Ccg \), and the Botobolar Granite \( Cbg \) (Figs 2.3 and 2.4). The silicic pyroclastics, volcanoclastics and epiclastics of the Early Permian (292 Ma: Shaw et al. 1989) Rylstone Volcanics \( Pr \) are overlain unconformably by sub-horizontal marine to terrestrial strata forming the bulk of the Permian-Triassic Sydney Basin sequence \( (Ps, Pi, Rn) \) and flank the region to the east, in addition to occurring widely as outliers further west (Fig. 2.3). Mesozoic alkaline intrusions \( Ma \) (trachyte, teschenite, phonolite) and Tertiary basaltic intrusions and flows \( Tb \) (Wellman and McDougall 1974) crop out widely in the east but are of only minor areal extent (Day 1961; Matson 1975). Cainozoic deposits \( (Qa, Cza) \) of unconsolidated clay, silt, sand and gravel occur along rivers and valley floors over much of the area. Large areas of such deposits between Mudgee and Gulgong have yielded economically significant quantities of gold, diamonds and clays (Taylor 1878; Matson 1975).

### 2.3 TECTONIC HISTORY, FACIES DISTRIBUTION AND PALAEOGEOGRAPHIES OF THE NORTHEAST LACHLAN FOLD BELT

#### 2.3.1 Introduction

The Lachlan Fold Belt provides a record of marked changes in palaeogeography and lithological facies during its Cambrian to Early Carboniferous tectonic history, reflecting the development of an active plate margin (Scheibner 1987). Currently, the knowledge of the tectonic and palaeogeographic evolution of the fold belt is a mixture of old beliefs and some partially-developed new theories devised within the framework of global tectonics (Talent 1988). No widely-accepted, comprehensive model has been proposed and the summary attempted herein briefly considers most of the old and new theories.

The northeast Lachlan Fold Belt (an area encompassed approximately by Fig. 2.3) is located wholly within the Molong-Monaro Terrane of Scheibner (1985, 1987) and Leitch and Scheibner (1987) and the Benambra Terrane of Fergusson et al. (1986). With one exception (Fergusson and Vandenberg 1990; Section 2.3.2), current theories assume this area of the Lachlan Fold Belt remained relatively intact throughout its geological history; consequently, terrane analysis concepts are not considered here. As the oldest rocks encountered in the region are Ordovician, the tectonic history, facies and palaeogeographies can be considered in four time slices (after Cas 1983):

#### 2.3.2 Ordovician to Early Silurian

Ordovician to Early Silurian strata of the northeastern Lachlan Fold Belt are characterised by two fundamentally different lithofacies associations (Fergusson and Coney 1992):

(a) a monotonous Early Ordovician succession of quartz-rich turbidites and chert, at least several kilometres thick, which occurs widely throughout much of Victoria and southern to central New South Wales, and is known in much of the eastern Lachlan Fold Belt as the Adaminaby Group (VandenBerg and Stewart 1992). The Adaminaby Group occurs extensively in the Capertee Zone in the
Capertee-Palmers Oakey area (Pogson and Wyborn 1994) and has been recently documented further north, to the east of Mudgee (Fig. 2.3); these latter exposures represent the oldest rocks in the region and the most northeasterly outcrop of the quartz turbidite facies in the Lachlan Fold Belt (Fergusson and Colquhoun 1996; Supporting Paper 9). The quartz turbidite succession has been interpreted as the deposits of a huge submarine fan of comparable dimensions to the present Bengal-Nicobar submarine fan complex (Cas et al. 1980; Powell 1984). A passive margin setting for this fan was proposed by VandenBerg and Stewart (1992). Strontium isotopic ratios of the quartz sandstone suggests probable derivation from Cambrian sedimentary rocks of the Kanmantoo Group in South Australia or Cambrian igneous rocks in western Tasmania (Gray and Webb 1995).

(b) a mainly Late Ordovician sequence of mafic volcanics, volcaniclastics and limestone which outcrops in several meridional belts in the northeastern Lachlan Fold Belt (Packham 1969, 1987; Powell 1984; Wyborn 1992a, 1992b). Contacts between the two sequences are commonly faulted; however, where conformable, the contacts tend to be sharp with only rare evidence of interdigitation (e.g. Fergusson 1979).

Traditionally, the Late Ordovician palaeogeography of the northeast Lachlan Fold Belt has been thought to consist of a near meridional, east-facing mafic volcanic island chain, with a forearc basin lying to the east and a marginal sea flanking the area to the west - a configuration similar to the present-day Andaman-Nicobar system of the northwest Indian Ocean (Cas et al. 1980; Powell 1983, 1984) (Fig. 2.6). The arc was characterised by recurrent episodes of basaltic volcanism alternating with volcanically quiet intervals during which carbonates accumulated and the mafic volcanic piles were locally reworked (Powell 1976). In the forearc and backarc basins, deep marine sedimentation of arc-derived, volcanic detritus dominated (Fig. 2.6). The Ordovician palaeogeography is discussed in more detail by Fergusson and Colquhoun (1996; Supporting Paper 9).

The geochemical character of the Ordovician volcanism is problematic: Owen and Wyborn (1979), Clarke (1990), and Wyborn (1992a, 1992b) considered the volcanism to be almost entirely shoshonitic, with minor calcalkaline characteristics. Pemberton (1990), Pemberton and Offler (1985), and Pemberton et al (1994) found the volcanics of the Capertee Zone (Rockley Volcanics, Sofala Volcanics, Coomber Formation) to be basaltic to andesitic, including both calcalkaline and shoshonitic episodes. In the Burranah Formation, shoshonitic volcanics dominate (Watkins 1996) whereas calcalkaline compositions dominate the Sofala Volcanics (Pemberton 1990).

Most authors have envisaged a convergent margin setting for the Late Ordovician, with the salient features being: (a) west-dipping Mariana B-type subduction causing the development of a (b) volcanic arc (Molong Volcanic Arc) on a fragment of Proterozoic craton (Molong microcontinent), with (c) backarc spreading and a marginal sea (Wagga Marginal Sea) separating the arc from (d) cratonic areas in the west (Powell 1983, 1984; Scheibner 1985 1987; Packham 1987). Variations on this theory have been suggested: (1) based largely on local geochemical trends, Scheibner (1987; 1989)
Figure 2.6 Block diagram of the Middle to Late Ordovician geography of the eastern Lachlan Fold Belt. Arrows indicate palaeoflow directions (from Powell 1984).

Figure 2.7 Geological cross-section and time-space diagram across the northeast Lachlan Fold Belt (see Fig. 2.8 for location of section) (from Powell 1984).
and Clarke (1990) proposed that the collision of a hypothetical oceanic plateau in the Late Ordovician possibly choked the subduction zone, causing it to flip to east-dipping subduction on the west side of the arc, thereby consuming the backarc basin; (2) Wyborn (1992a, 1992b) doubted the validity of a subduction model altogether and viewed the Ordovician volcanism as representing a broad belt of discrete volcanic centres unrelated to subduction. Instead, he proposed partial melting of subcontinental upper mantle after delamination and foundering of cold, dense subcontinental lithosphere to explain the volcanism. Based on the currently contentious affinities of shoshonite suites, this theory has been disputed on numerous grounds by Scheibner (1989) and Fergusson and Colquhoun (1996; Supporting Paper 9) and does not currently have wide support.

The Late Ordovician to earliest Silurian 'Benambran Orogeny' caused widespread deformation in the Lachlan Fold Belt (Packham 1969) and has been variously attributed to: (a) a change in plate convergence from decoupled Marinas-type to coupled Chilean-type (Powell 1983, 1984); (b) the accretion of the Molong-Monaro Terrane, following collision with the Victorian microcontinent and closure of the intervening Wagga Marginal Sea (Scheibner 1985, 1987); or (c) the amalgamation of the Parkes and Benambra Terranes (Fergusson and VandenBerg 1990). Packham (1987) viewed the event as representing a change from subduction to sinistral strike-slip tectonics due to the subduction of a spreading ridge from the southeast; he further considered transtensional forces created after this change in tectonics caused the Molong Volcanic Arc to split into 3 major and 10 minor segments which were subsequently displaced by a regional sinistral shear regime which operated until the Early Devonian.

Irrespective of its precise cause, the 'Benambran Orogeny' created significant north-south shortening, and produced mainly east-west trending folds, devoid of cleavage, which have been reported in the northeast of the fold belt at Sofala (Gilfillan 1976; Powell et al. 1978, but disputed after recent mapping; Wyborn et al. 1994), Wellington (Powell 1984), and near Capertee (Fergusson 1979; Fell 1984). A weak, poorly time-constrained early deformation has also been recorded from the Mudgee area (Fergusson and Colquhoun 1996; Supporting Paper 9) and around Orange (Glen and Watkins 1994) and may be related to the Benambran deformation.

2.3.3 Middle Silurian to Middle Devonian

Major changes in facies and palaeogeography followed the Benambran deformation (Cas 1983). Sedimentation recommenced in the region in the Middle Silurian when strong extensional forces caused widespread rifting and the formation of a 400 km wide zone of narrow horsts and grabens in the eastern Lachlan Fold Belt (Powell 1983) (Fig. 2.7). Concurrent with this sedimentation, voluminous intrusion and effusion of silicic magmatic rocks reflected a change to dominantly continental basement (Cas 1983).

In the northeast, the Molong Volcanic Arc split, with Ordovician rocks forming the basement of two horsts (the Molong and Capertee Highs) which flanked the newly-formed Hill End Trough
Most workers subscribe to this simultaneous formation of the Hill End Trough and its marginal highs in the Middle Silurian (Packham 1960; Scheibner 1973, 1976; Cas and Jones 1979); however, Powell (1984) and Packham (1987) argue the Capertee High was not present until the latest Silurian (Pridolian) to earliest Devonian (Lochkovian). Shallow marine and evaporitic strata have been documented from the late Wenlock-early Ludlow Willow Glen Formation and Millsville Formation on the Capertee High (Jones et al. 1987; Pemberton 1990) and support creation of the high during the Middle Silurian rifting episode; evidence of local shallow marine environments in the Late Silurian Dungeree Volcanics provide further support (Colquhoun et al. 1997).

The dominant facies on the Capertee and Molong Highs throughout the Middle Silurian to Middle Devonian period were shallow subaqueous and lesser subaerial silicic volcanics, terrigenous clastics and shallow water carbonates (Pemberton et al. 1994) (Fig. 2.7). Sedimentation patterns record numerous transgressions and regressions, which were, in part, eustatically-related (Talent and Yolkin 1987; Colquhoun 1995b; Section 5.2).

By contrast, coeval sedimentation in the Hill End Trough was characterised by deep water turbiditic sedimentation, of alternating silicic volcanic, quartzose, and lithic provenance, reflecting volcanism and sea-level movements on the adjacent highs. Thick mudstone intervals, reflecting periods of volcanic quiescence and/or high sea-level, and local deep water silicic lavas and sills also occur throughout the trough fill (Cas 1978, 1979). Marginal trough facies are typically proximal mass flow sequences and include olistostromes and slump deposits (Crook and Powell 1976). The facies distribution and palaeogeography outlined above compares favourably with that of the southern Havre Trough - a modern ensialic interarc basin to the north of New Zealand in the Tonga-Kermadec arc system (Cas and Jones 1979) (Fig. 2.9). Siluro-Devonian facies and palaeogeographies are discussed further in light of the results of the present study in Chapter 5.

The latest Silurian-earliest Devonian 'Bowning Orogeny' caused widespread deformation, metamorphism and granite intrusion in the southern Lachlan Fold Belt (Cas 1983; Scheibner 1989), and Fergusson and VandenBerg (1990) have proposed almost continuous deformation throughout the entire Middle Silurian-Middle Devonian period in the Bungonia-Delegate region. By contrast, the northeast Lachlan Fold Belt contains no appreciable deformation of this age (Powell 1976; Cas 1983) and sedimentation continued largely uninterrupted until the early Middle Devonian (e.g. at Mt Frome; Pickett 1987; Figs 2.4 and 2.7). However, a number of Siluro-Devonian disconformities occur on the Capertee High and may be related to a very mild Bowning deformation and/or sea-level fall (Pemberton 1989, 1990; Pemberton et al. 1994; Colquhoun et al. 1996).

The 'Tabberabberan Orogeny' cratonised much of the Lachlan Fold Belt in the Middle Devonian; however, the effects of this deformation are comparatively mild in the northeast. Powell and Edgecombe (1978), Millsteed (1985, 1992), Killick (1987) and Colquhoun et al. (1997) reported low angle unconformities between Early-early Middle and Late Devonian strata and concluded that uplift
Figure 2.8 Middle Silurian to Middle Devonian tectonic elements of the Lachlan Fold Belt (from Powell 1984).

Figure 2.9 Schematic block diagram of the palaeogeography of the Molong High, Hill End Trough, Capertee High, and Canberra Magmatic Province during the latest Silurian to Early Devonian (from Cas and Jones 1979).
and broad folding or tilting had occurred in the Middle Devonian, with the effects decreasing to the north and east of the region (Figs 2.4 and 2.7). Fergusson and Coney (1992) suggested that the Ordovician volcanic basement in the northeast acted as a rigid buttress and protected the overlying sediments from more significant deformation.

Plate tectonic models of the Middle Silurian-Middle Devonian interval are somewhat speculative and a large number of complex extensional-compressional settings have been proposed. These include: (a) an intra-oceanic arc setting with multiple forearc basins (Crook 1980); (b) a sinistral shear regime resulting in progressive tripling of the Ordovician arc and backarc basin (Packham 1987); (c) a dextral transform plate margin setting similar to the Basin and Range Province of western North America (Cas 1983; Powell 1983, 1984); (d) early arc evolution followed by arc-Proterozoic oceanic plateau collision, tectonic underplating and partial melting of the underplated crust (Fergusson and VandenBerg 1990; Collins and Vernon 1994); and (e) a backarc setting behind a frontal arc located in the ancestral New England region, characterised by west-dipping, oblique Mariana-B type subduction (Scheibner 1973, 1987, 1989; Fergusson et al. 1986; Fergusson and Coney 1992).

2.3.4 Late Devonian to Middle Carboniferous

The period of subaqueous horst and graben sedimentation and silicic magmatism, which characterised the previous interval, ceased after the Middle Devonian deformation. In the Late Devonian, deposition of quartz-rich mollassic sediments of the Catombal, Lambie and Mt Knowles Groups commenced diachronously (from west to east) across the northeast Lachlan Fold Belt (Connolly 1969; Webby 1972; Killick 1987) (Fig. 2.7). The Late Devonian sequence is usually preserved in large superimposed synclinoria (Scheibner 1976; e.g. Pine Ridge syncline, Mt Dulabree syncline). In the Mudgee district, the Late Devonian sequence was placed in the Lambie Group by Connolly (1969), Offenberg et al. (1971), and Killick (1987); however, the succession is substantially different from the type area east of Bathurst, and the Mudgee sequence has consequently been assigned to a new group, the Mt Knowles Group (Colquhoun et al. 1997; Supporting Paper 8). The Mt Knowles Group commences with a basal fluvial conglomerate unit, overlain by thick, shallow marine sequences of quartz sandstone and shale deposited in westerly transgressing conditions to the east of the uplifted Hill End Zone (Killick 1987). Following regression and a lowering of relief in the west, the upper part of the Lambie Group (restricted to the Bathurst area) was characterised by southwest progradation of volcanilithic detritus from a magmatic arc to the east, interfingering, in fluvial environments, with cratonic detritus supplied from the southwest (Killick 1987). The proposed tectonic setting for deposition is a foreland basin, located to the west of a westerly subducting Andean-type continental margin arc on the eastern edge of the Lachlan Fold Belt (Powell 1984; Scheibner 1989; Fergusson and Coney 1992).

Deposition concluded in the Fammenian and was followed in the Early Carboniferous by the 'Kanimblan Orogeny' (Packham 1969). This was the pre-eminent deformation ($D_2$) of the region and
caused the cratonisation of the northeast Lachlan Fold Belt. Deformation produced open to tight, north-south trending folds with axial surface cleavage and regional metamorphism to sub-greenschist and greenschist facies (250-300°C and 1.5-3.5 kbar; Pemberton 1990) (Figs 2.4, 2.7). The effects of this deformation increase towards the northeast of the region (Powell 1984). The 'Kanimblan Orogeny' has been attributed to high relative plate motions causing backarc deformation and is synchronous with the development of a large subduction complex in the New England Orogen (Fergusson and Coney 1992).

Post-deformational, I-type granitoids were subsequently emplaced in the east of the region during the Middle Carboniferous (Powell 1976; Vickary 1983). A mild north-south compressive deformation immediately postdating granite intrusions has been documented near Bathurst (Kosaka 1994), near Mudgee (Colquhoun et al. 1996, 1997; Supporting Papers 7 and 8), and elsewhere in New South Wales (Powell et al. 1985).

2.3.5 Late Carboniferous to Triassic

The 'Kanimblan Orogeny' was followed in the Late Carboniferous to Early Permian by a continental and alpine glaciation which is said to have affected most of Australia (Powell 1984). Little evidence apparently remains of this glaciation in the northeast of the Lachlan Fold Belt, although Early Permian fluvio-glacial sediments (Pu) occur in valleys (possibly glacially carved) around Mudgee (Dulhunty and Packham 1962; Dulhunty 1964). Silicic to basic magmatism occurred in a wide belt from north Queensland to eastern New South Wales during the latest Carboniferous to Early Permian (Scheibner 1989). A local expression of this activity is the Early Permian (292 Ma) Rylstone Volcanics - an extensive caldera-related sequence of silicic pyroclastics, lava and volcaniclastics in the Rylstone-Botobolar-Dunedoo district (Ranocchia 1981; Langworthy 1986; Shaw et al. 1989) (Figs 2.3 and 2.4). The tectonic affinities of this magmatism are uncertain, although it is possibly related to the formation of the Sydney-Bowen Basin, a Permian-Triassic retroarc foreland basin which developed largely to the east of the Lachlan Fold Belt (Murray 1985). Deposition of Middle Permian to Triassic, shallow marine to terrestrial strata (including extensive coal measures) occurred over a relatively planar erosional surface (Dulhunty 1964; Bembrick 1983). These sequences remain largely undeformed.
CHAPTER THREE
EARLY DEVONIAN STRATIGRAPHY

3.1 INTRODUCTION

This chapter aims firstly, to present a unifying scheme for the Early Devonian stratigraphy of the Capertee High, concentrating on the Kandos Group in the Rylstone-Cudgegong-Ilford district; and secondly, to document the petrography, sedimentology, conodont faunas and palaeoenvironments of each unit - with special emphasis on the Roxburgh Formation.

3.2 SUBDIVISION OF EARLY DEVONIAN SEQUENCES

Broadly, there is a twofold subdivision of Early Devonian rocks associated with the Capertee High (Colquhoun 1995b; 1996; Supporting Papers 5 and 6) (Fig. 3.2).

(1) Platform Sequences: shallow marine and subaerial units which were deposited on the platform areas of the Capertee High, and have been collectively defined as the Kandos Group (Pemberton et al. 1991; Supporting Paper 1) (Fig. 3.2). The Kandos Group is an essentially conformable sequence of fine- to coarse-grained clastics, carbonates, and silicic volcanics, deposited in various shallow marine to subaerial environments (Pemberton et al. 1991, 1994; Wright et al. 1994).

(2) Western Platform Margin Sequences: hemipelagic and coarse- to fine-grained mass flow units which were deposited in slope and base-of-slope settings along the western margin of the Capertee High. These sequences are known in the Mudgee-Ilford region as the Queens Pinch Group and as the Limekilns Group in the Limekilns area to the north of Bathurst. These units are transitional to deep-water turbiditic sequences of the eastern Hill End Trough.

3.3 KANDOS GROUP

3.3.1 Introduction and Definition

The Kandos Group crops out most extensively in the Rylstone-Cudgegong-Kandos district as an approximately 4000 m thick sequence, largely in faulted contact with Silurian and Late Devonian units and overlying the Late Silurian Dungeree Volcanics just east of Carwell Creek (Fig. 1) with a disconformity to low-angle unconformity (Pemberton et al. 1991; 1994). Work during this study has focussed on these exposures of the Kandos Group. Other areas of Kandos Group outcrop include (Fig. 3.2): (a) the east, central and northwest Capertee Valley as inliers surrounded by basal Permian sediments (Bembrick 1964, Wright 1966, Bembrick et al. 1969; Kennedy 1976; Bracken 1977; Colquhoun, 1995b, unpublished data); (b) the flanks of Mt Frome southeast of Mudgee (Wright 1966; Pickett 1978); (c) a small area just north of Ilford (Booth 1990, this study); (d) a small fault-bounded block
Figure 3.1 (a) Summary of previous stratigraphic nomenclature schemes for the Early Devonian strata of the Rylstone-Cudgegong district; (b) Composite stratigraphic column of the Early Devonian Kandos Group in the Rylstone-Cudgegong district as proposed by this study (first published by Pemberton et al. 1994). Asterisks indicate conodont-bearing horizons which have been correlated with the standard international conodont zonation for the Early Devonian.
Figure 3.2 Simplified geological map of the northeast margin of the Lachlan Fold Belt, highlighting the distribution of Early Devonian sequences of the Capertee High.
near Windamere Dam (Wright et al. in prep.; Supporting Paper 11).

Various stratigraphic schemes have been proposed for the Early Devonian strata of the Rylstone-Cudgegong district (Fig. 3.1a). Early schemes (Sussmilch 1934; Lavers 1960; Offenberg et al. 1971; Brown 1974) were broadly-defined and commonly suffered through inaccurate lithological mapping and/or the assignment of incorrect ages. A more reliable stratigraphy was proposed by Pemberton (1980a) and was subsequently modified and expanded by Campbell (1981), Millsteed (1985) and Cook (1988a) (Fig. 3.1a). Mapping during this study confirmed and modified the relationships proposed by these authors (Fig. 3.1b): Campbell's Clandulla Shale was found to be a shale member in the Late Silurian Dungeree Volcanics and the name has been suppressed, whereas Campbell's un-named unit has been assigned to the Yellowmans Creek Formation (Section 3.3.4); the Kandos Limestone has been renamed the Clandulla Limestone (Section 3.3.3); and the Glendale Limestone Member has been defined in the Carwell Creek Formation (Section 3.3.8). Definitions of most of the other units have also been modified to some degree. In addition, as all the Early Devonian units are closely related and contain only minor disconformities, they have been collectively defined as the Kandos Group, following Sussmilch's (1934) Kandos Series (Fig. 3.1a). This revised stratigraphy was first published in Pemberton et al. (1994; Supporting Paper 2) and has been summarised in Colquhoun et al. (1997; Supporting Paper 8). The stratigraphic symbols from the Mudgee 1:100 000 Sheet (Fig. 3) are given in italics after the unit name.

The stratigraphic column presented in Figure 3.1b is idealised, as the constituent formations of the Kandos Group are characterised by rapid lateral changes in thickness and facies which locally alter superpositional relationships (Fig. 3.3). All sedimentary units of the Kandos Group in the Rylstone-Cudgegong district thicken markedly to the south and west and pinch out on, or towards, the Silurian-Devonian unconformity in the northeast. However, the Riversdale Volcanics thin rapidly to the north, east and west about a very thick southern area (Fig. 3.3). The Kandos Group progressively onlaps the Late Silurian Dungeree Volcanics near Carwell Creek (GR 772015 6364800), so that only the uppermost member of the Group (the Carwell Creek Formation) is present north of here; it has been mapped almost continuously to the Havilah area, 30 km to the north, where it is offset by the Havilah Fault and continues north of the Lue Road in the Mt Knowles-Buckaroo-Eurundury district (Figs 1 and 3). A schematic representation of these thickness trends and stratigraphic relationships is presented in Figure 3.4. These trends and their significance are discussed further in the relevant sections.

3.3.2 WARRAH CONGLOMERATE Daw (Campbell 1981)

3.3.2.1 Synonymy and Definition

Sussmilch (1934) and Game (1935) described a thin persistent horizon of conglomerate and limestone breccia underlying the 'Quarry Limestone' near Kandos Quarries. Sussmilch included these rocks in the middle parts of his Middle Devonian Kandos Series whereas Game considered them to be
Figure 3.3 Thickness trends of the Early Devonian stratigraphic units of the Rylstone-Cudgegong district (After Pemberton 1980a, Millsteed 1985, 1992, Cook 1988a, 1990, and this study).
the basal Middle Devonian strata. Similarly, Lavers (1960) included these rocks as the basal unit of his Early Devonian Kandos Formation, and Brown (1974) assigned them to the basal parts of the Carwell Creek beds (Fig. 3.1a). Offenberg et al. (1971) included these rocks in the undifferentiated Siluro-Devonian Gulgamree beds, a name assigned by these authors to a broad tract of strata, but now discarded even in its type area where it was found to be equivalent to the Mullamuddy Formation. In the vicinity of Kandos Quarries, the unit is often informally referred to as the 'Basal Conglomerate' or 'Basal Breccia' by quarry workers.

Campbell (1981) defined these strata as the Early Devonian Warrah Conglomerate, a name derived from the now abandoned property of 'Warrah' (GR 771100 6363900). Campbell's definition is herein retained, with slight modification, following the recognition of additional lithologies and an unconformable lower contact during this study.
3.3.2.2 Distribution, Thickness, and Areal Extent

Approximately 4 km² of the Warrah Conglomerate crops out in the Carwell Creek valley as a narrow north-northwest-trending strip which extends along strike from GR 772100 6365100 in the north, to the southern end of Charbon Quarry (7 km south), where it is truncated by the Carwell Creek Fault (Figs 1 and 2a). In addition, further extensive outcrops occur to the north of Ilford, approximately 10 km to the southwest, around GR 765000 6352500 (Fig. 2b).

The unit typically forms low, bouldery outcrops which are intensely weathered and cleaved; however, exposures of moderate to occasionally good quality occur north of the Cudgegong Road at GR 771900 6364700, to the northeast of Charbon Quarry (GR 773200 6359400), and in the northeast of Kandos No. 2 Quarry (GR 772600 6361700). The latter exposures are frequently ephemeral due to the rapid rate of quarrying. The best exposures north of Ilford occur as well-washed pavements and benches along Cunningham Creek (GR 765500 6350900).

The unit attains a maximum thickness of 145 m to the east of Charbon Quarry and thins steadily towards the north, reaching 70 m in Kandos No.2 Quarry, 20 m at GR 771900 6364700 and finally wedging out against the Siluro-Devonian unconformity at GR 772100 6365100 (Figs 3.3 and 3.4). To the north of Ilford, the thickness of the unit is more difficult to estimate due to a paucity of structural readings; available data suggest the unit has a maximum preserved thickness of approximately 200 m in this area.

3.3.2.3 Structure and Stratigraphic Relationships

Over the majority of its outcrop, the unit is characterised by a consistent northwest-southeast strike and a low to steep westerly dip (Figs 1 and 2a). Bedding, however, was often extremely difficult to identify in the conglomerate and is derived dominantly from sandstone and limestone interbeds. In Kandos Quarries, the unit is folded into a number of northwest-trending folds, which generally plunge to the north at shallow angles. In addition, abundant small-displacement normal and reverse faults are apparent in exposures at Kandos No.2 Quarry.

Between GR 772100 6365100 and GR 773100 6359700, the Warrah Conglomerate overlies shale and silicic volcanics of the Late Silurian Dungeree Volcanics with a disconformity or low-angle unconformity. This contact is not exposed but is thought to be disconformable based on: faunal and geochronological evidence of a latest Silurian hiatus (Section 3.3.2.5), an abundance of basement clasts in the basal parts of the Warrah Conglomerate (Section 3.3.2.4), and a lack of any discernible angular discordance between the two units. To the south of Kandos No.1 Quarry, the unit is faulted against silicic volcanics and volcanioclastics of the Riversdale Volcanics by the Carwell Creek Fault (Fig. 2a). This fault truncates the east-dipping limb of an anticline at Kandos No.1 Quarry (Campbell 1981) and appears to cut slightly across strike at the south end of Charbon Quarry.
Figure 3.5 (a) Gradational contact between the Warrah Conglomerate and the Clandulla Limestone showing thin, persistent interbeds of coarse sandstone rhythmically interbedded with limestone. Exposure is northeast of Charbon Quarry at GR 772900 6359300. (b) Gradational contact between the Warrah Conglomerate and the Clandulla Limestone showing sharp, erosive-based beds of graded pebble conglomerate and coarse lithic sandstone interbedded with black shale of the basal Clandulla Limestone. Note low-angle tabular cross-beds in the sandstone. Exposure is in Kandos No. 1 Quarry (GR 773540 6359980). (c) Photomicrograph of a granule conglomerate from the basal 10 m of the Warrah Conglomerate displaying tightly-packed clasts of volcaniclastic (dacitic) sandstone (V) and radiolarian chert (R) in a fine cherty matrix. Crossed nicols. (d) Tightly-packed, moderate to well sorted, oligomict conglomerate from the basal parts of the Warrah Conglomerate, displaying strong tectonic alignment and distortion of clast shape. (e) Facies B: interbedded, normally graded to massive, granule conglomerate-coarse sandstone and shale. (f) Poorly sorted, clast-supported conglomerate from the middle parts of the Warrah Conglomerate at GR 772550 6361410 showing angular to subrounded clasts of intraformational limestone (blue-grey) and silicic volcanics (white). (g) Poorly sorted, matrix- and clast-supported polymict conglomerate from the upper Warrah Conglomerate at GR 772600 6361400, displaying rounded extraformational clasts of lithic sandstone (khaki), silicic volcanics (white) and shale (brown-grey) and angular, irregularly-shaped clasts of intraformational limestone (blue-grey) and calcareous shale (black).
The top of the Warrah Conglomerate passes conformably and usually gradationally into the basal limestone of the Clandulla Limestone. Good exposures of this contact occur at GR 773540 6359980, where upward-thining beds of conglomerate pass over several metres into the black shale of the basal Clandulla Limestone (Fig. 3.5b) and at GR 772900 6359300, where calcareous shales with pebbly interbeds pass upwards over several metres through silty limestone with thin pebbly interbeds into limestone (Fig. 3.5a). To the north of Ilford, the Warrah Conglomerate is apparently overlain by the Roxburgh Formation. This contact is poorly exposed; however, the absence of the Clandulla Limestone and Yellowmans Creek Formation suggest this contact is a fault or an unconformity.

### 3.3.2.4 Petrography and Sedimentology

The Warrah Conglomerate consists of an upward-fining sequence of pebble-granule conglomerate, pebbly sandstone, sandstone, shale, and lenses of limestone. Three facies have been recognised based on lithology and sedimentary structures (Table 3.1).

**Facies A: Conglomerate-pebbly sandstone**

Clast- to matrix-supported, pebble and granule conglomerate to pebbly sandstone is the dominant lithology of the unit (Table 3.1). The conglomerate displays marked vertical and lateral variations in textures and clast lithologies, indicating subfacies are undoubtedly present, but poor outcrop prevents adequate description of their characteristics.

In the basal parts of the unit, massive, tightly-packed, moderate to well-sorted, clast-supported, mature oligomict conglomerate dominates (Fig. 3.5d); best exposures occur to the east of Charbon Quarry. Subrounded to well-rounded clasts, of low sphericity, average 2-5 cm in diameter (rarely reaching 20 cm in the most basal parts) and are composed of contiguous basement lithologies, including: white-grey, aphyric to quartz-phyric volcanics; light grey shale; and radiolarian chert (Fig. 3.5c). Clasts typically display rotation and alignment parallel to cleavage and show some distortion in shape. The matrix (<20%) is friable and highly ferruginised and consists of moderately-sorted, subangular to rounded, fine sand-sized grains of quartz, chert, and feldspar with authigenic clays, micas, and hematite (Fig. 3.5d). Slightly higher in the sequence, conglomerate clasts include additional basement lithologies such as grey lithic sandstone, red jasper, and mafic volcanic rock fragments. Thin, well-sorted interbeds of pebbly sandstone - displaying diffuse parallel and wavy laminae, and low-angle cross beds - appear at a similar level.

These features (Table 3.1) are consistent with many of Ethridge and Wescott's (1984) criteria for wave-worked gravels deposited in a beachface or shoreface setting. However, poor sorting, sporadic angular boulders, and a general decrease in clast roundness in the most basal strata suggests a transition to fluvial conditions.
<table>
<thead>
<tr>
<th>Name</th>
<th>Lithology</th>
<th>Percentage of total exposure</th>
<th>Bed thickness and geometry</th>
<th>Boundaries</th>
<th>Structures</th>
<th>Fauna</th>
<th>Interpretation</th>
</tr>
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<tbody>
<tr>
<td><strong>Facies A:</strong> Conglomerate and pebbly sandstone</td>
<td>Tightly-packed clast- to rarely matrix-supported, oligomict and polymict, granule and pebble conglomerate to pebbly sandstone</td>
<td>90%</td>
<td>Difficult to determine due to poor outcrop; 20 cm to at least 1.5 m; laterally continuous to lensoidal over several metres</td>
<td>Planar to gently undulating basal and upper contacts where rarely exposed.</td>
<td>Basal to middle conglomerates are clast-supported, and either massive or with crude to well-defined horizontal stratification and clast alignment. Pebby sandstones show diffuse parallel laminae &amp; rare low-angle cross beds. Conglomerate in upper parts of unit is commonly matrix-supported, poorly-sorted, massive, disorganised &amp; unstratified or crudely stratified.</td>
<td>Rare corals (Favosites sp and Cladopora sp.) and stromatoproids (Amphipora sp. and indeterminate) in limestone clasts.</td>
<td>Variable: basal to middle parts represent tractional beachface and shoreface deposition, transitional to dominantly cohesive debris flows in the upper parts of the unit.</td>
</tr>
<tr>
<td><strong>Facies B:</strong> Conglomerate with thin interbeds of sandstone, shale or limestone</td>
<td>Granule conglomerate to pebbly sandstone rhythmically interbedded with shale, fine grained lithic sandstone, and near the top of unit, limestone. Gradational with the overlying Clandulla Limestone</td>
<td>7% to 8%</td>
<td>Conglomerate beds are max. 20 cm thick (avg. 3 to 4 cm), laterally persistent. Sandstone-shale layers 10-50 cm thick, laterally persistent</td>
<td>Conglomerate beds show sharp, loaded bases and sharp, loaded tops.</td>
<td>Conglomerate layers are massive or normally graded. Sandstone-shale layers are massive or show diffuse parallel, wavy or rare wave ripple laminae.</td>
<td>Abundant poorly-preserved burrowing trace fossils in the fine sandstones and massive shales.</td>
<td>Thin debris flows to high-density turbidites of coarse detritus generated by storms or floods, interbedded with fine-grained suspension deposited clasts rarely affected by traction currents.</td>
</tr>
<tr>
<td><strong>Facies C:</strong> Limestone</td>
<td>Dark to light grey micrite or biomicrite with variable amounts of coarse terrigenous impurities.</td>
<td>2% to 3%</td>
<td>2 m to 5 m thick; lensoidal over 100 m to 400 m</td>
<td>Very sharp, planar contacts with conglomerate and pebbly sandstone.</td>
<td>Massive, unbedded</td>
<td>Sparse to abundantly fossiliferous with stromatoproids (Amphipora and indeterminate), crinoids, brachiopods (Isorthis sp. and Cystina sp.) and bryozoans.</td>
<td>Small carbonate accumulations in sheltered parts of the fan delta. Fossiliferous limestones may represent small biostromes; sparsely fossiliferous limestones may be mounds or banks of lime mud.</td>
</tr>
</tbody>
</table>
In the middle to upper parts of the unit, around Kandos Quarries, angular to subangular, granule- to boulder-size clasts of intraformational micritic to biomicritic limestone and calcareous shale appear (Figs 3.5f and 3.5g). Sorting decreases to poor or very poor; sporadic angular clasts of limestone up to 25 cm across occur; matrix-supported fabrics predominate; and beds are massive, ungraded, and disorganised or show crude bedding-horizontal clast alignment (Fig. 3.5g). The fine sand- to clay-size matrix (25-40%) consists of a poorly-sorted, texturally immature mixture of angular to subrounded grains of clast lithologies, along with clays and micas. A carbonate cement is commonly present. Thin, petrographically-similar sandstone beds are locally interbedded with the conglomerate. Towards the top of the unit, massive, very poorly-sorted, matrix-supported beds composed entirely of angular limestone clasts of granule to boulder size occur - probably the limestone breccias of earlier authors.

The textural and sedimentary characteristics of these middle to upper conglomerates suggest deposition by subaqueous cohesive debris flows (Nemec and Steel 1984) which evolved from slumps of intra- and extraformational detritus. In places, these conglomerates appear to pass laterally into well-sorted, polymict conglomerates similar to those in lower parts of the sequence, indicating the debris flows were of limited lateral extent.

In summary, ascending through the Facies A conglomerates: sorting and average grainsize decrease, matrix-supported fabrics become dominant, the proportion of intraformational clasts increase, and deposition by mass-flow processes becomes more common.

**Facies B: Conglomerate with thin interbeds of sandstone, shale or limestone**

Confined to the uppermost parts of the unit, this facies consists of thin, massive or graded beds of granule conglomerate to coarse sandstone interbedded with fine-grained, bioturbated sandstone (showing rare traction current structures), calcareous shale or limestone (Fig. 3.5e; Table 3.1). Good exposures occur in Carwell Creek at GR 772300 6362700 and north-northeast of Charbon Quarry at GR 772900 6359300.

In the core of the main anticline in the Kandos No.1 Quarry, the contact between the Warrah Conglomerate and the Clandulla Limestone is well displayed (Section 3.3.2.3). The upper Warrah Conglomerate at this locality comprises pebble to granule conglomerate and fine sandstone in moderate to thick, sharp-based beds. Cross-bedding is visible in coarse sandstone horizons indicating sediment transport towards the southwest. Conglomerate clasts are very rich in black calcareous siltstone and quartz (Fig. 3.5b).

The facies represents deposition by distal debris flows or high-density turbidity currents of coarse detritus - evolved from slumps, storm-related currents or flash floods - interbedded with finer grained, suspension-deposited sediments. Rare traction current structures in the latter suggest deposition above the storm-wave base, probably in a transitional offshore setting.


**Facies C: Limestone**

In the middle to upper parts of the Warrah Conglomerate, a number of lenses of sparsely fossiliferous limestone occur (Table 3.1). Best exposures of these occur beside the Cudgegong Road at GR 772400 6363600. They are interpreted as small biostromes and/or mounds of lime mud which developed in times or areas of reduced coarse clastic influx.

### 3.3.2.5 Fauna and Age

The limestone lenses near Cudgegong Road (GR 772300 6362400) contain a sparse to abundant, low-diversity, silicified and non-silicified fauna which closely resembles that found in the lower parts of the Clandulla Limestone. The fauna comprises hemispherical stromatoporoids and the branching stromatoporoid *Amphipora* sp., bryozoan fragments, crinoid ossicles; and the brachiopods *Isorthis* sp. and *Cyrtina* sp. In addition, moderate to intense bioturbation of indeterminate origin is common in the finer grained lithologies at the top of unit. The limestone lenses and limestone clasts in the conglomerate were barren of conodonts (samples C9, C10, and C11).

The fauna indicates only a broad Early Devonian age for the unit; however, the conformably overlying Clandulla Limestone contains a conodont fauna indicative of the early Lochkovian *woschmidtii* Zone towards its base (Section 3.3.3.5). A similar to slightly older age (possibly extending into the latest Pridoli) is inferred for the Warrah Conglomerate (Colquhoun 1995b; Supporting Paper 5). The underlying Dungeree Volcanics contain a late Wenlock to Ludlow fauna (Pemberton *et al.* 1994; Supporting Paper 2) and SHRIMP U-Pb ages range from 415 ± 7 to 419.2 ± 4.8 Ma (Colquhoun *et al.* 1997; Supporting Paper 8), supporting a probable latest Silurian hiatus between the two units.

### 3.3.2.6 Environment of Deposition

Lenses of limestone, wave ripple lamination, and common bioturbation indicate that at least the middle to upper parts of the unit were deposited in a nearshore, shallow marine environment. The thickness of the unit, together with the variety of extraformational clast lithologies, indicates a significant fluvial input (Nemec and Steel 1984), probably by high-gradient streams which drained areas of local basement to the immediate east and built fan deltas into the shallow marine setting (Fig. 3.6). Sedimentation in the lower to middle parts of the unit reflects high-energy, nearshore reworking of this detritus in a beachface to shoreface setting, perhaps transitional to a fluvial environment in the earliest stages. The apparent paucity of fluvial sediment suggests high wave energies and abundant reworking; furthermore, evidence of storm-dominated sedimentation is common in other formations of the Kandos Group.

Higher in the sequence, small banks or mounds of lime mud formed in shallow, protected areas on the fan delta during times of reduced siliciclastic supply. These local carbonate buildups were smothered by subsequent influxes of coarse clastic detritus; however, their talus was locally
remobilised, together with variable amounts of extraformational detritus, and redeposited to slightly
deeper water by subaqueous debris flows. These debris flows were possibly initiated by slumping down the relatively steep fan delta fronts. Sedimentation in the uppermost parts represents open marine, sub-
fairweather wave-base deposition of the distal equivalents of these mass flows, interbedded with suspension deposits and limestones. However, in some areas, the fan deltaic deposits apparently prograded into a sheltered dysaerobic lagoon and were interbedded with the black shale facies of the basal Clandulla Limestone; continued transgression eventually cut off the terrigenous clastic supply and led to carbonate deposition dominated by stromatoporoid-*Thamnopora* communities.

The facies trends, together with the position of the unit overlying an unconformity and passing upwards into mostly biostromal limestone, indicates an upward-deepening trend through the unit. The thickness of the unit suggests a relatively slow transgression which only gradually outpaced concurrent subsidence and sediment supply.

In summary, the Warrah Conglomerate is a transgressive, ?subaerial to nearshore shallow marine sequence of traction- and mass flow-deposited conglomerate, pebbly sandstone, sandstone, shale and limestone, which probably accumulated in a wave-dominated fan delta setting (Fig. 3.6).
3.3.3 CLANDULLA LIMESTONE Dal (Pemberton et al. 1994)

3.3.3.1 Synonymy and Definition

The unit was first described by Sussmilch (1934) who also reported a small fauna which he considered to be Middle Devonian in age. Game (1935) mapped the unit from the south of Charbon Quarry to north of the Cudgegong Road and provided the first detailed descriptions. Lavers (1960) included these rocks in his broadly defined Wells Formation. Offenberg et al. (1971) included these strata in their Siluro-Devonian Gulgamree beds, whereas Brown (1974) regarded the unit as one of the basal members of the Carwell Creek beds. The unit is often informally referred to in reports on Kandos Quarries as the "Carwell Creek Limestone"; however, this name is clearly invalid as the Carwell Creek beds/Formation (Brunker and Rose 1969) has priority. Campbell (1981) mapped the unit in detail around Kandos Quarries and named it the Kandos Limestone; this nomenclature was followed by Cook (1988b). Further mapping was undertaken by the author who suggested the name Clandulla Limestone (as the Kandos Group had priority; Section 3.3.1); this name was subsequently published and defined by Pemberton et al. (1994). The name is taken from the village of Clandulla, 6 km south of Kandos, and the nearby Clandulla State Forest. The type section defined by Pemberton et al. (1994) was between GR 773500 6360000 to 773500 6359800.

The Clandulla Limestone is quarried for cement at Kandos Quarries by Kandos Quarries Australia Ltd. (formerly Australian and Portland Cement Ltd.) who have quarried the limestone here for around 100 years. The area consists of three main quarries (Fig. 2a): the older No. 1 Quarry which ceased production a number of years ago and now houses processing facilities, including the start of a 6 km long aerial skipway to the town of Kandos (Fig. 3.7f); the larger No. 2 Quarry from which all the current production is being obtained; and the old Charbon Quarry which ceased production many years ago and is directly along strike to the south from Kandos No. 2 quarry (Fig. 2a). Current production from the Kandos No. 2 quarry has an average grade of about 60% CaCO₃, reaching 97.5% CaCO₃ in very pure samples (Lishmund et al. 1986).

3.3.3.2 Distribution, Areal Extent and Thickness

The unit occurs dominantly as a narrow north-northwest trending strip extending 7 km from Clandulla State Forest in the south (GR 773100 6358340), where it is terminated by the Carwell Creek Fault (Fig. 3.7e), to near the "Lucknow" property, north of the Cudgegong Road (GR 771950 6365300). A number of smaller outcrops occur to the east of this main belt, including those in Kandos No. 1 Quarry (GR 773500 6359700) and a number of smaller, partially fault-bounded outcrops near the road into Kandos Quarries (GR 773150 6360450 and GR 773500 6360800) (Fig. 2a).

The Clandulla Limestone attains a thickness of 170 m in the No. 1 Quarry, 110 m in the No. 2 Quarry (Cook 1988b) and 120 m in Charbon Quarry. Around GR 772900 6359400, the unit is only 50
Figure 3.7 (a) Thin-bedded limestone and calcareous mudstone from just north of Charbon Quarry at GR 772500 6361500, 27 metres above the base of the Clandulla Limestone. Limestone beds are rich in silicified Amphiopora sp., and lesser branching tabulate corals (Cladopora sp.) and rare brachiopods (Isorthis sp., Salopina sp., and Machaeraná sp.). (b) Limestone rich in silicified Cladopora sp. and lesser brachiopods (mostly Isorthis sp.) from the banks of Carwell Creek near the "Lucknow" farmhouse (GR 771660 6363910). (c) Limestone horizon from the 'Amphipora biofacies' rich in non-silicified specimens of the stromatoporoid Amphipora sp. Outcrop is in the northern end of Kandos No. 2 Quarry (GR 772320 6361850). (d) Limestone conglomerate/breccia composed of massive, sparsely fossiliferous (rare favositids corals and ovoid stromatoporoids) clasts of blue grey limestone in a brown sandstone matrix. Outcrop is in Charbon Quarry at GR 773100 6358450. (e) Faulted contact between highly sheared and altered silicic volcanics of the Riversdale Volcanics (RV) and the Clandulla Limestone (CL) exposed at the southern end of Charbon Quarry at GR 773090 6358390. Dotted line marks the trace of the fault. (f) Open, gently south-plunging syncline (left) and anticline (centre) exposed in the north wall of Kandos No. 1 Quarry. The Warrah Conglomerate is exposed in the core of the anticline.
40

m thick in contrast to its much greater thickness in the quarries, and appears to represent mainly biostromal deposition without development of reefs. North of the Cudgegong Road, the unit thins steadily, to 30 m thick around GR 7719250 6364000 and 0 m at GR 771950 6365300 where it laps onto the Siluro-Devonian unconformity.

3.3.3.3 Structure and Stratigraphic Relationships

Over most of its outcrop area, the Clandulla Limestone strikes north-northwest-south-southeast and dips west at moderate to low angles (Figs 1 and 2a). At Kandos No.1 Quarry, the limestone is folded into a large anticline-syncline couplet which is well displayed on the north face of the quarry (Fig. 3.7f). Just north of the No.1 Quarry, the Clandulla Limestone occurs on the limbs of a faulted anticline and in the core of a small syncline (Fig. 2a).

The basal contact with the Warrah Conglomerate is conformable and gradational over several metres and has been described in Section 3.3.2.3.

The upper contact with the Yellowmans Creek Formation is also conformable and gradational and is well exposed on the southeastern top bench of Kandos No. 1 Quarry (GR 773640 6359710), just south of Carwell Creek (GR 772850 6359200), on the western top benches of Kandos No.2 Quarry (772310 6361600), and in Charbon Quarry. In these localities, massive coralline-stromatoporoidal limestone of the Clandulla Limestone passes up over several metres into flaggy allodapic limestone, richly fossiliferous calcareous shale, and volcanic breccia/sandstone of the basal Yellowmans Creek Formation. To the south, west and southeast of Kandos No.1 Quarry, the Clandulla Limestone exhibits complex faulted contacts with the Riversdale Volcanics (Fig. 2a). The Carwell Creek Fault cuts slightly across strike to the south of the No. 1 Quarry; this faulted contact is well displayed at the southern end of the Charbon Quarry (Fig. 3.7e). Similarly, near GR 773500 6360800, the Limestone is apparently faulted against the slate member of the Late Silurian Dungeree Volcanics (Fig. 3).

To the north of GR 771750 6364750, the Yellowmans Creek Formation lenses out and the Clandulla Limestone is apparently overlain disconformably by thin volcanic sandstones and conglomerates of the Riversdale Volcanics. North of GR 771810 6365010, this thin unit lenses out and the Glen­dale Limestone Member of the Carwell Creek Formation disconformably overlies the Clandulla Lime­stone; however, mapping an accurate contact between these two units is impossible because of the similarity of lithologies.

3.3.3.4 Petrography and Sedimentology

The west wall of the Kandos No. 2 Quarry and Charbon Quarry contain the best sections through the Clandulla Limestone. Cook (1988b) has divided the limestone in the No. 2 quarry into six biofacies; the following description of the No.2 quarry sequence is modified from his work. The
sequence begins with the Black Shale Biofacies which overlies the Warrah Conglomerate with a contact similar to that seen in Kandos No. 1 Quarry (see above). The Black Shale Biofacies lacks a fauna and contains extensive pyritisation, suggesting quiet, dysaerobic lagoonal conditions at the beginning of deposition. This is overlain by a Stromatoporoid Biofacies which represents development of stromatoporoid biostromes and bioherms, dominated by hemispherical stromatoporoids, deposited in a slightly higher energy, better oxygenated environment. The overlying *Thamnopora* Biofacies is characterised by biostromal development of *Thamnopora* thickets, with *Thamnopora* acting as sediment bafflers in the quiet water conditions. The Reef Biofacies is a reef deposit with the main reef building organism being the ramose branching stromatoporoid, *Amphipora* (Fig. 3.7c). The Biofacies shows a distinctive mound-like geometry which is visible in the western faces of the Kandos No.2 and Charbon Quarries. Such a reef building role for *Amphipora* is only possible in quiet, restricted waters. The Flank Biofacies represents carbonate mud and fine carbonate sand (composed of *Amphipora*, *Thamnopora*, and brachiopods and microsparry calcite) deposited on the reef margins. Some degree of reef orientation is suggested by lithological zonation with this biofacies in Quarry No.2 with fine carbonaceous muds to the south and slightly coarser carbonate sands to the north of the reef structure indicating back- and fore-reef deposits respectively. The Flank Biofacies is overlain by calcaeous shales, allo-dapic limestones and volcaniclastic breccia/sandstone of the basal Yellowmans Creek Formation, well exposed on the top two benches of the western side.

Similar facies are exposed in the Kandos No.1 and Charbon Quarries but are often inaccessible due to the dangerous nature of the quarry faces. The footwall of the Charbon quarry is composed of black carbonaceous shale which passes up into massive, thick bedded *Thamnopora* Biofacies in the middle of the unit. In the south end of Charbon quarry, there are several lensoidal breccia beds, up to 5 m thick, consisting of distorted clasts of massive micritic limestone supported in a brown mud matrix (Fig. 3.7d). These may represent small slumps transitional to debris flows which developed along the edges of oversteepened mud mounds or small bioherms. The Charbon quarry exposures contain only rare evidence of probable reef development - in contrast to the No.2 quarry - and are mainly dominated by the biostromal Stromatoporoid and *Thamnopora* biofacies with small, very local bioherms of hemispherical stromatoporoids developed.

To the south of Carwell Creek at GR 772500 6361500, the Clandulla Limestone is only around 50 m thick in contrast to its much greater thickness in the quarries, and appears to represent mainly biostromal deposition without development of reefs. The section is condensed and much more richly fossiliferous than the quarry sections and silicification is present throughout. The limestone lacks basal black shale and consists of mainly thick-bedded biomicrite at the base dominated by high density but low diversity stromatoporoid (mainly *Amphipora*) and coral (mainly favositid) faunas, suggesting a shallow, somewhat restricted environment (Fig. 3.7b). Towards the top of the Clandulla Limestone, the limestones become more thin bedded and brachiopods appear in increasing numbers, reflecting
deepening of the environment and more open marine conditions (Fig. 3.7a). In the upper 2 m of the Clandulla Limestone, thin beds of densely packed brachiopods (Cyrtina, Eoschuchertella, Machaeraria, Schizophoria) occur, possibly representing condensed faunal sections during times of low carbonate accumulation. Above the brachiopod beds, the Yellowmans Creek Formation begins.

To the north of the Lucknow property, the limestone is only approximately 30 m thick - again, lacking black shale at the base - and intensely dolomitized and stylolitised, consisting of mostly dolomite or dolomitic limestone with patches of more pure limestone. The carbonates are massive and moderate to thick bedded with rare silicified horizons of densely packed branching tabulate corals (Cladopora) and brachiopods (Salopina, Machaeraria and Isorthis).

3.3.3.5 Fauna and Age

The unit contains abundant non-silicified stromatoporoid-Thamnopora communities. In addition, several horizons have yielded large silicified shelly faunas and conodonts. Sampling was along two sections across the unit, and several spot samples. Section 1 is located northeast of the "Lucknow" property between GR 771925 6364640 (base) and GR 771820 6364680 (top) where the limestone is approximately 30 m thick and intensely dolomitized. Six samples were taken across the unit (Table 3.2; Fig. 3.8). Of these, sample C30 yielded Icriodus woschmidti woschmidti, Ozarkodina remscheidensis remscheidensis, and Oulodus sp.; icriodid fragments (Icriodus sp.) and O. r. remscheidensis were encountered in sample C30b, higher in the unit (Table 3.2; Fig. 3.8). These samples yielded a low diversity silicified macrofauna dominated by branching tabulate corals (Cladopora sp.) and brachiopods (Machaeraria catombalensis, Salopina sp., Isorthis sp.; identifications by A.J. Wright).

The second section was taken just south of Carwell Creek, to the southwest of Kandos No.1 Quarry between GR 772950 6359220 (base) and GR 772850 6359200 (top); this was proposed as the type section of the unit (then the Kandos Limestone) by Campbell (1981). In the lower to middle parts of the unit, macrofaunas are very sparse and microfaunas are dominated by scolecodont and the conodonts Oulodus sp., and Ozarkodina remscheidensis remscheidensis, in addition to simple cones (Table 3.3; Fig. 3.8). Macrofauna from the lower part of the section is mostly Amphipora, corals (Cladopora sp. and favositids indet.) and rare brachiopods (Isorthis sp. and Salopina sp.; identifications by A.J. Wright). Brachiopod rich beds (with abundant Cyrtina sp., Eoschuchertella sp., Machaeraria catombalensis, and Schizophoria sp.; identifications by A.J. Wright) in the upper 5 m of the Clandulla Limestone yielded abundant O. r. remscheidensis, O. excavata excavata, Amydrotaxis cf. sexidentata, rare Icriodus postwoschmidti, Oulodus aclys, and abundant fish plates (Table 3.3; Fig. 3.8).

Spot samples C8 and C9 just north of the Cudgegong Road at GR 772180 6362430, approximately 15 m above the base of the unit, yielded very similar faunas of O. r. remscheidensis, Oulodus sp. and abundant scolecodonts.
Table 3.2 Conodont faunas from the Clandulla Limestone, Lucknow section.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Metres above base</th>
<th>Fauna</th>
</tr>
</thead>
</table>
| C30        | 12                | Icriodus woschmidtii woschmidtii  
Ozarkodina remscbeidensis remscbeidensis  
Oulodus sp. |
| C30a       | 15                | Ozarkodina remscbeidensis remscbeidensis  
Oulodus cf. aclys |
| C30b       | 18                | Icriodus sp.  
Ozarkodina remscbeidensis remscbeidensis  
Oulodus sp. |
| C30c       | 21                | Barren |
| C30d       | 25                | Ozarkodina remscbeidensis remscbeidensis  
Panderodus sp. |

Table 3.3 Conodont faunas from the Clandulla Limestone, Carwell Creek section.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Metres above base</th>
<th>Fauna</th>
</tr>
</thead>
<tbody>
<tr>
<td>C3a</td>
<td>2</td>
<td>Panderodus unicostatus</td>
</tr>
</tbody>
</table>
| C3         | 12                | Ozarkodina remscbeidensis remscbeidensis  
Oulodus cf. aclys  
Panderodus sp.  
abundant scolecodonts |
| C3b        | 15                | Ozarkodina remscbeidensis remscbeidensis  
Oulodus aclys  
Panderodus unicostatus |
| C4         | 24                | Ozarkodina remscbeidensis remscbeidensis |
| C5         | 45                | Icriodus postwoschmidtii  
Ozarkodina remscbeidensis remscbeidensis  
Oulodus aclys |
| C5a        | 48                | Icriodus postwoschmidtii  
Ozarkodina remscbeidensis remscbeidensis  
Oulodus sp. |
Figure 3.8 Conodont fauna from the Clandulla Limestone (see also Fig. 3.11).

*Ozarkodina remschaidensis remschaidensis* ZIEGLER, 1960.
a. Pa element, lateral view of AMF 101510; C3a, x36
b. Pa element, lateral view of AMF 96369; C3b, x52
c. Pa element, upper view of AMF 101511; C3a, x36.
d. Pa element, lower view of AMF 96369; C3b, x52.
f. Pa element, lateral view of AMF 101512; C4, x52.
h. Sb element, AMF 101513; C3a, x32.
i. Sc element, AMF 101514; C4, x28.

*Amydrotaxis* cf. *sexidentata* MURPHY & MATTI, 1982

j. Pa element, lateral view of AMF 101515; C5, x52.
g. Pa element, upper view of AMF 101516; C5, x52.

*Icriodus woschmidtii woschmidtii* ZIEGLER, 1969

j. Pa element, upper view of AMF 96391; C30.
k. Pb element, AMF 96408; C30, x89.
l. M element, AMF 96407; C30, x89.

*Oulodus* sp.

m. Pa element, AMF 96389; C30 (wrongly labelled W57 in Colquhoun 1995b), x64.
The presence of *I. w. woschmidti* in the lower part of the unit north of "Lucknow" and *Icriodus postwoschmidti* near the top indicates that the Clandulla Limestone spans the *woschmidti-eurekaensis* boundary. A more precise assessment of the position of the contact within the limestone is not possible as the two zonal species occur in different sections of unequal thickness; however, the contact is provisionally placed in the middle to lower third of the unit.

### 3.3.3.6 Environment of Deposition

The Clandulla Limestone represents a shallow marine carbonate platform deposit characterised by mainly biostromal and local biohermal carbonate deposition. Evidence of bioherms is largely restricted to the thickest sequences in the three quarries. Cook (1988b) regarded the unit here as a quiet water reef deposit with local biohermal buildups dominated by hemispherical stromatoporoids and *Amphipora*, and basal black shales probably representing dysaerobic lagoonal conditions which developed behind the major bioherms. *Thamnopora* dominated biostromes possibly represented inter-reef deposits and carbonate sands occur in foreslope areas. Local foreslope breccias were developed to seaward of the main platform; however, their paucity suggests that, in general, fairly gentle slopes into basinal areas prevailed.

In areas away from the three quarries, the Clandulla Limestone lacks evidence of bioherms and appears to represent biostromal carbonate deposition throughout. In these areas, the unit is condensed, with its thickness characteristically less than half that in the quarries, and lacks the variety of facies. However, depositional environments varied both laterally and vertically through the unit with very prolific low diversity stromatoporoid-coral faunas at the base suggesting shallow, perhaps oxygen-poor waters. These are replaced up sequence by *Cladopora*, favositid, stromatoporoid and brachiopod facies, and finally by dense brachiopod-coral dominated facies close to the top of the unit, suggesting a transition to slightly deeper, better circulated waters in a quiet open shelf environment. To the north of "Lucknow", rich *Cladopora* and brachiopod facies occur throughout the sequence, suggesting a slightly deeper, or less restricted, environment was maintained throughout deposition of the unit.

The facies and faunal succession indicates a transgression throughout the deposition of the unit, with very shallow, restricted stromatoporoid-coral faunas at the base passing upwards into thin-bedded brachiopod-dominated facies and finally into allogenic limestone of the basal Yellowmans Creek Formation (Section 3.3.4.4). This succession indicates that transgression eventually outpaced carbonate deposition, leading to drowning of the Clandulla carbonate platform.

In summary, the Clandulla Limestone represents a transgressive carbonate platform which was dominated by biostromal carbonate deposition with local quiet water reefs, the latter with associated fore-, back-reef and lagoonal deposits. Inter-reef areas were dominated by much thinner biostromal deposits.
3.3.4 YELLOWMANS CREEK FORMATION Day (Millsteed 1992)

3.3.4.1 Synonymy and Definition

Sussmilch (1934) described over 500 m of claystones and acid tuffs to the west of Kandos Quarries, assigning them to the uppermost part of his Middle Devonian Kandos Series. Similarly, Game (1935) recorded 640 m of Middle Devonian clay-slates and acid tuffs from this area. Exposures of similar rocks to the east of Cudgegong were, however, regarded by Game (1935) as Late Devonian in age. Lavers (1960) included shales from west of Kandos Quarries in his broadly-defined Early Devonian Kandos Formation (Table 3.1). Wright (1966) included similar shales cropping out near the Cudgegong Road west of Carwell Creek in the Late Devonian Lawsons Creek Shale - as did Offenberg et al. (1971). Brown (1974), however, included these rocks within both the Lawsons Creek Shale and the Early Devonian Carwell Creek beds, separated by a faulted contact. Likewise, Campbell (1981) included shales from west of Kandos Quarries within an Early Devonian unnamed unit faulted against the Lawsons Creek Shale.

Mapping during this study and by Cook (1988a) found no evidence to substantiate the proposed faulting and, furthermore, found the above lithologies consistently underlie the Early Devonian Roxburgh Formation, indicating they are equivalents of the Early Devonian Yellowmans Creek beds - a unit proposed by Millsteed (1985, 1992) to the east of Cudgegong. The unit is raised to formation status (first published by Pemberton et al. 1994) based on increased knowledge of its boundaries, age, and areal extent. The representative section for the Yellowmans Creek Formation is taken from a creek west of the Kandos No.2 Quarry between GR 772300 6361600 and GR 771900 6361300 (Fig. 3.9). The base of the unit is defined as the first appearance of shaly sediment at the top of the Clandulla Limestone (Section 3.3.3.3). This differs from the proposal of Cook (1988b) who suggested a volcanic-rich breccia-sandstone bed, higher in the sequence (Fig. 3.9), as the base of the Formation - thereby including the underlying outcrops of calcareous shale in the Clandulla Limestone (formerly his lower Kandos Limestone). However, this breccia-sandstone unit is not laterally persistent and often occurs as up to 5 discrete lenses separated by calcareous shales. Cook's (1988b) calcareous shale biofacies of the Clandulla Limestone is therefore included in the basal Yellowmans Creek Formation.

3.3.4.2 Distribution, Areal Extent and Thickness

The Yellowmans Creek Formation crops out extensively in the Carwell Creek valley, extending as a southward-widening strip approximately 7.7 km from GR 771800 6364650 in the north, to GR 773000 635700 in Clandulla State Forest in the south (Fig. 2a). Further outcrops occur to the west of Mt Hope (GR 765500 6367500), east of the former position of Cudgegong (Fig. 3). Estimated total areal extent of the unit is approximately 8 km².
The Formation thickens substantially to the south and west - a characteristic feature of the Early Devonian units (Figs 3.3, 3.4). Unit thickness increases rapidly from 0 m at GR 771800 6364600 and 25 m at GR 771800 6364200, to >340 m west of "Lucknow" (around GR 771200 6363450), reaching 450 m in the type section and approximately 600 m to the west of Charbon Quarry. The base is not exposed to the west of Mt Hope, yet at least 200 m of the unit is present here.

3.3.4.3 Structure and Stratigraphic Relationships

Over much of its area of outcrop, the Yellowmans Creek Formation has a consistent northwest-southeast strike and dips to the west at 35° to 65°. To the north and south of the Cudgegong Road (near GR 770500 6364200), and again to the north of Kandos No. 2 Quarry and in Kandos No. 1 Quarry and west of Charbon Quarry, the unit is folded, along with the other Early Devonian units, into a series of open to close, north- and south-plunging folds (Figs 1, 2a, and 3). Exposures to the west of Mt Hope dip mainly east at moderate to high angles.
<table>
<thead>
<tr>
<th>Facies A: Thinly interbedded shale and sandstone</th>
<th>Shale: Light green, khaki, or brown. Very fissile to moderately indurated. Bioturbation is frequently visible in thin section, disrupting colour and grain size laminations. Composed of angular to subrounded, well-sorted framework grains of medium to fine silt-size in a supporting matrix of highly aligned clay minerals, chlorite, patches of authigenic hematite, and thin stringers of carbonaceous material. Framework grains comprise: monocrystalline quartz (20-25%, straight to undulose extinction); cherty rock fragments (3-5%, possibly aphanitic silicic volcanic fragments); plagioclase (2-3%); biotite and muscovite (1-2%) (Fig. 3.10b). XRD analysis indicates the presence of chlorite, muscovite/illite, kaolinite, quartz, hematite and biotite. Sandstone: Grey to brown, tightly packed, poorly to moderately-well sorted and immature. Composed of angular to subrounded, moderate sphericity framework grains of fine to very fine sand-size in a supporting matrix of clay minerals, chert, and chlorite. Framework grains comprise monocrystalline quartz (45-50%, undulose extinction); silicic volcanic rock fragments (5-8%, aphyric and porphyritic); biotite and muscovite (3-4%) and rare plagioclase. A quartz cement is occasionally present. Accessory hematite, zircon, sphene, tourmaline and ?pyroxene. Common mottled textures, due to bioturbation.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Facies B: Calcareous shale with interbeds of allodapic limestone</td>
<td>Shale: Dark grey to green, moderately fissile, well-bedded, richly fossiliferous locally. Composed of poorly-sorted, angular to subrounded, silt-sized framework grains of volcanic quartz, calcite, fossil fragments, and rare plagioclase - in highly variable proportions - in a submicroscopic matrix of micrite, clay minerals and carbonaceous material. Pyrite and muscovite are occasionally present as accessories. XRD analysis by Thomson (1988) indicates calcite, quartz, muscovite/illite, minor pyrite, kaolinite, and montmorillinite. Assays indicate CaCO₃ of 60% to 85% (Thomson, 1988). Allodapic Limestone: Light grey to black, massive, unfossiliferous micrite. Rare silt laminations and bioclastic textures.</td>
</tr>
<tr>
<td>Facies C: Volcaniclastic sandstone to breccia</td>
<td>Sandstone: Light to dark grey or black depending on the calcareous content. Very poorly sorted, tightly packed, and texturally immature. Angular to very angular fine to very coarse sand-sized framework grains comprise: volcanic quartz (20-25%, straight extinction, embayments and bipyramidal terminations); silicic volcanic rock fragments (25-30%, porphyritic, vitriclastic textures); calcite (0 to 20%, massive micrite or fossil fragments); shale (3-4%) and rare felspar (&lt;1%, heavily altered) and very rare metamorphic quartz (Fig. 3.10e). Matrix is supporting and generally consists of mainly calcareous silt or clay, carbonaceous material and chert. However, in some beds the matrix consists entirely of chert and authigenic clays which define flowage around framework grains and assume shard-like shapes in places (cf. pseudo-eutaxitic texture of Cas and Wright, 1987) (Fig. 3.10e). Breccia: Gradational to sandstone. Very poorly sorted, angular to subangular clasts of low sphericity supported in a matrix identical to the above sandstone. Clasts generally 1 to 10 cm across; Cook (1988b) reports rare megaclasts to 1.2m. Clasts comprise limestone, shale, calcareous shale, silicic volcanics, lithic sandstone and fossil fragments in highly variable proportions.</td>
</tr>
<tr>
<td>Facies D: Dolomite and Limestone</td>
<td>Mostly massive, brown, intensely jointed and stylolitised micritic dolomite. Intervals of well-bedded biomicritic limestone occur at the top of one lens and consist of up to 45% of well preserved fossils including crinoid stems and ossicles, brachiopods (mostly articulated), bryozoans, corals, and ostracodes, in a matrix of micritic calcite with patches of diagenetic dolomite.</td>
</tr>
</tbody>
</table>

Table 3.4 Summary of petrography of the Yellowmans Creek Formation
The calcareous shale at the base of the Formation is conformable and gradational over 5 m to 8 m with the upper limestone of the Clandulla Limestone (Fig. 3.10c). Excellent exposures of this contact occur on the western side of Kandos No. 2 Quarry and Charbon Quarry, and on the top bench of Kandos No. 1 Quarry. In some areas, however, this contact is less clearly defined; for example, north of the Cudgegong Road at GR 771700 6362700, where abundant interbeds of volcanioclastic sandstone occur near the contact, and west of Charbon Quarry, where common thick interbeds of limestone persist for a greater distance into the shales. South of GR 730500 6368310, the base of the Yellowmans Creek Formation is not seen as the unit is faulted against the Riversdale Volcanics along the Carwell Creek Fault; the fault cuts across strike, eventually truncating the entire unit at GR 773000 6356950 in Clandulla State Forest (Fig. 2a). The base of the Yellowmans Creek Formation is not seen to the west of Mt Hope, and the unit is faulted against the Late Silurian Willow Glen Formation and Windamere Volcanics by the Cudgegong Fault (Fig. 3).

The top of the Formation is also conformable and gradational, with the proportion and thickness of sandstone interbeds increasing gradually towards the basal sandstones of the Roxburgh Formation (Fig. 3.9). Good exposures of this contact occur beside Carwell Creek at GR 771100 6364100 and to the west of Kandos No. 2 Quarry (GR 771900 6361300). A similar gradational contact has been recorded to the immediate west by Cook (1988a). To the west of Mt Hope, this contact is poorly exposed, although Millsteed (1985, 1992) regarded it as sharp and non-gradational. To the north of "Lucknow" and east of GR 771300 6364300, the Roxburgh Formation lenses out, and the Yellowmans Creek Formation is overlain with probable disconformity by the Riversdale Volcanics (Fig. 1). Similarly, to the south of GR 772500 6357700, the disconformity cuts out all of the Roxburgh Formation, and for 300 m south, the Riversdale Volcanics disconformably overlies the Yellowmans Creek Formation (Fig. 2a).

### 3.3.4.4 Petrography and Sedimentology

On the basis of lithological and sedimentary features, the Yellowmans Creek Formation has been divided into four facies (Fig. 3.9). The petrography of each facies is described in Table 3.4 and the major characteristics and depositional processes are outlined in Table 3.5. These features are summarised below.

#### Facies A: Thinly Interbedded Shale and Sandstone

**Description:** Massive to laminated, green, brown, or rarely black shales containing thin, sharp-based interbeds of graded sublitharenite are the dominant facies (Tables 3.4, 3.5; Fig. 3.10a, b). Best exposures occur on the banks of Carwell Creek at GR 771400 6369300 and GR 772300 6359700 and in several creeks 200 m to 300 m west of Kandos No. 2 Quarry and Charbon Quarry.
<table>
<thead>
<tr>
<th>Name</th>
<th>Lithology</th>
<th>Percentage of total exposure</th>
<th>Bed thickness and geometry</th>
<th>Boundaries</th>
<th>Structure</th>
<th>Fauna</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Facies A: Thinly Interbedded Shale &amp; Sandstone</td>
<td>Shale interbedded with fine sandstone to silty sandstone. Shale generally dominant with sand: shale ratios avg. 0.3:1. Amount of sandstone increases up-section.</td>
<td>80% to 85%</td>
<td>Shale: 5 cm to 60 cm, laterally persistent. Sandstone: 0.2 cm to 25 cm, generally laterally persistent, planar sheets; rarely lensoidal over several metres.</td>
<td>Sandstone beds show sharp lower bed contacts with sharp or gradational upper bed contacts. Bed bases may be loaded, flame cast, or show shallow scour s.</td>
<td>Sandstone: massive or graded throughout. Occasional parallel lamination and rare ripple cross lamination in the upper parts of beds. Possible hummocky cross-stratification noted near top of unit. Shale: massive or finely laminated.</td>
<td>Macrofauna rare: brachiopod moulds, and crinoid stems occur sporadically in the shale; preservation is very poor. Bioturbation is absent to intense; identifiable forms include Chondrites and Planolites.</td>
<td>Low-energy suspension deposited muds interbedded with distal storm sandstone beds.</td>
</tr>
<tr>
<td>Facies B: Calcareous Shale with Thin Interbeds of Allodapic Limestone</td>
<td>Calcareous shale with thin interbeds of massive, flaggy limestone. Transitional to underlying Clandulla Limestone</td>
<td>10-15%</td>
<td>Shale: 5-30 cm thick; laterally persistent beds. Limestone: 2-60 cm thick; laterally persistent tabular beds.</td>
<td>Limestone beds show sharp and planar upper and lower contacts with calcareous shale.</td>
<td>Shale: massive or planar to wavy laminated. Limestone: massive or parallel laminated, or with silty laminae.</td>
<td>Rich fauna in basal calcareous shale and limestone, including: brachiopods, molluscs, trilobites, ostracodes, rostroconchs, bryozoans, sponges and abundant conodonts. Sporadic brachiopods, ostracodes and conodonts in limestone beds higher above base of unit.</td>
<td>Low energy, subtidal suspension deposition adjacent to an area of carbonate production, interrupted repeatedly by turbidity currents of carbonate detritus.</td>
</tr>
<tr>
<td>Facies C: Volcaniclastic Sandstone and Breccia.</td>
<td>Fine to coarse-grained, very poorly sorted volcanic-rich sandstone gradational to matrix-supported, granule to boulder breccia.</td>
<td>2% to 4%</td>
<td>0.5 m to 12 m thick beds. Laterally continuous tabular sheets which pinch and swell slightly along strike. Typically lensoidal over 10's of metres.</td>
<td>Sharp, planar to gently undulating base and sharp, planar top contacts with Facies B.</td>
<td>Typically massive, ungraded, disorganised, and poorly-sorted. Normal, coarse-tail, and distribution grading from breccia to sandstone occasionally present.</td>
<td>Barren in outcrops north of Cudgegong Road. Core from Kandos No. 2 Quarry reveals abundant fragments of brachiopods, crinoids and corals.</td>
<td>Deposition from subaqueous, cohesive debris flows and/or slurry flows which evolved from syn-sedimentary slumps or slides, or directly from subaerial pyroclastic flows passing into water.</td>
</tr>
<tr>
<td>Facies D: Dolomite and Limestone</td>
<td>Massive, sparsely fossiliferous dolomite to richly fossiliferous limestone</td>
<td>&lt;1%</td>
<td>50 cm to 1 m at base of lens; 10-25 cm at top of lens</td>
<td>Sharp and planar bed bases and tops.</td>
<td>Massive and poorly bedded throughout except at the top of one lens where thin bedded richly fossiliferous limestone occurs.</td>
<td>Rare stromatoporoids and tabulate corals. Rich fauna of brachiopods, trilobites, corals and gastropods in upper parts.</td>
<td>Intertidal to subtidal carbonate deposited in restricted conditions; more open marine conditions present locally.</td>
</tr>
</tbody>
</table>
Figure 3.10 (a) Facies A: brown to green shale with very thin, fine sandstone interbeds representing distal storm beds; these display normal grading, load casts, basal scouring and ripple cross lamination. (b) Facies A: Photomicrograph of thinly interbedded, fine-grained, sublithic sandstone (dominated by rounded to subangular quartz and silicic volcanic grains) and mudstone. Crossed polars. (c) Rhythmically interbedded calcareous shale and allodapic limestone of Facies B near the base of the Yellowmans Creek Formation. Outcrop is on the western bench of Kandos No. 2 Quarry (GR 772250 6361410) Hammer for scale. (d) Volcaniclastic sandstone-breccia bed of Facies C displaying abundant elongated and deformed clasts of intraformational calcareous shale supported in a matrix of coarse-grained, crystal-rich volcaniclastic sandstone. Outcrop is in Charbon Quarry at GR 772900 6358700. (e) Photomicrograph of a crystal-rich volcaniclastic sandstone of Facies C, displaying abundant angular quartz crystal fragments (Q), lithic fragments with vitriclastic textures (L) and platy to rarely cuspate relict shards (S) in a very fine matrix. Plane polarised light. (f) Thinly bedded dolomitic limestone of Facies D showing abundant silicified brachiopods, crinoids and corals. Outcrop is at GR 765480 6367500.
**Interpretation:** The shales show features indicative of low-energy, suspension sedimentation in a tranquil subtidal setting. The paucity of macrofauna, the variable amount of bioturbation, and the occurrence of intercalated carbonaceous material in these shales suggests this suspension sedimentation was generally rapid. The sharp-based graded sandstone interbeds have features characteristic of distal storm beds deposited in a middle to outer shelf setting (Brenchley 1985). The paucity of wave ripples on the tops of these beds indicates deposition was dominantly below storm-wave base; possible hummocky cross-stratification and rippling of bed tops near the contact with the Roxburgh Formation suggests deposition between the fairweather- and storm-wave bases in the upper parts of the unit. Rare palaeocurrent data, derived from 3 ripple cross beds in the sandstone interbeds (Fig. 3.10a), indicate a mean current flowing towards 208°.

**Facies B: Calcareous Shale with Thin Limestone Interbeds**

**Description:** Dark grey to green, thinly bedded, massive to laminated calcareous shales are common in the lower part of sequence (Fig. 3.9), particularly south of the Cudgegong Road (Tables 3.4, 3.5). These shales are richly fossiliferous in places, variably calcareous, locally pyritic, and contain abundant thin (10 to 50 cm thick), flaggy interbeds of massive, bioclastic limestone rich in crinoid and brachiopod debris (rarely silicified), occasionally graded and parallel laminated. The limestone beds increase in abundance and thickness towards the base of the unit (Fig. 3.10c). Excellent exposures of this lithology occur on the western walls of Kandos No. 2 Quarry (Plate 3.11) and Charbon Quarry and in Kandos No. 1 Quarry. The facies is confined to the basal 100-150 m of the unit and can be mapped as a separate member (Fig. 2a); it passes gradationally up-section into non-calcareous shales of Facies A. Prolific silicified shelly faunas and conodont faunas occur in thin bedded limestones in the basal 5 m of the facies on the top bench of the Kandos No. 1 Quarry and to the north of Charbon Quarry (Section 3.3.4.5). Further descriptions of this facies are given by Cook (1988b) and Thomson (1988).

**Interpretation:** The facies represents fairly rapid suspension sedimentation of carbonate and terrigenous mud derived from adjacent carbonate buildups and land sources in a low-energy subtidal setting. The rich, low-diversity fauna (Cook 1988b), and a lack of bioturbation suggests some ecological stress in the environment, perhaps caused by this fairly rapid sedimentation. However, abundant carbonaceous material and pyrite indicates bottom waters and/or substrate may have been locally dysaerobic. The thin limestone beds are interpreted as allochthonous limestones deposited by low density turbidity currents of calcareous detritus derived from the Clandulla Limestone carbonate platform. Such a facies is typical of fore reef areas and the deeper parts of carbonate ramps (Tucker and Wright 1990).
**Facies C: Volcaniclastic Sandstone/Breccia**

*Description:* Poorly-sorted, massive to graded, tabular beds of volcaniclastic, and lesser extra- and intraformational, detritus of fine sand to boulder dimensions are common in the lower parts of the Yellowmans Creek Formation (Figs 3.9, 3.10d, e; Tables 3.4, 3.5). Excellent exposures of this facies occur on the western walls of Charbon Quarry and Kandos No. 2 Quarry (Fig. 3.10d). The facies can be represented by a single bed or by up to 5 beds separated by the calcareous shales of Facies 2. The calcareous content of the facies varies considerably: from absent in the some of the most northern outcrops, to 30-50% in outcrops west of Kandos No.2 Quarry (Table 3.4). Furthermore, the calcareous content tends to decrease upward through each bed.

*Interpretation:* The structure, geometry and sorting of the facies suggests deposition generally by cold, subaqueous, cohesive debris flows to slurry flows (Cas and Wright 1987) which evolved from slumps or slides of rapidly-deposited volcaniclastic, extraformational and local intraformational detritus. The presence of such deposits implies the presence of a palaeoslope, albeit gentle (>1°: Stow 1986), in the depositional setting. The presence of vitriclastic textures, common fracturing of quartz grains, and relic glass shards in the matrix of some beds, indicates initial pyroclastic eruption of the volcaniclastic material. Following Cas and Wright (1987), these beds are not attributed to subaqueous pyroclastic flows unless welding can be proven; however, Cortis (1976) recorded rare thin beds of welded ashflow tuff interbedded with the calcareous shales at Kandos Quarries. This suggests at least some of the volcanic detritus was brought to the setting directly by pyroclastic, rather than epiclastic, processes. Some of the debris flows/slurry flows may therefore have evolved directly from subaerial pyroclastic flows passing into the sea - similar to the model proposed by Cas and Wright (1987, p. 283).

**Facies D: Dolomite and Limestone**

*Description:* Three lenses of dolomite, dolomitic limestone and limestone occur near the top of the Yellowmans Creek Formation to the west of Mt Hope, on the shores of Lake Windamere (around GR 765600 6367500). The carbonate lenses are up to 150 m long and 20-30 m thick and consist mostly of thick bedded dolomite or dolomitic limestone, with sparse non-silicified faunas of stromatoporoids and rare favositid corals. Rare outcrops of thin bedded, richly fossiliferous limestone occur at GR 765680 6367510 and these have yielded rich brachiopod, trilobite, coral and conodont faunas (Fig. 3.10f); these faunas are detailed in Section 3.3.4.5.

*Interpretation:* Facies D represents small carbonate accumulations which developed in sheltered areas of the muddy shelf. The thick bedded and sparsely fossiliferous character of these carbonates suggests a very shallow, restricted environment. The appearance of thin bedded, less dolomitised, and more fossiliferous carbonates in the upper parts of these lenses suggests deeper, more open marine conditions, indicating an upward-deepening trend in each lens. These lenses may therefore represent
mostly intertidal to subtidal carbonates deposited in sheltered embayments during a brief regression, grading upwards to more open marine carbonates and muddy shelf deposits deposited in an ensuing slow transgression.

3.3.4.5 Fauna and Age

Bioturbation and trace fossils (Chondrites sp. and Planolites sp.) were locally abundant in the Facies A shales, particularly at GR 771500 6364200 and GR 772000 6361500 (Table 3.3). Prolific, low-diversity microfaunas were obtained from allodapic limestone and calcareous shales (Facies B) in the basal Yellowmans Creek Formation from the top bench of the Kandos No.1 Quarry (GR 772660 6359710). These are characterised by abundant conodonts (Ozarkodina remscheidensis remscheidensis, Icriodus postmoschmidtii, Oulodus acyls) and fish plates, with lesser numbers of the conodonts Ozarkodina excavata excavata, Amydrotaxis cf. sexidentata, and simple cones (Panderodus sp.); rare scolecodonts were also present in the residues (Fig. 3.11). This conodont fauna is unusual in containing large numbers of partial and complete icriodid specimens; this feature is discussed further in Section 5.3 and in Colquhoun (1995b; Supporting Paper 3). Mawson (in Cook 1988b) listed O. r. remscheidensis, O. e. excavata, Oulodus acyls, and ?Amydrotaxis sexidentata from this locality. Cook (1988b) recorded rich, silicified and dolomitised, non-diverse shelly faunas from the same locality (Table 3.3). Cook (1988b) listed: brachiopods, Isorthis sp. cf. I. festiva, Cyrtina sp., Eoschuchertella sp.; the rostroconch ?Mulceodens sp.; trilobites Apocalymene sp.; ostracodes, bryozoans, gastropods, sponges, and rare tabulate corals. Further sampling for macrofauna by A.J. Wright and the author yielded: brachiopods, Isorthis sp., Howellella sp., Salopina sp., Cyrtina sp., ?Resserella sp., Machaeraria catombalensis; corals, ?Tryplasma sp., favositids; poorly preserved trilobite fragments, gastropods and crinoid ossicles.

A further series of samples (C58 to C62) was taken in thinly bedded flaggy allodapic limestones which crop out to the northwest of Charbon Quarry between GR 772850 6359200 (base) and GR 772700 6359195 (top) in the basal 80 m of the formation. These proved to be mostly barren apart from C58 (approximately 80 m above the base) which yielded Panderodus unicostatus and a single juvenile specimen of Ozarkodina cf. remscheidensis remscheidensis. Sample C59 (approximately 65 m above the base) also yielded Panderodus unicostatus and a single specimen (later broken) which was tentatively referred by Colquhoun (1995b) to Ancryodelloides omus.

Several carbonate lenses (Facies D; see Section 3.3.4.4) occur in the upper 100 m of the Yellowmans Creek Formation just to the northeast of Lake Windamere (around GR 765600 6367500). The lenses are largely thick-bedded, unfossiliferous or stromatoporoidal dolomite and are barren of conodonts (samples C17d, C17e). However, one horizon of thinly bedded limestone at GR 765680 6367510 (Samples C17, C17a to C17c) yielded conodonts including abundant Panderodus unicostatus, P. recurvatus, P. sp., and P. cf. valgus and lesser Amydrotaxis praejohnsoni, cf. O. r.
remschdienis, O. e. excavata, ?Ozarkodina sp., and Ambryodelloides delta (Fig. 3.12) (Colquhoun 1995b; Supporting Paper 5). A large silicified shelly fauna was also obtained from this horizon which was very similar to that listed by Millsteed (1985, 1992) and includes: brachiopods, Atrypa sp., Howellella sp., Eoschuchertella sp., Dolerorthis sp., Dicaelosia sp., Cyrtina sp., Anastrophia sp., Skenediodes sp., pentamerids gen. indet., dalmanellids gen. indet.; trilobites Apocalymene sp.; corals Rhizophyllum sp.; and gastropods (identifications by A.J. Wright, 1994).

Two samples (C39, C39a) were taken in sparsely fossiliferous limestones on the old Riversdale Road north of Lake Windamere (GR 768360 6366970). These yielded rare brachiopods (Cyrtina), solitary corals, and the conodonts O. r. remscheidensis, Amydrotaxis praejohnsoni, and Panderodus sp. This area was mapped as Carwell Creek Formation by Millsteed (1992), but the conodont fauna and abundance of shale in the surrounding outcrops suggest probable structural complexities in this area (see Fig. 3 for a revised map of this area).

The abundance of Icriodus postwoschmidti, together with Amydrotaxis sexidentata indicates the basal 15 m of the Yellowmans Creek Formation is of eurekaensis Zone age. The appearance of possible Ancryodelloides omus 65 m above the base suggests an early delta Zone age (Murphy and Matti 1983), indicating the eurekaensis-delta boundary may lie somewhere between the 15 and 65 metre levels of the unit to the west of Charbon Quarry. The appearance of Amydrotaxis praejohnsoni and ?Ancryodelloides delta indicates the Yellowmans Creek Formation is still within the delta Zone 100 m from the top, although Klapper & Murphy (1980) recorded Amydrotaxis praejohnsoni ranging into the early pesavis Zone.

The shelly faunas are mostly characterised by long-ranging forms which provide no further information on age. The basal fauna from the Kandos No.1 Quarry top bench appears to belong to the long-ranging Machaeraria catombalensis fauna which has been widely recorded in Lochkovian, and possibly younger, strata in New South Wales (Wright et al. in prep.; Supporting Paper 11).

3.3.4.6 Environment of Deposition

A shallow marine setting for deposition is indicated by the fauna; the patchy, though occasionally abundant, bioturbation; and the gradational lower contact with the Clandulla Limestone. The setting is envisaged as a quiet, subtidal, dominantly sub-storm-wave base environment on the middle to outer portions of a gently sloping shelf; rare cross beds suggest the slope was probably towards the southwest. Sedimentation was dominated by fairweather suspension deposition of fine-grained siliciclastic and carbonate detritus. Periodically, this was interrupted by distal, storm-generated incursions of quartz-rich sand and, in the lower parts of the unit, by debris flows/slurry flows composed primarily of volcaniclastic detritus; some of the later may have developed directly from subaerial pyroclastic flows. Relatively rapid sedimentation and locally restricted oxygen conditions would account for the low density and diversity of body and trace fossils.
Figure 3.11 Conodont fauna from alodapic limestones of the basal Yellowmans Creek Formation, top bench of Kandos No.1 Quarry (GR 772660 6359710).

*Icriodus postwoschmidti* MASHKOVA, 1968
a. Pa element, upper view of AMF 96362; C2a, x47
b. Pa element, upper view of AMF 96361; C5 (upper Clandulla Limestone), x47
c. Pa element, upper view of AMF 101517; C2a, x52.
d. Pa element, upper view of AMF 101518; C2b, x47.
i. Pb element, AMF 96406; C2, x103.
j. Pb element, AMF 101519; C2, x86.
e. M element, AMF 96405; C2, x102.
f. M element, AMF 101520; C2, x52.
g. M element, AMF 96403; C2, x103.
h. M element, AMF 101521; C2, x103.
k. M element, AMF 96404; C2, x103.
s. Sa element, AMF 96399; C2, x86.

*Ozarkodina remscheidensis remscheidensis* ZIEGLER, 1960.
l. Pa element, lateral view of AMF 96368; C2, x51.
o. Pa element, upper view of AMF 96368; C2, x52.
m. Pa element, lateral view of AMF 101522; C2, x56.
p. Pa element, upper view of AMF 101522; C2, x56.
n. Pb element, AMF 101523; C2b, x69.
v. Pb element, AMF 101524; C2a, x34.
y. Pb element, AMF 101525; C2, x56.
t. M element, AMF 101526; C2b, x52.
u. Sa element, AMF 101527; C2a, x56.
q. Sc element, AMF 101528; C2, x43.
z. Sc element, AMF 101529; C2a, x39.

*Oulodus aclys* MAWSON, 1986.
w. Pa element, AMF 963393; C2, x39.
r. M element, AMF 101530; C2b, x56.

Unidentified fish plate
a1. AMF 101531; C2, x47.
Figure 3.12 Conodont fauna from limestone lenses (Facies D) from near the top of the Yellowmans Creek Formation at GR 765680 6367510 (Samples C17, C17a to C17c) and the old Riversdale Road north of Lake Windamere at GR 768360 6366970 (Samples C39, C39a), and from the basal Yellowmans Creek Formation from the top bench of the Kandos No.1 Quarry (GR 772660 6359710) (Sample C2b).

  a. Pa element, lateral view of AMF 96387; C17a, x69.
  e. Pa element, lower view of AMF 96387; C17a, x69.
  p. Pa element, lateral view of AMF 96364; C17, x52.
  s. Pa element, upper view of AMF 96364; C17, x52.

Amydrotaxis praejohnsoni KLAPPER, 1969.
  m. Pa element, lateral view of AMF 96392; C39, x43.
  j. Pa element, upper view of AMF 96392; C39, x43.
  t. Pb element, AMF 96393; C17a, x52.
  w. Sb element, AMF 96360; C17a, x52.
  y. Sb element, AMF 101532; C17a, x52.

  b. Pa element, lateral view of AMF 101533; C17a, x73.
  f. Pa element, upper view of AMF 101533; C17a, x73.

  i. Pa element, lateral view of AMF 101534; C2b, x65.
  l. Pa element, upper view of AMF 101534; C2b, x78.

Ozarkodina excavata excavata BRANSON & MEHL, 1933.
  u. Pa element, lateral view of AMF 101535; C17, x69.
  v. Pa element, lateral view of AMF 101536; C17, x69.

  c. Pa element, upper view of AMF 101537; C17, x95.
  d. Pa element, lateral view of AMF 101537; C17, x95.
  r. Pb element, AMF 101538; C17a, x86.

?Ozarkodina sp. (broken specimen)
  h. Pa element, lateral view of AMF 101539; C17a, x43.
  k. Pa element, upper view of AMF 101539; C17a, x43.

Bellodella resima PHILIP, 1965.
  z. AMF 96377; C17a, x69.

  o. AMF 96412; C17a, x86.

Panderodus recurvatus RHODES, 1953.
  x. AMF 96398; C17, x60.

Unassigned element
  q. AMF 101540; C17a, x78.
The deposition of muddy shelf deposits of the Yellowmans Creek Formation over the shallow marine carbonates of the Clandulla Limestone suggests a relative deepening of the environment (Johnson et al. 1985), indicating a continuation of the transgression seen through the Warrah Conglomerate and Clandulla Limestone into the basal parts of the Yellowmans Creek Formation. The transition from the fossiliferous calcareous shales to barren grey-brown shales at the 100-150 m level of the unit's representative section, and the disappearance of allodapic limestones at this level, suggests drowning of the Clandulla Limestone carbonate platform due to continuing transgression, thereby cutting off the supply of calcareous detritus to the basinal areas and allowing mud to accumulate by suspension deposition.

The gradational contact with the siliciclastic shelf and nearshore sediments of the overlying Roxburgh Formation, together with the appearance of lenses of dolomite (Facies D) in the upper parts of the Yellowmans Creek Formation, indicates relative shallowing and a switch to regressive conditions in the upper parts of the unit.

In summary, the Yellowmans Creek Formation represents a transgressive to regressive sequence of shale, calcareous shale, sublitharenite, and volcanic-rich sandstone and breccia which was deposited on a low-energy, storm-influenced, muddy shelf during a period of active silicic pyroclastic volcanism.

3.3.5 ROXBURGH FORMATION Dar (Pemberton 1980a)

3.3.5.1 Introduction

The Roxburgh Formation was the subject of detailed study by the author, focussing on the sedimentology, palaeoenvironments and taphonomy of the faunas. The results of these investigations were published in Sedimentary Geology in June 1995 and a reprint of this paper is included as Supporting Paper 4 (Colquhoun 1995a). A small conodont fauna obtained from the Roxburgh Formation was described in Colquhoun 1995b (Supporting Paper 5). A synthesis of these papers, along with some additional data, is presented here.

3.3.5.2 Synonymy and Definition

Game (1935) described a thick sequence of Upper Devonian shallow marine quartzites to the east of Cudgegong. Offenberg et al. (1971) mapped these strata as part of the Gulgamree beds, a poorly defined Siluro-Devonian unit. Pemberton (1980a) proposed the name Roxburgh Formation after the County of Roxburgh for a thick sequence of quartz sandstone and shale which crops out to the southeast of Cudgegong. Subsequent mapping by Millsteed (1985, 1992), Cook (1988a, 1990), Booth (1990) and Colquhoun (1995a) considerably expanded the known outcrop of the unit.

Pemberton (1980a) suggested a representative section along Oakey Creek between GR 766700 6364000 and GR 765900 6364150; however, the base of the unit is not exposed here. No type section
was originally defined for the Roxburgh Formation. A good section through the unit occurs to the west of Kandos No. 2 Quarry (between GR 771600 6361550 and GR 770700 6361200) where the top and base of the unit crop out and the unit is entirely west-dipping. This section is suggested herein as a suitable type section.

### 3.3.5.3 Distribution, Areal Extent and Thickness

The Roxburgh Formation crops out prominently in three separate areas (Fig. 3.13). Between Rylstone and Cudgegong the unit occurs in two elongate belts, one just to the east of Cudgegong (largely fault-bounded) and the other in the Carwell Creek valley further to the east (Figs 1, 2a, and 3). Excellent exposures occur in a 3.5 km stretch of road cuttings along the Cudgegong Road near Lake Windamere (Figs 3.13, 3.15) and in several major creeks to the south of the road. Further outcrops of the Roxburgh Formation occur to the northwest of Ilford, approximately 10 km to the south of Cudgegong (Booth 1990; this study). Total outcrop area is estimated at 16.5 km².
The Roxburgh Formation thins dramatically to the east across the district from a maximum of >750 m southeast of Cudgegong, to 125 metres at GR 770500 6362300, to zero just east of Carwell Creek where it forms an offlap contact with the Yellowmans Creek Formation (Fig. 3.3). Isopachs suggest a depositional strike/palacoshoreline trend of approximately north-south; however, due to the locally disconformable upper contact, it is uncertain how much section has been removed and the overall effect this has on unit isopachs. Nevertheless, the regional trend of the Capertee High (assessed by delineating a line separating shallow marine and subaerial sequences, such as the Roxburgh Formation, in the east, from slope and basin sequences deposited on the western margin of the High) was north-northwest during the Early Devonian, supporting an approximately north-south shoreline trend for the Roxburgh Formation.

3.3.5.4 Stratigraphic Relationships

The Roxburgh Formation overlies the Yellowmans Creek Formation with a conformable and gradational contact (Section 3.3.4.3). The nature of the contact between the Warrah Conglomerate and the Roxburgh Formation to the north of Ilford is uncertain and unexposed, but is presumably disconformable to paraconformable.

The Roxburgh Formation is overlain by the Riversdale Volcanics with the contact varying from disconformable in the east of the district to conformable or paraconformable further west. The nature and significance of this contact is documented further in Section 3.3.6.3.

3.3.5.5 Petrography and Provenance

The Roxburgh Formation is composed of 70% to 80% very fine-grained to rarely coarse-grained, moderate to poorly sorted, and immature to submature sandstone, together with 20% to 30% shale and mudstone. Angular to subrounded framework grains indicate a threefold provenance for the unit: (1) a compositionally mature sedimentary basement source (composite quartz, sedimentary rock fragments, and jasper); (2) a compositionally immature silicic volcanic source (embayed quartz, plagioclase, and porphyritic to aphanitic silicic volcanic rock fragments); and (3) locally abundant, but minor intrabasinal framework grains (limestone, mudstone intraclasts, and fossils) (Cook 1990, Millsteed 1992; this study). Sublitharenite compositions (after Folk 1974) dominate (Fig. 3.19b), but a complete gradation exists between quartzarenite (Fig. 3.19a) and volcanic litharenite depending on the relative dominance of the two main source areas (Fig. 3.14). Sandstone composition often varies markedly between beds, reflecting sudden influxes of coarser volcaniclastic sediment; few consistent compositional trends are therefore apparent, although, in general, sandstones become increasingly volcanic rich up-section and in the east of the study area, reflecting increasing proximity to the volcanic source (Fig. 3.21). Subordinate lithologies (<1%) are described below and these include volcaniclastic conglomerate, accretionary lapilli and ashfall tuff, and limestone.
3.3.5.6 Facies analysis

Ten recurring lithofacies have been recognised in the Roxburgh Formation based on lithology, bed thickness, internal sedimentary structures, and faunal content (Table 3.6). Approximately 550 m of detailed sedimentary section was measured mainly in road cuttings on the Cudgegong Road and in cliff sections (Figs 3.15, 3.16); these were augmented by descriptions of other exposures throughout the district. Faulting and mesoscopic folding are common in the sections, although fault displacements were usually small and easily accounted for; however, some larger displacements caused pronounced breaks in some sections. Similarly, reliable correlation between many roadcut sections is not possible, restricting reconstruction of the overall vertical trends within the formation. Section RC10, however, shows a nearly complete exposure of the upper 175 m of the formation (Fig. 3.16).

3.3.5.6.1 Facies A: Massive Mudstone to Very Fine Sandstone

Description: Facies A consists of brown-grey mudstone to very fine silty sandstone with rare interbeds of fine- to medium-grained sandstone. Typically, the facies occurs as 2 to 23 m thick units at the base of major coarsening-upward cycles (Fig. 3.16, section RC3, 70-85 m; RC7, 37-60 m; RC5, 8-10 m). Individual mudstone beds are 10 to 190 cm thick, laterally continuous at outcrop scale, and lack internal sedimentary structures apart from rare horizontal lamination. Rare, 0.3-4 cm thick, laterally persistent siltstone beds with sharp, loaded bases, and normal grading appear irregularly throughout the facies. Bioturbation varies from sparse to locally intense and taxa are typical of the Cruziana ichnofacies (Pemberton et al. 1992) with Zoophycos, Chondrites, and Planolites dominant.

Taphonomy: Facies A contains the majority of the well-preserved fossils of the Roxburgh Formation; abundant brachiopods, crinoids, trilobites, tabulate and rugose corals, molluscs, stromatoporoids and bryozoans are usually supported in a very fine sand or coarse silt matrix. Typically, the fossils (now moulds) are concentrated in thin (5-15 cm thick) laterally persistent layers; less commonly, fossils occur evenly distributed throughout a bed or are disrupted by intense post-mortem bioturbation. Fossils are mostly unbrashed and uncorroded, ungraded, and display few indications of very strong current activity or extensive transport. Brachiopods (Cyrtina, 'Dolerorthis', Mesodouvilleina, Isorthis, Howellella, Salopina, Eoschuchertella, Gypidula, Spinatrypa, Iridistrophia and rhynchonellids indet.) dominate, and are generally disarticulated, poorly size-sorted, and 60-80% have concave-down orientations, with pedicle valves outnumbering brachial valves by approximately 3 to 1. The shells vary from loosely packed to tightly packed, in the latter case forming rare shell pavements (e.g. Fig. 3.16, section RC7, 44 m). Local patches of articulated brachiopods occur with the commissures usually at varying angles to bedding. Crinoids occur as patches of ossicles or columnnals up to 3 cm long. The distribution of other fossils is similarly patchy: well-preserved thickets of delicate fenestellid bryozoans, branching tabulate corals (?Cladopora) and stromatoporoids (Amphipora) occur parallel to bedding, with local
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Description</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Thickly bedded, massive mudstone to very fine sandstone. Variably bioturbated (<em>Zoophycos, Chondrites, Planolites</em>). Rich benthic fauna (brachiopods, corals, trilobites, crinoids, bryozoans, stromatoporoids).</td>
<td>Suspension sedimentation on a well-oxygenated outer shelf close to storm wave base.</td>
</tr>
<tr>
<td>B</td>
<td>Thin, graded beds of very fine to fine sandstone interbedded with green, grey or rarely black shales. Bioturbation variable but locally intense (<em>Planolites, Chondrites, Zoophycos, ?Palaeophycus</em>).</td>
<td>Middle to outer shelf setting with sandstone beds deposited by distal, storm-generated unidirectional flows alternating with fairweather suspension deposition of mudstone.</td>
</tr>
<tr>
<td>C</td>
<td>Thin, rarely graded beds of accretionary lapilli and silicic ashfall tuff interbedded with mudstone/shale.</td>
<td>Direct settling from suspension of pyroclastic airfalls through the water column to sub-storm wave base depths.</td>
</tr>
<tr>
<td>D</td>
<td>Medium to thick tabular beds of very fine- to medium-grained sandstone interbedded with variably bioturbated mudstones. Sandstone beds display HCS, parallel lamination, tabular cross beds, sole marks. Rich, thin shell beds dominated by disarticulated brachiopods are locally present.</td>
<td>Inner shelf deposition with sandstone beds generated by obliquely offshore-directed, combined flow storm currents alternating with fairweather suspension deposition of mudstone.</td>
</tr>
<tr>
<td>E</td>
<td>Thick, amalgamated beds of fine-grained, hummocky cross-stratified and parallel laminated sandstone; rare shale interbeds.</td>
<td>Storm-generated combined flow currents on the lower shoreface.</td>
</tr>
<tr>
<td>F</td>
<td>Thick bedded, fine- to medium-grained sandstone, with small- to medium-scale bimodal trough cross-bedding and parallel lamination. Rare HCS, shale interbeds.</td>
<td>Middle to upper shoreface dominated by onshore and longshore currents, generated by asymmetric shoaling waves and oblique wave approach.</td>
</tr>
<tr>
<td>G</td>
<td>Medium to thick bedded, fine- to medium-grained sandstone dominated by bimodal-bipolar tabular (lesser trough) cross bedding and parallel lamination. Reactivation surfaces, herringbone cross bedding, erosional surfaces, and mud-draped foresets are present.</td>
<td>Tide-dominated shoreface to ?tidal inlet complex dominated by southeast-directed flood tidal flows; lesser northwest ebb currents and rare south-directed longshore flows.</td>
</tr>
<tr>
<td>H</td>
<td>Thick, planar to wedge-shaped beds of parallel laminated sandstone. Rare tabular cross beds.</td>
<td>Beach-foreshore setting dominated by wave swash.</td>
</tr>
<tr>
<td>I</td>
<td>Rare lenses of fossiliferous limestone to barren dolomite, and beds of contorted limestone in a fine arenite matrix.</td>
<td>Small biostromal carbonate accumulations; occasionally smothered by storm-related influxes of sand, followed by post-depositional slumping.</td>
</tr>
<tr>
<td>J</td>
<td>Lenses of moderate to thick bedded, volcaniclastic, pebble-granule conglomerate; normally graded to coarse arenite or massive; rare clast imbrication and cross bedding. Massive to graded beds of coarse to pebbly volcaniclastic arenite in the upper parts of the Roxburgh Formation.</td>
<td>Sheetflows and grainflows on small, gravelly to sandy deltas composed of mainly volcaniclastic detritus, which prograded into the nearshore zone probably following eruptions.</td>
</tr>
</tbody>
</table>

Table 3.6 Summary of facies of the Roxburgh Formation.
Figure 3.14 QRF diagram (after Folk, 1974) for arenites of the Roxburgh Formation. Data from Pemberton (1980), Cook (1990), Millsteed (1992), and this study.

Figure 3.15 (a) Legend of symbols used in the measured sections (Fig. 3.16); (b) Location of measured sections (Fig. 3.16) along the Cudgegong Road and around Lake Windamere with simplified structural data for the sections (see Fig. 1c of Supporting Paper 4 for general location).
Figure 3.16 Vertical sections through portions of the Roxburgh Formation. Legend and location of sections is indicated in Fig. 3.15. Facies A to J and facies sequences are described in the text. Not all beds are shown in Facies B and D, only representative examples. Vertical scale is in metres.
Figure 3.16 (continued)
alignment of their long axes. These forms occasionally encrust disarticulated brachiopod valves, although they often appear to have colonised the sandy-silty substrate or an algal mat that was not preserved. Solitary cystimorph rugose corals, favositid and auloporid colonies locally occur in growth positions (Fig. 3.16, section RC5 10 m; Fig. 3.17a) or parallel to bedding. Trilobites (Apocalymene, Scutellum, and Crotalocephalus) are rarely preserved as complete individuals, or, more commonly, as well-preserved pygidia and cephalas.

**Interpretation and Palaeoecology:** The lithologies, textures, sedimentary structures, and Cruziana ichnofacies bioturbation (Pemberton et al. 1992) indicate predominantly low energy suspension sedimentation on a shelf that was generally below storm wave base. The variation in the intensity of bioturbation probably reflects fluctuating rapid and slow rates of supply of suspended sediment. The diverse fauna suggests a well-oxygenated shallow marine environment with abundant food, and the presence of delicate sessile benthic organisms such as corals and bryozoans indicates generally low levels of suspended sediment and very quiet water conditions. Taphonomic relations, however, indicate somewhat varied environmental conditions. Brachiopods that are disarticulated yet unabraded suggest long term accumulation in a low energy environment (Brett and Baird 1986). The predominantly concave-down orientations indicate some current activity, probably due to maximum storm wave impingement on the sea floor or distal storm-generated flows. However, areas of essentially random brachiopod orientations indicate intervals of very quiet conditions, as the concave-up arrangement is unstable, even in quite gentle currents (<10 cm/sec: Brett and Baird 1986). The dominance of pedicle valves has traditionally been used to indicate allochthonous faunas (Boucot 1981). However, recent studies (Brett and Baird 1986 and references therein) have shown that brachial valves are often destroyed preferentially by current activity, even in para-autochthonous faunas. The brachiopods locally formed shell pavements during times of low sedimentation; these shells were frequently colonised by encrusting sessile epifauna (corals, bryozoans). This created well-preserved, para-autochthonous fossil assemblages in distinct high-density beds due to successional buildup of epifauna (cf. taphonomic feedback of Kidwell and Jablonski 1983). The occurrence of articulated brachiopods, rare complete trilobites, and corals in growth positions suggests these assemblages were smothered by sediment influxes that were occasionally rapid and probably related to distal storm-generated flows. Local sporadically or evenly distributed fossils, presumably indicate a higher average, less episodic, supply of sediment, with patchy colonisation of the soft substrate.

The lack of evidence for significant transport of these fossil assemblages suggests some vestige of the original biocenotic communities may be preserved. The patchy distribution of most of the sessile organisms, together with evidence for rapid burial, suggest they reflect the original distribution of these organisms within the community, possibly as sub-communities (e.g. ?Cladopora and crinoid colonies, bryozoan thickets).
3.3.5.6.2 Facies B: Thinly Interbedded Sandstone and Mudstone/Shale

**Description:** Facies B consists of sharp-based, planar, fine- to very fine-grained sandstone beds (0.5 to 12 cm thick, av. 3.5 cm) interbedded with grey-green or rarely laminated black shale or bioturbated, fine sandy mudstone (Fig. 3.17b). Facies B is common, occurring in every measured section as intervals 0.5 to 4 m thick; however, it is best exposed in the fault-disrupted RC2 cutting (Fig. 3.15b). Sand/shale ratios in the measured sections vary from 0.25 to 2.3 and numerous small-scale (0.5 m to 3 m) thickening- and thinning-upwards trends in bed thickness occur. The sandstone beds display sharp, scoured bases, normal grading, parallel lamination and, more rarely, current ripple lamination. Zones of slumping were rarely recorded. The interbedded shales are weakly bioturbated in general with thin, intensely bioturbated intervals (often where sandstone beds are thin and widely spaced), leaving a homogenised sandy mudstone. Recognisable trace fossils are rare, but include Zoophycos, Chondrites, and Planolites. Body fossils are sparse with only infrequent brachiopods, corals, bryozoans and crinoids.

**Interpretation:** Facies B represents deposition of thin, graded sandstone beds from waning, storm-generated suspension currents, interbedded with fairweather suspension-deposited mudstone. The lack of wave-produced sedimentary structures suggests deposition below storm-wave base. The presence and intensity of *Cruziana* ichnofacies bioturbation suggests a shelf environment with generally well-oxygenated bottom conditions that could support an infauna which was locally capable of homogenising the sediment, although alternating zones of weak and intense burrowing again suggest fluctuating, but generally high, rates of suspended sediment supply (Pemberton *et al*. 1992). Rare dark, organic-rich, laminated shales may indicate locally dysaerobic conditions where even very thin storm beds (<1 cm) were preserved due to the lack of bioturbation. Palaeocurrent data from current ripple laminations indicate flows towards the southwest and west, whereas sole marks display bipolar northeast-southwest trends (Fig. 3.18).

Similar thin, normally graded beds transitional to proximal storm beds (see Facies D) have been widely recorded from ancient storm-dominated shelves, especially from the Mesozoic sequences of Canada (Hamblin and Walker 1979; Walker 1984) and the southwest U.S.A. (Swift *et al*. 1987), where they are typically interpreted to occur in middle to outer shelf settings. Such beds can only be reliably distinguished from deep water turbidites by their shallow marine body and trace fossils and their association with demonstrably shallow marine facies (Brenchley 1985). Absolute depths of deposition for Facies B are difficult to quantify. Comparison with analogous modern facies suggests depths greater than 20 m: Aigner and Reineck (1982) document thin distal storm beds (<5 cm thick) in depths of 20 m, 50 km offshore in the southeast North Sea, and Nelson (1982) found very similar 1-5 cm Bouma-like graded storm beds in 20-25 m of water, 150 km off the Yukon Delta in the northeast Bering Sea.
Figure 3.17 (a) Facies A; bioturbated green-brown mudstone with abundant disarticulated brachiopods, crinoid ossicles and branching tabulate corals (*Cladopora* sp.) with large solitary cystimorph corals in growth position. Hammer handle indicates bedding (section RC5, 9 m). (b) Facies B; bioturbated sandy mudstone interbedded with thin, graded beds of quartz sandstone displaying small-scale cross-lamination (cutting RC2; Fig. 3.15b for location). (c) Well developed, high angle (34° maximum dip) HCS in amalgamated hummocky cross-stratified sandstone of Facies E (north of Lake Windamere, GR 767600 6366700). (d) Facies D; massive to wavy laminated sandstone (with rare mudstone intraclasts) interbedded with bioturbated shale. Base of sandstone bed drapes ripples in the underlying shale, and the top is marked by asymmetric combined flow ripples (current from right to left) (section RC5, 21 m). (e) Facies D; small- to medium-scale HCS developed in a thick sandstone bed. (Oakey Creek, GR 766400 6362800). (f) Pavement of straight-crested symmetrical ripple marks of Facies G (RC10, 29 m). (g) Bedding plane exposure of well preserved, disarticulated, mainly concave-down brachiopods (*Isorthis, Mesodouvillina, and Howellella*) developed near the top of a Facies D storm bed (section RC10, 145 m).
Facies C: Accretionary Lapilli Tuff and Silicic Ashfall Tuff

Description: Thin (4 cm to 50 cm) beds of accretionary lapilli tuff occur in a 1 m to 2 m thick horizon at the base of the Roxburgh Formation (Cook 1990) and as isolated beds higher in the sequence, predominantly interbedded with Facies A or Facies B mudstones. The beds are composed of whole or fragmental, spherical to ellipsoidal accretionary lapilli, 0.2 cm to 1.7 cm in diameter supported in a matrix of ash; accretionary lapilli to ash matrix ratios commonly range from 0.5/1 to 1/1. (Fig. 6A of Supporting Paper 4). Beds are laterally persistent - over at least 2.5 km in one case - and display flat, non-erosive bases, sharp to bioturbated tops, and rare normal grading.

Scattered sporadically throughout the Facies A and B mudstones are numerous 2-25 cm thick, sharp-based beds of structureless or parallel-laminated volcaniclastic mudstone. These resemble the matrix of the accretionary lapilli beds in consisting of quartz, plagioclase, devitrified volcanic detritus (including probable glass shards) and clays.

Interpretation: Although accretionary lapilli may form in a number of volcanic settings (Cas and Wright 1987) the thickness and lateral persistence of these beds, together with the total dominance of accretionary lapilli and volcanic detritus, and the lack of sedimentary structures (apart from rare normal grading) suggests that Facies C represents direct fallout from the ash cloud of a major phreatic or phreatomagmatic eruption, followed by passive settling through shallow water to a sub-storm wave base depth. The preservation of accretionary lapilli in a water-settled fall deposit suggests the lapilli entered the water as hardened pellets, possibly due to bake hardening in the eruption column (Cas 1992). Furthermore, the above characteristics suggest the presence of an intermittently active centre of explosive silicic volcanism (Cas 1992), probably within 10's of kilometres of the depositional site (Moore and Peck 1962; Williams and McEirney 1979). Occasional increases in fragmentation of the lapilli suggest minor current reworking and resedimentation. Normal grading in similar deposits has been attributed to waning of the eruption and post-eruption settling of detritus (Ledbetter and Sparks 1979). Similarly, the volcaniclastic mudstone beds are interpreted to represent direct suspension settling of fine volcanic ash.

Alternatively, similar deposits can be formed by subaqueous gravity flows following slumping of subaerially-deposited pyroclastic flows, or through transformation of subaerial to subaqueous pyroclastic flows (Cas and Wright 1991); normal grading in such deposits would reflect waning flow velocities. However, such deposits might be expected to show some incorporation of intraclasts of basin floor sediments, which has not been observed in Facies C.

Facies D: Moderate to Thickly Interbedded Sandstone and Shale

Description: The most abundant facies of the Roxburgh Formation, Facies D occurs in every measured section as sandstone beds 6 cm to 70 cm thick (av. 17.3 cm) rhythmically interbedded with green-grey mudstone or shale, with sandstone/shale ratios commonly greater than 1:1. Facies D
intervals vary from 0.75 to 18 m thick (averaging 2 to 3 m) and the facies usually passes gradationally into Facies E and Facies B (Fig. 3.16). Interbedded mudstones are variably bioturbated (Planolites and Chondrites) and often rich in volcanioclastic debris. Sandstone bases are sharp and vary from planar through slightly undulating to locally scoured to depths of 5 cm (Fig. 3.17d). Sole marks are common and usually comprise groove and tool marks showing consistent bipolar northeast-southwest trends (Fig. 3.18); rare flute marks show a sense to the southwest.

Bed tops are sharp and commonly planar to hummocky. Some display well developed symmetric to slightly asymmetric ripple marks, typically straight-crested or bifurcating, with wavelengths of 6 to 21 cm; crests trend generally north-northeast-south-southwest (Fig. 3.18). Asymmetric combined flow ripples are less common and display southeast dipping foresets (Fig. 3.17d). Deep scours (to 15 cm) are common at the top of sandstone beds thicker than 40 cm. Internal structures show considerable variation and are summarised in Table 3.7.

Fossils in Facies D consist of either a graded basal lag (to 20 cm thick) in the sandstone beds composed of crinoid ossicles and abraded, disarticulated brachiopods, or well-preserved shell beds at the tops of sandstone beds (Fig. 3.17g). The latter exhibit brachiopod-dominated faunas similar to Facies A, with Isorthis, Mesodouvillina, Howellella, and indeterminate atrypids and rhynchonellids the dominant taxa. The faunas, however, show signs of higher current activity compared to Facies A: over 90% of shells are concave-down; fragmentation is more common, with well-preserved shells often occurring on a lag of intensely fragmented and winnowed shell material. Size sorting between horizons, lineation of beaks, and local imbrication of shells is common. Encrusting sessile epifauna vary from rare to abundant and typically comprise twiglike tabulate corals (Cladopora) and fenestellid bryozoans.

**Interpretation:** Rhythmically interstratified bioturbated shales and sharp-based sandstones displaying hummocky cross-stratification (HCS) reflect the alternation of slow fairweather mudstone deposition with storm-emplaced sands on the inner shelf between the fairweather and storm wave bases (Dott and Bourgeois 1982; Walker 1984; Brenchley 1985). The massive graded beds (Type 1 of Table 3.7; Fig. 3.17d) reflect rapid sedimentation from suspension fallout that prevented the development of tractional sedimentary structures. Common parallel-laminated to undulating laminated beds (Type 2 of Table 3.7) resemble those described by Arnott (1993) who attributed them to high energy combined flow conditions characterised by a long-period high-velocity oscillatory component and a weak to strong unidirectional component. Hummocky cross-stratified beds (Type 3 of Table 3.7; Fig. 3.17e) are similar to the idealised sequence (and its variants) proposed by Dott and Bourgeois (1982). HCS is considered to form under conditions of strong storm wave oscillatory flow with a superimposed unidirectional geostrophic current. Disagreement, however, exists as to the relative magnitudes of these currents (Nottvedt and Kreisa 1987; Duke et al. 1991), although the majority of recent experimental and field work favours either purely oscillatory flows or very oscillatory-dominant combined flows (Duke et al. 1991; DeCelles and Cavazza 1992).
<table>
<thead>
<tr>
<th>Type</th>
<th>Description</th>
<th>Lithology</th>
<th>Abundance</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Massive and commonly graded throughout with rare mudstone intraclasts towards base (Fig. 3.17d). Bases sharp and undulating (sole marks and load casts rarely developed); tops sharp, usually planar, and often bioturbated; less commonly, tops display form discordant symmetrical wave ripples or asymmetric combined flow ripples (Fig. 3.17d). Bed tops rarely hummocky ($\lambda = 0.5-1$ m) or wave scoured. Bed thickness typically $&gt;10$ cm.</td>
<td>Fine-grained quartz or sublithic sandstone</td>
<td>Common to abundant</td>
</tr>
<tr>
<td>2.</td>
<td>Similar to type 1 but with a thin basal lag of tightly packed crinoid ossicles and brachiopods, or mudstone intraclasts.</td>
<td>Fine- to medium-grained quartz or sublithic sandstone</td>
<td>Uncommon to rare</td>
</tr>
<tr>
<td>3.</td>
<td>Planar parallel laminated or with gently undulating lamination throughout, or passing upwards from a thin massive basal layer. Numerous internal mud drapes (to 1 cm thick) are common. Common sole marks on base; tops either planar or with well developed form discordant wave ripples. Variations on the 'classic' hummocky sequence of Dott and Bourgeois (1982) and Brenchley (1985). Basal graded interval is usually absent and lower parallel laminated zone varies from thick (transitional to 3.) or absent. HCS is generally low to moderate angle, $\lambda = 25$ cm to 1.8 m (Fig. 3.17e); sometimes comprises entire bed. Bed tops planar or with shallow wave scour, rare form discordant ripples.</td>
<td>Fine- to very fine-grained quartz to sublithic sandstone</td>
<td>Common</td>
</tr>
<tr>
<td>4.</td>
<td>Massive, sometimes graded or vaguely parallel laminated throughout, or with moderate-angle planar cross-beds. Bed bases sharp, frequently scoured; tops, sharp and planar to undulating. Thick (&lt; 50 cm) graded beds showing complete to nearly complete Bouma sequences (Tabcd or Tbcde). Shale drapes are common on parallel laminations and ripple foresets (Fig. 6D; Appendix 4). Climbing current ripples common in Tc.</td>
<td>Medium- to coarse-grained volcaniclastic sandstone</td>
<td>Rare to very rare</td>
</tr>
<tr>
<td>5.</td>
<td>Convolute to contorted bedding with dewatering structures, rarely capped by a zone of climbing current ripples.</td>
<td>Fine-grained, often calcareous sandstone</td>
<td>Rare</td>
</tr>
</tbody>
</table>

Table 3.7 Summary of Facies D sedimentary structures, faunas, and lithology.

Symmetrical ripples capping storm beds indicate wave propagation in a west-northwest to east-southeast direction (Fig. 3.18). Although storms and waves can approach the coast from different directions, oscillatory flow in nearshore, wave-dominated settings is often oriented perpendicular to shore, as storm waves approaching the coast will refract until the bottom paths of their wave orbitals are nearly perpendicular to shelf isobaths (Komar 1976). The orientation of wave-ripple crests on the inner shelf can provide a first approximation of local shoreline trend, especially on relatively straight
coasts with cuspatc deltas forming subdued headlands, typical of wave-dominated, sand-rich shorelines
(Leckie and Krystinik 1989). Ripple crest trends, together with the shoreline trends indicated by
regional palaeogeography and unit isopachs outlined earlier (Section 3.3.5.3), suggest an approximate­
ly north-northeast-south-southwest trending palaeoshoreline.

Infrequent asymmetric combined flow ripples on bed tops indicate eastward (ie. onshore) translation of the vortex ripples. Leckie and Krystinik (1989) found shoreline normal, offshore-directed orientations for similar asymmetric ripples and attributed this to waning combined flows. However, onshore orientations have been noted by Duke (1990) who attributed them to orbital asymmetry beneath shoaling storm waves, creating a slight net landward current.

As noted by workers elsewhere, classic HCS storm beds and their variants appear to be largely restricted to the fine to very fine sand fractions (Duke 1990 and references therein), and medium to coarse volcanioclastic beds (Type 5 of Table 3.7; Fig. 6D of Supporting Paper 4) display partial to complete Bouma sequences, suggesting deposition by waning, low density suspension currents (Lowe 1982). Frequent mud drapes in these beds indicate fluctuations in the waning current; similar features have been noted by Chakraborty and Bose (1992). Other coarse-grained beds dominated by high-angle tabular cross-beds (Type 6 of Table 3.7) indicate obliquely offshore migrating current ripples to small sandwaves. The presence of rare convolute-laminated beds with dewatering structures (Type 7 of Table 3.7) attests to rapid vertical expulsion of pore water through loosely compacted sands, indicating rapid deposition and high initial porosities.

3.3.5.6.5 Facies E: Amalgamated Hummocky Cross-stratified Sandstone

Description: Facies E is represented by units of very fine- to medium-grained sandstone up to 9 m thick that commonly occur at the top of coarsening-upward sequences or pass upwards into Facies F and G, and pass gradationally downwards into proximal storm beds of Facies D (Fig. 3.16). Rare, low aspect (40 cm deep by 4 m wide) channels are incised where the upper contact of the facies occurs with Facies B (e.g. Fig. 3.16, section RC7, 28 m level); these are lined by shale and filled by massive or parallel-laminated sandstone and trend on average to 340°-160°. The facies is unfossiliferous, apart from sparse crinoid ossicles, and bioturbation is restricted to rare, poorly preserved horizontal burrows (?Planolites). Beds are 0.3 to 2.1 m thick with upper and lower contacts that vary from planar to highly erosive. Zones of parallel lamination and massive sandstone pass laterally and vertically into HCS (Fig. 3.17c) and swaley cross-stratification (SCS). Morphology of the HCS and SCS varies greatly, yet falls into several distinct categories: (1) thin zones of low amplitude HCS transitional to wavy lamination (Fig. 3.16, section RC7, 35 m), (2) zones dominated entirely by SCS (Fig. 3.16, section RC3, 45-47 m) with 1 to 3 m wide swales displaying gently dipping (<20°) laminations; (3) high-angle HCS with dips approaching 35° and wavelengths less than 2 metres (Fig. 3.17c), and (4) Moderate- to large-scale, low-angle HCS with wavelengths of 2 to 5 metres mainly of the scour and drape variety (after
<table>
<thead>
<tr>
<th>Facies B</th>
<th>Current Ripple Foresets</th>
<th>Sole Marks (Tools, Grooves, Flutes)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N = 7</td>
<td>N = 9</td>
</tr>
<tr>
<td></td>
<td>Sd = 11.72</td>
<td>Sd = 8.56</td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 265^\circ$</td>
<td>$\bar{X} = 057^\circ - 237^\circ$</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies D</th>
<th>Wave Ripple Crests</th>
<th>Sole Marks (Flutes, Grooves)</th>
<th>Current Ripple Foresets</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N = 128</td>
<td>N = 46</td>
<td>N = 18</td>
</tr>
<tr>
<td></td>
<td>Sd = 25.65</td>
<td>Sd = 32.68</td>
<td>Sd = 32.11</td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 21^\circ - 201^\circ$</td>
<td>$\bar{X} = 068^\circ - 248^\circ$</td>
<td>$\bar{X} = 270^\circ$</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies G</th>
<th>Cross Beds (mainly tabular)</th>
<th>Wave Ripple Crests</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N = 42</td>
<td>N = 38</td>
</tr>
<tr>
<td></td>
<td>Sd = 15.15</td>
<td>Sd = 17.56</td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 289^\circ$</td>
<td>$\bar{X} = 016^\circ - 196^\circ$</td>
</tr>
<tr>
<td></td>
<td>N = 110</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sd = 36.21</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 137^\circ$</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies F</th>
<th>Trough Cross beds</th>
<th>Pebble Imbrication</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N = 19</td>
<td>N = 5</td>
</tr>
<tr>
<td></td>
<td>Sd = 24.84</td>
<td>Sd = 14.01</td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 213^\circ$</td>
<td>$\bar{X} = 310^\circ$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies J</th>
<th>Pebble Imbrication</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N = 5</td>
</tr>
<tr>
<td></td>
<td>Sd = 14.01</td>
</tr>
<tr>
<td></td>
<td>$\bar{X} = 113^\circ$</td>
</tr>
</tbody>
</table>

Figure 3.18 Summary of palaeocurrent data for the Roxburgh Formation. $N =$ number of readings, $Sd =$ standard deviation, $X =$ vector mean of readings or mode.
Brenchley 1985), where laminae thicken into the swales (Fig. 3.16, sections RC10, 107-110 m; RC3, 25 m). Zones of deformed bedding occasionally disrupt beds.

**Interpretation:** Amalgamation of sandstone beds containing parallel lamination and HCS has been widely recorded and is considered to represent frequent episodes of high energy storm deposition above storm-wave base (Dott and Bourgeois 1982), typically on the lower shoreface or offshore transition zone (Brenchley 1985), close to the fairweather wave base (commonly 5 to 15 m deep: Walker 1984). Facies E preserves little evidence of fairweather deposition (i.e. no ripple marks or angle-of-repose cross-bedding) and the paucity of bioturbation or fauna together with the rarity of mudstone indicate water depths at which storms of average intensity would erode the bottom deeply enough to destroy any evidence of everyday infaunal activity (Bourgeois 1980) and fairweather deposition. Zones dominated by SCS probably represent slightly shallower deposition, as SCS-dominant intervals are usually found above HCS-dominant intervals in progradational sequences and are considered to represent storm-dominated deposition on the upper shoreface (Tillman 1985). The rare channels are oriented oblique (approximately 40-50°) to the probable palaeoshoreline and may represent either fluvial channels incised during brief regressive intervals, or channels cut by strong, obliquely offshore-directed storm rip currents and filled by mainly suspension-deposited sandstone. The lack of cross-bedding appears to preclude a tidal origin for these channels.

### 3.3.5.6.6 Facies F: Trough Cross-bedded Sandstone

**Description:** Facies F is well exposed in sections RC4, RC5, and LW1 where it forms units up to 8 m thick of very fine- to medium-grained, well-sorted, unfossiliferous, quartzarenites to rare sublitharenites (Fig. 3.16). Rare shale horizons, lenses of mud chip breccia, and 1-3 cm thick, laterally persistent bands of coarse volcanic litharenite occur sporadically. Bed thicknesses vary from 0.6 to 1.2 m, and parallel lamination, very low-angle planar cross-stratification, and small- to medium-scale trough cross-beds (10 cm to 50 cm sets) are the dominant sedimentary structures (Fig. 3.19c). Palaeocurrent readings from the latter show a bimodal pattern with a major mode to the east and east-southeast, and a lesser mode to the south and southwest (Fig. 3.18). Rare bioturbation and symmetrical wave ripples also occur, and zones dominated by small-scale HCS and SCS occur throughout the facies.

**Interpretation:** The abundance of angle-of-repose cross stratification and parallel lamination, the paucity of mudstone, and the presence of rare bioturbation suggests a high energy marine environment above the fairweather wave base, probably on the middle to upper shoreface (Walker 1984). Facies F palaeocurrents record the migration of numerous small- to medium-scale lunate megaripples in onshore and lesser longshore and oblique offshore directions. The facies lacks any evidence of tidal activity, and the onshore trends are approximately normal to the general ripple crest trend (Fig. 3.18), suggesting the megaripples were driven landwards by asymmetric oscillatory wave surge generated by
Figure 3.19 (a) fine-grained, well sorted quartzarenite from RC6 (13 m) dominated by quartz grains with rare plagioclase grains (P) and sedimentary lithic fragments (S) with authigenic biotite and a quartz cement. Crossed polars. (b) fine-grained, quartz-cemented sublitharenite from RC10 (35 m). Note silicic volcanic grains (V). Crossed polars. (c) Facies F; small-to medium-scale trough cross bed sets developed in fine-grained sandstone. Hammer is 33 cm long (section RC4, 12 m). (d) Bimodal tabular cross-bed sets in Facies G sandstones. Note reactivation surface on the lower set (arrowed), the mud draped foresets on the upper set (arrowed), the strong erosional surface (beneath hammer) and the syn-depositional faulting on the lower cross-bed set (section RC10, 70 m). See Fig. 8 of Supporting Paper 4 for a sketch of the photograph. (e) Thick-bedded, parallel laminated sandstone (with rare small-scale cross-bedding); Facies H. Hammer (circled) for scale (section RC10, 124 m). (f) Facies I; thin, contorted beds of limestone, favositid corals,stromatoporoids and brachiopods in a quartz sandstone matrix (RC6, 22 m). (g) Facies J; volcaniclastic pebble conglomerate with clasts of porphyritic dacite (light) and minor basement lithologies (jasper, shale) in a fine sandstone matrix. Lens cap is 5.5 cm across. (north of Lake Windamere, GR 766300 6368500).
large breaking waves (Clifton et al. 1971). Under such conditions, the near-bottom oscillatory currents are modified, with the shoreward flow under the wave crest becoming stronger and of shorter duration than that under the wave trough, resulting in a net landward current (Komar 1976). Nummedal (1991) suggested that upwelling wind-induced currents or non-linear interactions between currents and wave-induced bed shear stress may also cause onshore transport. These conditions were occasionally replaced by southerly-directed longshore currents, probably generated by oblique wave approach from the north. Obliquely seaward-directed cross strata suggest the action of rare rip currents. The rare occurrence of HCS and SCS indicates oscillatory-dominant storm currents were occasionally prevalent over unidirectional flows (Duke et al. 1991) and horizons rich in mud chips indicate erosion of offshore muds by lowering of the wave base. The paucity of biotic activity indicates that prolonged low-energy fairweather conditions are largely absent from this facies, which probably accumulated rapidly during storm, or waning storm, conditions.

Similar deposits have been described from modern high-energy barred and non-barred near-shore settings, typically in water depths just seaward of the breaker zone (usually 2 to 4 m in fair-weather conditions) (Clifton et al. 1971; Clifton 1976; Hunter et al. 1979; Greenwood and Mittler 1985). Clifton et al. (1971), for example, document a lunate megaripple zone dominated by obliquely onshore facing lunate megaripples and lesser offshore directed megaripples from the high-energy non-barred coastline of Oregon. Short (1984) recorded similar facies from the inner nearshore regions of low-gradient, intermediate to dissipative beach profiles where waves break several hundred metres out from shore. Such profiles are typically present after storms or on consistently high energy beaches and are characterised by the slow shoreward migration of multiple bars during waning storm and post-storm conditions with associated rips and longshore currents. Theoretical calculations for the Californian coast by Clifton (1988), based on Stokes' second-order wave theory, suggest that lunate landward-facing bedforms can potentially form in depths from 15 m (yearly) to 45-50 m (1000 year storm), although the character of the substrate is likely to restrict this to shallower depths; Vincent (1986) regarded shoreward bedload transport due to wave asymmetry, as only important in water less than 10 m deep. Analogous facies are also documented in ancient deposits (e.g. Bourgeois 1980; Dupré 1984; DeCelles 1987; Martino and Curran 1990).

3.3.5.6.7 FACIES G: TABULAR CROSS-BEDDED SANDSTONE

Description: Well-sorted, fine- to coarse-grained quartzarenites of Facies G occur in units up to 8 m thick that are best exposed in sections RC10 and RC3 (Fig. 3.16) and in other exposures, dominantly towards the top of the Roxburgh Formation. Thin shale beds and horizons of mud chip breccia also occur and beds of coarse to pebbly volcanic litharenite (to 40 cm), often overlying minor erosional surfaces, become very common near the contact with the overlying Riversdale Volcanics. Dominant structures are parallel lamination, alternating with 10 to 40 cm thick sets of high- to moderate-angle
tabular, and lesser trough, cross-beds (Fig. 3.19d). Overall, these show a bimodal and slightly oblique bipolar pattern with a minor mode towards the south (Fig. 3.18). Locally, however, trends can vary from stacked unimodal cosets trending to the dominant southeast mode, to alternating southeast and northwest trending sets either separated by erosional surfaces (Fig. 3.19d) or rarely arranged in a herringbone pattern. East to southeast palaeocurrents in section RC10 show a consistent shift towards the south proceeding up-section. Zones of unimodal current ripple laminated (2-4 cm thick sets) sandstone up to 3 m thick occur, trending to the dominant southeast mode (Fig. 3.16, section RC10, 88-91 m). There are also rare instances of medium-scale (0.5-0.75 m thick) low angle cross-stratification that may represent lateral accretion surfaces. Mud drapes and reactivation surfaces are occasionally visible on foresets (Fig. 3.19d) and rare instances of sigmoidal cross-bedding were noted in the southeast-trending sets. Symmetrical to slightly asymmetrical ripple marks are locally well preserved (Fig. 3.17f), and display approximately north-northeast-south-southwest crest trends (Fig. 3.18). Pronounced erosional surfaces are common, often marking contacts with Facies E; these are planar to gently undulating or highly erosive and hummocky with relief up to 70 cm. Some erosional surfaces exhibit a low aspect channel morphology, typically >5 m wide and < 1 m deep. Fossils are rare and occur in poorly sorted, laterally persistent layers that mark parallel laminae or the foresets of cross-beds. They comprise highly abraded brachiopods (*Isorthis, Howellella, and atrypids*), finely comminuted crinoids, and rare tabulate corals. Rare *Skolithis* bioturbation occurs, typically near bed tops.

**Interpretation:** A marine setting above fairweather wave base is suggested by the fauna, *Skolithos*-type bioturbation, and the abundance of angle-of-repose cross stratification. The setting was characterised by the migration of mainly straight-crested bars and lesser lunate dunes. The palaeocurrents indicate alternating unidirectional currents approximately perpendicular to the ancient shoreline, the dominant southeast mode being onshore and the subordinate northwest mode offshore. These bimodal, slightly oblique bipolar palaeocurrents may be attributable to either alternating episodes of onshore transport due to asymmetric shoaling waves and offshore transport by rip currents (Clifton *et al.* 1971), or reversing tidal currents (Visser 1980). Sporadic evidence for tidal flows (rare mud drapes, herringbone cross-beds, sigmoidal cross-beds and reactivation surfaces) suggests the latter is more likely, with the dominant southeast mode representing flood tidal bedforms and the northwest mode representing ebb tidal structures. Minor southerly flows represent either longshore drift due to oblique wave approach, or shore-parallel tidal currents; the similarity to the direction of longshore currents of Facies F suggests the former. Local reactivation surfaces and mud draped foresets indicate episodic bedform migration in a time-velocity asymmetric tidal current (Visser 1980), although definite bundling of foreset thickness was not observed; in contrast, rare herringbone cross-lamination suggests the occasional influence of ebb and flood tidal currents of approximately equal magnitude. The presence of frequent basal erosional surfaces and rare evidence of a channelised base for the facies, together with the development of lag conglomerates and rare lateral accretion surfaces, suggests tidal flows may have
been channelized, possibly in a flood-dominated to mixed dominance tidal inlet complex in a barrier island setting. The facies bears some resemblance to the microtidal inlet and delta deposits of Israel et al. (1987), and the flood-dominated to mixed influence tidal inlet complexes described by Cheel and Leckie (1990) which also developed on a wave-dominated coast, similarly passing downwards, below an erosional contact, into storm-dominated lower shoreface deposits. The various subenvironments of a tidal inlet setting (inlet, delta, channel fill) could not be delineated with confidence in Facies G exposures. The abundance of symmetrical wave ripples, parallel laminated zones (possibly representing swash bars) and possible HCS indicate the frequent presence of ocean waves in the setting.

3.3.5.6.1 Facies H: Parallel-laminated Sandstone

**Description:** Facies H comprises unfossiliferous, rarely bioturbated, fine- to medium-grained sandstone units up to 4.5 m thick that occur at the top of coarsening-upward sequences (Fig. 3.16, section RC10, 125 m level) close to the top of the Roxburgh Formation. The facies grades downwards into cross-bedded sandstones of Facies G and is rarely interbedded with Facies J conglomerate lenses (Fig. 3.20). Dominant sedimentary structures are parallel to very low-angle (3° to 5°) laminations in tabular to wedge-shaped beds up to 50 cm thick (Fig. 3.19e). Heavy mineral laminae are locally developed and rare parting current lineation (with a general east-west trend) is visible on some laminae. Low-angle tabular cross-beds rarely occur and have identical bi-directional orientations to those of the underlying Facies G.

**Interpretation:** Facies H represents mainly upper flat bed conditions and lacks associated HCS or SCS which would suggest a lower shoreface setting. The abundant planar laminations are therefore interpreted as representing wave swash in a relatively flat beach foreshore zone. Parting lineations suggest general east-west directions of wave swash and backwash, whereas rare cross beds indicate interaction with bimodal tidal currents, similar to those of Facies G.

3.3.5.6.8 Facies I: Limestone and 'Limestone/sandstone breccia'

**Description:** Rare, poorly exposed bodies of massive unfossiliferous dolomitic limestone crop out in the upper parts of the Roxburgh Formation on the north side of Lake Windamere (Millstead 1992) (GR 766050 6369940). Another probable occurrence includes several 2 to 3 m thick lenses of sparsely fossiliferous limestone in volcanlastic sandstone and conglomerate of the upper 50 m transition zone (see Facies J) of the Roxburgh Formation just north of Lake Windamere (GR 767250 6366230); fossils present here include articulated brachiopods, favositid corals, stromatoporoids (*Amphipora*), and rare rugose corals. A possibly related facies occurs in section RC6 (Fig. 3.16, section RC6, 22 m and 25 m) and in several other outcrops where 1 to 2 metre thick horizons of limestone 'breccia' with a fine quartzarenite matrix occur in association with Facies E amalgamated HCS sandstones. The limestone fragments are from <1 cm to 45 cm in diameter and well-rounded to highly
contorted in shape (Fig. 3.19f). The limestone varies from unfossiliferous sparite (with small lenses of gypsum) to tightly packed biosparite with abundant stromatoporoids (*Amphipora*), favositid corals, crinoid stems (to 4 cm long), and well preserved, mostly articulated brachiopods (mainly low-diversity assemblages dominated by *Isorthis*, and rhynchonellids). Similar, well-preserved fossils are also common outside the limestone clasts.

**Interpretation:** The limestones lack mound or reef-like characteristics and are interpreted as small biostromal carbonate accumulations that developed in areas sheltered from clastic supply and storm waves, possibly in small embayments. Faunas and limestone textures suggest a variety of conditions during deposition, ranging from supratidal flats with evaporites, through restricted shallow marine with low-diversity faunas, to open subtidal conditions with more diverse faunas. The 'breccia' variety of the facies represents rapid smothering of semi-consolidated limestone and well-preserved, para-autochthonous fossils by thick, storm-deposited fine- to medium-grained sand, possibly accompanied by some storm-related erosion of limestone. This was followed, in some instances, by varying degrees of post-depositional liquefaction and flowage, as indicated by the highly contorted limestone fragments. Slumping and contortion of laminae is common in thick, amalgamated storm beds (Dott and Bourgeois 1982), although the association with smothered carbonates is apparently rare.

### 3.5.6.9 Facies J: Volcaniclastic Conglomerate and Sandstone

**Description:** Several lenses of volcaniclastic conglomerate and sandstone, up to 650 m long and 55 m thick (Fig. 3.20), crop out in the middle parts of the Roxburgh Formation (Pemberton 1980a; Cook 1990; Millsteed 1992; Colquhoun, unpublished data). These lenses typically occur within, and pass laterally into, fossiliferous parallel-laminated and cross-bedded sandstone of Facies G and Facies H. However, conglomerate lenses in the Oakey Creek area sharply overlie interbedded sandstone and mudstone of Facies D and Facies B. The conglomerate comprises moderate- to poorly-sorted, subangular to well-rounded clasts, typically of pebble-granule dimensions (max. 25 cm) (Fig. 3.19g). Clast-supported fabrics dominate and the matrix consists of poorly-sorted, medium to coarse volcanic litharenite containing angular to subangular grains of volcanic quartz, plagioclase, and dacitic rock fragments in a matrix of chert, chlorite, hematite, and clays. Clast types include: porphyritic to aphanitic dacite, with eutaxitic textures and flow layering (85-90%); immature volcanic litharenite (5-10%); and rare sedimentary basement lithologies (jasper, mudstone, quartz-lithic sandstone) (Fig. 3.19g). Conglomerate beds vary from 10 cm to 3.5 m thick (av. 80 cm to 90 cm) with erosive to planar bases, sharp tops and sheet-like, rarely lensoidal, geometries (Fig. 3.20). Internally, the beds can be massive and unorganised or horizontally stratified throughout but they more commonly grade upwards into massive or parallel-laminated (rarely cross-bedded) coarse volcanic litharenite, similar in composition to the matrix. The stratified beds rarely show clast imbrication (long axis parallel to current) and are occasionally capped by unfossiliferous, siliceous shales. Interbedded with the conglomerate-volcanic
litharenite are horizons up to 8 m thick composed of fossiliferous, parallel-laminated and cross-bedded quartzarenite identical to Facies G and Facies H (Fig. 3.20). These display shallow pebble-filled scours and laterally persistent, 1 to 5 cm thick beds of granule conglomerate. Conglomerate beds underlying these horizons are thinner, dominated by granules, and often matrix-supported and normally graded.

In the upper 20 to 40 m of the Roxburgh Formation at Oakey Creek (GR 766150 6363210), thick (max. 1.5 m), massive or graded beds of coarse to pebbly dacitic sandstone with rare interbeds of unbioturbated shale occur. These pass gradationally upwards into dacitic ignimbrites, airfalls, and volcaniclastics of the Riversdale Volcanics. This facies is absent, however, where the contact is locally disconformable.

**Interpretation:** The sedimentary characteristics and lensoidal shape of this facies, passing laterally into foreshore and shoreface facies, suggests Facies J represents the deposits of small, gravelly to sandy deltas which prograded into the nearshore zone. The normally graded, erosively-based conglomerates probably represent the stacked deposits of heavily sediment-laden sheetflows generated by waning ephemeral currents, such as unchannelised flash floods (Nemec and Steel 1984; Flint and Turner 1988). Sparse clast imbrication readings show a wide scatter but indicate general currents to the northwest (ie. obliquely offshore) (Fig. 3.18). The presence of authigenic hematite in the matrix and the absence of sedimentary structures indicative of a subaqueous setting suggest deposition mainly on the subaerial part of the delta. However, the thinner graded granule conglomerate beds interbedded with shoreface sandstones may represent subaqueous sheetflows or turbulent mass flows typical of delta front facies (Nemec and Steel 1984). The predominance of juvenile volcanic detritus in the conglomerate suggests rapid resedimentation of unconsolidated pyroclastic and epiclastic detritus following a dacitic pyroclastic eruption. The repetitive interbedding of the conglomerate with quartz-rich shoreface and foreshore sandstones suggests numerous episodes of progradation of the alluvial part of the deltas, separated by intervals when the deltas were transgressed and reworked by the shallow marine facies. This may reflect pulses of progradation associated with periods of uplift and active volcanism in the source terrain followed by subsidence and transgression during inactive periods; alternatively, they may reflect high frequency eustatically-related movements of the strandline.

The unfossiliferous volcanic litharenites of the upper transition zone reflect the onset of silicic volcanism more proximal to the shoreline. Large volumes of immature volcanic detritus were rapidly eroded from the incipient volcanoes and delivered to swiftly prograding deltas and shorelines as high density sandy to gravelly sheetfloods or grainflows (Lowe 1982). Alternatively, similar thick, massive to graded volcaniclastic beds can be created by ignimbrites which entered shallow water and transformed into water-supported mass flows of pyroclastic detritus (Cas and Wright 1987). With continued regression these transitional rocks pass upwards into genuine welded ignimbrites, airfall tuffs, and volcanic litharenite of the Riversdale Volcanics, deposited in generally subaerial conditions.
Figure 3.20 Measured section through a Facies J volcaniclastic conglomerate lens north of Lake Windamere (GR 766300 6368500).
3.3.5.7 Facies Sequences

Overall, the Roxburgh Formation passes upwards from a basal contact with the muddy shelf deposits of the Yellowmans Creek Formation into the overlying, largely subaerial, Riversdale Volcanics, suggesting a regressive or upward-shallowing trend. This regression is discussed further, along with the other sea-level movements of the Kandos Group, in Section 5.2.

Superimposed upon this general shallowing-upward trend are dozens of asymmetric coarsening- and shallowing-upward sequences which vary from 3.5 to 14.4 metres thick (av. 7.4 m) and are separated by non-depositional disconformities or, less commonly, erosional surfaces (Fig. 3.16). The facies sequences of these cycles are usually partial shelf progradational units, typically commencing with the deeper shelf facies (Facies A or B) and shallowing upward and ending in one of the nearshore facies (Facies E, F, G, or H) (Fig. 3.16). The characteristics of these cycles fit the description of the regressive, high sediment supply parasequences of Mitchum and Van Wagoner (1991) and Swift et al. (1991) which have been documented widely in siliciclastic sequences. Such cycles indicate repeated episodes of rapid deepening and landward shifts of the strandline followed by slow progradation of facies under static or slowly falling relative sea level. As the continuity of these cycles both within and outside the depositional basin is unknown, it is not possible to determine the driving mechanism for these parasequences. Mechanisms proposed for similar cycles include: (a) repeated episodes of delta lobe progradation, abandonment, and rapid compaction of prodelta muds; (b) repeated episodes of rapid tectonic subsidence followed by progradation; and (c) high frequency oscillations of absolute sea level (Swift et al. 1991; Mitchum and Van Wagoner 1991).

Although much of the Roxburgh Formation appears to be readily divisible into metre scale coarsening-upward cycles, in some areas this cyclicity breaks down. For example, in sections RC4 (16-26 m), RC7 (0-18 m), and RC8 (Fig. 3.16), thick sections of Facies E, D and B show numerous small-scale thickening and thinning trends, yet lack consistent overall shallowing or deepening trends. Similarly, thick sections of Facies A mudstones occur in section RC7 which also lack any indication of consistent shallowing or deepening.

3.3.5.8 Depositional Environment, Palaeogeography, and Shelf Dynamics

Integration of the facies variation, faunal data, palaeocurrents, petrography and unit isopachs suggests the depositional setting for the Roxburgh Formation was a west-sloping open marine shelf flanked to the east by a moderate relief source terrain which was not preserved, due to latter erosion composed of sedimentary basement rocks, and active silicic volcanoes (Fig. 3.21). The silicic volcanoes generally remained distal to the shallow marine environments, contributing only sporadic airfalls and influxes of volcanioclastic sediment, until late in the history of the Roxburgh Formation when more voluminous influxes of volcanic detritus choked the shoreline and coastal plain systems, heralding the onset of major subaerial dacitic pyroclastic eruptions of the overlying Riversdale Volcanics. Sediment
input into nearshore areas remained fairly high throughout the history of the Roxburgh Formation and
sedimentation was therefore almost entirely clastic, apart from rare biostromal limestone accumulations
in sheltered nearshore areas (Facies I). Unit isopachs, regional palaeogeography and nearshore symmet­
rical ripple marks (Facies D and G) suggest average strandline trends of north-northeast-south-
west-southwest, although this probably varied slightly throughout the history of the unit. Palinspastic resto­
ration of the Capertee High suggests a maximum shelf width of 30 to 40 km before slope facies of the
eastern Hill End Trough commenced.

Ripple crest and asymmetry trends over the inner and middle shelf (Facies D) indicate major
storm waves approached from the northwest, probably shoaling slightly obliquely to the coast and
setting up a dominant south-directed longshore flow (Fig. 3.21). During post-storm conditions, asym­
metric shoaling waves on the shoreface drove lunate megaripples (3D dunes) east-southeastward,
towards the shoreline, balanced by oblique offshore rip currents and longshore currents (Facies F).
Unidirectional and combined flow currents were directed offshore to the west and southwest over the
shelf and shoreface; as these currents waned, thick to thin storm beds were rapidly deposited from
suspension or moulded into HCS or continued on to below the storm wave base. Similar offshore-
directed palaeocurrents have been widely documented in ancient shelf sequences (Leckie and Krystinik
1989 and references therein). Various mechanisms have been invoked to account for this current
including storm-surge ebb currents (Brenchley et al. 1979; Mount 1982) and shelf turbidity currents
created by substrate liquefaction due to storm wave loading (Walker 1984); however, neither of these
mechanisms has been convincingly observed in modern oceans, although shelf turbidites remain a
theoretical possibility, especially on relatively high gradient shelves (Nummedal 1991). Present unifor­
mitarian interpretations, based on modern oceanographic studies, favour wind-driven currents caused
by coastal set-up, whereby a buildup of water at the coast, due to onshore-directed winds and low
offshore barometric pressures, creates an offshore-decreasing hydrostatic pressure gradient which is
compensated by downwelling to seaward of the surf zone and the development of offshore-directed
bottom return currents (Duke 1990; Duke et al. 1991; Nummedal 1991). These waning currents have
been observed to deflect steadily, due to Coriolis Force, until they are approximately parallel to the
shelf isobaths in modern distal shelf areas, thereby creating a longshore current on the outer shelf
(Leckie and Krystinik 1989; Duke 1990). This geostrophic veering may be apparent in the Roxburgh
Formation as all shelf palaeocurrent indicators are deflected significantly to the left of shore-normal
orientations, which is the expected deflection for the southern hemisphere (Fig. 3.18). Duke (1990) and
Duke et al. (1991), however, caution that sole marking and partings may form in response to peak,
near-bottom instantaneous flow conditions associated with wave orbital motion, rather than reflecting
steady current flow. Nevertheless, rare current cross laminations in Facies B (and Facies D) show simi­
lar orientations, albeit with a high variance due to the low number of readings. Distal currents (ie.
Facies B) should deflect to nearly parallel to the shelf isobaths under the geostrophic model; however,
Figure 3.21 Block diagram summarising the depositional environments, processes and palaeogeography of the Roxburgh Formation. (N.B.: not all processes or facies were coeval; see text for details).

Figure 3.22 Probable Lochkovian storm tracks in southeastern Australia superimposed on a Lochkovian palaeogeographic and tectonic reconstruction of northeast Gondwana-land (Australia and Antarctica). Palaeolatitudes from Scotese and McKerrow (1990) (full lines) and Li et al. (1990) (broken lines), tectonic elements and palaeogeography from Powell (1984, Fig. 217) and storm tracks based on Marsaglia and Klein (1983).
Facies B still shows mainly oblique offshore currents. This may be due to the low palaeolatitude (12-15°; see below) which would create a low value of Coriolis Force, possibly not strong enough to cause consistently complete deflection of even weak distal currents (Duke 1990). Relatively shallow, friction-dominated shelf conditions may also contribute to the suppression of geostrophic veering (Duke 1990).

Facies G records the local influence of tides on the shoreface, possibly in tidal inlet complexes (Fig. 3.21). The preservation of tide-dominated deposits mainly in the upper parts of the otherwise storm- and wave-dominated Roxburgh Formation may be explained by: (a) a lessening of wave energy; (b) a reduction in the intensity of southerly longshore currents; (c) amplification of tidal processes by a particular morphology of the depositional basin during this time. At other times, the wave-dominated Facies F appears to have occupied the same position on the middle to upper shoreface (Fig. 3.16).

3.3.5.9 Storm type and Regional Palaeogeography

Australia occupied low palaeolatitudes during the Early Devonian; the reconstruction of Scotese and McKerrow (1990) indicated a palaeolatitude of 12° to 15° S for the Capertee High during the Lochkovian (Fig. 3.22). However, Li et al. (1990), using only palaeomagnetic data from Australia, indicated a higher palaeolatitude during the Lochkovian, in the vicinity of 33°S (Fig. 3.22).

In modern seas, belts dominated by hurricanes and summer tropical storms occur between 5° and 45° latitude, commonly on the western edge of major oceans, whereas winter storms are common at latitudes above 25° (Marsaglia and Klein 1983; Duke 1985; Barron 1989). Barron (1989), however, cautioned that these belts may have shifted during warmer and cooler periods of geological time and may be influenced by the local palaeogeography and different palaeocontinent configurations. The palaeolatitudes of Scotese and McKerrow (1990) indicate that hurricanes were probably the major storm type recorded in the Roxburgh Formation, providing the Hill End Trough was open at its northern end for a sufficient distance to allow the generation of large storm waves; Mesozoic cover rocks of the Great Australian Basin prevent this from being observed directly (Fig. 2.2). Inferred tropical storm tracks for the Early Devonian (Marsaglia and Klein 1983) would probably have followed similar paths to those encountered by the present-day northern Australia and Queensland coast (i.e. they would commence near the equator and assume anticlockwise curving paths as they proceeded into higher latitudes). This indicates that the Capertee High area may have been exposed to frequent hurricane generated storm-waves approaching mainly from the north and northwest (Fig. 3.22) as they swept down the Hill End Trough, refracting slightly as they crossed the shelf areas of the High. This is in broad agreement with the palaeocurrent data and local palaeogeography previously outlined.

It is possible, given the higher palaeolatitudes of Li et al. (1990), that winter storms emanating from low pressure systems around 40°S and approaching from the south and southwest, may also have contributed. However, fetch distances in this direction were probably not high, due to large areas of
Figure 3.23. Conodont fauna from a limestone horizon (Facies I) in section RC6 of the Roxburgh Formation (GR 766570 6365120).

Ozarkodina cf. remscheidensis remscheidensis ZIEGLER, 1960
a. Pb element, AMF 101541; RC6, x43.
b. Pa element (broken), AMF 101542; RC6, x78.

Pandorinellina exigua philipi KLAPPER, 1969
2. Pa element, lateral view of AMF 96382; RC6, x48.
3. Pa element, upper view of AMF 96382; RC6, x52.
land in the southeastern Lachlan Fold Belt and Australia's attachment to Antarctica at this time (Fig. 3.22).

3.3.5.10 Fauna and Age

Fine sandstones to mudstones in the unit contain a diverse Early Devonian shallow marine fauna of brachiopods (Cyclina, 'Dolerorthis', Mesodouvillina, Isorthis, Howellella, Salopina, Eoschu-chertella, Gypidula, Spinatrypa, Iridistrophia and rhynchonellids indet.), trilobites (Apocalyymene, Scutellum, and Croculocephalus), branching tabulate corals (?Cladopora), favositids and auloporids, and stromatoporoids (Amphipora) (Sections 3.3.5.6.1 and 3.3.5.6.4). Several samples of rare limestones in the Roxburgh Formation were processed for conodonts. Of these, sample C150 (GR 766050 6369796) was barren, and sample C79 (GR 767120 6366100) contained only rare Panderodus sp. One limestone block from a road cutting on the Cudgegong Road (sample RC6, GR 766570 6365120) yielded a small fauna including Pandorinellina exigua philipi, Ozarkodina cf. remscheidensis remscheidensis and several ozarkodinan Pb and Sb elements (Colquhoun 1995b, Supporting Paper 5) (Fig. 3.23).

Pandorinellina exigua philipi first appears in the pesavis Zone in eastern Australia (Bischoff and Argent 1990; Wilson 1989; Sorentino 1989) and O. r. remscheidensis ranges from high in the Pridoli up to at least to the sulcatus Zone (Klapper and Johnson 1980; Wilson 1989; Sorentino 1989); this suggests a pesavis-sulcatus (late Lochkovian-early Pragian) age for the fauna. The Roxburgh Formation therefore ranges from the delta Zone at its base (Section 3.3.4.5) to either the pesavis or sulcatus Zone.

3.3.6 RIVERSDALE VOLCANICS Dav (Brunker and Rose 1969 following Wright 1966)

3.3.6.1 Synonymy and Definition

Extensive Upper Devonian acid tuffs and intrusive quartz porphyry and felsite were first noted several kilometres to the east of Cudgegong by Game (1934). Lavers (1960) included approximately 230 m of acid tuffs and volcaniclastic sandstone from the same area in his Devonian Wells Formation (Fig. 3.1a). The name Riversdale Rhyolite (derived from the now abandoned 'Riversdale' property) was assigned to these rocks by Wright (1966), who considered them Early Devonian in age and probably equivalent to similar lithologies in the Capertee Valley. Brunker and Rose (1969) and Offenberg et al. (1971) modified this name to Riversdale Volcanics and opted for a broad Siluro-Devonian age – as did Brown (1974) and Talent et al. (1975) (Fig. 3.1a). Pemberton (1977, 1980a) established an Early Devonian age for the unit on stratigraphic grounds and subsequent studies (Millsteed 1985; Cook 1988a; this study) have both confirmed and refined this age, and considerably expanded the known areal extent.
No type section was originally defined for the unit. Pemberton et al. (1994) suggested a representative section between GR 770900 6360700 and GR 769000 6360300.

3.3.6.2 Distribution, Areal Extent and Thickness

The Riversdale Volcanics crop out prominently between Rylstone and Cudgegong or Ilford. To the north of Lake Windamere around GR 767300 6369300, the unit is present as a 2 km long, 500 m wide strip which is faulted against the Late Devonian sequence along its western side (Fig. 3). Further south, the unit is present as a thin, persistent reworked horizon to the north of the 'Lucknow' property (Fig. 1). The main body of outcrop of the unit, however, extends 12 km as a discontinuous, southward-widening belt, from the Cudgegong Road near GR 768000 6365000 in the north, through complexly folded and faulted sequences in the "Glendale" area (Cook 1988a) and around Oakey Creek, to the extensive outcrops in the rugged Clandulla State Forest, in the upper part of the Carwell Creek valley (Figs 2a and 3). Further south near the "Ophir" property (GR 774900 6350750), the Riversdale Volcanics is present as an elongate inlier surrounded by basal Permian strata.

Along Hurlstone Creek (GR 771450 6359400 to GR 769750 6358100), the unit attains a thickness of at least 1100 m (Fig. 3.24a); however, bedding readings in this area are fairly sparse. To the north of here, just south of "Glendale", Cook (1988a) recorded a maximum thickness of 890 m (1250 m composite thickness) (Fig. 3.24c). The basal and upper contacts are not exposed in the extensive outcrops in Clandulla State Forest, but available bedding data suggest at least 1000 m of the unit is present. Further north and west, the unit is markedly thinner, attaining approximately 200 m around the nose of the syncline southwest of "Glendale" (Fig. 3.3), yet thinning on the western limb of the syncline south of "Glendale" (GR 767400 6362500) to only 70 m thick, and, further to the south, to only 30 m on the limbs of an anticline around GR 765900 6360600 (Pemberton 1980a). Outcrops to the north of Lake Windamere at GR 767300 6369300 and GR 767500 6365000 have faulted contacts and insufficient bedding data from which to infer thickness. The thin reworked horizon north of "Lucknow" attains a maximum thickness of 22 m at GR 770700 6364200 before lensing out to the east and terminating against the Riversdale Fault in the north.

In summary, the Riversdale Volcanics thin markedly to the north, west, and especially to the east, away from a thick southern portion centred in Clandulla State Forest, west of Clandulla (Fig. 3.3).

3.3.6.3 Structure and Stratigraphic Relationships

Bedding is often difficult to determine in the ignimbrites of the Riversdale Volcanics and is restricted to rare pumice lenticle foliations and columnar jointing. Between "The Basin" (GR 772600 6355800) and "Glendale", the unit has a consistent northwest-southeast strike and a moderate westerly dip; at the "The Basin" a large syncline is present within the unit (Figs 2a, 3). To the southwest of
Figure 3.24 (a) section through the Riversdale Volcanics along Hurlstone Creek between GR 771450 6359400 and GR 769750 6358100; (b) section through the Riversdale Volcanics adjacent to Carwell Creek (GR 7707050 6364200 to GR 7707100 6364200); (c) composite stratigraphic section through the Riversdale Volcanics in the 'Glendale' area (from Cook 1988a).
"Glendale", the unit crops out on the limbs of a number of open, gently south-plunging folds (Fig. 3). The exposures north of " Lucknow" are folded, along with contiguous Early Devonian units, into a series of open, gently north-plunging folds (Fig. 1). Elsewhere, bedding data are often scarce and determining structure or younging directions is frequently impossible.

The Riversdale Volcanics generally overlie the Roxburgh Formation, although the nature of the contact varies throughout the district. An excellent disconformable contact is exposed in the north, beside Carwell Creek (GR 770700 6364200) (Fig. 3.25). The contact here is clearly erosional and gently undulating with local areas of sharp relief of up to 1 m. The average of 16 measurements of the surface, restored to horizontal on a stereonet, indicates a palaeoslope of 5°-7° towards 315° - providing there was no tectonic tilting during the brief hiatus between the units. To the immediate west, Cook (1988a) recorded both disconformable and conformable contacts. South of GR 771420 6359000, the contact is clearly disconformable and cuts across the strike of the Roxburgh Formation so that south of GR 772500 6357700 the Riversdale Volcanics disconformably overlies the Yellowmans Creek Formation (Fig. 2a). Further west, Pemberton (1977, 1980a) recorded a sharp, apparently conformable contact, and exposures of the contact along Oakey Creek and in roadcuttings on the Cudgegong Road appear to be conformable (Section 3.3.5.6.9). The contact is also well exposed around Tin Hat Creek (GR 770280 6362280) where there is a conformable and gradational transition from the Roxburgh Formation to the Riversdale Volcanics. The variable relationship between the units may be due to a combination of the following factors: (a) the probably short hiatus between the units permitted only minor erosion prior to volcanism commencing; (b) the shelf, during deposition of the Roxburgh Formation, sloped to the west, indicating areas in the east would have been exposed longer to subaerial erosion during regression, especially if regression was relatively slow and roughly kept pace with subsidence on the outer shelf, further west (cf. Type 2 unconformity of Vail et al. 1984); and (c) the base of the Riversdale Volcanics is grossly diachronous (e.g. the reworked horizon which overlies the Roxburgh Formation east of Carwell Creek, is underlain by thick volcanic sequences further west).

North of "Lucknow" and east of GR 771200 6364300, the Roxburgh Formation lenses out and the reworked horizon disconformably overlies the Yellowmans Creek Formation (Fig. 3.3). Further east, the Yellowmans Creek Formation also lenses out and the reworked Riversdale Volcanics disconformably overlie the Clandulla Limestone and cuts across bedding in the limestone at a low angle towards the north (Figs 1 and 3.3).

The upper reworked horizon passes conformably and gradationally into the micritic limestone or crinoid-bearing sandstone of the Carwell Creek Formation (best exposed at GR 770700 6364200 and GR 769500 6364800). A similar gradational, reworked contact was recorded by Rogis (1974), Brown (1974), Millsteed (1985), and Pemberton (1977). Cook (1988a) documented the contact as sharp, gradational or locally faulted.
Figure 3.25. Sketch (a) and photograph (b) of the disconformable contact between well-bedded, parallel-laminated, quartz sandstone of the Roxburgh Formation and massive dacitic boulder conglomerate of the Riversdale Volcanics. Outcrop is close to Carwell Creek at GR 770720 6364320.
The Riversdale Volcanics is generally overlain by the Permian Shoalhaven Group in the Clандulla State Forest area. Many spectacular exposures of the angular unconformity between the two units occur in this area (Fig. 3.26a)

3.3.6.4 Petrography, Volcanology, and Sedimentology

3.3.6.4.1 Facies A: Dacitic to Rhyolitic Ignimbrite

Description: The most common facies of the Riversdale Volcanics, Facies A, crops out extensively in the Carwell Creek valley and is particularly well exposed along Hurlstone Creek, south of Charbon Quarry, and in several un-named creeks south of "Glendale" (Fig. 2a). The facies is composed of welded and lesser non-welded ignimbrites which are characteristically dark purple in colour but weather to green-grey and white. Outcrops are often massive and highly weathered; however, a pumice lenticle foliation defined by fiamme is commonly visible (Fig. 3.26g). Columnar jointing is often well developed where fiamme are visible and is exposed in thick welded ignimbrite sequences at GR 773500 6355650 and GR 773080 6354950 (Figs 3.26e and 3.26f). Locally, the ignimbrites are non-welded and variably crystal rich (55%) to crystal poor (<20%). The best exposures suggest multiple flow units which were at least 10 m thick and massive and generally ungraded; basal ignimbrite facies such as ground layers or layer 2a (Sparks et al. 1973) were not detected and intervening ashfalls are rare (See Facies G). The rugged terrain, blocky outcrop, and complex structure prevented the mapping of individual flow units within the facies.

In thin section, the ignimbrite commonly has a crystal content ranging from 20% to 32% and consists of quartz (to 5 mm, fractured, embayed, rare bipyramidal shapes); K-feldspar (to 2 mm, subhedral); and plagioclase (to 1.5 mm, subhedral, albite) with rare biotite. Cognate lithic fragments of porphyritic dacite up to 2 cm are common, and pumice lapilli range from absent to abundant. The groundmass consists of submicroscopic ash and glass shards, which range from non-welded through partially-welded to densely-welded (Figs 3.26b and 3.26c). Shards are often well preserved in the non-welded ignimbrite and display mainly cuspate and Y shapes. Devitrification of the shards to finely crystalline quartz is ubiquitous with axiolitic, spherulitic and granular textures present. Hematite and zircon are present as accessories. Chemical analyses by the author (3 samples submitted through the Geological Survey of New South Wales), Millsteed (1985) and Cook (1988a) indicate rhyolitic to dacitic compositions with SiO₂ commonly between 64% and 85%, with some rock compositions modified by silicification and probable fines depletion.

Interpretation: The predominance of densely welded sequences of dacitic to rhyolitic ignimbrite suggests deposition by voluminous subaerial pyroclastic flows. The scarcity of intervening ashfalls or reworked horizons suggests long periods of almost continuous eruption of pyroclastic flows. Most of the ignimbrites in the Clандulla State Forest area are densely welded and represent multiple compound cooling units at least 10's of metres thick; these characteristics suggest a caldera-related
source for these ignimbrites. Alternatively, some of these thick sequences may represent small-volume eruptive products of small composite cones, which were ponded in canyons to thicknesses sufficient to cause welding. Thinner, less crystal-rich, and more variably welded sequences to the north and west of these thick southern exposures may represent more distal outflow ignimbrites from the possible caldera. The depositional setting of the Riversdale Volcanics is discussed in more detail in Section 3.3.6.6.

3.3.6.4.2 Facies B: Volcaniclastic Sandstone and Pebble Conglomerate

Description: The second most common facies in the Riversdale Volcanics, Facies B was mapped as a separate unit, commonly as lenses within Facies A ignimbrites, and becomes more common towards the top of the Volcanics (Figs 2a and 3). Best exposures of the unit occur in creeks to the south of the "Glendale" property (GR 769800 6361000), along Hurlstone Creek (GR 770200 6358250) and at "The Basin" (GR 772600 6355800).

In thin section, these rocks are tightly packed with closed frameworks, poor to very poorly sorted, and texturally immature with angular to subrounded grains of volcanic quartz, plagioclase, porphyritic silicic volcanic grains, aphyric silicic volcanic grains, and rare K-feldspar, fragmental accretionary lapilli, and highly altered ferromagnesian minerals (Fig. 3.28b). Matrix consists of very fine siliceous ash and a quartz or, less commonly, carbonate cement is developed. Pervasive alteration to chlorite, epidote, sericite and hematite is common. The crystal composition suggests a dacitic to rhyodacitic volcanic provenance; rocks of this facies plot dominantly as volcanic litharenites (after Folk 1974) (Fig. 3.27).

In outcrop, the facies comprises medium-grained sandstone through very coarse-grained sand­stone to monomict pebble-granule conglomerate with subrounded clasts of white to purple dacite or rhyodacite (Fig. 3.28a). Beds are 1 to 3 m thick with common erosional bases, and are lensoidal over 10's of m. Beds are commonly massive or display crude horizontal laminae; small-scale tabular cross beds were noted in 2 outcrops near GR 769950 6358310 and these indicated flows to the east-northeast (Fig. 3.28d). Normal distribution grading is infrequently developed, particularly in the lower parts of beds (Fig. 3.28a). Shale interbeds are generally rare and thin (<10 cm) except at "The Basin" (GR 772650 6355800), where the facies fines grossly upwards and contains interbeds of grey, massive and bioturbated shale up to 1.5 m thick before passing into limestone (Facies F).

Interpretation: The sedimentary characteristics, lensoidal shape of the facies, and rare association with limestone suggest that Facies B represents the deposits of gravelly to sandy fan-deltas or braid-deltas, which, in some areas, prograded into the nearshore zone. The massive to normally graded, erosively-based conglomerate to sandstone probably represent the stacked deposits of heavily sediment-laden sheetflows generated by waning ephemeral currents, such as unchannelised flash floods (Nemec and Steel 1984; Flint and Turner 1988). Sparse palaeocurrent readings suggest flows to the east-northeast. The presence of authigenic hematite in the matrix and the absence of sedimentary structures
indicative of a subaqueous setting suggest deposition mainly on the subaerial part of the delta. However, the thinner graded granule conglomerate beds interbedded with bioturbated shale at "The Basin" suggest subaqueous sheetflows or turbulent mass flows typical of delta front facies (Nemec and Steel 1984). The predominance of juvenile volcanic detritus in the conglomerate suggests rapid resedimentation of unconsolidated pyroclastic and epiclastic detritus following a dacitic pyroclastic eruption. Large volumes of immature volcanic detritus were rapidly eroded from the volcanoes and delivered to swiftly prograding deltas and shorelines as high density sandy to gravelly sheetfloods or grainflows (Lowe 1982). Alternatively, similar thick, massive to graded volcaniclastic beds can be created by ignimbrites which entered shallow water and transformed into water-supported mass flows of pyroclastic detritus (Cas and Wright 1987).

### 3.3.6.4.3 Facies C: Dacitic Boulder Conglomerate to Dacitic Sandstone

**Description:** This facies is consistently present as a 1 m to 22 m thick horizon of volcanic conglomerate and sandstone at the top of the Riversdale Volcanics; it is well exposed at GR 770700 6364200 and GR 769300 6365300. The horizon displays a gross fining-upward trend and passes gradationally into the Carwell Creek Formation (Fig. 3.24b). The basal parts are dominated by very poorly-sorted, clast- to matrix-supported, oligomict conglomerate displaying clasts of welded and non-welded, purple dacitic ignimbrite, and rare dacitic arenite. Clasts vary from well-rounded to subangular, with low sphericities, and range from granule to boulder dimensions (Fig 3.28c); a positive correlation exists between maximum clast size to bed thickness. The matrix and sandstone interbeds are loosely-packed, poorly sorted and immature, and comprise medium sand- to granule-sized framework grains of volcanic quartz, welded and non-welded dacitic ignimbrite fragments, and rare plagioclase and hornblende in a supporting matrix of chert, and authigenic clays, micas and hematite.

The conglomerate beds are 30 cm to 2.5 m thick and laterally continuous, with sharp and planar to scoured boundaries. Basal beds are typically massive, ungraded, unstratified and disorganised. Higher in the sequence (Fig 3.24c), thinner beds of pebbly conglomerate to coarse-grained sandstone – containing sporadic interbeds of shale to 30 cm thick – display crude to well-defined horizontal stratification and pebble alignment, normal and reverse grading, clast imbrication, and rare low-angle cross stratification; the latter two indicate a northwest-dipping palaeoslope (Fig. 3.24b) similar to that recorded on the disconformity surface (Section 3.3.6.3). Thin interbeds of impure micritic limestone occur towards the top of the horizon and the calcareous content of the sandstone increases, both as carbonate cement and as angular framework grains of calcite. In the uppermost beds, the conglomerate and sandstone are well-sorted and display 10-50 cm thick trough and tabular cross-bed sets showing identical orientations to those of the basal Carwell Creek Formation.
Figure 3.26 (a) Angular unconformity between steeply dipping welded ignimbrite of the Riversdale Volcanics and sub-horizontal conglomerate of the Permian Shoalhaven Group. Hammer (circled) for scale. Outcrop is in Red Springs Creek, Clandulla State Forest at GR 772000 6354150. (b) Photomicrograph of a moderate to densely welded ignimbrite from GR 774000 6357520. Compacted and deformed shards (S), in a matrix of fine ash, define a foliation which deflected around quartz crystals (Q). Cuspate shard shapes (CS) are locally present. Plane polarised light. (c) Photomicrograph of a crystal-and lithic-rich, devitrified, welded ignimbrite from GR 770280 6362280 with crystal fragments of quartz (Q) and K-feldspar (K), and volcanic lithic fragments (L) in a matrix of formerly glassy shards (S) and fine ash. Plane polarised light. (d) Photomicrograph of a rhyolite ?lava from GR 766925 6369439 showing euhedral K-feldspar phenocrysts in a fine quartz-feldspar groundmass. Crossed polars. (e) Columnar jointing, in cross-section, developed in a welded dacitic ignimbrite at GR 773500 6355650, Clandulla State Forest. (f) Columnar jointing in a welded dacitic ignimbrite near Clandulla Picnic Grounds at GR 773080 6354950. (g) Well developed eutaxitic texture in a densely welded rhyodacitic ignimbrite at GR 774750 6357310, displaying very strong compaction of pumice lapilli.
Figure 3.27 QRF diagram (after Folk, 1974) for sandstones of Facies B (circles) and Facies D (crosses) of the Riversdale Volcanics. Data from this study and Cook (1988a, 1990).

Interpretation: The basal boulder-cobble conglomerate has features suggesting deposition by subaerial cohesive debris flows (Nemec and Steel 1984), whereas conglomerate and sandstone higher in the sequence suggest shallow subaqueous deposition by a variety of mass-flow processes: cohesive debris flows, slurry flows, and possibly transitional to high-density turbidity currents in the graded and cross-stratified beds. The interbedding of these mass-flow deposits with shales and micritic limestones, indicates they were emplaced in a very quiet, shallow marine setting – probably a lagoon. The uppermost beds record nearshore reworking of the detritus by west- and west-southwest-directed longshore currents. The complex interplay of mass-flow processes and marine traction currents in the upper sequence is typical of the fan-delta front facies documented by many workers (e.g. Nemec and Steel 1984).

In summary, this facies represents a transgressive subaerial to shallow subaqueous accumulation of boulder to granule volcanic conglomerate and sandstone derived from a high-relief volcanic terrain to the south and deposited on northwest-sloping fan deltas which built out into a generally quiet lagoonal environment (Fig. 3.30).
**3.3.6.4 Facies D: Quartz to Sublithic Sandstone**

*Description:* White, medium- to fine-grained quartz to sublithic sandstone occurs in the Riversdale Volcanics in several areas, most notably in a creek north of "The Basin" at GR 771320 6357650 and near the "Wyalda" property (GR 771700 6352150) where it is well exposed in a road cutting. At the latter locality, the sandstone conformably overlies volcaniclastic sandstone of Facies B. and dacitic ignimbrites of Facies A, and becomes enriched in volcanic detritus close to the contact with these facies. In thin section, the sandstone is dominated by well rounded to subrounded grains of quartz, sedimentary lithic fragments (mostly shale), and variable amounts of silicic volcanic grains, with a quartz cement commonly well developed; the sandstones mostly plot as sublitharenites (after Folk 1974) (Fig. 3.27). Facies D sandstones are commonly medium to thick-bedded (0.4 m to 1 m thick) and interbedded with grey shale (1 to 5 cm thick). Beds display parallel lamination and rare low angle tabular and trough cross bedding and symmetrical ripple marks (Fig. 3.28e). Fauna is limited to rare *Skolithos* burrows and locally abundant crinoid ossicles.

*Interpretation:* The facies represents high energy shallow marine deposition probably in a foreshore to middle shoreface environment. The dominance of quartz and sedimentary lithics suggests derivation mainly from exposed areas of quartzose sedimentary basement during intervals of volcanic quiescence.

**3.3.6.4.5 Facies E: Carbonate**

*Description:* Massive unfossiliferous, medium- to thick-bedded dolomite occurs in the upper part of Rocky Waterhole Creek, east of Ilford at GR 772250 6352250. This dolomite lens is 10-15 m thick and 70 m long and passes laterally and vertically into fine-grained volcanic sandstone and shale (Facies B), becoming impure with silicic volcanic detritus towards the margins (Fig. 3.28f).

At "The Basin" in Clandulla State Forest (GR 772300 6355800), a 700 m long limestone lens crops out within volcaniclastic sandstone and conglomerate of Facies B. Bedding readings in the adjacent clastics suggest the limestone occurs in the core of a small syncline and is possibly folded internally. In addition, several large sink holes in this area indicate the presence of a cave system. The limestone is typically blue-grey micrite and massive or with alternating light and dark bands and possible birdseye textures (ie. small cavities filled with sparite). The limestone is thickly bedded with rare tabulate corals (favositids) and stromatoporoids, and generally becomes impure with volcanic detritus towards the base. Rare horizons have widespread silicification, normally of stromatoporoids and favositid corals. At GR 772450 6355420, the limestone is thin-bedded and contains an abundant silicified fauna of brachiopods, corals, crinoids and conodonts (Section 3.3.6.5).

*Interpretation:* These horizons are interpreted as small autochthonous carbonate accumulations which developed in shallow marine conditions during volcanically quiescent intervals. The limestone facies suggests a fairly shallow, somewhat restricted environment, and the laminated facies with
Figure 3.28 (a) Facies B: part of a thick bed of volcanic-rich sandstone to granule conglomerate exposed in Carwell Creek south of Clandulla Picnic Grounds at GR 773000 6353250. Outcrop shows three thin, normal graded to parallel-laminated units indicating repeated episodes of waning flow events, possibly sheetfloods. Lens cap for scale. (b) Photomicrograph of a poorly sorted, immature, volcanic rich, coarse-grained sandstone to granule conglomerate of Facies B from GR 770500 6358010. Framework grains are dominated by angular to subangular quartz crystals (Q) and silicic volcanic rock fragments (V) displaying vitriclastic textures including abundant glass shards (variably welded) and broken quartz crystals. Matrix is rich in authigenic clays and hematite. Plane polarised light. (c) Facies C: Massive, ungraded bed of clast- to matrix-supported, pebble to cobble conglomerate composed of mostly rounded clasts of purple dacitic ignimbrite. Outcrop is at GR 769300 6365300. (d) Facies B: fine- to medium-grained immature volcanic-rich sandstone from GR 769950 6358310 displaying climbing ripple lamination passing upwards into low-angle tabular cross beds. (e) Facies D: roadcutting exposure of fine-grained parallel-laminated quartz sandstone at GR 771700 6352150. Hammer (circled) for scale. (f) Facies E: a small lens of massive, thick-bedded brown dolomite exposed in the upper part of Rocky Waterhole Creek, east of Ilford at GR 772250 6352250.
possible birdseye textures suggests local intertidal conditions. However, rare well-bedded limestone, with rich diverse faunas suggests deeper and more open marine environments were present higher in the sequence.

3.3.6.4.6 Facies F: Volcanic Ash

**Description:** Thin (1 to 3 m thick), laterally persistent horizons of very fine, white to cream volcanic ash occur throughout the Riversdale Volcanics as lenses within Facies A. The best exposures occur along Carwell Creek near Clandulla Picnic Grounds at GR 772850 6354500, near Great Western Dam (GR 774400 6349650) and further north in the "Glendale" area at GR 767600 6362100, close to the base of the Volcanics. The facies is usually massive and uniformly fine-grained in outcrop with rare diffuse lamination. A 2 cm thick, ungraded horizon rich in whole and fragmental, spherical to ellipsoidal accretionary lapilli, 0.2 cm to 1.7 cm in diameter supported in an ash matrix was noted at GR 772850 6354500; Cook (1988a) noted similar accretionary lapilli-rich horizons at GR 766600 6360600. Thin sections reveal very fine grains (<0.05 mm) of quartz, plagioclase, devitrified volcanic detritus (including probable glass shards) and clays. Rare bioturbation and poorly-preserved brachiopods were noted in a thin white ash horizon near Great Western Dam (GR 774400 6349650).

**Interpretation:** The close association with Facies A, together with the very fine-grained nature and the minor thickness of many horizons, suggests Facies F represents co-ignimbrite ashfalls (the fallout of a dilute ash cloud accompanying a pyroclastic flow: layer 3 of Sheridan 1979). Some of the thicker deposits with abundant accretionary lapilli are more likely to represent the direct fallout from the ash cloud of a major phreatic or phreatomagmatic eruption. The deposits contain only rare evidence of current activity or bioturbation, suggesting a generally subaerial deposition environment, with the exception being the thin, brachiopod-bearing horizon at Great Western Dam which probably represents a water-laid airfall tuff deposited in a quiet shallow marine environment.

3.3.6.4.7 Facies G: Polymict Conglomerate

**Description:** Restricted lensoidal outcrops (normally < 20 m thick) of clast- to matrix-supported, polymict, cobble to pebble conglomerate or pebbly sandstone occur sporadically throughout the Riversdale Volcanics interbedded with Facies A; best exposure occurs at GR 772800 6356050. Beds are typically 20 cm to 50 cm thick, laterally continuous in outcrop, and massive to parallel laminated. Normal grading of clasts is rarely developed and rare thin beds of unbioturbated grey mudstone occur. The clast population is diverse and consists of green lithic sandstone, quartz sandstone, quartz-feldspar phryic volcanic clasts, and dark chert. Clasts are dominantly subrounded to well rounded and the matrix is composed of medium to coarse sand size quartz crystals and lithic fragments.

**Interpretation:** Facies characteristics and association with Facies A suggest this facies represents mainly rapid subaerial deposition by fluvial traction currents or rare sheetfloods, probably in
alluvial fan or braidplain settings during volcanically quiet periods. Clast lithologies suggest derivation mainly from areas of exposed basement.

3.3.6.4.8 Facies H: Flow-banded Rhyolite Lava

**Description:** This rare facies has only been noted to the north of Lake Windamere, around GR 767400 6369100 and at GR 766925 6369439, where it crops out in thin horizons (<10 m thick) which pass laterally and vertically into Facies A ignimbrites. In outcrop, the rock is white, evenly quartz- and feldspar-phryc, and characterised by planar to gently undulating flow-banding. Thin sections show 24-35% phenocrysts of euhedral K-feldspar (often in glomeroporphyritic aggregates), and embayed, frequently bipyramidal quartz in a microcrystalline groundmass of quartz, feldspar, sericite and minor chlorite (Fig. 3.26d).

**Interpretation:** The limited extent of the rhyolite lava and association with Facies A ignimbrites suggests Facies H represents small subaerial flows localised around minor vents. Williams and Mc Birney (1979) and Cas and Wright (1987) regard subaerial rhyolite lava as a near-vent facies, normally within 3 km.

3.3.6.5 Fauna and Age

The limestone lens at "The Basin" in Clandulla State Forest, locally (at GR 772450 6355420) contains a thin horizon from which a silicified fauna of brachiopods (Cyrtina sp., rhynchonellids indet.), corals (Rhizophyllum sp., Cladopora sp., favositids) and crinoids has been obtained. Conodonts from this horizon (Samples C28, C28a, C28b) include Ozarkodina excavata excavata, O. pseudomiae, Pandorinellina cf. optima, P. steinhornensis miae, Eognathus sulcatus ssp. (1 specimen), and Oulodus sp, Panderodus unicoostatus, and P. sp. (Fig. 3.29) (Colquhoun 1995b; Supporting Paper 5). As the subspecies of Eognathus sulcatus cannot be identified due to the broken nature of the specimen, the fauna can only be assigned a general sulcatus-kindlei Zone (early to mid Pragian) age. The other species have been documented from both of these zones.

The only other fossils noted in the Riversdale Volcanics were poorly-preserved crinoid ossicles near the upper contact with the Carwell Creek Formation. To the south of "Glendale" at GR 768700 636200, Cook (1988a) reported the brachiopod Iridistrophia sp., an unidentified mollusc, and some crinoid stems from a sublitharenite horizon in the upper parts of the Riversdale Volcanics; however, this horizon was remapped by the author and is now considered to be part of the Roxburgh Formation which has been repeated by faulting. Bradbury (1987) reported minor brachiopods and a poorly-preserved bryozoan fauna from an ashfall tuff at GR 774400 63495650; attempts by the author to collect further material from this site yielded only rare unidentifiable brachiopods.

Three samples from the Riversdale Volcanics were submitted by the Geological Survey of New South Wales for U/Pb dating by SHRIMP. Dacitic ignimbrites from close to the base of the unit (GR
Figure 3.29 Conodont fauna from a small limestone lens in the Riversdale Volcanics at "The Basin" in Clandulla State Forest (GR 772300 6355800).

*Pandorinellina steinhornensis miae* BULTYNCK 1971.

- a. Pa element, lateral view of AMF 96385; 28a, x56.
- c. Pa element, upper view of AMF 96385; 28a, x56.


- b. Pa element, lateral view of AMF 96384; 28a, x43.
- d. Pa element, upper view of AMF 96384; 28a, x47.


- e. Pa element, lateral view of AMF 96376; 28, x43.
- g. Pa element, upper view of AMF 96376; 28, x43.

*Ozarkodina excavata excavata* BRANSON & MEHL 1933.

- f. Pa element, lateral view of AMF 96375; 28, x30.
- h. Sc element, AMF 101543; 28a, x48.

*Eognathus sulcatus* ssp. PHILIP 1966.

- j. Pa element, upper view of AMF96370; 28a, x61.
- i Sc element, AMF 101544; 28a, x43.
770280 6362280) gave an age of 410 ± 8 Ma, whereas purple dacitic ignimbrites close to the top of the unit (GR 765870 6363000) gave 389 ± 8 Ma. Due to the large error caused by multiple zircon populations, these dates only confirm a general Early Devonian age for the Riversdale Volcanics. The third sample from the unit at GR 769520 6360380 gave an anomalous age of 418.8 ± 5.5 Ma, suggesting inheritance of zircons from Late Silurian volcanics. The presence of numerous volcanic lithics in the Facies A ignimbrites (Section 3.3.6.4.1) supports such an interpretation.

In addition, a possible equivalent of the unit in the Capertee Valley (the Huntingdale Volcanics) has been assigned a similar age by Bracken (1977), and other volcaniclastic units marginal to the Capertee High and in the Hill End Trough suggest voluminous silicic volcanism on the Capertee High during the Pragian (Section 5.4.3).

3.3.6.6 Environment of Deposition and Discussion

The Riversdale Volcanics was considered to be shallow marine – based mainly on stratigraphic relationships – by Pemberton (1977, 1980a) and Millsteed (1985); however, Cook (1988a) proposed dominantly subaerial conditions due to thickness considerations and the presence of well-preserved accretionary lapilli and unfossiliferous, immature epiclastic interbeds. In addition, the abundance of welding also suggests a subaerial or very shallow subaqueous setting (Cas and Wright 1987). However, local marine incursions occur towards the top of the Volcanics and the uppermost horizons record transgressive reworking.

The predominance of Facies A suggests rapid and almost continuous voluminous eruptions of dacitic to rhyolitic pyroclastic flows; thin syn-eruptive resedimented deposits (ie. small lenses of Facies B) and airfall tuff horizons (Facies F) record sporadic short intervals of volcanic quiescence. Coherent volcanic facies were restricted to rare near-vent accumulations of flow-banded rhyolite (Facies H). Resedimented pyroclastic deposits of Facies B became more common late in the unit's depositional history, suggesting a waning of volcanism and rapid redeposition of pyroclastic detritus by a combination of mass-flow, hyperconcentrated flow and traction current processes in alluvial fan to fan-deltaic environments. The uppermost parts of the Volcanics (Facies C) record epiclastic reworking and deposition by mass-flow and traction currents in transgressive fan-deltas which prograded into a sheltered, restricted lagoon (Fig. 3.30). Quartz sandstone (Facies D) and polymict conglomerate (Facies G) indicate exposure and erosion of small areas of pre-Riversdale basement at various times during the depositional history. Facies E records a break in volcanic activity in the upper parts of the unit, with associated subsidence, transgression, and deposition of carbonates in a generally shallow, restricted marine environment.

The most striking, yet puzzling, feature of the unit is its rapid lateral changes in thickness and facies (Fig. 3.3). The Volcanics attain a thickness of up to 1100 m of mainly ignimbrites to the south of the Cudgegong Road, yet thin to 22 m of reworked equivalents (Facies C) only 4 km to the northeast.
Similarly, the unit thins – albeit less rapidly – to the north and west (Fig. 3.3). The rapidity of some of these variations may have been accentuated to some degree by crustal shortening across thrust faults; however, the location and relatively small displacements of these faults indicate they alone were not sufficient to cause the observed thickness trends. The abundance of densely welded ignimbrites, together with the thickness, volume and areal extent of the Volcanics indicates a likely association with a small to medium volume subaerial silicic caldera (Smith 1979). Such calderas typically produce very thick, ponded intracaldera ignimbrites (up to 2 km thick) which pass laterally, outside the caldera, into coeval outflow ignimbrites which are normally an order of magnitude thinner (Cas and Wright 1987). The thick ignimbrites of the central area (Fig. 3.3) may represent an intracaldera facies; furthermore, features documented in this study and by Cook (1988a) are consistent with such an interpretation (e.g. abundance of welding, thick interbedded sequences of very immature epiclastics, relatively high percentages of phenocrysts and lithics, and a paucity of airfall deposits). In addition, many features documented in this study and by Millsteed (1985) and Pemberton (1977, 1980a) to the north and west are characteristic of outflow ignimbrites (e.g. absent or rare welding; variable, but often low, phenocryst contents; geochemical evidence of fines depletion (Millsteed 1985); thinner sequences with more regular geometry and less rapid thickness variations). The particularly rapid thinning to the east may indicate that most of the outflow ignimbrites were directed down a palaeoslope to the north and west. The lavas (Facies H) recorded to the north of the possible caldera are contrary to standard facies models, which view subaerial silicic lavas as a largely intracaldera facies (McPhie et al. 1993). Such lavas may represent only local accumulations around small, extracaldera vents.

Caldera margins are typified by steep, inward-sloping faults and common collapse breccias (Cas and Wright 1987) and are often extremely difficult to locate in ancient, folded sequences. A number of possible caldera margin faults exist in the thick central area south of the Cudgegong Road where mapping during this study and by Cook (1988a) has delineated numerous complex faults within the Riversdale Volcanics and surrounding units, some of which appear syn-volcanic (Figs 2a and 3).

Alternatively, thick ignimbrites may result from ponding against other high-relief topographic features (e.g. valley walls). The creation of such features, however, would require a period of prolonged subaerial erosion following deposition of the Roxburgh Formation – and such a hiatus is not supported by the available faunal evidence.

In summary, the Riversdale Volcanics represent an accumulation of: dominantly subaerial, welded and lesser non-welded dacitic ignimbrites; subaerial to subaqueously deposited, immature epiclastics; minor rhyolite lavas; thin airfall tuffs; rare carbonates and basement-derived elastics. Thickness and facies variations suggest the unit may represent a thick intracaldera sequence centred in, and to the south, of the district, passing to the north, west, and east into thinner outflow sequences (Fig. 3.30). Volcanism waned towards the top of the unit and the uppermost horizons indicate reworking of volcanic detritus in a paralic, probably fan delta, setting under the influence of a transgressing sea.
3.3.7 CARWELL CREEK FORMATION *Dac* (Brunker and Rose 1969)

3.3.7.1 Synonymy and Definition

Fossiliferous Devonian sandstone, shale and limestone were first noted outcropping around the confluence of Carwell Creek and the Cudgegong River by Clarke (1878) and Wilkinson (1884). Carne (1913) produced the earliest geological map of these lithologies from the vicinity of the Cudgegong Cinnabar Mine. Carne and Jones (1919) described and reported chemical analyses for several large bodies of Devonian limestone, dolomitic limestone and dolomite north of the Cudgegong River, between Rylstone and Cudgegong. Rocks included in the present Carwell Creek Formation were also described by Game (1934, 1935) as Middle Devonian limestone and Upper Devonian quartzite, claystone, acid tuff, grit and limestone.

Wright (1966) considered calcareous sandstone, conglomerate, grit, shale and limestone between Cudgegong and Rylstone, and in the Capertee Valley, to correlate with the Melrose Formation – a Middle Devonian unit he defined to the west of Mt Frome (now known as the Boogledie Formation; Offenberg *et al.* 1971). The name Carwell Creek beds was assigned to a substantially similar group of lithologies in this area by Brunker and Rose (1969) on the Sydney 1:500 000 sheet; however, they regarded the unit as Silurian. Subsequent published references to the unit (Offenberg *et al.* 1971; Talent *et al.* 1975; Pickett 1982a; Cas 1983) largely followed Brunker and Rose's opinions. An Early
Devonian age was later established on stratigraphic (Brown 1974; Pemberton 1977, 1980a) and eventually faunal (Millsteed 1985, 1992) grounds. Mapping during this study has considerably refined knowledge of the outcrop area and stratigraphic relationships.

Clastics and carbonates of the Mt Knowles-Buckaroo and the Brogans Creek areas were correlated with the Carwell Creek Formation by Pemberton et al. (1994; Supporting Paper 2) and Colquhoun et al. (1997; Supporting Paper 8) and are described herein. The Mt Knowles-Buckaroo outcrops were previously included in the Boogledie Formation by Offenberg et al. (1971), the youngest known unit of the Kandos Group and presumably early Middle Devonian (Eifelian) in age (Colquhoun et al. 1996; Supporting Paper 7). The Boogledie Formation was redefined by Colquhoun et al. (1997) to include only the clastics which overlie the Mt Frome Limestone in the Mt Frome area (Section 4.4.1.1).

The Carwell Creek beds have been elevated to formation status in recent studies (Cook 1988a, 1990; Pemberton et al. 1994; Colquhoun et al. 1997; this study), as the lithologies, extent, boundaries and age of the unit are now well known. It should be noted that the name Carwell Creek Formation was proposed by Lavers (1960) to describe an 'Early Devonian sequence of shale, sandstone and conglomerate. However, Lavers' unit transects numerous age and lithological boundaries recognised by subsequent authors and bears scant resemblance to the unit currently defined, which contains elements of Lavers' Kandos Formation, Wells Formation and Carwell Creek Formation. As Brunker and Rose (1969) designated no type section for the Carwell Creek beds, a suitable type section for the Carwell Creek Formation is proposed from GR 767800 6368700 (base) to GR 768500 6368900 (preserved top).

Laterally discontinuous bodies of basal limestone, dolomitic limestone and dolomite have been defined as the Glendale Limestone Member, based on outcrops on the "Glendale" property (GR 769500 6363200), by Pemberton et al. (1994). Previously, these rocks were informally referred to as the 'Upper Kandos Limestone' by Cook (1988a, 1990) and Booth (1990). No type section was originally defined for the unit, but Pemberton et al. (1994) proposed a representative section between GR 768100 6368500 and GR 768800 6368700.

3.3.7.2 Distribution, Areal Extent and Thickness

The Carwell Creek Formation has the widest distribution of any Early Devonian unit on the Capertee High; total areal extent of outcrop is estimated at 33 km². The unit crops out extensively in the core of a synclinorium to the east of Carwell Creek (Figs 1, 3), and also to the south of the Cudgegong Road as complexly folded and faulted exposures to the west and south of "Glendale" (Fig. 3). Less extensive occurrences are: (a) a small inlier 1 km west of Charbon, surrounded by Permian Shoalhaven Group (Fig. 2a); (b) another small inlier 2 km northwest of Kandos at GR 776000 6362300, surrounded by the Early Permian Rylstone Volcanics and the Shoalhaven Group (Fig. 2a); (c) extensive outcrops in the Brogans Creek area (around GR 776000 6346000) in the extreme southeast corner of
The Mudgee 1:100 000 Sheet (Figs 2c, 3); (d) a narrow belt of outcrop, partially covered by basal Permian strata, extends north from the Pinnacle Swamp area (6.5 km northwest of Rylstone; GR 770500 6371300) to south of "Havilah" (GR 757300 6383200), where the unit is displaced by the Havilah Fault (Fig. 3).

The Carwell Creek Formation occurs again to the north of "Havilah", extending from the Mt Knowles area (GR 755200 6388000) to the Buckaroo-Eurundury district where it is folded around the nose of the Pine Ridge Syncline (GR 749200 6401500) and the Eurundury Anticline (Fig. 3). A small fault-bounded outcrop occurs on the "Erang" property (Fig. 3.34) to the east of Windamere Dam (GR 761560 6374390). Similarly, a very small exposure of Carwell Creek Formation, surrounded by Late Devonian units, occurs north of Middle Creek at GR 763650 6376810 (Fig. 3).

The Carwell Creek Formation displays great changes in preserved thickness due, in part, to the amount of erosion on its unconformable upper contact (Fig. 3.3). The unit attains a maximum preserved thickness of approximately 500 m around the lower reaches of Carwell Creek. Millsteed (1985, 1992) estimated the thickness here as 1200 m; however, this is probably an overestimate as several folds were mapped by the author in the section from which Millsteed obtained this estimate. A composite measured section in the Carwell Creek area (Fig. 3.31b) indicates a thickness of 475 m. Cook (1988a, 1990) obtained a composite thickness of 260 m from exposures around "Glendale", south of Cudgegong Road and Pemberton (1977, 1980a) estimated the maximum thickness as 350 m from exposures southeast of Cudgegong (Fig. 3.3). North of the Rylstone-Cudgegong district, the unit thins: approximately 180 m is present in the Pinnacle Swamp area (Fig. 3.31a), whereas southwest of "Havilah" the unit has thinned to between 60 m and 110 m. The base is not exposed in the Mt Knowles and Eurundury areas and available bedding data suggest at least 220 m of the unit is present. Thickness is difficult to estimate in the Brogans Creek area as neither the top nor base of the unit is exposed and tight folding is present.

3.3.7.3 Structure and Stratigraphic Relationships

The Carwell Creek Formation conformably overlies the Riversdale Volcanics; this contact has been described in Section 3.3.6.3. North of GR 771950 6365320, the Carwell Creek Formation onlaps the Silurian-Devonian unconformity and overlies the Dungeree Volcanics with disconformity or slight angular unconformity. Previous authors regarded this Siluro-Devonian contact as conformable (Sussmilch 1934; Game 1935; Brown 1974). The contact truncates several bedded horizons in the Dungeree Volcanics at a number of localities (Fig. 1); for example, at GR 772000 6364850 the base of the Carwell Creek Formation cuts across a shale-volcanics contact in the Dungeree Volcanics. Furthermore, at GR 771830 6363300, crinoidal sandstone of the Carwell Creek Formation overlies silicic volcanics of the Dungeree Volcanics. The unconformity surface is sharp and relatively planar, a common feature of marine transgressive unconformities (Shanmugan 1988). Further east at GR
Figure 3.31 (a) section through the Carwell Creek Formation along Pinnacle Swamp Creek between GR 770620 6372020 and GR 769880 6372260; (b) composite stratigraphic section through the Carwell Creek Formation in the Carwell Creek area (compiled from sections between GR 769040 6366090 and GR 769870 6366790; GR 769190 6367480 and GR 770160 6367810; GR 771710 6368880 and GR 772070 6367110). Vertical units are in metres.
772700 6366400, a thin horizon of basal volcanioclastic conglomerate in the Carwell Creek Formation overlies the unconformity and has been described in Section 3.3.7.4.1. (Facies A).

The amount of angular discordance across the Siluro-Devonian contact is difficult to gauge due to a paucity of bedding in the Dungeree Volcanics. The 6 closest bedding pairs (50-200 m outcrop separation), restored on a stereonet, show no more than 8° disparity in dips across the contact, suggesting either a disconformity or very low-angle unconformity. This low structural discordance suggests the unconformity was probably due mainly to eustatic sea-level fall, rather than tectonic uplift (Shanmugan 1988).

The Carwell Creek Formation is overlain unconformably by the Late Devonian (?Frasnian) Buckaroo Conglomerate. This contact is rarely exposed, but can often be located to within 2 m and the degree of angular discordance is always less than 20° and usually less than 10° (Colquhoun et al. 1997). Millsteed (1985, 1992) also noted a low-angle unconformity between these units to the west of Rylstone. Powell and Edgecombe (1978) found a low-angle discordance of 5° to 25° between the early Middle Devonian Boogledie Formation and the Buckaroo Conglomerate in a structural study of the Pine Ridge syncline (their 'Mt Frome syncline'). They attributed this angular unconformity to a period of erosion, uplift and broad tilting during the Middle Devonian (Section 2.3.3).

In the Carwell Creek south area (Fig. 2a), the Carwell Creek Formation is unconformably overlain at numerous localities by flat-lying sediments of the Permian Shoalhaven Group. Spectacular cliff exposures of this angular unconformity occur throughout this area (Fig. 3.35h).

3.3.7.4 Petrography and Facies

The dominant clastic lithologies of the formation are conglomerate, pebbly sandstone and sandstone. Sandstones are fine- to coarse-grained and texturally submature with subrounded to subangular, moderate- to poorly-sorted, tightly-packed sand-sized grains averaging 0.25-0.55 mm in diameter. A gradation in composition is present, from volcanic-rich litharenite to quartzarenite (after Folk 1974; Fig. 3.32). In general, the sandstone becomes more quartz-rich and finer grained with increasing stratigraphic height and the majority of coarse lithic sandstones are confined to Facies A and C in the basal portions of the Carwell Creek Formation. Bioclastic content of sandstones also tends to decrease away from the base of the unit. The petrography of specific facies is discussed in the relevant sections.

Eight facies were recognised in the Carwell Creek Formation based on lithology and sedimentary structures. Two detailed sections were measured through the unit in the Carwell Creek area (Fig. 3.31). As the sequence has an overall transgressive character, successive facies onlap the Siluro-Devonian unconformity east of the Carwell Creek. The facies succession and associations are discussed further in Section 3.3.7.6.
3.3.7.4.1 Facies A: Basal volcanic conglomerate

**Description:** Facies A is present sporadically at the base of the Carwell Creek Formation where it unconformably overlies the Late Silurian Dungeree Volcanics. It is best developed as an 8.5 m thick, 400 m long lens north of the Cudgegong River at GR 772750 6166560; smaller lenses occur overlying the unconformity for 400 m east of here. The facies comprises tightly-packed, poorly-sorted pebble to granule conglomerate to pebbly sandstone composed of angular to subrounded clasts of silicic volcanics and, more rarely, jasper. The matrix is composed of fine- to very fine-grained volcanic-rich sandstone. Bed thickness is difficult to discern due to poor outcrop and varies from 40 cm to over 1.5 m. Facies A is generally massive and disorganised or displays crude horizontal lamination and clast alignment, normal, coarse-tail grading is rarely present. The conglomerate of Facies A passes gradationally upwards over several metres into coarse crinoidal sandstone of Facies C (Fig. 3.33a).

**Interpretation:** The conformable position of Facies A at the base of a transgressive shallow marine sequence suggests a shallow marine, alluvial fan, or fan-delta depositional environment – or a mixture of these. The thickness, massive character, poor sorting, open frameworks, and lack of bioturbation are not indicative of transgressive lag deposits developed at the base of shallow marine successions (Elliot 1986). These features suggest deposition by subaerial debris flows or
hyperconcentrated flood flows (Lowe 1982). The highly lensoidal geometry of the facies suggests these deposits were probably channelised in valleys (in the underlying Dungeree Volcanics) which developed prior to transgression. The highly gradational upper contact with Facies C indicates at least the upper part of Facies A was transitional to a nearshore marine system – possibly resembling a fan-delta setting. Clast lithotypes and matrix petrography indicate derivation from a volcanic and sedimentary basement source area and minimal transportation, consistent with the above interpretations.

3.3.7.4.2 FACIES B: GLENDALE LIMESTONE MEMBER

Description: The Glendale Limestone Member (Daeg of Fig. 3) occurs at the base of the Carwell Creek Formation between Rylstone and Cudgegong as numerous exposures of limestone, dolomite limestone and dolomite overltying the Riversdale Volcanics and possibly the Clandulla Limestone and Dungeree Volcanics (see Section 3.3.7.3). The upper contact with basal crinoidal sandstone of Facies C is commonly highly gradational, with carbonates of the member becoming gradually more impure and displaying increasing evidence of current activity such as cross-beds or parallel lamination. This contact is well exposed in numerous localities in folded sequences to the north of "Lucknow" (e.g. GR 770540 6364760). The Glendale Limestone Member varies from less than 10 m thick to a maximum of 180 m in "The Dolomites" area to the north of Lake Windamere.

Carbonates in the member are quite variable. Near "Erang" (GR 761500 6374420; Fig. 3.34), northeast of the Windamere Dam wall, Facies B consists of yellow to grey biomicrite to micrite with abundant, occasionally silicified, brachiopods, corals and trilobites (Section 3.3.7.5). Elsewhere, Facies B is characteristically thick-bedded, massive, and very sparsely fossiliferous with only rare ovoid to hemispherical stromatoporoids and favositid corals present. In rare instances, a more diverse silicified fauna is present; for example, near "The Dolomites" (GR 768380 6367890) where a fauna of corals (Rhizophyllum, favositids), stromatoporoids and rare brachiopods (cf. Spinella sp.) occurs. Recrystallisation is often extensive and stylolites are typically well developed. Between GR 771780 6365790 (north of "Lucknow") and an area known as "The Dolomites", the Glendale Limestone Member is intensely dolomitised. Millsteed (1985) studied the dolomitisation of the unit in this area, concluding that it was an early diagenetic feature. "The Dolomites" area contains one of the largest deposits of dolomite in the state (Ringis 1966).

In the upper reaches of Cumber Melon Creek (around GR 774850 6156450), a thick sequence of massive blue-grey limestone outcrops in the core of an anticline whose axis parallels the creek bed. The limestone, which has been quarried in several locations, is thick-bedded and very sparsely fossiliferous with rare favositid corals and hemispherical stromatoporoids.

In a detailed measured section through Facies B (Fig. 3.31b, 0-45 m), the carbonate was extensively dolomitised, and highly impure towards the base with numerous angular grains of volcanic quartz and silicic volcanic rock fragments (Fig. 3.33d). Thin sections from the basal 35 m reveal
occasionally abundant, small (< 1 mm) grains of gypsum; Millsteed (1985) also noted minor quantities of gypsum in dolomite north of the Cudgegong River. The dolomite is very sparsely fossiliferous and contains only rare ovoid stromatoporoids. Normally graded to massive beds of clast-supported dacitic pebble-granule conglomerate and sandstone (10 cm to 1.5 m thick) are present and become more common towards the basal contact with the Riversdale Volcanics. Contrasting with these beds are numerous 10-50 cm thick beds of fine- to very fine-grained quartz sandstone with parallel lamination and diffuse, southwest-dipping, low-angle cross-beds which occur sporadically throughout Facies B, becoming more common towards the upper contact with Facies C. Rare thin interbeds of dark mudstone are common around the 20 m level of the section.

In the "Glendale" area, the facies outcrops in a small syncline and comprises fairly pure grey limestone which has been extensively recrystallised to microspar calcite. The limestone is sporadically fossiliferous, particularly towards the north, and contains rare favositid corals, stromatoporoids and bryozoans.

Extensive limestone outcrops occur near the Cudgegong Road (around GR 768000 6365000); these are typically blue-grey, medium- to thin-bedded biosparite with abundant interbeds of well bioturbated brown shale up to 35 cm thick. In road cuttings along the Cudgegong Road, the limestone is well bedded and abundantly fossiliferous with numerous crinoid stems and ossicles, tabulate corals, bryozoans and very rare brachiopods (Fig. 3.33b).

Limestone (or marble) quarries at GR 765810 6362750 and GR 768300 6363910 display very thick-bedded, massive and very sparsely fossiliferous limestone which has been extensively recrystallised. The limestone is generally fairly pure and white in colour but locally contains up to 3% fine volcanic quartz grains and variable quantities of secondary hematite and goethite. The carbonates at these localities have been quarried for dimension stone marble for over 100 years with the quarries yielding two varieties known colloquially as 'Cudgegong Ivory' and 'Cudgegong Gold' — reflecting the iron oxide content. More recently, the marble from these quarries has been crushed nearby for terrazzo slabs and tiles.

In summary, the Glendale Limestone Member (Facies B) is a highly variable series of discontinuous lensoidal carbonate bodies at the base of the Carwell Creek Formation.

**Interpretation:** A shallow, subtidal restricted lagoonal environment with limited clastic input is indicated by the carbonate facies (including rare evaporites) and rare, low diversity fauna. The overall paucity of fauna suggests generally restricted conditions combined with hypersalinity (indicated by evidence of local evaporitic conditions). The presence of rare highly fossiliferous intervals suggest small areas (possibly inlets) or periods of better oxygenated, normal salinity water which allowed benthic communities to establish. Although frequently high density, the latter faunas have a characteristically low diversity, suggesting some environmental stress. Well-bedded bioclastic limestone
Figure 3.33 (a) Facies A and C; medium- to coarse-grained volcanic sandstone passing upwards into fine-grained crinoidal lithic sandstone characterised by parallel lamination and low-angle swash cross-bedding. Outcrop is north of the Cudgegong River at GR 772830 6366840. Hammer (circled) for scale. (b) Facies B; bioclastic limestone from the Glendale Limestone Member (GR 768050 6364950) displaying abundant crinoid stems and ossicles, favositid corals, and disarticulated brachiopods. (c) Photomicrograph of a medium- to coarse-grained crinoidal lithic sandstone from GR 770750 6364660. The thin section shows subrounded to subangular grains of quartz (Q), silicic volcanic rock fragments (V) and crinoid ossicles (C) with a carbonate cement. Crossed polars. (d) Photomicrograph of an impure dolomitic limestone from the basal Glendale Limestone Member (GR 768750 6366840). The thin section shows a mosaic of massive recrystallised dolomite and calcite with large grains of volcanic quartz (Q) and silicic volcanic rock fragments (V). Crossed polars. (e) Facies C; well developed parallel lamination and low-angle swash cross-bedding in medium-grained crinoidal lithic sandstone from the basal Carwell Creek Formation at GR 770580 6364810. Hammer for scale. (f) Medium-scale bimodal cross-bedding in crinoidal quartz-lithic sandstone (Facies C) on the banks of the Cudgegong River at GR 771420 6366540. Arrow points to lower cross-bed set.
interbedded with mudstone (Fig. 3.33b) resembles calcareous storm beds (Aigner 1985), suggesting that occasional more open marine conditions (possibly gaps in the barrier island) allowed waves to rework the carbonate platform. The predominance of carbonate in a lagoon setting indicates a low clastic input into the lagoon (Elliot 1986).

Sporadic intervals of high clastic influx, indicated by beds of graded volcanic-rich conglomerate, are interpreted as subaqueous debris flows (possibly transitional to fully turbulent subaqueous mass flows) which developed near several volcanic fan-deltas prograding into the lagoon; these mass flows are similar to those in Facies C of the upper Riversdale Volcanics, close to the transitional boundary with the Carwell Creek Formation (see Section 3.3.6.4.3). Similarly, quartz sandstone interbeds in Facies B fit the description of barrier washovers from modern and ancient settings (cf. McKie 1990). Such deposits are created by large waves that, combined with storm tides, break through the barrier carrying sediment and water into the lagoon where it is deposited as a small fan. The predominance of planar lamination suggests sediment transport in washovers was mainly by sheetflow conditions, with upper flow regime plane bed conditions prevailing. The presence of several of these quartz sandstone beds, closely spaced and interbedded with dolomite suggests the barrier was breached in the same place during successive storms, reactivating the distributary system.

Facies B carbonates have undergone extensive early diagenetic dolomitisation particularly in the east of the district, suggesting dolomitisation processes were restricted to near the edge of the depositional basin. The thickest and most continuous occurrences of the Glendale Limestone Member are restricted to the east of the district (Fig. 3) further suggesting that lagoonal conditions were more prevalent and prolonged in this landward direction. The discontinuous nature of the facies probably reflects the original discontinuous geometry of the barrier island system.

3.3.7.4.3 FACIES C: CRINOIDAL LITHIC SANDSTONE

Description: Facies C is a common and distinctive component of the basal portions of the Carwell Creek Formation, occurring in intervals 20 m to 40 m thick immediately overlying Facies B or, where Facies B is absent, overlying the Riversdale Volcanics (e.g. GR 767590 6369390 and GR 768050 6362100) or unconformably overlying the Late Silurian Dungeree Volcanics (e.g. GR 771810 6366430). The lower contact — with Facies B (described in Section 3.3.7.4.2) and the upper Riversdale Volcanics — is highly gradational, as is the upper contact with Facies D or F.

Facies C consists of fine- to very coarse-grained lithic sandstone. The unweathered sandstone is highly calcareous and comprises up to 60% crinoid ossicles (up to 15 mm in size), along with angular quartz grains, angular to subangular silicic volcanic rock fragments, and minor shale fragments and plagioclase grains (Fig. 3.33c). The matrix is an interstitial mixture of clays and micas. Two phases of carbonate cement are present: an early ferroan calcite cement which has been overprinted by a later
Figure 3.34 Geological map of the "Erang" area, east of Windamere Dam. From Wright et al. (in prep.; Supporting Paper 11).
neomorphic spar calcite (Millsteed 1985). Dolomitisation varies from patchy to pervasive, particularly in the east of the district. The calcareous content decreases steadily away from the base of the facies and the sandstone becomes correspondingly better sorted and compositionally, and texturally, more mature; however, most still plot as litharenites (after Folk 1974, Fig. 3.32).

The sandstone is mainly medium- to thick-bedded and beds are laterally continuous at outcrop scale with sharp to undulating upper and basal contacts. Dominant sedimentary structures are parallel to very low-angle (3° to 5°) laminations in tabular to wedge-shaped beds up to 50 cm thick (Fig. 3.33e) alternating with 10 to 40 cm thick sets of high- to moderate-angle tabular, and lesser trough, cross-beds (Figs 3.33f, 3.35e). Overall, these cross-beds show a bimodal and slightly oblique bipolar pattern with a minor mode towards the southeast (Fig. 3.36). Reactivation surfaces are occasionally visible on foresets. Symmetrical to slightly asymmetrical ripple marks are locally well preserved, although often intensely bioturbated (Fig. 3.35b), and display approximately north-northwest-south-southeast crest trends. Thin zones of convolute lamination were rarely recorded.

**Interpretation:** A marine setting above the fairweather wave base is suggested by the fauna and the abundance of angle-of-repose cross stratification. Wave ripples suggest exposure to an open marine environment and wave propagation in a west-southwest to east-northeast direction (Fig. 3.36). Although storms and waves can approach the coast from different directions, oscillatory flow in near-shore, wave-dominated settings is often oriented perpendicular to shore, as storm waves approaching the coast will refract until the bottom paths of their wave orbitals are nearly perpendicular to shelf isobaths (Komar 1976). The orientation of wave ripple crests on the inner shelf can provide a first approximation of local shoreline trend, especially on relatively straight coasts with cuspate deltas forming subdued headlands, typical of wave-dominated, sand-rich shorelines (Leckie and Krystinik 1989). Ripple crest trends, together with the shoreline trends indicated by regional palaeogeography and direction of onlap outlined earlier (Section 3.3.7.3), suggest an approximately north-northwest to south-southeast trending palaeoshoreline with east being the onshore direction.

The bimodal (slightly oblique bipolar) palaeocurrent pattern for this facies suggests a location exposed to opposing fairweather tidal streams operating approximately at right angles to the coastline. There is no evidence suggesting channelling of the facies and its widespread, almost sheet-like, nature suggests a tidal inlet setting was unlikely. The setting is therefore interpreted as a wave- and tide-influenced shoreface to foreshore characterised by the fairweather migration of mainly straight-crested bars and lesser lunate dunes under slightly velocity-asymmetric tidal currents. These conditions were occasionally replaced by upper flow regime flat bed conditions probably caused by wave-enhanced tidal currents or, particularly towards the base of the facies, wave swash in a foreshore setting. A minor current mode to the southeast (Fig. 3.36) may indicate longshore wave-driven currents. The presence of rare convolute-laminated beds with dewatering structures attests to rapid vertical expulsion of pore water through loosely compacted sands, indicating rapid deposition and high initial porosities. The
stratigraphic position of the Facies C suggests it constituted the barrier island system which protected Facies B carbonates from open marine conditions.

The abundance of the crinoid ossicles in the facies suggests an oxygen- and food-rich environment in which crinoid colonies flourished below the fairweather wave base. These colonies were periodically destroyed by lowering of the wave base during storms and the crinoidal debris was subsequently reworked by tidal currents.

3.3.7.4.4 Facies D: Moderate to Thin Bedded Quartz Sandstone and Shale

**Description:** Facies D consists of sharp-based, planar, fine- to very fine-grained sandstone beds (0.5 to 42 cm thick, av. 10 cm) interbedded with grey-green bioturbated, fine sandy mudstone or, more rarely, laminated black mudstone. Facies D commonly occurs in the measured sections as 5 to 17 m thick intervals which normally overlie Facies C or F with highly gradational contacts (Fig. 3.31). Sand/shale ratios in the measured sections vary from 0.35 to 1.7 and numerous small-scale (0.5 m to 3 m) thickening- and thinning-upwards trends in bed thickness occur (Fig. 3.35d). The sandstone beds display: sharp, scoured bases; normal grading; parallel lamination; rare current ripple lamination grading to starved ripple lamination in very thin beds; and rare hummocky cross-stratification. Symmetrical to combined flow ripples are common on bed tops. The interbedded shales are moderately bioturbated in general with thin, intensely bioturbated intervals, leaving a homogenised sandy mudstone. Recognisable trace fossils are rare, but include *Chondrites* sp. and *Planolites* sp., typical of the *Cruziana* ichnofacies of Pemberton *et al.* (1992). Body fossils include only infrequent crinoids and bryozoans.

**Interpretation:** Facies D represents deposition of thin, graded sandstone beds from waning, storm-generated suspension currents, interbedded with fairweather suspension-deposited mudstone. The presence of wave-produced sedimentary structures suggests deposition in an offshore environment between the fairweather and storm-wave bases. Hummocky cross-stratification is considered to form in such environments under conditions of strong storm-wave oscillatory flow with a superimposed unidirectional geostrophic current. Disagreement, however, exists as to the relative magnitudes of these currents (Nottvedt and Kreisa 1987; Duke *et al.* 1991), although the majority of recent experimental and field work favours either purely oscillatory flows or very oscillatory-dominant combined flows (Duke *et al.* 1991; DeCelles and Cavazza 1992).

The presence and intensity of *Cruziana* ichnofacies bioturbation suggests an shelf environment with generally well-oxygenated bottom conditions that could support an infauna which was locally capable of homogenising the sediment, although zones of weak and intense burrowing again suggest fluctuating, but generally high, rates of suspended sediment supply (Pemberton *et al.* 1992).

The facies resembles Facies B and D of the Roxburgh Formation (Sections 3.3.5.6.2 and 3.3.5.6.4). Similar thin, normally graded beds transitional to shoreface facies have been widely recorded from ancient storm-dominated shelves, especially from the Mesozoic sequences of Canada.
(Hamblin and Walker 1979; Walker 1984) and the southwest U.S.A. (Swift et al. 1987), where they are typically interpreted to occur in inner to middle shelf settings. Such beds can only be reliably distinguished from deep water turbidites by their shallow marine body and trace fossils, the occurrence of hummocky cross-stratification, and the close association with demonstrably shallow marine facies (Brenchley 1985).

3.3.7.4.5 Facies E: Polymict Conglomerate

**Description:** Polymict conglomerate (Dacc of Fig. 3) forms a prominent 5 to 20 m thick marker horizon in the Carwell Creek Formation, around the 140 to 180 m level (Fig. 3.31) which can be traced nearly 6 km from GR 766310 6371780 in the northwest (where it terminates against a fault) south around several folds in the Carwell Creek area to where it is obscured by the Rylstone Volcanics north of GR 772780 6367240. The facies reappears along strike in exposures along Stony Creek, west of Tongbong Mountain (GR 770610 6371910; Fig. 3). A small lens, 350 m long and 5 m thick at GR 767910 6375850 (west of Lue) represents the most northerly occurrence of Facies E. The facies is underlain by either a thin interval of foreshore deposits of Facies F (Fig. 3.31) or a disconformity with Facies D.

The facies is typically present as lenses of pebble conglomerate separated by medium- to coarse-grained, locally pebbly sandstone. Conglomerate beds are 30 to 80 cm thick, clast-supported, moderately well-sorted and sharply segregated from intervening sandstone beds (Fig. 3.35c). Beds are massive or horizontally stratified. Clasts are well rounded to subangular with low to moderate sphericity. Imbrication is locally well developed with pebbles dipping uniformly to the east-southeast (indicating currents to the west-northwest). Clast lithologies are varied and include (in order of decreasing abundance): porphyritic silicic volcanics (many with vitriclastic textures), aphyric silicic volcanics, vein quartz, fine-grained green lithic sandstone, rare shale, black chert and jasper. The matrix is composed of fine-grained mature sublithic sandstone with rounded to subangular framework grains of quartz and rock fragments of similar lithology to the clasts (Fig. 3.35g). A quartz cement is occasionally present.

**Interpretation:** The sedimentary and textural features described above are consistent with many of the criteria Ethridge and Wescott (1984) proposed for wave-worked gravels deposited in a beachface or shoreface setting. The lateral persistence and gross geometry of the facies also favours a beach or shoreface setting. However, poor sorting, sporadic angular boulders, and a general decrease in clast roundness in some areas suggests a transition locally to fluvial conditions. The thickness of the unit, together with the variety of extraformational clast lithologies, indicates a significant fluvial input (Nemec and Steel 1984), probably by high-gradient streams which drained areas of local basement to the immediate east. Deposition was mainly by wave swash, producing a seaward-dipping imbricate fabric, and by longshore currents. Clast lithotypes suggest derivation mainly from the underlying
Riversdale Volcanics with minor input from Ordovician (lithic sandstone, black chert) and Silurian (aphryic silicic volcanics, jasper) sources. The facies occurs above a minor disconformity within the Carwell Creek Formation, suggesting rapid uplift of source areas and the shoreline followed by regression. Polymict detritus was rapidly eroded from the source areas and delivered to the shoreline by high-gradient streams where it was redistributed by wave-related processes.

3.3.7.4.6 FACIES F: MEDIUM- TO THICK-BEDDED, CROSS-BEDDED QUARTZ-LITHIC SANDSTONE

Description: Facies F is common in the middle to upper parts of the Carwell Creek Formation as a thick monotonous sequence of quartz-rich sandstone and minor shale. North of GR 770710 6371550, Facies F forms the basal facies of the Carwell Creek Formation and can be traced from here to northeast of Mudgee (GR 744700 6401700) as the dominant facies. It exhibits highly gradational upper and lower contacts with Facies E, D or C (Fig. 3.31). The facies consists of medium- to thick-beded quartz to sublithic sandstone with thin interbeds of mudstone. Light brown, grey and purple sublitharenite and quartzarenite (after Folk 1974) are the dominant lithologies (Fig. 3.32) although more lithic sandstones were rarely recorded from the facies in the Brogans Creek area.

In thin section, the sandstones are tightly packed, poorly to moderately sorted, and submature. Subangular to subrounded framework grains (0.1-0.6 mm) are composed of volcanic and metamorphic quartz (72-93%) and lithic fragments (12-18%) of silicic volcanics, chert, and rare shale. The matrix is interstitial and is composed of chert and authigenic clays, micas and hematite. A quartz cement is occasionally present as overgrowths around the detrital grains. Lithic content fluctuates sporadically throughout the facies. Body fossils are very sparse and only rare crinoid ossicles and poorly preserved brachiopods were recorded.

In outcrop, the facies occurs in moderate to thick tabular beds with rare thin to moderate interbeds of grey shale. Beds are generally massive with sharp and planar to undulating bases and tops. Sandstone is typically fine- to medium-grained and sedimentary structures include (in order of decreasing abundance): parallel lamination in sets up to 30 cm thick, mainly near the base of the formation; symmetrical ripple marks with approximately meridional crest trends (Fig. 3.36); thin intervals of abundant silt-lined vertical burrows (probably Skolithos); lenses of mud chip breccia; and 10-35 cm sets of low- to moderate-angle tabular, and lesser trough, cross-beds. Overall, the latter show a bimodal pattern with mean currents to the northeast and south (Fig. 3.36). Locally, however, trends can vary from stacked unimodal cosets trending to the dominant south mode, to alternating southeast and northwest trending sets separated by prominent erosional surfaces. Zones of unimodal current ripple laminated (2-4 cm thick sets) sandstone up to 3 m thick occur, trending towards the dominant south mode. Hummocky cross-stratification was noted at GR 772420 6367270 and frequent lenses of unfossiliferous mudstone (5-20 cm thick) appear in the upper parts of the unit. A spaced cleavage is frequently well developed throughout the facies, particularly in outcrops to the west of Lue.
Figure 3.35 (a) Bedding plane exposure of Planolites sp. in fine-grained quartz-lithic sandstone (Facies F) adjacent to Cumber Melon Creek (GR 776210 6357610). (b) Intensely bioturbated symmetrical ripple marks in a calcareous lithic sandstone (Facies C), Cumber Melon Creek (GR 775940 6357650). (c) Facies E; tightly-packed beds of clast-supported pebble conglomerate (dominated by chert and silicic volcanic clasts) interbedded with medium- to fine grained quartz-lithic sandstone displaying low-angle tabular cross-bedding. Outcrop is south of the Cudgegong River at GR 771090 6365720. (d) Facies D; medium- to thinly-interbedded quartz-lithic sandstone and mudstone at GR 771390 6365940. Sandstone beds display load cast bases, symmetrical wave-ripped bed tops, and parallel-laminated or rarely hummocky cross-stratified internally; intervening mudstones are intensely bioturbated. These rocks are interpreted as storm deposits below the fairweather wave base. (e) Crinoidal lithic sandstone (Facies C) with small-scale tabular cross-bedding and parallel lamination. Outcrop in east of Brogans Creek Quarry at GR 776960 6344860. Hammer handle for scale. (f) Photomicrograph of a fine-to very fine-grained sublithic sandstone (Facies F) from the upper Carwell Creek Formation at GR 768910 6368740, displaying quartz (Q), silicic volcanic rock fragments, feldspar (F), in a clay-rich interstitial matrix with patchy quartz cement. Crossed polars. (g) Photomicrograph of coarse-grained quartz-lithic sandstone to granule conglomerate (Facies E) from GR 769610 6365680 dominated by grains of quartz (Q) and quartz-phyric silicic volcanic fragments (V). Crossed polars. (h) Unconformity between steeply dipping, thinly-bedded sandstone and mudstone of the upper Carwell Creek Formation (Facies D) and sub-horizontal pebble-cobble conglomerate of the Shoalhaven Group (Snapper Point Formation). Outcrop is east of Cumber Melon Creek at GR 776760 6357840.
Interpretation: The abundance of angle-of-repose cross-stratification and parallel lamination, the paucity of mudstone, and the presence of rare bioturbation and marine body fossils suggest a high energy marine environment above the fairweather wave base, probably on the foreshore to middle shoreface (Walker 1984). Facies F palaeocurrents record the migration of numerous small- to medium-scale, straight crested to lunate megaripples in longshore and lesser offshore (and rare onshore) directions (Fig. 3.36). The offshore trends are approximately normal to the general ripple crest trend (Fig. 3.36), suggesting the megaripples were driven offshore by wave-related rip currents, possibly enhanced by tidal currents. These conditions were occasionally replaced by southerly-directed longshore currents, probably generated by oblique wave approach from the north. Similar oblique wave approach from the north has been noted in palaeocurrents of the Roxburgh Formation (Section 3.3.5.6.7). Wave ripples capping beds support the interpretation that all sandstone facies were modified by, or were a product of, wave activity. Parallel-laminated sandstone was generated by combined-flow processes and/or intense bed shear due to storm wave activity (Nottvedt and Kreisa 1987) or swash processes. The intercalated mudstone and siltstone layers represent deposition of finer-grained sediment during periods of low energy or fairweather conditions. The rare occurrence of hummocky cross-stratification and long amplitude wave ripples indicates oscillatory-dominant storm currents were occasionally prevalent over unidirectional flows (Duke et al. 1991).

3.3.7.4.7 Facies G: Carbonate

Description: Limestone and dolomite (Dacl of Fig. 3) are present at numerous stratigraphic levels well above the base of the Carwell Creek Formation. In the Rylstone-Cudgegong district, several lenses of limestone up to 150 m long and approximately 50 m thick crop out close to the Cudgegong River (Fig. 1; GR 769610 6366590; 770080 6367500; 771250 6367210) at the 300, 350 and 450 m level of the sequence (Fig. 3.31b). These limestones tend to be thick-bedded, massive, and composed primarily of blue-grey micrite. Typically, the limestones in this area are unfossiliferous or sparsely fossiliferous, with rare non-silicified crinoid stems, ovoid stromatoporoids, gastropods and favositid corals. Millsteed (1985) recorded rare corals, ostracodes, bryozoans, brachiopods and crinoids from limestone at GR 770080 6367500. Horizons rich in bulbous stromatoporoids were noted at GR 769620 6366580. Recrystallisation is frequently well developed, creating a granoblastic mosaic of sparite. Dolomitisation is patchy, occurring mainly close to joint planes, and interbeds of grey shale, up to 25 cm thick, appear towards the top of the each lens.

At Brogans Creek (Fig. 2c), limestone is developed as three belts interbedded with Facies F and minor Facies C and D. The most extensive is the easterly belt centred around the main Brogans Creek quarry (GR 776480 6344550) which occupies the core and limbs of an anticline (Fig. 2c). The sequence here commences with shale which is disconformably overlain by very thick-bedded, massive limestone. The sequence fines, and bedding thins, grossly upwards through moderate- to thick-bedded
Figure 3.36 Palaeocurrent rose diagrams for facies of the Carwell Creek Formation.

blue-grey limestone with sparse stromatoporoids and favositids, to dark, thin-bedded, richly fossiliferous muddy limestone. The latter have yielded rich silicified shelly faunas (*Coelospira* cf. *dayi*, *Salopina*, *Leptostrophia*, *Howellella*, along with the stromatoporoid *Amphipora* sp., and indeterminate gastropods, crinoids, tentaculitids, and receptaculitids; identifications by A.J. Wright 1994) and conodonts (Section 3.3.7.5; Table 3.8). The limestone sequence grades upwards through crinoidal sandstone (Facies C) into massive shale and volcanic-rich sandstone.
Further west, around GR 775200 6345900, numerous small bodies of blue-grey limestone occur. The limestone is very thick-bedded and massive and dominated by dense coral and stromatoporoid communities at the base. This is overlain by a buff to pink dolomitic limestone which passes upwards into shale.

In the far west of the Brogans Creek valley, Facies G crops out in three areas to the north of the "Boori" property (Fig. 2c). Carbonates here comprise relatively pure, thick-bedded limestone with stylolites and silicification well developed throughout, particularly in the north (GR 776750 6344810). Faunas are largely dominated by tabulate and rugose corals in decreasing order of abundance (favositids indet., heliolitids indet., Cladopora sp., Alveolites sp., and ?Xystriphyllum sp.). A low diversity conodont fauna was obtained from this area (Table 3.8).

In the Mt Knowles district, 13 km east of Mudgee (around GR 754000 6390000), several carbonate lenses occur in the upper part of the Carwell Creek Formation. Detailed maps of the carbonates in this area have been produced by Wright (1966) and McClung (1969). Carbonates in this area were informally termed the "Mt Knowles Member" by McClung (1969) and Pemberton et al. (1994); however, the name Mt Knowles has subsequently been published as a Group name for the Late Devonian clastic sequence in the same area and is therefore not available (Colquhoun et al. 1997). The carbonates at Mt Knowles are present as two main bands. The most easterly (centred around GR 754050 6390250; Fig. 3) comprises approximately 60 m of massive, thick-bedded unfossiliferous dolomite with local birdseye textures preserved; a large dolomite quarry is developed on the unit at GR 754050 6390250. The second belt crops out extensively on the west side of Dolomite Lane between GR 753900 6390000 and GR 754400 6388500 and comprises three separate lenses of limestone up to 60 m thick and 300-450 m long (Fig. 3). The limestone lenses show two thinning- and fining-upwards cycles, commencing with massive thick-bedded blue-grey biomicrite with a low diversity tabulate coral, stromatoporoid, and receptaculitid fauna. Bedding becomes thinner and brachiopods and rare tetracorals become increasingly common up-sequence reaching a peak in thin beds at the top of each cycle (one adjacent to the Late Devonian unconformity) which are densely packed with brachiopods; important brachiopods include Spinella sp. and Malurostrophia sp. (Section 3.3.7.5). Silicification is characteristically well developed throughout.

The Buckaroo limestone deposits (the 'Archer Limestone Member' of Wright 1960) occur approximately 11 km northeast of Mudgee at GR 750750 6393200, along strike from the aforementioned Mt Knowles carbonates. Limestone occurs as three lenses (two of which have been quarried) up to 70 m thick and between 50 and 200 m long; the limestone is overlain unconformably by the Late Devonian Buckaroo Conglomerate. Thick-bedded, massive, unfossiliferous limestone with abundant spar-filled cavities resembling Stromactactis is the dominant lithology. At GR 750760 6393250, a horizon rich in gastropods, crinoids, favositid corals, and Syringopora and lamellar stromatoporoids was noted. Conodont samples from this locality were barren.
Interpretation: Carbonates of Facies G represent highly variable depositional settings although, in general, less restricted and better oxygenated conditions prevailed than those of the Glen­dale Limestone Member (Facies B). The carbonates lack mound or reef-like characteristics and are interpreted as autochthonous biostromal carbonate accumulations that developed in areas sheltered from clastic supply and storm waves, possibly in small embayments or, in some cases, lagoons. Faunas and limestone textures suggest a variety of conditions during deposition, ranging from sheltered, mud-rich, restricted lagoonal settings (Carwell Creek area); restricted intertidal flats (Mt Knowles dolomite lens); through restricted subtidal to intertidal shallow marine with low-diversity faunas (Buckaroo limestone deposits); to open, well-oxygenated, normal salinity subtidal conditions with more diverse faunas (most of Brogans Creek and Mt Knowles [west] carbonates). Facies sequences at Brogans Creek and Mt Knowles indicate up to 2 transgressive events occurred during the perbonus Zone; these transgressions are discussed further in Section 5.2. Dolomitisation is patchy and less pervasive than in the older Facies B and is only locally well developed at Mt Knowles.

3.3.7.4.8 Facies H: Massive Volcanic Sandstone

Description: This facies is largely restricted to one main area, identified by Cook (1988a), approximately 4 km south of the "Glendale" farmhouse (around GR 768240 6361200). The facies also occurs sporadically in the Brogans Creek area but was not studied in detail there. Facies H varies between approximately 20 and 40 m thick and persists along strike for approximately 400 m in the "Glendale" outcrops. It consists of massive or graded beds of fine- to medium-grained, crystal-rich dacitic volcanic sandstone (up to 1.5 m thick) with rare interbeds of unbioturbated shale. Beds are internally structureless apart from rare diffuse horizontal laminae. In thin section, the sandstone is poorly-sorted, immature, and composed of angular to very angular framework grains of quartz, plagioclase, silicic volcanic rock fragments (with vitriclastic textures), and rare glass shards in an interstitial matrix rich in clay minerals, chlorite and opaque oxides. The basal contact with Facies F is sharp and the facies appears to pass gradationally upwards into cross-bedded sandstone of Facies F, with the volcanic content diminishing rapidly away from the contact.

Interpretation: Cook (1988a) interpreted these rocks as non-welded ignimbrites. The association with Facies F suggests deposition took place in a nearshore environment and the lack of pumice or welding suggests subaqueous conditions. Two interpretations are possible for this facies: (a) rapid erosion of large volumes of immature volcanic detritus from the incipient volcanoes and deposition on swiftly prograding deltas and shorelines as high-density sandy sheetfloods or grainflows (Lowe 1982); or (b) similar thick, massive to graded volcaniclastic beds can be created by ignimbrites which entered shallow water and were transformed into water-supported mass flows of pyroclastic detritus (Cas and Wright 1987). In either case, the unfossiliferous volcanic sandstone of Facies H is interpreted to reflect
the onset of silicic volcanism proximal to the shoreline. This represents the youngest documented volcanic episode on the platform areas of the Capertee High.

3.3.7.5 Fauna and Age

Marine macrofossils are abundant in the basal Carwell Creek Formation and include crinoid ossicles, rare brachiopods, and poorly-preserved trilobites in the sandstone, along with bulbous stromatoporoids, bryozoans, gastropods, crinoids, rare favositid and rugose corals, and brachiopods in the carbonates (Millsteed 1985; Cook 1988a; and this study). However, age-diagnostic macrofossils are extremely rare. Abundant crinoid ossicles and rare brachiopods (Spinella sp.) were noted in crinoidal lithic sandstone at GR 769520 6365320 during this study. Cook (1988a) recorded the brachiopod Howellella sp. and an unidentified spiriferid from GR 769700 6362800, and Millsteed (1985) documented sparse shelly faunas from GR 771250 6367210, containing, inter alia, the brachiopods Buchanathyris sp. and Spinella sp. A sparse, low diversity, partially silicified fauna of corals (favositids, Rhizophyllum) and ovoid to lamellar stromatoporoids with very rare brachiopods (cf. Spinella sp.) was recorded in the Glendale Limestone Member east of "The Dolomites" (GR 768500 6367830). Conodonts are sparse and non age-diagnostic. Eight spot samples from these basal carbonates from the Rylstone-Cudgegong district were processed and all were barren. Sample C37 from silicified coralline and stromatoporoidal dolomite from "The Dolomites", north of Lake Windamere (GR 768500 6367830) contained a very sparse conodont fauna of an ozarkodinan Pb element and Oulodus sp. (Colquhoun 1995b).

The limestone at "Erang", near Windamere Dam, tentatively assigned to the Glendale Limestone Member (Section 3.3.7.4, Facies B), contains a moderately diverse silicified brachiopod and trilobite fauna which is described by Wright et al. (in prep; Supporting Paper 11). The "Erang" fauna contains a high proportion of new brachiopod and trilobite taxa. Common species include the trilobites Cyphaspis windamerensis sp. nov., Maurotarion erangensis sp. nov., Apocalymene sp. and Dudleyaspis sp. and the brachiopods Skenidioides johnsoni sp. nov., Aulacella boucoti sp. nov., Resserella rhodesi sp. nov., and Mesodouvillina savagei sp. nov., and the tetracoral Sociophyllum wilsoni sp. nov. Less common and generally poorly preserved taxa left in open nomenclature include: Salopina sp., Douvillina sp., Apocalymene sp., and Dudleyaspis sp. Occasional poorly preserved leptenid, chonetid, orthotetid, pentamerid, atrypid, spiriferid (?Cyrtina) and rhychonellid brachiopods also occur, as well as stromatoporoids, the tetracoral Rhizophyllum sp. and tabulate corals (Wright et al. in prep.; Supporting Paper 11). Several conodont samples (C41 and C41a from GR 761560 6374390 and C41b and C41c from GR 761460 6374380) were processed from the "Erang" limestone, all yielding a varied and reasonably preserved conodont fauna, despite the level of deformation. All samples contained similar rich, low diversity faunas (Fig. 3.38) dominated by simple cones (Panderodus unicostatus, P. sp., P. recurvatus, and P. valgus) as well as abundant Ozarkodina pseudomiae, O. buchanensis and less
common *Pandorinellina steinhornensis miae*, *Ozarkodina excavata excavata*, *Amydrotaxis* sp. and *O*. sp. (Colquhoun 1995b; Supporting Paper 5). In addition, limestone was intersected nearby during exploratory drilling for the construction of Windamere Dam. From borehole W37 (GR 759700 6375000), Pickett (1974) recorded mainly conodont faunas from the intervals 42.4 m-59.4 m and 85.3 m-85.9 m. In particular, the lower level yielded *Spathognathodus remscheidensis?* as well as silicified ostracodes; the conodont occurrence was used by Pickett to indicate a provisional Gedinnian age. Pickett (1976) noted the occurrence of corals (species indeterminate) in cores from borehole DDH 102 (GR 759500 6375200). It is likely, given the proximity and lithological similarity, that these limestones represent further subsurface occurrences of the "Erang" limestone (i.e. Facies B of the Carwell Creek Formation). If this is the case, *Spathognathodus remscheidensis?* may actually be referable to *Ozarkodina buchanensis*.

The paucity of conodonts near the base of the Carwell Creek Formation restricts inferences as to the age of the lower contact with the Riversdale Volcanics. The "Erang" fauna, although varied and reasonably well preserved, lacks polygnathids and the ranges (after Mawson et al. 1992) of *O. pseudomiae*, *O. buchanensis*, *O. e. excavata*, and *Pandorinellina steinhornensis miae*, suggest a broad *pireneas* to *perbonus* Zone age (late Pragian to early Emsian). The presence of the brachiopod *Spinella* sp. in crinoidal sandstone near the base of the Carwell Creek Formation suggests a *dehiscens* Zone age (Garratt and Wright 1988) but it has been suggested that this genus extends downwards at least to the late Pragian *pireneas* Zone in Victoria (Mawson et al. 1992) and also occurs, albeit less commonly, in the *perbonus* Zone. Hence the base of the Carwell Creek Formation is thought to lie either in the *pireneas* or *dehiscens* zones and cannot be located more accurately by conodont or macrofossil data at this time.

Carbonates higher in the sequence at Brogans Creek and Mt Knowles locally contain abundant silicified macrofaunas. At Brogans Creek, the fauna includes silicified brachiopods including *Coelospira* cf. *dayi*, *Salopina*, *Leptostrophia*, *Howellella*, along with the stromatoporoid *Amphipora* sp., and indeterminate gastropods, crinoids, tentaculitids, and receptaculitids (identifications by A.J. Wright 1994). In addition, *Xystriphyllum* sp., *Spinella* sp. and *Malurostrophia* sp. have also been recorded from the Brogans Creek limestone (Wright, pers. comm. 1994). Conodont faunas from Facies G limestone at Brogans Creek (Fig. 3.37; Table 3.8) are dominated by simple cones and *Pandorinellina exigua exigua*, with subordinate *Ozarkodina linearis*, *Oulodus* sp., *Pandorinellina exigua* cf. *philipi*, and *Polygnathus nothoperbonus* (1 specimen). At Mt Knowles, similar macrofaunas occur and include: *Donia* sp., *Lyrielasma* sp., *Receptaculites australis*, *Cymostrophia* sp., *Nadiastrophia* sp., *Acrospirifer* sp., *Spinella* sp., *Anoplotheca* sp. and *Cyrtina* sp. (Wright 1966, 1969). Conodont faunas at Mt Knowles are characterised by a great abundance of simple cones (*Panderodus* sp. and *P. unicostatus*) and *Pandorinellina exigua exigua* (Plate 2, Figs 5 and 16 of Supporting Paper 5) with lesser *Ozarkodina linearis* (Plate 1, Fig. 19, Plate 3, Figs 15, 18 and 19 of Supporting Paper
Figure 3.37. Conodont fauna from the Carwell Creek Formation in the Brogans Creek area and the Myrtle Grove Formation in the Capertee Valley. See Table 3.9 and text for grid references of sample localities.

**Pandorinellina exigua exigua** PHILIP 1966.
  a. Pa element, lateral view of AMF 96381; C72, x 69.
  b. Pa element, upper view of AMF 96381; C72, x 69.
  u. Pa element, lateral view of AMF 96291; C49a, x61.
  v. Pa element, lateral view of AMF 101545; C75, x56.
  y. Pa element, lower view of AMF 96291; C49a, x61.
  z. Pa element, upper view of AMF 101545; C75, x56.
  s. M element, AMF 101547; C67, x80.
  t. Sb element, AMF 101548; C67, x48.

**Pandorinellina cf. exigua philipi** KLAPPER 1969.
  n. Pa element, lateral view of AMF 101549; W57, x48.
  q. Pa element, upper view of AMF 101549; W57, x48.

**Ozarkodina buchanensis** PHILIP 1965.
  c. Pa element, lateral view of AMF 96371; C49a, x65.
  f. Pa element, upper view of AMF 96371; C49a, x65.
  g. Pa element, lateral view of AMF 101550; C49a, x52.
  j. Pa element, lower view of AMF 101550; C49a, x52.
  i. Pa element, lateral view of AMF 101551; W57, x56.
  l. Pa element, upper view of AMF 101551; W57, x56.

**Ozarkodina linearis** PHILIP 1966.
  e. Pa element, lateral view of AMF 96373; W57, x43.
  h. Pa element, upper view of AMF 96373; W57, x43.
  p. Pb element, AMF 101552; W57, x43.
  r. Sa element, AMF 101553; W57, x52.
  k. Sc element, AMF 101554; W57, x39.

**Polygnathus nothoperbonus** MAWSON 1987.
  w. upper view of AMF 96366; C67, x95.

**Oulodus** sp.
  m. AMF 101555; C67, x52.

Unidentified fish pieces
  o. AMF 101556; C72, x86.
  x. AMF 101557; C72, x65.
Figure 3.38 Conodont fauna from the Carwell Creek Formation at "Erang" near Windamere Dam (GR 761560 6374390).

**Ozarkodina pseudomiae MAWSON ET AL. 1992.**
- a. Pa element, lateral view of AMF 96379; C41a, x48.
- d. Pa element, upper view of AMF 96379; C41a, x48.
- o. M element, AMF 101559; C41a, x86.

**Ozarkodina cf. pseudomiae MAWSON ET AL. 1992.**
- g. Pa element, lateral view of AMF 101560; C41b, x69.
- j. Pa element, upper view of AMF 101560; C41b, x69.
- x. Sc element, AMF 101561; C41a, x35.

**Ozarkodina buchanensis PHILIP 1965.**
- c. Pa element, lateral view of AMF 96374; C41a, x65.
- f. Pa element, upper view of AMF 96374; C41a, x65.
- r. Sc element, AMF 101562; C41c, x39.
- t. Pb element, AMF 101563; C41b, x43.
- y. Sa element, AMF 101564; C41b, x61.

**Ozarkodina cf. buchanensis PHILIP 1965.**
- b. Pa element, lateral view of AMF 101565; C41a, x56.
- e. Pa element, lower view of AMF 101565; C41a, x56.

**Ozarkodina excavata excavata BRANSON & MEHL 1933.**
- v. Pb element, AMF 101566; C41a, x39

**Pandorinellina steinhornensis miae BULTYNCK 1971.**
- h. Pa element, lateral view of AMF 96386; C41a, x43.
- k. Pa element, upper view of AMF 96386; C41a, x43.
- m. Pa element, lateral view of AMF 101567; C41c, x43.
- q. Pa element, upper view of AMF 101567; C41c, x43.

**Amydrotaxis sp.**
- i. Pa element, lateral view of AMF 101568; C41d, x78.
- l. Pa element, upper view of AMF 101568; C41d, x78.
- w. Sa element, AMF 101569; C41d, x48.

**Panderodus valgus PHILIP 1965.**
- n. AMF 101570; C41a, x65.

**Panderodus recurvatus RHODES 1953.**
- p. AMF 101571; C41b, x78.

**Oulodus sp.**
- z. Pb element, AMF 101572; C41c, x52.

Unassigned elements.
- s. Pb element, AMF 101573; C41c, x52
- u. AMF 101574; C41a, x39.
Table 3.8 Spot sample localities and conodont faunas from the Brogans Creek area.

5), Oulodus sp., and rare Polygnathus nothoperbonus (Plate 1, Fig. 8 of Supporting Paper 5). Similar conodont faunas from the Mt Knowles carbonates were recovered by McClung (1969); furthermore Pickett (1972) noted "Polygnathus foveolatus" from limestones at Mt Knowles.

The rare presence of Polygnathus nothoperbonus in the upper parts of the Carwell Creek Formation suggests a perbonus Zone age (Mawson 1987), at least at this level. How much further up
into the Emsian the Carwell Creek Formation ranges probably varies from area to area, depending on the amount of erosion below the sub-Late Devonian unconformity. At Mt Knowles the unconformity occurs only 30 m above the highest occurrence of *Polygnathus nothoperbonus* whereas, at Brogans Creek, a considerable thickness of sediment is apparently present above its occurrence, suggesting the unit may range up into a higher zone.

3.3.7.6 Environments of Deposition

The Carwell Creek Formation represents a thick accumulation of shallow marine clastics and carbonates deposited in environments ranging from storm-dominated inner to middle shelf to a sheltered, restricted lagoon. The setting for deposition and palaeogeography is depicted in Fig. 3.39.

Deposition commenced in the latest Pragian (\( ? \) pirenae Zone) probably during a widespread rise in eustatic sea-level (Section 5.2). The basal portions of the Carwell Creek Formation represent a retrograding barrier island complex with an approximately meridional-trending shoreline (Fig. 3.39). The sequence progressively onlapped Late Silurian basement east of Carwell Creek. The barrier island sequence was composed, in ascending order, of: basal volcaniclastic deposits, sheltered restricted lagoon, barrier and shoreface sands, and offshore environments. The back barrier environments are represented by the Glendale Limestone Member (Facies B) and were deposited in a sheltered, hypersaline lagoon in generally subtidal conditions. The lagoon received sporadic inputs of volcanic detritus from rivers and influxes of more quartz-rich detritus from storm washovers on the barrier. The barrier (Facies C) was composed of sand rich in crinoid debris and lithic detritus and preserved environments of deposition include foreshore and tide-dominated shoreface. These pass upwards into a storm-dominated middle to inner shelf clastic succession (Facies D).

The retrograding barrier island configuration outlined above was terminated prior to deposition of Facies E by a rapid regression which was probably caused by uplift of the source areas (to the east) and the shoreline. Facies E represents a storm-dominated gravelly beach-shoreface deposit where coarse detritus from newly uplifted source areas was brought to the shoreline by high-gradient rivers and redistributed by longshore wave currents and wave swash. The supply of coarse detritus quickly diminished and Facies E was succeeded by Facies F, a thick sequence of quartz and sublithic sandstone deposited in wave-dominated, tide-influenced shoreface and offshore settings. The thickness of Facies F and the lack of any discernible trends in bed thickness or structures suggest a delicate balance of subsidence and sediment input was maintained, leading to a long period of facies aggradation punctuated by only brief pulses of progradation (Fig. 3.31). Brief periods of volcanism occurred in the Carwell Creek area, resulting in influxes of immature volcanic detritus to the shoreline (Facies H). Clastic input was reduced into the shelf setting in about the *perbonus* Zone and a series of carbonate bodies (Facies G) were deposited in highly variable environments ranging from sheltered, restricted lagoons (suggesting a return, in some areas, to a barrier island configuration) to well-oxygenated open
marine conditions with normal salinity. Wave-dominated clastic deposition resumed, probably late in the *perbonus* Zone, and continued for an unknown length of time into at least the middle Emsian.

In summary, the Carwell Creek Formation comprises an approximately 500 m thick sequence of lithic to quartz sandstone, limestone, dolomite, and shale deposited initially in a transgressive wave- and tide-influenced barrier island setting, and later in wave-dominated shoreface and offshore settings.

3.4 WESTERN PLATFORM MARGIN SEQUENCES (QUEENS PINCH GROUP AND LIMEKILNS GROUP)

3.4.1 Introduction

Strata deposited along the western margin of the Early Devonian Capertee High include (Figs 3, 3.2, and 4.1): (a) sequences of the Queens Pinch Group which occur in a largely fault-bounded belt
centred 16 km south of Mudgee (Wright 1966; McCracken 1990); (b) sequences of the Limekilns area, 17 km south-southeast of Sofala (Packham 1968; Voorhoeve 1986; Watkins and Pogson 1994); (c) a small area north of Ilford (this study); (d) minor outcrops in the Palmers Oakey area (Fergusson 1979; Bischoff and Fergusson 1982); and (e) around Flirtation Hill and north of Mudgee (Galloway 1961; Chatterton and Wright 1986; Colquhoun et al. 1997). All units in these areas are assigned to the Queens Pinch Group (Colquhoun et al. 1997) with the exception of the Limekilns sequence which is assigned to the Limekilns Group (Watkins and Pogson 1994) (Fig. 4.1). Only one unit, the Kingsford Formation, is described here. Further description of the Queens Pinch Group and Limekilns Group units, and correlation of these units with the Kandos Group and Hill End Trough units, is contained in Chapter Four.

3.4.2 Kingsford Formation

3.4.2.1 Synonymy and Definition

Limestone of possible Devonian age was first noted adjacent to the Mudgee Road, just north of Ilford, by Carne and Jones (1919). Day (1951) mapped two units in this area, the Westwood Limestone and Terri Hi Hi Limestone, and described Early Devonian faunas from both; the surrounding elastics were not described in detail. The original Dubbo 1:250 000 Sheet of Offenberg et al. (1971) included these rocks in undifferentiated Devonian strata. Booth (1990) carried out mapping in the area and obtained a large Early Devonian shelly fauna from elastics; she correlated these rocks with the Carwell Creek Formation of the Rylstone-Cudgegong district. Further mapping was carried out during this study and conodont faunas were obtained from the limestone (documented in Colquhoun 1995b; Supporting Paper 5). The conodonts confirmed the unit as older than the Carwell Creek Formation and the new name Kingsford Formation was suggested by Pemberton et al. (1994; Supporting Paper 2). The unit takes its name from the "Kingsford" property to the north of Ilford (GR 765210 6355300). No type section was suggested by Pemberton et al. (1994). A suitable representative section from GR 766050 6352320 to GR 767200 6353300 includes a traverse along Cunningham Creek.

3.4.2.2 Distribution, Areal Extent and Thickness

The Kingsford Formation occurs in three small areas on either side of the Sydney-Mudgee Road between 2 and 3 km to the north of Ilford (Fig. 2b). Total areal extent of the unit is approximately 1.6 km².

Thickness of the unit is difficult to estimate. At least 250 m is exposed in the representative section to the south of the "Hillcrest" property; however, the base is nowhere exposed and the top is probably faulted (Section 3.4.2.3).
3.4.2.3 Structure and Stratigraphic Relationships

Throughout most of its outcrop, the Kingsford Formation dips and youngs to the west-southwest at moderate angles. Local east dips are present close to the upper contact of the unit; however, mesoscopic folding was not observed. A strong planar slaty cleavage is commonly present in most lithologies. Some allochthonous limestones within the unit can be recognised as they exhibit anomalous east-west strikes of bedding at marked angles to bedding in the surrounding clastics.

The base of the unit is not exposed and the eastern limits are obscured by the overlapping Permian Shoalhaven Group. The western margin of the unit is in contact with quartz sandstones of the Roxburgh Formation of the Kandos Group. This contact is probably faulted, because: (a) bedding and younging in the Kingsford Formation are to the west-southwest whereas bedding and younging in the Roxburgh Formation are predominantly to the east (Fig. 2b), suggesting the contact may be a faulted fold hinge; and (b) local outcrops of brecciated mudstone and ironstone mark the line of contact between the two units.

3.4.2.4 Petrography and Facies

The Kingsford Formation has been subdivided into three sedimentary facies.

Facies A: Volcaniclastic Turbidites

**Description:** In thin section, these rocks are fine- to coarse-grained, submature volcanic litharenites (after Folk 1974), composed entirely of angular quartz and plagioclase crystals, and silicic volcanic rock fragments. Beds are 2 cm to 2 m thick, laterally continuous, and possess sharp and planar bases and planar to gently undulating tops. Bed thickness and internal structures show considerable variation. In the east (around GR 767100 6353320), the sandstone contains angular limestone clasts (to 20 cm across) and is thick-beded, very coarse-grained, massive, and ungraded. Further west (around GR 766550 6352940), beds are still thick (50 cm-2 m), massive and ungraded, but are now composed of medium-fine sand and occasionally show sand volcanoes on their upper surfaces and vague dish structures internally (cf. S3 bed of Lowe 1982). Small channels (<50 cm wide), filled with rounded granules of silicic volcanics, occasionally cut through these beds. Along Cunningham Creek, towards the western extremity of outcrop (GR 766540 6352320), beds are thinner (2-30 cm), interbedded with shales, and variably massive to normally graded throughout (S3 beds; Lowe 1982), or passing upwards into parallel and wavy laminations or rare current ripple lamination (Tabc partial Bouma sequence). The latter were not exposed in three dimensions, thereby preventing accurate measurement; however, they suggest a general flow to the west. Clasts or small pods of richly fossiliferous limestone are locally present and the finer sandstones yielded a rich shelly fauna (Fig. 3.40a, b) (Section 3.4.2.5).
Interpretation: The sandstone lacks evidence of hot emplacement (e.g. welding, columnar jointing) and is interpreted as a sub-storm-wave base deposit of cold, high- to low-density turbidity currents (Lowe 1982). Sandstone provenance, together with meagre palaeocurrent data, are consistent with derivation from the coeval silicic volcanics to the west. Bedding thins upwards and sedimentary structures suggest more distal settings up-sequence; this trend may be due to autocyclic processes, such as channel infilling or gradual lateral lobe shifting, or allocyclic processes, such as a rise in sea-level (Walker 1992). The angular nature of the detritus indicates limited reworking and probable derivation from slumping of unstable silicic volcanic detritus. Well-rounded granules/pebbles in the conglomerate may suggest more prolonged erosion of the volcanics in a nearshore or fluvial environment, and the presence of angular limestone clasts indicates erosion of shelf-edge limestones and incorporation into the turbidity currents.

Facies B: Limestone blocks and limestone breccia

Description: Limestone blocks ranging from 5 m to 200 m across crop out prominently to the south of the "Hillcrest" property and just north of "Old Westwood" (Fig. 2b). The limestone blocks and breccia typically comprise massive, light coloured limestone composed of tabulate corals (mainly favositids) and lamellar to ovoid stromatoporoids (Fig. 3.40e) or, less commonly, massive unfossiliferous dolomitic limestone to dolomite. Many of these limestone blocks display brecciation towards their margins, and show a transition from coherent limestone, through clast-supported breccia, to matrix-supported breccia with angular to subrounded, irregularly-shaped limestone clasts of granule to boulder dimensions (Fig. 3.40e). A large limestone block (200 m across) occurs to the south of the "Hillcrest" homestead and has been quarried at GR 766150 6352810. It consists of moderate- to thick-bedded, dark grey muddy limestone with abundant silicified twiglike stromatoporoids (Amphipora sp.) (Fig. 3.40c). This block also displays areas of limestone breccia, predominantly towards its margins (Fig. 3.40d). Bedding in the limestone strikes east-west, almost at right angles to bedding measured in nearby clastics.

Areas composed predominantly of limestone breccia occur around GR 766400 6352950. This breccia is poorly-sorted and composed of sparsely fossiliferous angular limestone clasts of granule to boulder dimensions in a calcareous sand matrix (Fig. 3.40f). The limestone breccia occurs as 10 cm to >8 m thick, disorganised and ungraded beds which are laterally persistent at outcrop scale. Bed bases are sharp and planar to gently undulating, whereas bed tops are either sharp and planar, and pass directly to another breccia bed, or, in about 50% of exposures, are capped by a graded calcareous sandstone layer. Internal deformation of the smaller limestone clasts is common.

Interpretation: The limestone blocks possess many features suggesting allochthoneity (e.g. brecciated margins, bedding discordant with surrounding sediments, association with turbidite facies) and are interpreted as allochthonous blocks emplaced by gravity sliding. The limestone facies suggest
Figure 3.40 (a) Fine- to medium-grained volcanic sandstone with 10-25 cm thick, richly fossiliferous layers with limestone clasts and abundant moulds of brachiopods, crinoids, tabulate and rugose corals and trilobites. Outcrop is 100 m west of the Sydney-Mudgee Road at GR 766540 6352320. (b) Close-up of the fossiliferous layers in (a) showing abundant irregularly shaped limestone clasts in a matrix of fine volcanic sandstone. (c) Dark grey to black limestone rich in non-silicified stromatoporoids (?Amphipora sp.). Outcrop is part of a large probably allochthonous block at GR 766150 6352810. (d) Limestone breccia composed of massive dark blue, unfossiliferous, angular limestone clasts of granule to cobble dimensions in an interstitial matrix of fine calcareous sandstone. Outcrop is towards the margin of a large limestone block at GR 766100 6352910. (e) Limestone breccia at the margin of a 5-6 m diameter allochthonous block composed of irregularly shaped light grey limestone clasts with large bulbous stromatoporoids and favositid corals in a matrix of fine brown calcareous sandstone. Breccia grades into unfragmented limestone towards the base of the photo. Outcrop is beside the Sydney-Mudgee Road at GR 766620 6352650. (f) Poorly-sorted limestone breccia composed of sparsely fossiliferous clasts of granule to boulder dimensions (GR 766400 6352950).
derivation from a variety of environments: blocks rich in colonial reef-building organisms (corals, stromatoporoids) were derived from lithified frontal reef facies, whereas the dark micritic limestone, dominated by twiglike stromatoporoids, was probably derived from lithified to partially-lithified back-reef or lagoonal facies.

The limestone breccia beds possess many features indicative of an unstable marine slope or base-of-slope setting adjacent to a carbonate platform. These include: conformability with other slope sediments, matrix-supported fabrics, and capping calcareous sandstone layers (Cook et al. 1972; Coniglio and Dix 1992). The depositional mechanisms by which the limestone clasts are transported downslope are poorly understood, but are thought to be a combination of debris flow and grain flow processes (Coniglio and Dix 1992). The matrix represents a mixture of clastic slope sediments and fine-grained breccia material. True debris flows have a mud and water matrix characterised by a combined strength and buoyancy that allows the flow to carry clasts that are denser than the bulk density of the flow itself (Lowe 1982). However, the presence of a non-cohesive granular matrix in these breccia beds suggests turbulence and grain interactions were important transport mechanisms and that deposition may have been mostly by grain flows.

Similar limestone breccia beds, capped by calcareous turbidites, have been widely documented (Krause and Oldershaw 1979), with the calcarenite interpreted as either separate events, or the dilute upper portions of the carbonate debris flows/grain flows (Coniglio and Dix 1992). Such flows are thought to reflect the catastrophic collapse of oversteepened fore-reef slopes, perhaps triggered by earthquakes or tsunamis (Cook et al. 1972).

**Facies C: Dark Mudstone**

*Description:* Dark grey to black mudstone occurs as 2-18 m thick units which crop out only sporadically and tend to weather recessively. Best exposures occur in a road cutting at GR 766560 6352610. The mudstone is typically massive or finely laminated, locally pyritic and generally devoid of body or trace fossils, with the exception of some minor poorly preserved bioturbation at GR 766720 6352910 which is probably referable to Zoophycos sp.

*Interpretation:* The facies represents slow settling from suspension of hemipelagic mud in a quiet, sub-storm-wave base environment. The paucity of body or trace fossils and the local abundance of pyrite, indicates that dysaerobic to anoxic bottom waters or substrate – perhaps combined with relatively rapid suspension sedimentation – created conditions unfavourable for the establishment of benthic communities. Sparse, high-density, low-diversity burrow systems are typical of such oxygen-depleted environments.
3.4.2.5 Fauna and Age

Limestones of the unit are dominated by macrofaunas with long age ranges. Most limestones are light coloured and dominated by favositid corals and stromatoporoids; Day (1951) recognised numerous favositid species in these limestones, although most of these species are no longer in general use. A sparse, non-age-diagnostic conodont fauna was obtained from one of these limestones (sample 52; GR 766650 6352540) and consisted of Panderodus sp. and Ozarkodina sp (Fig. 3.41). One limestone just south of the Hillcrest homestead (GR 766150 6352810) contained abundant silicified Amphipora but was barren of conodonts. A thin fossiliferous horizon in fine volcanioclastic sandstone beside the Mudgee Road (GR 766540 6352320) contained a rich benthic fauna from which Wright (in Booth 1990) identified trilobites (Paciphacops sp., Scutellum sp., Crotalocephalus sp., and Harpidella sp.), and the brachiopod Boucotia australis, in addition to indeterminate clams, gastropods, hyolithids, tentaculitids, echinoderm plates, plant fragments and isolated nautiloid septa. This horizon contains some minor pods of fossiliferous limestone (sample 56) from which the following conodonts have been recovered (Colquhoun 1995b): Eognathus sulcatus cf. kindlei (1 specimen), Ozarkodina remscheidensis repetitor, Ozarkodina sp., Pandorinellina exigua philipi, P. steinhornensis miae, P. sp., Pandorodus unicostatus, P. recurvatus, P. sp., and Bellodella resima, in addition to numerous scolecodonts and probable fish fragments (Fig. 3.41).

The macrofauna includes the brachiopod Boucotia australis, suggesting correlation with the 'late' Lochkovian australis Zone of Garratt and Wright (1988). The trilobite genera are also common in Lochkovian and Pragian strata of the Garra Limestone (Chatterton et al. 1979). The age of the conodont fauna relies mainly on a single, slightly broken specimen of Eognathus sulcatus which appears constricted in its basal cavity and is tentatively referred to E. sulcatus kindlei, suggesting a kindlei Zone (middle Pragian) age. However, as the specimen could be either E. s. sulcatus or E. s. kindlei, the Kingsford Formation can only be assigned with certainty to the sulcatus-kindlei intervals. The other species are known from these zones, although Ozarkodina remscheidensis repetitor is reported from the eurekaensis to pesavis Zones (Klapper and Johnson 1980), it extends up to at least to the sulcatus Zone in Australian sequences (Bischoff and Argent 1990).

3.4.2.6 Environment of Deposition

The Kingsford Formation is interpreted as a volcanioclastic/carbonate slope to base-of-slope deposit. The alternating predominance of mass-flow and hemipelagic deposition together with the marine fauna, suggests a marine slope environment which was consistently below storm wave base. The clastic sequence shows evidence of small- to medium-scale sedimentary cycles which suggest channel infilling or lobe migration, suggesting a slope fan setting rather than an apron. Salient elements of the depositional setting were:
Figure 3.41. Conodont fauna from minor limestone pods in volcaniclastic turbidites of the Kingsford Formation (GR 766540 6352320; sample 56 and GR 766650 6352540; sample 52).

*Eognathus sulcatus* cf. *kindlei*. (broken specimen)

  a. upper view of AMF 96365; 56, x78.

*Bellodella resima* PHILIP 1965.

  b. AMF 101575; 56, x104.

*Ozarkodina remschiedensis repetitor* CARLS & GANDL 1969.

  c. Pa element, lateral view of AMF 96372; 56, x110.

  d. Pa element, upper view of AMF 96372; 56, x110.

*Pandorinellina exigua philipi* KLAPPER 1969.

  e. Pa element, lateral view of AMF 96380; 56, x61.

  g. Pa element, upper view of AMF 96380; 56, x61.

*Ozarkodina* sp.

  f. Pa element, lateral view of AMF 101576; 52, x104

  h. Pa element, upper view of AMF 101576; 52, x104

*Panderodus unicostatus* BRANSON & MEHL 1933.

  i AMF 101577; 56, x80.
(a) a steep, possibly tectonically-controlled, west-dipping slope, backed by a high-relief dacitic volcanic terrain to the east.

(b) a narrow shelf, with fringing carbonate banks and shoals, incised by numerous fast-flowing rivers which deposited coarse volcaniclastic detritus onto the upper slope. Subsequently the volcanic detritus was remobilised by turbidity currents of low to high density, whereas the carbonate detritus was transported downslope by grain flows or debris flows. Numerous large allochthonous limestone blocks were displaced from the shelf and probably transported by gravity sliding (cf. Conaghan et al. 1976).

(c) steady offshore settling of hemipelagic mud in dysaerobic to anaerobic conditions.

Water depths of the slope are difficult to quantify, although a total lack of wave-formed sedimentary structures suggests deposition took place entirely below the maximum storm-wave base (10-204 m water depth in modern oceans; Elliot 1986). The trace fossil Zoophycos is generally taken to suggest bathyal depths; however, it is known from both shallow and deep water environments in the Palaeozoic (Ekdale 1988). The lack of autochthonous limestones, in situ body fossils, the evidence of oxygen depletion and appreciable slopes in the depositional setting, are indicative of a relatively deep water setting – probably greater than 200 m.

The unit is of similar age to the Riversdale Volcanics of the Kandos Group, and has a silicic volcanic provenance which suggests derivation from this unit. The Kingsford Formation is therefore considered to be a proximal slope equivalent of the Riversdale Volcanics and also correlates with the Mullamuddy Formation of the Queens Pinch Belt which was deposited in a similar environment at the same time (Section 4.3). Further correlations between the Capertee High and Hill End Trough sequences are discussed in Chapter Four.
CHAPTER FOUR
REGIONAL EARLY DEVONIAN CORRELATIONS

4.1 INTRODUCTION

The Kandos Group in the Rylstone-Cudgegong district contains the most complete and thoroughly studied Early Devonian sequence on the Capertee High, and is therefore the most suitable location from which to infer correlations. A comprehensive review of the relevant literature, and a number of new radiometric dates, have revealed numerous possible and probable correlatives, although the variability in the quality and amount of data from area to area frequently hampered the precision of correlations. This section concentrates on correlating the Kandos Group with other Early Devonian units of the Capertee High, its margins, and the Hill End Trough, and comparing Capertee High sequences with those of the Molong High. For convenience, correlations or comparisons are divided into 3 timespans: (1) Lochkovian; (2) Pragian; and (3) Emsian. These are further subdivided into: (a) platform sequences from lesser known parts of the Capertee High, possible extensions of the Capertee High, or the Molong High; (b) sequences along the western margin of the Capertee High; and (c) units of the Hill End Trough succession. Correlations have been synthesised in Fig. 4.1.

4.2 LOCHKOVIAN

4.2.1 Platform sequences

4.2.1.1 Capertee High

Only broad interregional correlations to the Lochkovian Kandos Group (Warrah Conglomerate to Roxburgh Formation) are possible as proven or suspected Lochkovian platform strata are unknown on the Capertee High outside the Rylstone-Cudgegong district.

In the Yerranderie-Windellama region, approximately 100 km directly south of the last Capertee High exposures, the 2.5 km thick Lochkovian-Pragian Bindook Volcanic Complex provides the fill for the Wollondilly Basin (Simpson 1990); the eastern margin of the Wollondilly Basin may represent a southerly extension of the Capertee High (Powell 1984). The Bindook Volcanic Complex resembles coeval parts of the Kandos Group, although individual units are not correlatable over such long distances. The Tangerang Formation, the mostly Lochkovian basal unit of the Bindook Volcanic Complex (Jones et al. 1984), comprises shallow marine shelf to subaerial volcanioclastic arenite, conglomerate, chert, shale, limestone and silicic pyroclastics and lava (Jones et al. 1984; Simpson 1990); these lithologies, depositional environments and provenance resemble the Lochkovian portion of the Kandos Group. The Tangerang Formation was characterised by shallowing up-section, culminating in a period of subaerial exposure, and the increasingly proximal influence of explosive silicic volcanism. In addition, the Windellama Limestone Member spans the eurekaensis and delta Zones (possibly
### Devonian

#### Early Devonian

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### References

1. Booth (1990)
2. Colquhoun (1995a)
3. Pemberton et al. (1994)
extending downwards into the uppermost Silurian; Mawson 1986) and shows macrofaunal affinities with the Clandulla Limestone (Mawson 1973; 1975).

4.2.1.2 Molong High

Lochkovian sequences on the Molong High, the western bounding element of the Hill End Trough, are less clastic and more carbonate dominated than coeval units of the Capertee High (Fig. 4.2). The Camelford Limestone crops out on the Molong High between Laras Lee and Cumnock and conformably overlies the Late Silurian Barnaby Hills Shale of the Mumbil Group. The limestone consists of a 300 m thick sequence of limestone conglomerate, calcisiltite, and coarse crinoidal calcarenite and contains a fauna of silicified stromatoporoids, crinoid stems, tabulate and rugose corals and brachiopods (Chatterton et al. 1979). The brachiopods include species of Skenidiodes, Dalmanella, Isorthis, Howellella, Atrypa, Anatrypa, Schuchertella, Howellites, and Schizoporia (Farrell 1992). Conodont data indicate the Camelford Limestone extended from the latest Silurian eosteinhornensis Zone to the Lochkovian eurekaensis Zone (Farrell 1996). The limestone is therefore of similar age to the Clandulla Limestone from the Capertee High and contains a similar, though more diverse, macrofauna.

The Cuga Burga Volcanics extend more than 60 km from north of Wellington to south of Kerrs Creek. The unit attains a maximum thickness of 800 m southwest of Stuart Town and thins both southwards along the Molong High and eastward into the western Hill End Trough. The formation is composed of breccia, conglomerate, and sandstone of mafic to intermediate volcanic detritus; mudstone; and basalt lava flows and shallow intrusions (Morton 1974). Facies and thickness variations can be explained in terms of growth, emergence and final decay of a marine volcanic pedestal centred near Wellington (Morton 1974). Recent conodont dating of the underlying Camelford Limestone and the overlying Garra Limestone constrains the age of the Cuga Burga Volcanics to the Lochkovian delta Zone (Farrell 1996). The Capertee High lacks a coeval sequence of mafic volcanics; however, geochemically similar basalt lava flows occur in the early Pragian Mullamuddy Formation (Section 4.3.2) and recent SHRIMP zircon dating by the Geological Survey of New South Wales has revealed numerous mafic to intermediate intrusions of Early Devonian age (404 Ma: diorite in theLate Ordovician Burrannah Formation; 397 Ma: dolerite in theLate Silurian Chesleigh Formation). This suggests a potentially prolonged period of bimodal mafic/silicic volcanism lasting throughout the Early Devonian with the mafic component often manifested as intrusions and rare flows close to the margins of the highs.

The Garra Limestone is a predominantly carbonate unit 900-1200 m thick which outcrops meridionally for over 100 km from Geurie in the north to Cudal in the south. The unit conformably overlies the Cuga Burga Volcanics in the north (near Wellington) and the Maradana Shale or the Berkley Formation (formerly the Mandagery Park Formation) around Manildra in the south (Mawson et al. 1988). The Maradana Shale is dominated by grey shale and thin limestone beds, and contains
abundant brachiopod faunas (Savage 1968) which have been correlated by Garratt and Wright (1988) with their 'early' Lochkovian *janaea* Zone. The unit has also yielded *eurekaensis* Zone conodonts (Mawson *et al.* 1988). The Berkley Formation conformably overlies the Maradana Shale and consists of well-bedded tuffaceous sandstone with minor siltstone, shale and fossiliferous limestone deposited in a high energy shoreline setting deepening upwards to a quiet open marine setting (Savage 1968). The Berkley Formation contains a rich brachiopod fauna (Savage 1974) and has yielded *delta* Zone conodonts (Mawson *et al.* 1988). The unit was regarded as a basal tongue of the Garra Limestone by Mawson *et al.* (1988) and Farrell (1992) and appears to be equivalent, at least in part, to the Camelford Limestone. The Maradana Shale and Berkley Formation are of similar age to the Clandulla Limestone and Yellowmans Creek Formation on the Capertee High which were deposited in similar settings and contain similar faunas.

The overlying *Garra Limestone* contains extensive coral, brachiopod and trilobite faunas (Chatterton *et al.* 1979; Lenz and Johnson 1985). Conodont studies indicate the base of the limestone is markedly diachronous, with lowest horizons yielding *delta* Zone conodonts at Eurimbla in the south (Sorentino 1989), and *pesavis* Zone conodonts farther north, near Wellington Caves (Wilson 1989). *Polygnathus dehiscens* has been recorded from the upper parts of the Garra Limestone (Philip and Pedder 1967) indicating possibly continuous carbonate sedimentation through 5 conodont zones, from the middle Lochkovian to the basal Emsian (Mawson and Talent 1992). This suggests the Molong High in this interval was a shallow marine platform characterised by very low clastic input and high levels of carbonate production. This contrasts markedly to the corresponding interval on the Capertee High where the equivalent sequence is dominated by thick sequences of shallow marine siliciclastics of the Roxburgh Formation and subaerial dacitic pyroclastics and epiclastics of the Riversdale Volcanics.

**4.2.2 Western Platform Margin Sequences**

Deeper water equivalents of the Lochkovian Kandos Group occur at Palmers Oakey and Queens Pinch. The limestone-mudstone breccia facies at Palmers Oakey is a post-middle Lochkovian mass-flow breccia with clasts of mudstone, fine quartzose siltstone, limestone (to 100 m across) and lesser quartzose sandstone and silicic volcanics (Bischoff and Fergusson 1982). The limestone clasts contain conodonts ranging in age from Wenlock to middle Lochkovian (early *delta* Zone). Recent mapping by the Geological Survey of New South Wales (Watkins and Pogson 1994) included these hitherto unnamed sequences in the Waterbeach Formation of the Crudine Group. The clast lithologies and the minimum age of the limestones suggest derivation from the lower Kandos Group (or unexposed equivalents) along with areas of Silurian silicic volcanics and limestone.

The basal parts of the Mullamuddy Formation of the Queens Pinch Group may extend down into the late Lochkovian based on macrofaunas; the lower fauna in the unit was correlated by Garratt and Wright (1988) with their 'late' Lochkovian *australis* Zone. However, the oldest conodonts obtained
from the unit are early Pragian (sulcatus Zone) in age. Correlations of this unit with Pragian platform and trough units are discussed in Section 4.3.

4.2.3 Hill End Trough Sequences

The Crudine Group of the Hill End Trough comprises approximately 2600 metres of turbiditic sandstone and shale of mainly silicic volcanioclastic provenance with rare concordant volcanic horizons. Stratigraphy of the Crudine Group has been discussed by Packham (1968, 1969), Glen and Watkins (1994), Pogson and Wyborn (1994), and Colquhoun et al. (1997). Correlating these widespread units with the lower Kandos Group, which only crops out in the Rylstone-Cudgegong-Ilford area, is difficult as there were undoubtedly other unpreserved or presently unexposed platform areas of the Capertee High which were contributing sediment to the Hill End Trough during the Lochkovian and influencing the composition of the Crudine Group. In addition, recent mapping has shown the Crudine Group to change markedly throughout the Trough, and some of its constituent units are difficult to recognise in the most northern parts of Hill End Trough.

The Cookman Formation consists of up to 400 m of slate and thinly interbedded fine- to medium-grained, quartz to sublithic sandstone turbidites and rare beds coarse-grained sandstone to conglomerate. Rare beds of volcanolithic sandstone occur at irregular intervals throughout the unit (D. Pogson pers. comm. 1995). Flute casts on the base of overturned beds in Cheshire's Creek indicate a palaeo-flow from the east-southeast, indicating derivation from the Capertee High (Packham 1968). The Cookman Formation thins steadily towards the north, and north of GR 741250 6369850 the Turondale Formation rests directly on the Chesleigh Formation. The Cookman Formation contains sparse, non-age-diagnostic plant fossils; the unit was placed in the earliest Lochkovian after recent work on the Bathurst 1:250 000 Sheet, although Packham (1968, 1969) regarded the unit as latest Silurian (Pridoli). The unit is therefore possibly equivalent to the basal Kandos Group (?Warrah Conglomerate) but is so different lithologically that no correlation can be attempted. More likely, the unit reflects a period of erosion on the Capertee High around the Silurian-Devonian boundary, indicated by a widespread Silurian-Devonian unconformity. Packham (1968) regarded the unit as derived from a diverse source terrain which included silicic volcanics (probably the Silurian Windamere Volcanics and Dungeree Volcanics), quartz-rich sediments (Early Ordovician Adaminaby Group), mafic volcanics (Late Ordovician Sofala Volcanics or Coomber Formation) and granite (?; no pre-Devonian granites are known from the Capertee Zone).

The Turondale Formation conformably overlies the Cookman Formation and comprises up to 900 m of thick-beded volcanioclastic and quartz-lithic turbidites of fine sand to pebble conglomerate size, along with tuffaceous mudstone and concordant rhyolitic porphyry (probably thick sills). A U/Pb SHRIMP age of 411.3 ± 3.4 Ma was obtained from a porphyry at the Turon River on the Bathurst 1:250 000 Sheet to the south, whereas zircons from sandstones in the upper part of the unit yielded a
SHRIMP age of 409.5 ± 4.2 Ma (E. Jagodzinski, written comm. 1995); these sills must have been approximately syn-sedimentary given the overlap in the error range (Jagodzinski et al. in prep.). On the Devonian timescale of Young (1995), this suggests a basal Lochkovian age; however, the base of the Devonian will be lower in the next timescale, probably at 415 Ma (G. Young, written comm. 1996).

The unit is largely unfossiliferous, although a coarse matrix-supported conglomerate from the basal parts of the Turondale type section contains silicic volcanic and limestone clasts; Packham (1968) recorded the corals *Favosites* sp. and *Tryplasma* sp. and pentamerid brachiopods from the limestone clasts. A sample of these clasts submitted by E. Jagodzinski was processed by the author for conodonts; however, only one fragment of a simple cone (*?Panderodus* sp.) was obtained and is not age-diagnostic. The presence of limestone clasts rich in pentamerid brachiopods, however, suggests derivation from Silurian, rather than coeval Lochkovian, limestone.

The type section of the Turondale Formation shows a roughly threefold subdivision (Packham 1968): the basal third is dominated by very thick-bedded volcaniclastic sandstone and conglomerate; the middle is mainly siltstone; and the upper third is dominantly lithic sandstone derived from a quart-zose and sedimentary basement provenance. However, this subdivision does not really hold outside the type area. Elizabeth Jagodzinski has measured 8 sections through the unit: four in the Sofala area and another four around the Hill End anticline. Her data suggest the Sofala area was more proximal to the source of the volcaniclastics, with the volcaniclastic horizons being thicker and generally coarser than the Hill End anticline area further west. Many of the thick-bedded volcaniclastic sandstones were probably deposited by voluminous high-density turbidites of silicic volcanic detritus which may have developed directly from subaerial pyroclastic flows passing into the ocean (Cas and Wright 1991; E. Jagodzinski pers. comm. 1996). In addition, Packham (1968) has recorded palaeocurrents flowing to the west-southwest, suggesting the Turondale Formation was derived from an active volcanic source on the Capertee High.

The Turondale Formation is difficult to correlate with any exposed volcanic unit on the Capertee High, particularly given the current state of flux of the Devonian timescale, and is probably equivalent to the lower Kandos Group (ie Warrah Conglomerate, Clandulla Limestone, Yellowmans Creek Formation). The Turondale Formation appears older than the Riversdale Volcanics (Section 3.3.6.5); however, it may be related to the Huntingdale Volcanics, a dacitic pyroclastic sequence which is a probable correlative of the Riversdale Volcanics and crops out extensively in the Capertee Valley to the east of Sofala (Section 4.3.1.1). No base for the Huntingdale Volcanics is exposed in the Capertee Valley and it is possible this unit may extend well down into the Lochkovian and the lower part could possibly be coeval with the Turondale Formation. This interpretation is consistent with the aforementioned palaeocurrent data and proximality trends indicating a source on the Capertee High east of Sofala. Furthermore, the Turondale Formation thins to the north and loses much of its coarser, juvenile volcanic component (E. Jagodzinski, pers comm. 1996), which also supports a source on the southern
part of the Capertee High. The presence of ?Silurian limestone clasts and quartz-lithic detritus indicates erosion of Silurian and Ordovician lithologies on the Capertee High also contributed sediment to the Turondale Formation.

The median SHRIMP U/Pb ages for the upper Turondale Formation and the basal Merrions Formation (Section 4.3.3) are very similar, although error bars in both dates leave time for a possible 2 to 4 Ma break between the two units, and sandstones of the two units are petrographically identical (E. Jagodzinski, pers comm. 1996). This suggests that both units were related to the one volcanic source (the Riversdale/Huntindale Volcanics; Section 4.3.3), with the intervening Waterbeach and Guroba Formations (see below) representing a quiescent interval in Riversdale eruption. This, in turn, would imply the base of the Devonian is older than 410 Ma - the age of the Devonian base on the timescale of Young (1995) - as the Riversdale Volcanics and Merrions Formation are both constrained by fossil evidence to the Pragian. This implies that the 409 Ma dates from the basal Merrions Formation and the upper Turondale Formation would be in the Pragian - and the base of the Pragian was placed at 404 Ma by Young 1995.

The Turondale Formation passes conformably into the overlying Waterbeach Formation. In type section, this approximately 600 m thick unit consists of 120 m of shale to siltstone at the base passing up into a thick sequence of shale interbedded with lithic-rich 30 cm to 3 m thick turbiditic sandstone beds. The Dunmoogin Formation, a unit mapped widely in the northern Hill End Trough by Offenberg et al. (1971), was shown by the recent mapping to be a junior synonym of the Waterbeach Formation and this name has been suppressed (Colquhoun et al. 1997). The Guroba Formation – a unit mapped extensively in the northern Hill End Trough by Offenberg et al. (1971) and defined to the north of Burrendong Dam – was found to occur widely to the north of GR 747100 6355510 as a distinctive sandy facies between the Waterbeach Formation and the Merrions Formation (Fig. 3); the unit has also been mapped widely on the Euchareena 1:100 000 Sheet to the immediate west. The Guroba Formation consists of massive to thin-bedded turbiditic sandstone of volcanic and basement lithic provenance, and mudstone.

At Limekilns (22 km south-southeast of Sofala; Fig. 2.5), the Waterbeach Formation contains graded siltstone and shale, minor lithic greywacke and conglomerate (Packham 1969). Wright (1966, 1967) recorded a (?transported) brachiopod fauna in this area at Paling Yards, close to the top of the Waterbeach Formation. The fauna comprises the brachiopods Dolerorthis sp., Skenidiodes sp, Isorthis sp., Schizophoria sp., Plectadonta sp., Notanoplia sp., ?Schelwienella sp., Eospirifer sp., ?Ivanothyris sp., Spinatrypa sp., Lissatrypa lenticulata and the coral Pleurodictyum sp. Garratt and Wright (1988) correlated the fauna to their 'late' Lochkovian australis Zone and the fauna contains many genera common to the Yellowmans Creek Formation and Roxburgh Formation of the Kandos Group. The lithological similarity with these units, and the presence of palaeocurrents indicating derivation from the east (Packham 1968), suggests the basal Waterbeach Formation may be a deeper water
Figure 4.2 Time-space plot of the Early Devonian sequences of the Molong High. Data from time-space plot on Bathurst 1:250 000 Sheet, Mawson et al. (1988), Mawson and Talent (1992), Farrell (1996), and E. Morgan (pers. comm. 1996).

<table>
<thead>
<tr>
<th>Late Silurian</th>
<th>Lochkovian</th>
<th>Pragian</th>
<th>Emsian</th>
<th>Middle Devonian</th>
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<tr>
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<td>Delta</td>
<td>Pesavus</td>
<td>Perisoma</td>
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<td>Euretidae</td>
<td>Pesavus</td>
<td>Sulcatus</td>
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Middle Devonian
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<th>Molong High North</th>
</tr>
</thead>
<tbody>
<tr>
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<td>Garra Limestone</td>
</tr>
<tr>
<td>Berkley Formation</td>
<td>Nubrigyn Formation</td>
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<tr>
<td>Maradana Shale</td>
<td>Tolga</td>
</tr>
<tr>
<td>Cuga Burga Volcanics</td>
<td>Calcarenite</td>
</tr>
<tr>
<td>Mafic Intrusions</td>
<td>&quot;Red Hill Fan&quot;</td>
</tr>
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<td>Wansey Formation</td>
<td>Camelford Limestone</td>
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<td></td>
<td>Hanover Formation</td>
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<tr>
<td></td>
<td>Barnaby Hills Shale</td>
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equivalent of the Yellowmans Creek Formation whereas the sandy quartz-lithic rich Guroba Formation may be equivalent to the Roxburgh Formation.

4.3 PRAGIAN

4.3.1 Platform Sequences

4.3.1.1 Capertee High

In the central and eastern Capertee Valley, over 1000 m of Early Devonian welded and non-welded silicic ashflow pyroclastics, and immature intraformational sandstone, conglomerate and shale crop out (Fig. 4.1). The similarity of these rocks to the Riversdale Volcanics was first noted by Wright (1966) and various names have been subsequently assigned to them; herein, these volcanics are referred to as the Huntingdale Volcanics after Bracken (1977). In lithology and stratigraphic position, the Huntingdale Volcanics are strikingly similar to the Riversdale Volcanics (Fig. 4.1). Kennedy (1976) and Bracken (1977) inferred a Pragian age based on fossils from overlying units; furthermore, conodonts obtained from the overlying Myrtle Grove Formation during this study strengthen this interpretation (Section 4.9.2.3). In addition, Kennedy (1976) and Bracken (1977) inferred deposition on a gently west-sloping shelf, with dominantly subaerial deposition proximal to the volcanic source in the east, passing westwards into thinner sequences emplaced and reworked in shallow to marginal marine settings; this interpretation closely parallels that of the Riversdale Volcanics (Section 3.3.6.6). The large thickness of ashflow tuffs also suggests proximal caldera-related emplacement, although mapping has not yet been sufficiently detailed to elucidate other supportive facies or trends.

The Huntingdale Volcanics pass conformably and gradationally upwards into approximately 350 m (maximum) of thick-bedded, structureless dacitic sandstone, pebble conglomerate with dacite clasts, and lesser quartz sandstone and interbedded shale (Genowlan Formation and Emu Swamp Formation of Bembrick et al. 1969; the Emu Swamp Formation of Kennedy, 1976; and the Watervale Formation of Bracken 1977) (Fig. 4.1). Wright (1967) recorded the brachiopods Spinella cf. Buchananycnis and cf. Buchanathyris sp. from this unit, suggesting a late Pragian to early Emsian (pirenea to dehiscens Zone) age. This unit can be correlated on age, stratigraphic, and lithologic grounds with the immature dacitic sandstone and conglomerate (Facies B and Facies C) which are common near the top of the Riversdale Volcanics (Sections 3.3.6.4.2 and 3.3.6.4.3). Kennedy (1976) and Bracken (1977) similarly interpreted the Emu Swamp/Genowlan Formation as representing rapid erosion of the underlying volcanics and a sedimentary source, and deposition on a west-sloping, shallow to marginal marine shelf during a period of marine transgression (compare with Section 3.3.6.6). The much greater thickness of the horizon in the Capertee Valley may indicate an earlier cessation of volcanism and more vigorous and prolonged erosion prior to, and during, transgression. This suggests that silicic volcanism on the Capertee High was diachronous, possibly commencing earlier (Section 4.2.3) and ending earlier in the Capertee Valley than in the Rylstone-Cudgegong district.
Silicic pyroclastic volcanism was particularly active in the northeast Lachlan Fold Belt during the Pragian. The Tangerang Formation (see Section 4.2.1.1) was replaced, probably during the late Lochkovian to early Pragian, by initially fluvial-lacustrine deposition and finally by major subaerial silicic ignimbrite eruptions, forming extensive volcanic plateaux and calderas of the upper Bindook Volcanic Complex in the Yerranderie-Windellama region (Simpson 1990). This period of time coincides with the transition from the mostly shallow marine siliciclastics of the Roxburgh Formation to the major subaerial silicic volcanism during deposition of Riversdale and Huntingdale Volcanics. Rapid erosion of these subaerial volcanics led to deposition of a thick mass-flow volcaniclastic sequence (the Kowmung Volcaniclastics of Cas et al. 1981) in deeper water to the north and west, similar to the correlation between the Riversdale/Huntingdale Volcanics of the Capertee High and the Merrians Formation of the Hill End Trough (Section 4.3.3).

4.3.1.2 Molong High

The Molong High during the Pragian was dominated by the shallow marine platform carbonates of the Garra Limestone (Section 4.2.1.2). In marked contrast to the Capertee High, the Molong High was volcanically quiet during the Pragian: the silicic pyroclastics of the Bulls Camp Volcanics were considered to be Pragian by Packham (1969). However, recent U/Pb SHRIMP dating of this unit by the Geological Survey of New South Wales yielded an age of 412 Ma and the unit has been placed in the Late Silurian Mumbil Group (M. Scott, pers. comm. 1996).

During the Pragian and Emsian, a number of carbonate debris flow lobes developed along the eastern margin of the Molong High, probably derived from erosion of the extensive platform carbonates of the Garra Formation during intervals of regression and platform exposure (Mawson and Talent 1992). Lithologically, these fans contrast dramatically with coeval deposits along the western margin of the Capertee High (Section 4.3.2) which are dominated by volcanic detritus shed from the Capertee High. The Tolga Calcarenite is a sequence of flaggy calcarenite (alloplanic limestone) in beds up to 1 m thick interbedded with grey shale, calcilutite and rare horizons of calcirudite (Conaghan et al. 1976). It unconformably overlies the Cuga Burga Volcanics and contains a coral fauna of probable Pragian age (Packham 1969). Detailed sampling of the unit for conodonts is currently being undertaken by Ruth Mawson and John Talent. Morton (1974) lists a 'Lochkovian' conodont fauna from the Tolga Calcarenite. The changes in conodont taxonomy and ranges since this time make it difficult to reassess the age of Morton's fauna, although the occurrence of 'Spathognathodus philipi' (= Pandorinellina exigua philipi) and 'Ozarkodina remsheidensis' (= O. remsheidensis remsheidensis) suggest a pesavis to sulcatus Zone age (ie. latest Lochkovian to early Pragian). A similar sequence referred to as the Red Hill Fan by Mawson and Talent (1992) (the Red Hill Formation of Conaghan et al. 1976) was deposited during the pireneae Zone of the latest Pragian (Mawson and Talent 1992).
4.3.2 Western Platform Margin Sequences

In areas marginal to the Capertee High, and in the Hill End Trough, there are a number of deeper water units of the Queens Pinch Group which were possibly derived from the Riversdale Volcanics and its Capertee Valley equivalents. In the Queens Pinch district (Fig. 2.2), the late Lochkovian-Pragian Mullamuddy Formation contains conglomerate and sandstone with clasts/grains of silicic pyroclastics and volcaniclastics, coeval and Silurian limestones, and vesicular basalt (Wright 1966; McCracken 1990). McCracken (1990) identified a sulcatus-kindlet Zone (early to mid Pragian) fauna from a conglomerate with volcanic and limestone clasts low in the Mullamuddy Formation; the fauna included *Eognathus sulcatus* eta morph (Murphy, Matti and Walliser), *E. sulcatus* kappa morph (*E. sulcatus juliae*), *Pedavis* n. sp. B (Klapper and Philip) and *O. e. excavata* and *Pandorinellina exigua philipi*. McCracken (1990) interpreted the Mullam HDDd Фо Formation as a deep marine volcani-

The Taylors Hill Formation conformably overlies the Mullamuddy Formation and comprises up to 240 m of dark mudstone, thin-bedded volcanic sandstone, lenses of basaltic conglomerate and fossiliferous, massive to well-bedded, allochthonous and possibly allogenic limestone (Wright 1966, 1969). Faunas were obtained by Wright (1966) from the base and near the faulted top of the unit. Both faunas can be correlated (due to the presence of *Nadiastrophia* and the tetracoral *Martinophyllum*) with the *Nadiastrophia-loyolensis* assemblage of Garratt and Wright (1988), which spans most of the Pragian. Garratt and Wright (1988) also recorded a sulcatus-kindleti Zone (early to middle Pragian) conodont fauna comprising *Eognathus* cf. *sulcatus*, and *Pedavis* sp.. Depositional environments differed from those of the Mullamuddy Formation with relatively prolonged periods of hemipelagic suspension deposition and only minor turbidity currents of coarser detritus (Pemberton *et al.* 1994; Supporting Paper 2). The decrease in volcanic material may indicate a cessation of volcanism, and
possibly higher sea levels on the platform areas of the Capertee High, suggesting the unit may be equivalent to the uppermost Riversdale Volcanics (Fig. 4.1).

The ?middle Pragian Kingsford Formation occurs in a small area to the north of Ilford where it is faulted against the Roxburgh Formation; it has been described in Section 3.4.2. It is composed of volcaniclastic turbidites, dark mudstone, and allochthonous limestone blocks and breccia and is considered to be a proximal slope equivalent of the Riversdale Volcanics and has been placed in the Queens Pinch Group (Colquhoun 1995b; Supporting Paper 5).

4.3.3 Hill End Trough Sequences

The Merrions Formation (formerly the Merrions Tuff) crops out extensively in the Hill End Trough and comprises approximately 450 m of thick-bedded volcaniclastic sandstone, shale, and concordant silicic porphyries interpreted as cold, subaqueous mass-flows; background hemipelagic sedimentation; and subaqueous lavas respectively (Cas 1977, 1978, 1979). Early isochron dating suggested only a broad Early Devonian age (395 Ma +/- 32 Ma; Cas et al. 1976). Suspected lavas in the unit yield zircons with a SHRIMP age near the base of the Devonian (411+/-2 Ma: Pogson and Wyborn 1994). However, several recent U/Pb SHRIMP dates put the Merrions Formation between 408 Ma and 405 Ma (Jagodzinski et al. in prep.), placing the unit in the late Lochkovian to early Pragian on the timescale of Young (1995) and Jones (1996).

In addition, the basal beds of the overlying Cunningham Formation have yielded a poorly preserved shelly fauna which may correlate with the late Pragian Taylors Hill Formation fauna of the Queens Pinch Group (Wright 1994), and the underlying Waterbeach Formation contains a "late" Lochkovian shelly fauna (see Section 4.2.3). Furthermore, the unit is overlain by the ?late Pragian-early Emsian "Rosedale Shale" in the Limekilns area (Section 4.4.2). The unit is therefore constrained biosstratigraphically mainly to the Pragian, suggesting a correlation with the Riversdale Volcanics and Huntingdale Volcanics (Table 4.1). This is supported by palaeocurrent data which indicate particularly the middle to upper parts of the unit were derived from coeval silicic volcanics to the east, northeast and southeast (Cas 1977).

Cas and Wright (1991) suggested many of the voluminous cold-state turbidites which were responsible for the thick-bedded volcaniclastic sandstones of the Merrions Formation may have developed directly from subaerial ignimbrite eruptions passing into the water column. Such an interpretation accords well with the depositional settings of the Riversdale and Huntingdale Volcanics: both sequences are dominated by subaerially deposited ignimbrites which show evidence of shallow marine incursions, especially in the west.
4.4 EMSIAN

4.4.1 Platform Sequences

4.4.1.1 Capertee High

Correlatives of the Emsian Carwell Creek Formation are very widespread on the Capertee High - a fact first noted by Wright (1966) and one which has important implications for the palaeogeography of this interval (Section 5.4.4). To the south, in the eastern and central Capertee Valley, a similar thick Emsian sequence of cross-bedded lithic sandstone, calcarenite, calcareous shale, biostromal and bioclastic limestone (often impure with volcanic detritus), and mudstone conformably overlies the Emu Swamp/Genowlan Formation. This unit has been referred to as the Myrtle Grove Formation by Bembrick et al. (1969) and Kennedy (1976), and the Carinya Formation by Bracken (1977). Macrofaunas from the limestone include the Spinella-Buchcmathyris brachiopod assemblage, with subordinate Howittia and Adrena (Wright 1966; Kennedy 1976) and the corals Xystriphyllum mitchelli, Trapezoophyllum and Martinophyllum (Wright 1966). Limestone occurs most abundantly in the middle to lower parts of the sequence. Six spot samples were processed from recent road cuttings along the Capertee-Glen Davis Road. Of these, only two contained conodonts: sample C49a (GR 728930 6329470) produced a small fauna of Ozarkodina buchanensis, O. linearis, Pandorinellina exigua exigua, and Panderodus sp.; sample C91 (GR 829430 6329460) had O. buchanensis and O. linearis (Fig. 3.36) (Colquhoun 1995b; Supporting Paper 5). In addition, Bischoff in Bracken (1977) recorded a fauna of Pandorinellina e. philipi, P. e. exigua, P. s. miae and Ozarkodina linearis from limestone in Coco Creek (GR 828260 6327160). Kennedy (1976) recorded Ozarkodina buchanensis from limestone at "Blue Rocks" (GR 831300 6322300) in the far south of the Capertee Valley. Ranges of conodonts suggest a late Pragian to early Emsian (pireneae to perbonus Zone) age - similar to the Carwell Creek Formation. Furthermore, the Myrtle Grove Formation shows facies variation comparable to the Carwell Creek Formation: limestones are mainly confined to the basal and middle parts of the sequence and thin towards the east with a concurrent increase in volcanic and calcareous lithic sandstone in the same direction (Kennedy 1976; Bracken 1977). The latter authors proposed deposition took place on a broad, west-sloping shallow marine shelf adjacent to an eroding volcanic source, with shallow, restricted intertidal conditions in the east grading to a deeper, open marine environment in the west - an interpretation, again, similar to the Carwell Creek Formation (Section 3.3.7.7).

The Carwell Creek Formation can be traced well to the north of its type area in the Rylstone-Cudgegong district - as far as the Eurundury-Pipeclay Creek area, northeast of Mudgee. The Mount Frome Limestone, a sequence of richly to sparsely fossiliferous limestone deposited in shallow subtidal to intertidal environments, occurs 5 km southeast of Mudgee (Fig. 3) on the eastern limb of the Pine Ridge syncline and occupies the same stratigraphic position as the Carwell Creek Formation in the Mt. Knowles area. The Mount Frome Limestone is between 178 m (Wright 1966, 1981) and 238 m (Pickett 1978, 1987) thick, depending on whether separate outcrops near the Cudgegong River are included.
The unit consists of beds of dark blue-grey calcarenite, from several cm to at least 6 m thick, separated by thick, sparsely fossiliferous, silty calcareous intervals. The Limestone contains rich tabulate and rugose coral and subordinate brachiopod faunas; Wright and Flory (1980) recognised seven successive faunal assemblages. A biostromal depositional environment, rich in food and oxygen, was envisaged by Wright (1966). Birdseye textures are present, typically in the middle and lower parts, indicating ephemeral intertidal or supratidal conditions. Conodont studies (McClung 1969; Philip 1974; Pickett 1978, 1987; Mawson et al. 1985; Wright 1981 and unpublished data, summarised in Garratt and Wright 1988; Strusz et al. 1988) collectively indicate the Mt Frome Limestone spans most of the Emsian with ages ranging from at least *perbonus* Zone (late early Emsian) for the basal beds, possibly extending into the earliest Eifelian (*partitus* Zone) in the uppermost 25 to 30 m. *Polygnathus perbonus* occurs in the lowest 20 m of outcrop (Pickett 1987; Garratt and Wright 1988) and Pickett (1987) recorded *Polygnathus inversus* in a thin unit just below unit A of Wright (1981). Garratt and Wright (1988) recorded *Polygnathus serotinus* throughout the basal 130 m of the main outcrop of the limestone. The upper 38 m of the limestone are dominated by *Eognathus bipennatus* and *Pandorinellina expansa*; polygnathids are rare in the upper parts of the limestone although Philip (1974) recorded *Polygnathus costatus patulus* and *Bipennatus bipennatus* from the uppermost beds.

Conodont data therefore indicate the basal parts of the Mt. Frome Limestone correlate with the Carwell Creek Formation. The upper parts (*serotinus* Zone and younger) are younger than any unit recognised in the Kandos Group in the Rylstone-Cudgegong district or Capertee Valley areas (Fig. 4.1).

The Mt Frome Limestone is overlain by the Boogledie Formation, a unit of crystal-rich felsic lithic sandstone (frequently crinoidal), shale, pebbly sandstone and conglomerate which was deposited in wave-dominated shallow marine conditions (Colquhoun et al. 1997; Supporting Paper 8) (Fig. 4.1). The Boogledie Formation is the youngest unit of the Kandos Group and presumably is early Middle Devonian (early Eifelian), although the unit lacks age-diagnostic fossils. Recent mapping by the Geological Survey (Colquhoun et al. 1997) has redefined the Boogledie Formation to include only the clastics which overly the Mt. Frome Limestone in the Mt. Frome area, and not the older (early Emsian) clastics and carbonates of the Mt. Knowles-Buckaroo area which are now thought to be part of the Carwell Creek Formation. The Boogledie Formation cannot be correlated outside its type area around Mt. Frome (Fig. 4.1).

### 4.4.1.2 Molong High

As mentioned in Section 4.2.1.2, the Garra Limestone extends up into the basal Emsian *dehiscens* Zone and, based on conodont studies of allochthonous carbonates along the eastern margin of the Molong High, Mawson and Talent (1992) concluded that carbonate sedimentation continued on the Molong Platform for three more conodont zones - until the late Emsian *serotinus* Zone.
The Nubrigyn Formation consists of approximately 400 m of allochthonous limestone blocks (up to several hundred metres in diameter), hemipelagic mudstone, allogenic carbonate and megabreccia with clasts of limestone, intermediate volcanics, quartzite, shale and mudstone (Conaghan et al. 1976). The unit intertongues eastward with lithic sandstone of the basal Cunningham Formation of the Hill End Trough sequence, and it overlies both the Cuga Burga Volcanics (Section 4.2.1.2) and the Tolga Calcarenite (Section 4.3.1.2). Conaghan et al. (1976) interpreted the Nubrigyn Formation as a debris flow-dominated fan derived from erosion of the Garra Limestone and the Cuga Burga Volcanics and transported east to slope settings along the eastern margin of the Molong High. Conodont studies of the matrix and clasts of the Nubrigyn Formation are ongoing (R. Mawson, pers. comm. 1996). Packham (1968) listed a macrofauna which has a general early Emsian aspect and conodonts listed included *Spathognathodus (= Ozarkodina) linearis*, a species not known for certain in pre-Emsian strata (Mawson et al. 1992). The Nubrigyn Formation has similar lithology, age and stratigraphic position to the Sutchers Creek Formation and Jesse Limestone Member which developed along the western margin of the Capertee High in the Emsian (Section 4.4.2); the same regressive event, or events, may have been responsible for erosion of platform carbonates on the Molong and Capertee High. However, any firm correlations must await publication of the conodont data from the Nubrigyn Formation.

### 4.4.2 Western Platform Margin Sequences

The Warratra Mudstone (*Dqw*; Fig. 3) was originally defined by Wright (1966) as 120 m of grey mudstone underlying the Sutchers Creek Formation. McCracken (1990) altered the definition to include limestone conglomerate, lithic sandstone and mudstone which Wright had placed in the overlying Sutchers Creek Formation. Here, I recognise the two formations in their original sense. The Warratra Mudstone has a sparse macrofauna including, in particular, such brachiopods as notanopliids and chonetids but especially rare *Spinella* sp.; the latter is important in indicating a *pireneae* to *dehiscens* Zone (late Pragian to earliest Emsian) age (Garratt and Wright 1988; Mawson et al. 1992). A very similar unit of mostly dark shale and mudstone, the Rosedale Shale, occurs in the Limekilns area, to the south (Fig. 4.1). The Rosedale Shale (Packham 1968; now the basal part of the Limekilns Formation of Pogson and Wyborn 1994) contains the graptolite *Monograptus yukonensis* (Packham, in Strusz 1972) and the dacryoconarid *Nowakia acuaria* (Wright and Haas 1990), suggesting a late Pragian age. The shale is strongly bioturbated locally, suggesting the sediment/water interface was not completely anaerobic but probably dysaerobic. Local hardgrounds, and condensed horizons rich in bioturbation and shelly debris alternate with comparatively barren mudstone sequences (A.J. Wright, pers. comm. 1996), indicating a quiet, sub-storm wave base environment characterised by slow hemipelagic deposition and sediment starvation over long periods. The Rosedale Shale is approximately coeval with, and lithologically similar to, the Warratra Mudstone; both units are approximately coeval with the upper Taylors Hill Formation (Fig. 4.1). Each of these units indicate an environment comparatively starved of
coarse detritus and limestone and appear to coincide with a major transgression on the platform areas of the high at the inception of the Carwell Creek Formation, which cut off supply of clastic detritus to basinal areas (Section 5.2).

There are a number of similar correlative units in the Western Platform Margin sequences between *perbonus* and *serotinus* Zone in age. Collectively, these units indicate at least two periods of erosion of platform sequences at this time.

The Sutchers Creek Formation (*Dsq*; Fig. 3) consists of up to 180 m of limestone conglomerate and blocks, usually commencing with a coarse turbidite unit; pebbly mudstone intervals and alloclastic limestone are equally characteristic of the unit, which conformably overlies the Warratra Mudstone (Wright 1966; McCracken 1990). Wright (1966, 1969) listed extensive tetracoral and brachiopod faunas. McCracken (1990) documented an early *perbonus* Zone conodont fauna of *Polygnathus dehiscens dehiscens* and *Polygnathus nothoperbonus* from alloclastic limestone in his "Warratra Formation" - a unit equivalent to the Warratra Mudstone and basal Sutchers Creek Formation of Wright (1966) and Offenberg *et al.* (1971). *Eognathus sulcatus* theta morph (Murphy, Matti, and Walliser) and *Polygnathus pireneae* have also been recorded in the clasts of carbonate breccias in the upper part of the unit (McCracken 1990), suggesting erosion of Pragian platform-edge limestones. The Sutchers Creek Formation contains abundant *perbonus* to *serotinus* Zone conodont faunas from several stratigraphic levels (McCracken 1990). Near the base of his Sutchers Creek Formation (which is different to Wright's (1966) original definition; see above), McCracken (1990) recorded a late *perbonus* Zone fauna which included *Polygnathus nothoperbonus* and *Polygnathus perbonus*. Towards the top of the unit, McCracken (1990) has recorded *serotinus* Zone faunas from alloclastic limestone; conodonts include *Polygnathus serotinus* delta morph, *P. pseudoserotinus*, *P. inversus*, and *Pandorinella cf. steinhornensis steinhornensis*. In addition, Garratt and Wright (1988) have recorded *Polygnathus perbonus* from the Sutchers Creek Formation.

Recent mapping has extended the outcrop of the Sutchers Creek Formation north from its type area to the southern outskirts of Mudgee (Fig. 2). At Flirtation Hill, on the outskirts of Mudgee, minor outcrops of limestone boulders in a mudstone matrix occur. The limestone clasts have yielded a fauna of silicified trilobites, rare brachiopods and the conodont *Polygnathus inversus* (Chatterton and Wright 1986). In addition, Mawson and Talent (in McCracken 1990) reported *Polygnathus serotinus* and *Polygnathus pseudoserotinus* and Mawson and Talent (1992) also documented *perbonus* Zone clasts.

Outcrops north of Mudgee formerly known as the Tinja Formation have been placed in the Sutchers Creek Formation (Colquhoun *et al.* 1997) (Fig. 3). The unit at this locality comprises shale, mudstone, volcanioclastic turbidites, pebbly mudstone, allochthonous limestone blocks and diorite (the latter of unknown age, but suspected to be similar to the diorite in the Burranah Formation which was dated at 404 Ma). A shelly fauna recovered from shale on Snakes Creek Road (GR 743500 6412500) has affinities with the fauna from the Sutchers Creek Formation, suggesting a middle to late Emsian
(perbonus to serotinus Zone) age (Wright 1995). Limestone blocks in the unit were recently sampled by the Geological Survey of New South Wales during the mapping of the Mudgee 1:100 000 sheet; the results have not yet been finalised but preliminary data indicate both Silurian and Early Devonian blocks are present. In addition, Exon (1962) recorded an Early Devonian coral fauna from these limestone blocks.

The Ingleburn Formation (Dqi; Figs 2 and 3) of the Queens Pinch Group crops out as a fault-bounded block which contains 550 to 600 m of basal dacite lava, fossiliferous mudstone, conglomerate (with dacite clasts), pale green andesite, cross-bedded quartz sandstone and conglomerate (with dacite and quartz sandstone clasts) at the top (Wright 1966, 1969). Coral and brachiopod faunas occur towards the middle of the mudstone interval. Original age determination based on the macrofauna (Wright 1966, 1969) suggested a position in the sequence between the Taylors Hill Formation and the Warratra Mudstone (ie. late Pragian-earliest Emsian). Garratt and Wright (1988) subsequently recorded inversus Zone (middle Emsian) conodonts, thus indicating partial equivalence with the Sutchers Creek Formation. Dacite lava from GR 754222 6371751 in the Ingleburn Formation yielded a U/Pb SHRIMP age of 403 ± 8 Ma (Colquhoun et al. 1997).

At Limekilns, the Jesse Limestone Member of the Limekilns Formation (Pogson and Wybom 1994) is a mass-flow breccia comprising clasts of limestone and lesser silicic volcanics in a sandy biomicrite matrix, along with allodapic limestone, and hemipelagic mudstone (Voorhoeve 1986; Wright and Chatterton 1988). Conodonts from the matrix are of late Emsian (serotinus Zone) age (Mawson et al. 1990; Mawson and Talent 1992), whereas the limestone clasts contain conodonts of early Emsian (dehiscens and perbonus Zones) age (Mawson et al. 1990); rare halysitid corals suggest a Silurian age for some of the clasts (Wright and Byrnes 1980). The age of the limestone clasts and matrix suggests derivation dominantly from lithified Carwell Creek Formation and Myrtle Grove Formation, along with Silurian limestones, and deposition on slopes marginal to the Capertee High late in the Emsian. The lithology and age indicate the Jesse Limestone Member is a partial correlative of the Sutchers Creek Formation.

4.4.3 Hill End Trough Sequences

The ungrouped Cunningham Formation (Dn Fig. 3) is the youngest unit of the Hill End Trough and consists of over 3000 m of mainly slate and thinly bedded volcaniclastic sandstone with horizons of pebble-granule conglomerate. It crops out extensively in tightly folded sequences in the southwest corner of the Mudgee 1:100 000 sheet and widely on the adjoining Euchareena 1:100 000 sheet and Bathurst 1:250 000 sheet. The unit has yielded a poorly preserved shelly fauna near its base which may correlate with the late Pragian Taylors Hill Formation fauna of the Queens Pinch Group (Wright 1994). The Jesse Limestone Member of the upper Limekilns Formation – believed to correlate
with the upper Cunningham Formation – extends up into the late Emsian *serotinus* Zone (Mawson and Talent 1992). In addition, Russell (1975) listed a conodont fauna obtained from a crinoidal limestone clasts in the lower parts of the Cunningham Formation; forms listed include 'Polygnathus foveolatus' (*Polygnathus perbonus*) and 'Spathognathodus exigus exigus' (= *Pandorinellina exigua exigua*) and 'Spathognathodus linearis' (= *Ozarkodina linearis*) indicating an early Emsian *perbonus* Zone age.

Collectively, these data suggest the Cunningham Formation extended from about the late Pragian to at least the late Emsian. The type section measured by Packham (1968) along the Turon River shows the formation commences with a 100 m thick package of featureless slate and siltstone; the lithology and stratigraphic position suggest this package correlates with the Rosedale Shale of Packham (1968) and presumably the Warratra Mudstone (Section 4.4.2). Away from the margins of the Hill End Trough where the Cunningham Formation interdigitates with the Nubrigyn Formation, the unit consists of a monotonous sequence of slate and lithic sandstone which is very poorly known and difficult to correlate to any of the many units deposited on the Highs through this interval.

4.5 SUMMARY OF CORRELATIONS

Possible equivalents of the Lochkovian-?Early Pragian lower Kandos Group occur in the Yerranderie-Windellama district on a probable southern extension of the Capertee High. Lochkovian western platform margin sequences are restricted to rare occurrences near Palmers Oakey. Recent SHRIMP isotopic ages from the Crudine Group of the Hill End Trough allow broad correlation with the Lochkovian Kandos Group, and enable speculation on the composition of unpreserved or presently unexposed parts of the Capertee High which were contributing sediment to the Trough during the Lochkovian. The Pragian Riversdale Volcanics can be correlated with the Huntingdale Volcanics and an overlying volcaniclastic horizon in the Capertee Valley, and to allochthonous deeper water deposits further west at Queens Pinch (Mullamuddy Formation) and in the Hill End Trough (Merrians Formation). Corresponding Pragian sequences on the Molong High are dramatically different and dominated by the Garra Limestone and several carbonate debris flow lobes derived from this unit. The Emsian Carwell Creek Formation can be confidently correlated with platform sequences which crop out almost continuously from the southern Capertee Valley (Myrtle Grove Formation) and to the basal Mt. Frome Limestone, east of Mudgee. Deeper water equivalents of the Carwell Creek Formation occur as hemipelagic mudstone and carbonate-clastic aprons which developed along the western margin of the Capertee High at Limekilns (Jesse Limestone), Queens Pinch and around Mudgee, and in the Cunningham Formation of the Hill End Trough. Similar carbonate-dominated fans occur adjacent to the Molong High on the western side of the Hill End Trough.
CHAPTER FIVE
DISCUSSION AND SYNTHESIS

5.1 INTRODUCTION

This chapter discusses various aspects of the Early Devonian history of the Capertee High, specifically the sea-level changes of the Kandos Group, the conodont biofacies, and the sedimentation and volcanic history of the High and its relation to the Hill End Trough. Finally, modern and ancient analogues for the Capertee High and Hill End Trough are discussed and a summary of the major thesis findings is presented.

5.2 SEA-LEVEL CHANGES OF THE KANDOS GROUP

Changes in relative sea-level are due to eustasy (ie. global movements of sea-level), local or regional tectonic movements, changes in sediment supply, or a combination of these effects (Struckmeyer and Brown 1990). If tectono-sedimentary effects can be eliminated or quantified in relative sea-level curves, a residual record of eustatic sea-level movements can be obtained, albeit with some distortion due to differential sediment compaction and loading. The Kandos Group succession provides an interesting record of transgressive and regressive events on the Capertee High during the Early Devonian. The new conodont data obtained in this study provide an opportunity to compare these trends with published studies of third order eustatic sea-level movements. In particular, the Kandos Group has valuable data to contribute on basal Lochkovian sea-level movements, not previously well-documented in Australia or, indeed, in many parts of the world (Johnson et al. 1985; Talent and Yolkin 1987). This section has been revised and updated from Colquhoun (1995b; Supporting Paper 5).

Deposition of the fan-deltaic Warrah Conglomerate above the basal Devonian unconformity marks the start of a Lochkovian transgressive cycle (probably in the \textit{woschmidtii} Zone) which continued through the biostromal to reefal carbonates of the Clandulla Limestone (\textit{woschmidtii} to \textit{eurekaensis} Zones) and into the lower to middle parts of the Yellowmans Creek Formation (probably \textit{delta} Zone), which record the drowning of the Clandulla carbonate platform, showing a transition from allodapic carbonates and calcareous shales to massive non-calcareous shales (Fig. 5.1). Furthermore, these units progressively onlap the Siluro-Devonian unconformity to the east of Carwell Creek (GR 772000 6364600).

Initial studies by Johnson \textit{et al.} (1985) concluded there were insufficient data in Euramerican sequences to infer confidently a sea-level curve for the basal Lochkovian. Based mainly on sequences in Virginia, Dennison (1985) added a Lochkovian section to the sea-level curve of Johnson \textit{et al.} (1985). Dennison's curve showed a slow early Lochkovian transgression, beginning in the \textit{woschmidtii} Zone and peaking (switching to regression) around the \textit{eurekaensis-delta} boundary (Fig. 5.2a). Based on Nevada
Figure 5.1 Transgression-regression curve for the Early Devonian Capertee High based on trends in the Kandos Group and Western Platform Margins sequences. Dotted lines indicate intervals of uncertainty. Data from this study, Talent and Yolkin (1987), Pickett (1987), Garratt and Wright (1988), McCracken (1990), and Mawson and Talent (1992).
sequences, Johnson and Sandberg (1988), documented a transgression which possibly began in the *eurekaensis* Zone and peaked in the *delta* Zone. Johnson and Klapper (1992) recorded, from the U.S.A. mid-west, a widespread transgressive event which began around the base of the *eurekaensis* Zone. In Australia, the Garra Formation and Windellama Limestone (Mawson 1986) show a transgressive cycle (Garra 1 of Talent and Yolkin 1987) which commenced not later than the late *eurekaensis* Zone and switched to regression during the *delta* Zone (Talent and Yolkin 1987; Talent 1989) (Fig. 5.2a and b). In addition, studies of Australian basins (Struckmeyer and Brown 1990) and Central Bohemia sequences (Chlupac and Kukal 1986) showed clear transgressive trends during the early to mid Lochkovian, although the curves of these studies were not calibrated by conodont zonation.

In summary, many sequences throughout the world show transgressions during the *woschmidtii*, *eurekaensis*, and ?early *delta* Zones, suggesting the basal Kandos Group transgression is due to a eustatic sea-level rise. Some disagreement exists, however, as to the actual time of inception of this transgression and its duration; this is probably a reflection of the paucity of data on this part of the Devonian.

Late Lochkovian regressive conditions commenced towards the top of the Yellowmans Creek Formation, with continued, shallowing throughout deposition of the overlying Roxburgh Formation culminating in subaerial exposure and local erosion at the top of the Roxburgh Formation (*pesavis* or *sulcatus* Zone) (Fig. 5.1). The lack of precise conodont dating for the Roxburgh Formation makes it difficult to match this regression to others with any certainty. The Garra 1 cycle shows shallowing up through the *delta* and early *pesavis* Zones, followed by a deepening event (Garra 2 cycle; Fig. 5.2b) in the late *pesavis* Zone (Talent and Yolkin 1987; Talent 1989; Talent pers. comm. in Young 1995). However, rapid regression was recorded through the *pesavis* Zone and into the early *sulcatus* Zone by Johnson and Sandberg (1988: Pre 1A cycle). Dennison (1985) recorded a regression in the *pesavis* Zone but also a deepening event in the late *delta* Zone, not documented elsewhere (Fig. 5.2a). Chlupac and Kukal (1986) and Struckmeyer and Brown (1990) also recorded general shallowing towards the top of the Lochkovian. Eustatic sea-level trends for the late Lochkovian are therefore not well-defined, although the *pesavis* interval appears to have a dominantly regressive character, suggesting the regression in the Roxburgh Formation may, at least in part, be eustatic in origin. It is likely, however, that local tectonic movements linked to increasingly proximal volcanism in the upper Roxburgh Formation would overshadow most eustatic sea-level changes (Colquhoun 1995a; Supporting Paper 4).

The subaerial Riversdale Volcanics indicates low relative sea-level stands during the Pragian, with volcanic detritus from the largely subaerial Capertee High being redeposited in proximal mass-flow dominated environments marginal to the High (Kingsford and Mullamuddy Formations), and the Merrions Formation of the Hill End Trough (Section 3.3). The presence of limestone clasts of *sulcatus* Zone age in the Mullamuddy Formation (McCracken 1990) and probable *kindlei* Zone clasts in the Kingsford Formation (this study) suggests platform exposure and erosion on the Capertee High during
Figure 5.2 Eustatic sea-level curves for the Early Devonian. (a) Qualitative eustatic curves for Australia and the Altai-Sayan region of Siberia (from Talent and Yolkin 1987) overlain on the Euramerican curve of Johnson et al. (1985) modified for the earliest Devonian by Dennison (1985). (b) Early Devonian section of the transgression-regression curve of Young (1995).
the *sulcatus-kindlei* interval. At least two brief marine incursions occurred in the middle to upper portions of the Riversdale Volcanics (Section 3.3.6.6) in the *sulcatus-kindlei* interval (Fig. 5.1). A lack of more precise dating prevents matching these transgressions with eustatic T-R curves. It is probable, however, that they may be linked to more local factors, such as caldera subsidence during a period of volcanic quiescence. The Australian sea-level curves of the Pragian period are vastly different to the Euramerican curve and the Siberian curve (Fig. 5.2a, b), suggesting the transgressions and regressions in the Australian curve may be strongly influenced by regional tectonic factors (Young 1995). Regression and platform exposure in the latest Pragian are indicated by the 'Red Hill submarine fan' which developed along the eastern margin of the Molong High and is aligned with the base of the *pireneae* Zone (Mawson and Talent 1992; Young 1995) (Fig. 5.2b).

The inception of the Carwell Creek Formation records a widespread sea-level rise on the Capertee High, drowning and reworking the pre-existing volcanic topography (Fig. 5.1). This transgression may correlate with a widespread earliest Emsian eustatic sea level rise (Johnson *et al.* 1985; Talent and Yolkin 1987), one of only two transgressions which Talent and Yolkin (1987) regarded as unequivocally eustatic in nature. Originally, this transgression was thought to have to have commenced in the early Emsian *dehiscens* Zone (Talent and Yolkin 1987; Talent 1989); more recent conodont data from Victorian sequences, however, suggests it probably commenced in the *pireneae* Zone, presumably early in this zone (Mawson *et al.* 1992) (Fig. 5.2b). In common with other sequences associated with this transgression, the basal carbonates of the Carwell Creek Formation and Myrtle Grove Formation are shallow and dolomitic, and contain faunas which are assigned only a general *pireneae* to *perbonus* Zone age (Section 3.3.7.6). Hence, these sequences cannot be aligned with certainty with this widespread transgression. In general, the Carwell Creek Formation/Myrtle Grove Formation show very similar facies throughout their great thickness, indicating a balance of subsidence and sediment supply with the sea-level rise.

Transgressions higher in the Carwell Creek Formation are recorded in limestones at Mount Knowles and Brogans Creek; these appear to have taken place during the *perbonus* Zone (Fig. 5.1). The Mount Frome Limestone seemingly records transgressions in the *perbonus* (or earlier), and late *inversus* to early *serotinus* Zones (Talent and Yolkin 1987; Pickett 1987). The Sutchers Creek Formation is dominated by clasts of *perbonus* and *serotinus* Zone carbonates (McCracken 1990), suggesting platform exposure and erosion during these intervals. The Jesse Limestone Member of the Limekilns Formation was apparently deposited entirely during a *serotinus* Zone regressive event (Mawson and Talent 1992). The Ingleburn Formation also contains limestone conglomerate with *inversus* Zone clasts (Garratt and Wright 1988). Therefore, in the Emsian, the Capertee High appears to record a basal transgression, possibly commencing in the latest Pragian *pireneae* Zone (see above), a *perbonus* Zone transgression followed by regression and platform exposure in the late *perbonus* and *inversus* Zones, a brief transgression in the early *serotinus* Zone or late *inversus* Zone, followed by regression and
erosion in the late *serotinus* Zone (Fig. 5.1). The *perbonus* Zone transgression appears to be too young to correlate with the basal Taravale transgression of Young (1995), although it may align with one of the small transgressions recorded in the basal Ib cycle of Johnson *et al.* (1985) (Fig. 5.2a). The late *inversus* to early *serotinus* Zone transgression may align with the Lockup Well transgression of Young (1995) (Fig. 5.2b) which appears to correlate with the upper part of cycle Ib of the Euramerican curve (Fig. 5.2a). The final regressive event on the Capertee High apparently concluded in the earliest Eifelian *partitus* Zone and was considered by Talent and Yolkin (1987) and Talent (1989) to represent the onset of the Tabberabberan Orogeny (Fig. 5.2a).

5.3 CONODONT BIOFACIES OF THE CAPERTEE HIGH

The relative abundance of icriodids in the uppermost Clandulla Limestone and basal Yellowmans Creek Formation is unexpected, as other authors have noted a general dearth of icriodids in Australian Early Devonian faunas. A review of the literature by Mawson and Talent (1989) revealed only 18 specimens, representing 3 species from 10 sections during the entire Early Devonian. Mawson and Talent (1994) have since documented a further 19 icriodid specimens from the Pragian of the Tyers-Boola area. By the late Emsian, large numbers of icriodids are apparent; Mawson *et al.* (1995), for example, have documented 162 icriodid Pa elements from the Tamworth and Yarrol belts of northeast New South Wales and east central Queensland respectively. Icriodids appear to be less numerous in the Lochkovian and the 20 complete Icriodid specimens and many fragments from 5 samples in the upper Clandulla Limestone and basal Yellowmans Creek Formation, represents one of the largest Lochkovian icriodid faunas in Australia; Farrell (1996) has documented similar numbers of icriodid specimens in Lochkovian units of the Molong High, mainly in the Camelford Limestone.

Mawson and Talent (1989) noted a general increase in icriodids throughout the Early Devonian and attributed this to zoogeographic factors, with icriodids not reaching the southern hemisphere in significant numbers until well after they were established in the northern hemisphere. The relative paucity of icriodids in the Lochkovian is thus in accord with their findings. Furthermore, they suggested that the nearshore Icriodid Biofacies of the northern hemisphere was probably replaced in the southern hemisphere by an Ozarkodinan Biofacies, dominated by simple cones and ozarkodinan elements. The icriodid faunas of the Clandulla Limestone and basal Yellowmans Creek Formation are most abundant in thinly bedded, richly fossiliferous, presumably relatively deep water facies of the upper Clandulla Limestone-basal Yellowmans Creek Formation and become extremely scarce in the middle to lower sections of Clandulla Limestone which are dominated by thick-bedded coralline and stromatoporoidal facies and mainly ozarkodinans and simple cones. This suggests nearshore areas were dominated by a Simple Cone-Ozarkodinan Biofacies perhaps passing offshore into an Icriodid Biofacies restricted to slightly deeper sequences.
Icriodids are again rare in the late Lochkovian, Pragian and Emsian carbonates of the Capertee High; this may also reflect biofacies control as a small number of icriodid specimens is known from the Taylors Hill Formation and one from the Ingleburn Formation (Fig. 4.1) of the Queens Pinch Belt on the western margin of the high (Wright, pers. comm. 1995).

In general, polygnathids are rare in basal Emsian (and suspected Emsian) strata in the thick-bedded, often unfossiliferous or stromatoporoidal-coraline carbonate facies which typically dominate platform areas of the Capertee High. The conodont faunas of the Carwell Creek Formation and its correlatives from the southern Capertee Valley, Brogans Creek, and "Erang" are characterised by a great preponderance of simple cones and pandorinellids and a rarity of polygnathids which only appear in dark, thinly bedded, richly fossiliferous, usually silicified facies; this scarcity has inhibited precise correlation of the platform sequences. In addition, no icriodids have been recorded in Emsian platform strata (in contrast to the Lochkovian; see above). This suggests most of the platform areas of the High during the early Emsian were within a niche occupied mainly by pandorinellids, ozarkodinans, and simple cones (cf. Pandorinellid-Simple Cone Biofacies of Mawson and Talent 1989), which was probably equivalent to the Icrioid biofacies of the northern hemisphere.

By contrast, polygnathids are relatively abundant in the carbonate sequences along the western margin of the High, at Limekilns and Queens Pinch. A Polygnathid Biofacies therefore clearly dominated these mainly allochthonous carbonate sequences along the western margin of the Capertee High. Only in the upper Mt Knowles sequence and at Mt. Frome did the Polygnathid Biofacies encroach onto the platform areas of the High to any significant extent, presumably due to deeper water conditions (Section 5.2).

5.4 REGIONAL PALAEOGEOGRAPHY, SEDIMENTATION AND VOLCANISM OF THE EARLY DEVONIAN CAPERTEE HIGH AND EASTERN HILL END TROUGH

5.4.1 Introduction

There are few palaeogeographic models of the Capertee High (Webby 1972; Cas and Jones 1979; Cas 1983; Powell 1984; McCracken 1990) and all suffer from the lack of published data from this part of the Lachlan Fold Belt. By comparison, the palaeogeography of the western bounding element of the Hill End Trough - the Molong High - is known in detail. Indeed the paucity of information on the Capertee High is such that it has led to controversy over the timing of its initiation and temporal duration, and even to doubts of its actual existence (Section 2.3). All evidence in this thesis and in other recent studies (e.g. Pemberton 1990; Pickett et al. 1996) indicates the Capertee High persisted as a shallow marine to terrestrial entity from at least the Late Silurian to the Middle Devonian; however, its nature probably changed with time. Cas and Jones (1979), Cas (1983), and McCracken (1990) viewed the Capertee High as a line of unconnected island volcanoes or plateau-like features separated by deeper water areas. Such a configuration may well have existed in the Late...
Silurian when there was apparently a close lateral juxtaposition of shallow marine-terrestrial units and deeper marine units (Colquhoun unpubl. data; Pemberton 1989).

By Early Devonian times - especially during the Pragian and Emsian - the high lateral continuity of facies and facies trends indicates the Capertee High had coalesced into a persistent shallow marine to terrestrial feature, able to be traced from the southern Capertee Valley to north of Mudgee. The Early Devonian evolution of the Capertee High is depicted here through three isometric block diagrams (Figs 5.3, 5.4, 5.5) representing the late Lochkovian, Pragian and Emsian. To make the diagrams more realistic, an attempt has been made to account for the approximately 50% of northeast to southwest crustal shortening which occurred during deformation (Colquhoun unpubl. data). This was done by making the northeast to southwest axis of the diagrams approximately twice present-day widths; however, the lack of precise data on the amount of crustal shortening in other areas of Capertee High presently prevents truly palinspastic reconstructions. In addition, an attempt has been made to correct for speculative large-scale rotations which occurred during deformation (e.g. the 35° of clockwise rotation around the Pyramul Lineament proposed by Powell et al. 1985). Areas of known and inferred sedimentation are also depicted, based largely on the correlations suggested in Chapter Four. Much of the regional data are open to a number of interpretations, and the synthesis offered here, while considered the most probable, is only one of many possible.

Sequence stratigraphy has been defined as 'the study of rock relationships within a chronostratigraphic framework of repetitive, genetically related strata bounded by surfaces of erosion or non-deposition, or their correlative conformities' (Van Wagoner et al. 1988). The technique provides a means for subdividing an apparently complex sedimentary succession into packages (sequences) that correspond to chronostratigraphically constrained depositional intervals. The use of sequence stratigraphic concepts to enhance understanding of geological relationships within a time-stratigraphic framework has gained widespread acceptance in recent years. Most successful application of these concepts has taken place in ancient, weakly deformed passive margin sequences where there is an abundance of drill hole and seismic data; however, the concepts are equally applicable to active margins providing local conditions such as tectonics, sediment flux and physiography – which are far more variable – are taken into account. However, application of these concepts to the Hill End Trough-Capertee High is hampered by the following factors: (a) the imprecision of correlations between the Platform/Western Platform Margin sequences, which are well constrained biostratigraphically, and the Hill End Trough, where biostratigraphic control is minimal and chronology relies largely on a scatter of SHRIMP dates (Chapter 4). There is commonly a 1% or greater error in isotopic ages in the Devonian and this error can encompass several conodont zones (Young 1995). Furthermore, Early Devonian geochronology is currently in a state of flux and the ages of many key intervals, particularly the base of the Devonian, are likely to alter dramatically in forthcoming versions of the timescale (G. Young pers comm 1996). (b) recognition of sequence bounding surfaces is often difficult given the constraints of
exposure and the occasional high levels of deformation. (c) discriminating between the tectonic and
eustatic signals on the High is not always possible (Section 5.2). Bearing these limitations in mind,
sequence stratigraphic concepts have been integrated into the following discussion and have provided
an important aid in understanding the sedimentation history of the Capertee High and Hill End Trough.

5.4.2 Lochkovian (Fig. 5.3).

As definite Lochkovian strata on the Capertee High are restricted to the Rylstone-Cudgegong
district, elucidating the regional palaeogeography of this interval is more difficult than for the Pragian
or Emsian. Salient palaeogeographic features of the High in the Rylstone-Cudgegong district were:
low- to moderate-relief subaerial areas of Silurian and Ordovician basement in the east; a southwest- to
west-southwest-sloping subaqueous shelf (of variable width due to extensive shoreline movements
during this time); and slopes marginal to the Hill End Trough in the west (Figs 3.6, 3.21). The Lochko­
vian began with a widespread type 1 unconformity (Van Wagoner et al. 1988) which separates the
Silurian and Devonian sequences on the Capertee High. This contact is a disconformity or very low
angle unconformity, suggesting it is dominantly due to sea-level lowstand, perhaps with very mild uplift
relating to widespread Siluro-Devonian tectonism in the Lachlan Fold Belt (the 'Bowning Orogeny' of
Packham 1969) (Section 2.3.3). Therefore, during the latest Silurian (?late Pridoli), the Capertee High
platform areas were exposed and sediment derived from erosion of the shelf was redeposited seaward to
form a lowstand turbidite wedge along the western margin of the High; this may be reflected in the
Cookman Formation which Packham (1968) regarded as being derived from a diverse source terrain to
the east of the Hill End Trough which included silicic volcanics (probably the Silurian Windamere
Volcanics and Dungeree Volcanics), quartz-rich sediments (Early Ordovician Adaminaby Group),
mafic volcanics (Sofala Volcanics or Coomber Formation) and granite (?; no pre-Carboniferous gran­
ites are known from the Capertee Zone).

Devonian sedimentation on the Capertee High began in the early Lochkovian woschmidti Zone
and comprised (in ascending order): a transgressive sequence of wave-dominated, fan-deltaic clastics
(Warrah Conglomerate); a unit of transgressive biostromal to biohermal limestones (Clandulla Lime­
stone; woschmidti to eurekaensis Zones); a transgressive to regressive muddy, storm-influenced shelf
deposit (Yellowmans Creek Formation; eurekaensis to delta Zones); and a thick regressive unit of
storm-dominated siliciclastics deposited in middle shelf to nearshore settings (Roxburgh Formation,
delta to pesavis or early sulcatus Zones) (Fig. 3.21). The sequence from the Warrah Conglomerate to
the lower parts of the Yellowmans Creek Formation represents a transgressive system tract which
developed due to a probable eustatic sea-level rise (Section 5.2). During a transgressive system tract,
brasinal areas such as the Hill End Trough, should be characterised by muddy condensed sequences as
the rising sea-level cuts off the supply of coarse sediment. However, this early Lochkovian transgres­sive system tract is difficult to correlate with the probably equivalent Turondale Formation of the Hill
LATE LOCHKOVIAN GEOGRAPHY AND FACIES - NORTHERN
CAPERTEE HIGH AND EASTERN HILL END TROUGH

Figure 5.3 Late Lochkovian geography and facies of the Capertee High and eastern Hill
End Trough (Fig. 4.1) which was receiving large volumes of silicic volcanic detritus that, palaeocur-
rents by Packham (1968) suggest, was derived from a source east of Sofala - probably the incipient
Huntingdale Volcanics (Section 4.2.3) - in addition to basement lithologies. This suggests local tectonic
and volcanic factors were strongly influencing sedimentation sources on the Capertee High at this time.
Thick sills of silicic porphyry were intruded into the Turondale Formation soon after deposition of the
clastic portion of the unit.

The inception of the Waterbeach Formation of the Hill End Trough possibly correlates with the
delta Zone Yellowmans Creek Formation, with both units representing late transgressive to early high-
stand system tract deposits; the Yellowmans Creek Formation was deposited in shelf areas whilst the
lower Waterbeach Formation was deposited in a basinal setting. The deposition of these units apparent-
ly coincided with a waning of volcanism in the southern volcanic source which supplied volcanic detri-
tus to the Turondale Formation. Condensed sequences often characterise shelf and basinal areas during
late transgressive and early highstand systems tracts (Posamentier et al. 1988). Condensed sections are
distinctively thin, often richly fossiliferous marine stratigraphic units consisting of pelagic to hemipe-
lagic sediment which displays signs of sediment starvation (Posamentier et al. 1988); however, no defi-
nite condensed sections have so far been identified from within the Yellowmans Creek Formation or
Waterbeach Formation.

The Roxburgh Formation was deposited in a late highstand system tract. The quartz- and
volcanic-rich sandy facies of the Roxburgh Formation prograded towards the edge of the high. This
may be reflected in the quartz-lithic sandy packages of the upper Waterbeach Formation and the Guro-
ba Formation in the Hill End Trough sequence (Fig. 4.1); palaeocurrents in these units support such an
interpretation (Section 4.2.3). Furthermore, the increase of volcanic detritus up-section through turbi-
dites of the upper Waterbeach and Guroba Formations (E. Jagodzinski, pers. comm. 1996) probably
reflects a similar increase in silicic volcanic detritus towards the top of the Roxburgh Formation
(Section 3.3.5.5).

As regression continued, erosion of platform carbonates and clastics occurred. These were
redeposited by mass-flows possibly in fans and aprons marginal to the Capertee High at Palmer's
Oakey (in the upper Waterbeach Formation) and possibly at Queens Pinch (?basal Mullamuddy
Formation) during relative sea-level lowstands during the latest Lochkovian pesavis Zone.

Depositional environments on the Molong High during the early Lochkovian woschmidti and
eurekaensis Zones were similar to those of the Capertee High: dominated by shallow marine clastics
and carbonates. However, the development of a mafic to intermediate volcanic pile in the delta Zone
(Cuga Burga Volcanics) followed by carbonates of the Garra Limestone, commencing diachronously in
the delta to pesavis Zones, contrasts markedly with coeval sedimentation on the Capertee High (Fig.
4.2).
Silicic volcanism on the Early Devonian Capertee High is first noticeable in the Rylstone-Cudgegong district near the top of the Clandulla Limestone and was intermittently active throughout the remainder of the Lochkovian, as demonstrated by volcanic debris flows in the basal Yellowmans Creek Formation, and accretionary lapilli and a partially volcanic provenance in the Roxburgh Formation. Volcanism was distal from the Rylstone-Cudgegong district and the proposed location of most of the volcanoes shown in the east (Fig. 5.3) is merely speculative as coeval volcanics of this age are lacking. As outlined above, voluminous volcanic activity was probably active on the southern parts of the Capertee High (the present-day Capertee Valley area) from the early Lochkovian on (Fig. 5.3).

Continued regression in the latest Lochkovian to earliest Pragian culminated in exposure and erosion at the top of the Roxburgh Formation mostly in the east of the Rylstone-Cudgegong district, and a conformable relationship further west. Such a contact resembles the Type 2 unconformity of Posamentier et al. (1988) which develops when some deposition is maintained landward of the shelf break at the time of maximum offlap. Under such conditions, there is subaerial exposure of part of the shelf, but deep valley erosion and a seaward shift in facies are absent. In the case of the Roxburgh Formation-Riversdale Volcanics contact, this was probably caused by gentle uplift of areas proximal to the developing silicic volcanoes in the east (Fig. 3.21). However, further uplift and regression eventually resulted in subaerial conditions over most of the Capertee High during deposition of the Riversdale Volcanics.

5.4.3 Pragian (Fig. 5.4)

Voluminous silicic pyroclastic volcanism dominated the Capertee High from north of Cudgegong to the south of the Capertee Valley during the early to mid Pragian sulcatus and kindlei Zones (Riversdale Volcanics and Huntingdale Volcanics). Medium-scale calderas centred southeast of the present-day locations of Cudgegong and the eastern Capertee Valley were flanked by extensive, gently-dipping sheets of silicic ignimbrites and epiclastics which probably overlapped, forming a continuous sheet. A fall in sea-level relative to Lochkovian times ensured volcanism was largely subaerial and a lowstand system tract (Van Wagoner et al. 1988) dominated. Epiclastic reworking of the volcanics was extremely rapid, and vast volumes of volcanic detritus were transported rapidly by fast-flowing rivers to a narrow shelf margin system tract in the west where, along with fringing shallow marine limestones and siliciclastics, they were redeposited westwards by mass-flow processes to lowstand fans and aprons. Proximal areas marginal to the Capertee High (Kingsford Formation, Mullamuddy Formation) were characterised by coarse volcaniclastic and carbonate mass-flow aprons, whereas more distal deposits in the Hill End Trough (Merrions Formation and ?basal Cunningham Formation) were dominated by thick volcaniclastic turbidites, along with hemipelagites (representing volcanically quiet intervals) and thick, subaqueously-deposited silicic lavas. The extreme thickness and volume of some turbidites in the Merrions Formation, led Cas (1979, 1992) to propose that some of the turbidity currents
may have developed directly from large subaerial ignimbrite eruptions; such an interpretation accords well with the setting outlined above. In addition, the presence of a secondary sedimentary source in the upper Riversdale Volcanics (this study) and Huntingdale Volcanics (Kennedy 1976), along with the presence of Silurian limestone clasts in the Mullamuddy Formation (Wright and Byrnes 1980), suggests a flanking belt of Silurian and older volcanics and sediments was exposed to the east – similar to the situation in the Lochkovian (Fig. 5.4).

Volcanism waned towards the end of the Pragian (?late kindlei to pireneae Zones), and at least two brief intervals of volcanic quiescence were followed by transgression and deposition of carbonates in intracaldera areas (Facies E of Riversdale Volcanics; Section 3.3.6.4.5). At the top of Riversdale/Huntingdale Volcanics, the silicic pyroclastic sequences were reworked in fluvial and nearshore settings. The much greater thickness of these deposits in the Capertee Valley sequences (see Section 4.3.1.1) suggests an earlier cessation of volcanism and more prolonged fluvial activity prior to transgression in the earliest Emsian or latest Pragian. The supply of volcaniclastic detritus to marginal high areas and the Hill End Trough diminished: this is recorded by the Taylors Hill Formation, which is dominated by mudstone, thin volcanic sandstone beds and allochthonous limestone blocks, the latter indicating continued erosion of shelf margin carbonates in the late Pragian. Furthermore, sections measured by Cas (1977) suggest that bedding becomes thinner and grain size decreases in the upper volcanic sandstones of the Merrions Formation, and that there is a transition to slates of the basal Cunningham Formation.

Magmatic activity during the Pragian and into the Emsian was apparently bimodal. Local sheet lava flows of subalkaline island-arc basalt occurred in the Mullamuddy Formation in the Queens Pinch district (McCracken 1990) and mafic to intermediate intrusive activity was apparently widespread during this time period (Section 4.2.1.2). In addition to the aforementioned silicic pyroclastic activity of the platform of the Capertee High, calcalkaline silicic to intermediate lavas also erupted in the Hill End Trough (Merrions Formation) during this time (Cas 1978).

Pragian sedimentation and palaeogeography of the Molong High were dramatically different to the Capertee High. The Molong High was volcanically quiet and isolated from sources of clastic detritus, and was dominated by thick shallow water carbonates in platform areas which were locally redeposited in debris flow aprons along the eastern margin of the High during brief regressive intervals.

5.4.4 Emsian (Fig. 5.5)

Volcanism on the Capertee High largely ceased at the end of the Pragian, although minor silicic lavas were extruded in the Ingleburn Formation of the Queens Pinch Group. A transgressive system tract architecture (Van Wagoner et al. 1988) then dominated the Capertee High and eastern Hill End Trough. A eustatic transgression in the latest Pragian to earliest Emsian (pireneae to dehiscens Zones) initially reworked then drowned the existing volcanic topography, creating a broad southwest-
Figure 5.4 Pragian geography and facies of the Capertee High and eastern Hill End Trough. Not all facies or geographic features were coeval, see text for details.
sloping shelf with local eroding volcanic islands. Sedimentation in this setting was dominated by a laterally and temporally extensive regime of storm- and tidally-influenced nearshore to shelf siliciclastic sequences deposited in barrier island settings with local biostromal and biohermal carbonate buildups deposited in lagoonal and open-marine environments (Carwell Creek Formation and Myrtle Grove Formation). Subsidence largely kept pace with, or exceeded, depositional rates, resulting in retrogradational to aggradational sequences; however, minor fluctuations in sea-level and sediment supply caused progradational pulses. The sequence gradually onlapped flanking low-relief areas of Silurian and ?older volcanics and sediments to the east.

The rising sea-level cut off the supply of coarse detritus from basinal areas and condensed mudstone sequences, typical of transgressive system tracts, developed on slope and basin areas marginal to the Capertee High; these are represented by the Warratra Mudstone (?dehiscens Zone) and Rosedale Shale (late Pragian; ?pireneae Zone). Both units contain features typical of condensed mudstone sequences including condensed horizons rich in bioturbation and shelly debris alternating with comparatively barren mudstone sequences, along with local hardgrounds in the Rosedale Shale (Wright, pers comm. 1996), indicating a quiet, sub-storm wave base environment characterised by slow hemipelagic deposition and sediment starvation over long periods. (Section 4.4.2). Furthermore, the basal slate-dominated 110 m of the Cunningham Formation probably correlate with this widespread earliest Emsian sea-level rise (Section 4.4.3).

A further transgressive pulse on the Capertee High is recorded in limestones at Mt Knowles and Brogans Creek and apparently took place in the early Emsian perbonus Zone. This was followed by a regressive interval in the late perbonus to inversus Zone resulting in platform exposure and erosion; this is recorded by the middle sections of the Mt. Frome Limestone which are dominated by birdseye textures, indicating intertidal conditions (Wright 1966), and an abundance of pireneae, dehiscens, and perbonus Zone clasts in the basal Sutchers Creek Formation (McCracken 1990). The dominantly carbonate platform detritus was redeposited by mass-flows on aprons and fans marginal to the Capertee High (lower Sutchers Creek Formation). Later, during deposition of the Ingleburn Formation of the inversus Zone (Fig. 4.1), these aprons/fans became dominated by mass-flows of coarse lithic detritus (dominantly dacitic) and, more rarely, quartzose detritus (Wright 1966), suggesting erosion of Silurian or Devonian silicic volcanics along with quartz-rich units such as the Early Ordovician Adaminaby Group or Early Devonian Roxburgh Formation. Dacite lava was locally extruded in marginal high areas at this time; a U/Pb SHRIMP age of 403 ± 8 Ma was obtained from these lavas (Colquhoun et al. 1997). In the Hill End Trough, deposition of lithic sandstone and mudstone of the Cunningham Formation continued; however, due to a lack of information on this unit, it is difficult to correlate other intervals in this unit with any of the aforementioned events on platform and platform margin areas of the Capertee High.
Figure 5.5 Emsian geography and facies of the Capertee High and eastern Hill End Trough. Not all facies or geographic features were coeval; see text for details.
A transgression resulted in further deposition of carbonates on the Capertee High during the serotinus Zone (Mt Frome Limestone and unpreserved equivalents). Regression later in the serotinus Zone resulted in erosion of some of these platform carbonates and redeposition in fans along the western margin of the Capertee High (Jesse Limestone and upper Sutchers Creek Formation) (Fig. 4.1). This regression appears to continue into the earliest Eifelian partitus Zone, when carbonates of the Mt Frome Limestone were replaced by wave-dominated shoreline and shelf clastics of the Boogledie Formation (Section 4.4.1.1). Deposition of this clastic sequence continued for an unknown length of time into the early Eifelian before it was terminated by uplift and regression related to a mildly manifested Middle Devonian deformation ('Tabberabberan Orogeny'; Packham 1969) which effectively ended the depositional history of the Capertee High.

5.5 MODERN AND ANCIENT PALAEOGEOGRAPHIC AND TECTONIC ANALOGUES OF THE CAPERTEE HIGH-HILL END TROUGH-MOLONG HIGH

5.5.1 Introduction

As outlined in Section 2.3, the Siluro-Devonian tectonic setting of the northeast Lachlan Fold Belt is poorly understood, although a number of models have been proposed. Examination of marginal basins of the present-day southwest Pacific region reveals possible palaeogeographic and tectonic analogues for the Capertee High-Hill End Trough-Molong High system during the Early Devonian. Modern backarc basins of the western Pacific display a variety of sediment types including coarse-grained mass-flow deposits, submarine fan turbidites and basinal turbidites, hemipelagic pelagic silts and clays, biogenic carbonates and biosiliceous sediments and various types of tephra deposits. The volume and nature of siliciclastic and volcaniclastic materials are controlled by arc volcanism and the presence and relief of nearby land sources.

Many backarc basins of the western Pacific have been investigated during the most recent phase of the Ocean Drilling Program. Three modern analogues from the western Pacific are discussed here, along with one ancient sequence from the North Island of New Zealand.

5.5.2 Modern Analogues

5.5.2.1 Sumisu Rift, Izu-Bonin Arc (Western Pacific)

Extension of the Izu-Bonin arc since the Pliocene has resulted in the formation of a series of asymmetric rift basins which are 20-60 km wide by 20-120 km long and located adjacent to the active volcanic arc (Fig. 5.6). The rifts are bounded by high-angle normal fault zones and are separated along strike by structural highs and chains of submarine volcanoes (Klaus et al. 1992). The Sumisu Rift is a 120 km long, 30-50 km wide, asymmetrical, normal fault-bounded rift basin located on the Izu-Bonin Arc to the south of Honshu, Japan (Fig. 5.6), interpreted as a nascent intraoceanic backarc basin by Marsaglia et al. (1995). The area has been the subject of detailed study by the Geological Survey of
Japan and the Hawaii Institute of Geophysics and is the most intensely and comprehensively studied intraoceanic arc rift segment (Taylor 1992).

Klaus et al. (1992) interpreted the Sumisu Rift to be in an early syn-rift stage of backarc basin formation; the age of rift inception was estimated by Taylor (1992) as approximately 2 Ma. On the model of backarc basin evolution proposed by Carey and Sigurdsson (1984), the Sumisu Rift is in Stage 1 - before the rift has developed a seafloor spreading phase. Volcanic products range from tholeitic basalt to sodic rhyolite (Marsaglia et al. 1995); however, after initiation of the rift, volcanism became primarily dacitic to rhyolitic. In general, mineral and whole-rock geochemical analyses indicate source rocks of island arc character (Hiscott and Gill 1992). Drilling indicates the Rift is floored by vesicular basalt reflecting the pre-rift island arc (Fig. 5.10).

The Sumisu Rift is located to the west of an active frontal arc containing several large, submarine, dacitic to rhyolitic calderas (South Sumisu and Sumisu calderas; Figs 5.7, 5.9 and 5.10) surrounded by volcaniclastic sequences; these calderas are presently 300 to 400 m below sea-level. The rift basin contains two major depocentres, the Kita and Minami Sumisu basins (Figs 5.7 and 5.8), separated by a fault zone which is marked by submarine mafic to dacitic volcanoes. The sub-basin floors are smooth, deep (2100-2275 m) plains that gently shallow to the north and south toward the volcaniclastic aprons that flank adjacent arc volcanoes and calderas (Klaus et al. 1992). The rift basin is bounded to the west by a remnant arc which rises 300-1500 m above the sedimented basin floor (Klaus et al. 1992). Basin fill is characterised by high sediment accumulation rates (>4 km/Ma) and has a wedgelike geometry, thickening towards the eastern margin (Fig. 5.10). Thick, coarse pumiceous mass flow units and direct deposition of fallout from periodic eruptions of the silicic calderas on the arc to the east constitute most of the basin fill; mass-flows were fed primarily into the northern and southern ends of the rift basin (Fig. 5.9). Nishimura et al. (1992) identified five large-scale, eruption-related, fining-upward sequences up to 118 m thick. These volcaniclastic sequences are separated by intervals of hemipelagic mud, predominantly in the basal parts of the basin fill.

The Sumisu Rift is a good modern analogue for the Early Devonian Capertee High-Hill End Trough-Molong High. Sumisu Rift dimensions are similar to slightly smaller to those of the Hill End Trough, once east-west crustal shortening of the Trough sequence is taken into account. The frontal arc is a close analogue for the Capertee High, particularly in the Pragian when it was probably marked by several large silicic calderas along its length (compare Fig. 5.4 with Fig. 5.9). Similar to the Early Devonian Molong High, the remnant arc contains volcanic flows and vents which are more restricted and concentrated along active rift faults. Both of the bounding highs to the Sumisu Rift apparently lack a comparable suite of extensive shallow marine siliciclastics and carbonate; however, this is probably due to a different balance of tectonics and eustasy than the Capertee High and Molong High, where a delicate equilibrium of these features apparently allowed shallow marine conditions to persist for long periods of time. In addition, the higher latitude of the Sumisu Rift (around 31°) possibly accounts for
Figure 5.6 Generalised tectonic map of the western Pacific showing the location of the Izu-Bonin arc-trench system (modified from Klaus et al. 1992).

Figure 5.7 Detailed bathymetric map of the Sumisu Rift region, showing the location of ODP Leg 126 sites and arc calderas. From Marsaglia et al. (1995).

Figure 5.8 Structural and volcanic map of the Sumisu Rift. Extrusive volcanic rocks are shaded black, active calderas (Sumisu, South Sumisu, and Torishima) are stippled, and pyroclastic cones are striped and stippled. CR = Central Ridge and SM = Shadow Mountain. From Klaus et al. (1992).

Figure 5.9 A block diagram of the Sumisu Rift illustrating the likely direction of sediment input into the rift from arc calderas (primary sediment input). From Klaus et al. (1992).
Figure 5.10 Diagram summarising the lithostratigraphy of Leg 126 sites 788, 790, and 791, Sumisu Rift (see Fig. 5.7 for location of drill holes). Units are not pictured, but unit boundaries are marked on the left side of the stratigraphic columns in metres below sea floor (mbsf). Relative grainsize variations are profiled on the right, from clay (c) to gravel (g). The schematic cross section at the base is simplified from a seismic profile that links the sites. From Taylor (1992).
the paucity of carbonates. It should be noted that drilling on the volcanic highs is in a preliminary stage.

Sumisu Rift fill is broadly comparable to the Early Devonian Hill End Trough. Both contain: thick volcaniclastic mass-flow sequences which were derived mainly from, and thicken towards, the more active eastern arc; thick hemipelagic intervals reflecting periods of volcanic quiescence; inputs of clastic debris reworked from the basement of the eastern and western flanking highs; and bimodal silicic/mafic volcanics in the basin fill (Fig. 5.8).

Plots of sandstone detrital modes for the Sumisu Rift (Marsaglia 1992) reveal far less free quartz and more mafic to intermediate rock fragments than Hill End Trough volcaniclastics which are dominated by felsitic rock fragments and free quartz (Packham 1968). This reflects the eruptive history of the nearby volcanic centres around the Sumisu Rift which became more silicic as rifting progressed (Klaus et al. 1992). The Hill End Trough, by comparison, was in a more mature stage during the Early Devonian. However, the Hill End Trough was probably still in Stage 1 of Carey and Sigurdsson's (1984) model as oceanic basaltic basement has not been identified, and it appears that seafloor spreading had not commenced before uplift terminated deposition in the Middle Devonian. The volcanoes on the Capertee High and, to a lesser extent, the Molong High were predominantly silicic from the Late Silurian on, accounting for the more silicic volcaniclastic fill of the Hill End Trough.

The configuration of sub-basins in the Sumisu Rift may also provide a close analogue for the Hill End Trough and explain some of the correlation problems within the Trough sequence. The Sumisu Rift contains two depocentres, the Kita and Minami Sumisu Basins, divided by a ridge of submarine basaltic to dacitic volcanoes (Fig. 5.7). The basins are unlinked and seismic profiles suggest profoundly different sequences, although preliminary drilling has not yet proved this to be the case. By comparison, Glen and Watkins (1994) have suggested the different sequences in the Hill End Trough to the north and south of the Bathurst Granite below the Cunningham Formation may indicate a basement high in the area presently occupied by the granite. This high was finally overwhelmed by sediment during deposition of the Cunningham Formation, which occurs both north and south of the Bathurst Granite. Furthermore, problems in correlating the Crudine Group in the northern Hill End Trough may have a similar explanation.

5.5.2.2 Lau Basin-Havre Trough, Tonga-Kermadec Arc (Western Pacific)

The Lau Basin-Havre Trough-Tonga-Kermadec Ridge area forms part of the southwest Pacific island arc-back-arc system (Fig. 5.11). The volcanic Tonga and Kermadec islands delineate the active volcanic arc, generated by the subduction of the Pacific Plate in the east beneath the Austral-Indian Plate in the west. To the west of the active volcanic arc, the back-arc basin system comprises the Lau Basin, Havre Trough and the Taupo Volcanic Zone (Cole 1984). On the west side of the Lau Basin and Havre Trough lies the remnant arc represented by the Lau and Colville Ridges (Karig 1970). The basins have an average depth of 2500 m (Cole 1984).
The Lau Basin is approximately 700 km long and varies in width from 400 km in the north to about 170 km in the south, at a distinct sill which separates it from the Havre Trough (Fig. 5.11). The Basin's flanking ridges to the east and west are largely submerged but emerge sporadically as islands of the Lau and Tonga Groups. The Tonga Ridge represents the active frontal arc and comprises a largely submerged chain of active basaltic to dacitic volcanoes flanked by aprons of volcanioclastic and rimmed by local shallow water carbonates (Marsaglia et al. 1995). The Lau Ridge is volcanically inactive and represents the remnant arc, consisting of local carbonate platforms on degraded volcanic pedestals.

The Lau Basin is floored by oceanic basalt crust. Recent drilling in the Lau Basin has revealed it to be characterised by a very thin average sediment cover and the basin is divided into numerous half-graben sub-basins which pond the supply of coarse sediment close to the arc in the east, leaving the basin centre virtually barren of sedimentary cover (Marsaglia et al. 1995). Each of the sub-basins shows a similar upward-fining sequence with basal coarse-grained, proximal deposits (volcanic sands and gravels) that grade upward into nannofossil-ooze dominated sections with rare airfall ash beds. The proximal facies record an apparent eastward shift in volcanism across the basin during the Pliocene and Pleistocene (Marsaglia et al. 1995).

As pointed out by Cas and Jones (1979), the Lau Basin has only limited value as an analogue for the Hill End Trough and its margins. Although the basic facies types and gross asymmetry of sediment fill are similar, there are many significant differences, notably: (a) the Early Devonian Capertee and Molong Highs were more persistent features than the isolated volcanic islands and flanking carbonate aprons of the Tonga and Lau Ridges (Section 5.4); (b) the Lau Basin is much longer and wider than the Hill End Trough; (c) thick sedimentary sequences and coarse clastic rocks occur everywhere in the Hill End Trough, not just marginally as in the Lau Basin; (d) volcanic products are on average more silicic and have higher total alkalis in the Hill End Trough.

The Havre Trough commences at 24°S at a sill which divides it from the Lau Basin. The Trough varies from 100 to 165 km wide and is bounded to the east by the volcanically active Kermadec Ridge and to the west by the Colville Ridge (Fig. 5.11). Like the Lau Basin to the north, the Havre Trough is an actively spreading, oceanic marginal basin (Lewis and Pantin 1984). Magnetic anomaly patterns led Malahoff et al. (1982) to conclude that backarc spreading had commenced in the Havre Trough about 2 Ma ago and the basin has had an average half spreading rate of 27 mm/yr. The Havre Trough terminates against the lower slope off the Bay of Plenty, North Island, New Zealand, and is replaced on the upper slope, and on land, by an ensialic marginal basin known as the Taupo Volcanic Zone (Fig. 5.12). The Taupo Volcanic Zone is offset about 50 km eastwards from the Havre Trough, by transform faults on the middle and upper slope (Lewis and Pantin 1984).

Little is known about the composition of sediments of the deeper Havre Trough. No drilling has been carried out and Cronan et al. (1984) interpreted seismic reflection profiles as indicating
Figure 5.11 Regional bathymetry of the Lau Basin-Havre Trough-New Zealand area of the southwest Pacific. From Cas and Jones (1979).

Figure 5.12 Bathymetry and simplified onshore geology of the Bay of Plenty. Offshore bathymetric contours are at 200 m intervals and the broken intervals and broken hatched line indicates the shelf break. Onshore geology, O = Matakaoa ophiolite; M = Mesozoic greywacke; T = Tertiary sediment; A = Miocene andesite; R = Miocene-Pleistocene rhyolite; I = Quaternary ignimbrites; Unstippled = Late Quaternary marine fluviatile deposits. From Lewis and Pantin (1984).
generally less than 100 m of sediment cover with few ponded areas of sediment up to 500 m thick. Along the Kermadec Ridge and its western flanks, however, sediment cover is up to 700 m thick (Cronan et al. 1984). In this area, many ridges, seamounts and scarps, some of which are probably fault-controlled, tend to restrict sediment input to the Havre Trough.

In the Bay of Plenty area, the southern Havre Trough has a far thicker sediment cover and is characterised by continuous, rapid, terrigenous supply from rising, frontal ridge mountains of Mesozoic greywacke and Tertiary marine sediments in the east, and an intermittent supply of mainly rhyolitic, pyroclastic debris from the Taupo Volcanic Zone to the south (Fig. 5.12).

Cas and Jones (1979) considered the Hill End Trough to be an ancient analogue of the southern termination of the Havre Trough where it impinges the Taupo Volcanic Zone in the Bay of Plenty. Here, both basins are: of similar lateral dimensions; flanked by platform highs with arc volcanics, the eastern arc being the active arc; and pass south into regions of intense silicic magmatism (Cas and Jones 1979). Sediment input into both basins was a mixture derived from terrigenous basement and coeval silicic volcanics. The main problem with the analogy is that it only applies to the southern ensialic Havre Trough and, proceeding northwards, the Havre Trough apparently takes on the character of a sediment starved intra-oceanic marginal basin, similar to the Lau Basin, and bears scant resemblance to the Hill End Trough. As palaeogeographic reconstructions in this study have shown, the Capertee High provided voluminous terrigenous and volcanic input from the east into the Hill End Trough, not just from the south as with the Havre Trough analogue. Due to the paucity of information on the Capertee High, Cas and Jones (1979) considered much of the volcanic input from the Hill End Trough to come from the south (the Canberra Magmatic Province) which was largely subaerial during the Early Devonian.

The Havre Trough is therefore, at best, a partial palaeographic and facies analogue for the Hill End Trough.

5.5.2.3 Sulu Basin, Borneo-The Philippines

The Sulu Sea, between Borneo and The Philippines, contains two distinct basins separated by the northeast-trending Cagayan Ridge (Fig. 5.13); the southeast Sulu Basin has an oceanic basement whereas the northwest Sulu Basin is apparently floored by continental crust (Marsaglia et al. 1995). The southeast basin is 4 to 5 km deep, with a sedimentary fill 1 to 2 km thick and is presently subducting southeastward beneath the Sulu Ridge, an active volcanic arc. Leg 124 of the Ocean Drilling Program drilled three sites in the southeast Sulu Basin; all terminated in oceanic basement (Figs 5.13 and 5.14). The basin opened in the early Miocene and andesitic to basaltic pyroclastic deposits with thin lava flows and pelagic sediments accumulated on the Cagayan Ridge (Fig. 5.14), whilst in the southeast Sulu Basin, rhyolitic to andesitic sands were deposited by a series of submarine gravity flows and turbidity currents fed by voluminous explosive submarine eruptions from the arc (Fig. 5.14). This
was followed by a prolonged period of mostly pelagic deposition in the Middle Miocene. During the Late Miocene, deposition in the southeast Sulu Basin was dominated by thin- to thick-bedded quartz turbidites probably derived from quartz-rich strata in northern Borneo. This period was apparently synchronous with a major sea-level lowstand that defines the base of the TB3 global sea-level super-cycle (Haq et al. 1984), and subdivisions of the turbidite sequence correlate with third-order eustatic cycles, indicating an important eustatic control on turbidite sedimentation in the basin (Betzler et al. 1991). Minor inputs of volcanic mass-flows derived from subaerial arc volcanism on the Sulu Ridge occurred towards the beginning of the Late Miocene. Sedimentation on the shallower Cagayan Ridge was unaffected by turbidite deposition and accumulated only pelagic and hemipelagic deposits. From the Pliocene to Recent, pelagic and hemipelagic deposits have dominated deposition throughout Sulu Sea, with rare airfall tuff layers.

The Sulu Basin may provide a partial palaeogeographic and sedimentation analogue for the Hill End Trough, with the Sulu Ridge and Cagayan Ridge equivalent to the Capertee and Molong Highs respectively. The undeformed Hill End Trough was of similar dimensions and shows a similar pattern of alternating dominance of volcaniclastic mass-flows dominated by silicic volcanic detritus from the Capertee High; intervals dominated by hemipelagic and airfall tuff deposits (although there are no true pelagic deposits in the Trough); and periods which were dominated by quartzose mass-flows derived from sedimentary basement exposed near the active frontal arc during sea-level lowstands (compare with the Cookman Formation in the Hill End Trough). No drilling has yet been carried out on the Sulu Ridge and it is therefore difficult to compare the sequence here to that of the Capertee High. However, the Cagayan Ridge sequence lacks shallow marine carbonates and clastics and is dominated by hemipelagic and pelagic sequences.

Tectonically, the Sulu Sea setting is complex. Silver and Rangin (1991) interpreted the southeast Sulu Basin opening behind an active Cagayan arc, with volcanism ceasing in early Middle Miocene time, concurrent with collision between the Cagayan arc and the continental Reed Bank-Palawan block. Subduction then shifted to the southeast, making the Sulu Ridge the active arc. The Hill End Trough had a similar pattern of shifting volcanism with much of its sediment in the Late Silurian being derived from volcanic sources on the Molong High to the west, and volcanism shifting mainly to the Capertee High during the Early Devonian (Packham 1968).

5.5.3 Ancient Analogue

5.5.3.1 Waitemata interarc basin, Auckland, New Zealand

The Waitemata interarc basin of northern New Zealand (Fig. 5.15) formed during the Middle to Late Oligocene and was one of the initial responses to the propagation through New Zealand of a convergent plate margin, following a prolonged period of tectonic stability (Ricketts et al. 1989). It consisted of a central deepwater basin 130 km long by 60 km wide, bounded to the north and south by
Figure 5.13 Location of Sulu Basin and the drillholes of Leg 124 of the Ocean Drilling Program. Barbed lines indicate subduction zones and the Cagayan Ridge is highlighted with vertical lines (from Rangin et al. 1990).

Figure 5.14 Simplified lithologic columns for ODP Leg 124 sites in the southeast Sulu Basin and Cagayan Ridge showing relative proportions of interbedded lithologies. Key to symbols used for other lithologies: (1) marl; (2) hemipelagic clay (stone) and silty clay (stone); (3) pelagic brown claystone; (4) terrigenous turbidites; (5) carbonate turbidites; (6) oceanic basement.
basement blocks on which shallow marine shelf facies accumulated, and to the east and west by basement ridges and active calcalkaline arcs, situated above a subduction zone that probably dipped west; the active volcanoes on the arc were around 100 km apart (Ballance 1974; Ricketts et al. 1989) (Fig. 5.16). The western volcanic arc built up a thick, fragmental volcanic pile deposited in largely shallow marine conditions. Lithic, basement-derived sediment entered the basin from the north, mingling with contemporaneous volcanic debris mainly from the western volcanic arc, yielding turbidite facies up to 1000 m thick variously rich in volcanic detritus, basement-derived lithics or a mixture of the two provenances (Ballance 1974) (Fig. 5.17). The basin contains a thin, basal shallow marine facies (Ricketts et al. 1989). Water depths of up to 2000 m were attained in the basin centre (Ballance 1974). Deposition was terminated by uplift in the Middle Miocene (Ballance 1974).

Jones et al. (1983, unpublished manuscript) noted close parallels between the facies, geography, and scale of the Waitemata interarc basin and the Hill End Trough, with the flanking western (Waitakere) and eastern (Coramandel) volcanic arcs corresponding to the Molong and Capertee Highs respectively (Fig. 5.15). The Waitemata interarc basin represents a good analogue for the Capertee High-Hill End Trough-Molong High system. Both basins display thick, rapidly deposited sedimentary successions derived from coeval silicic volcanism and erosion of basement exposed on the trough margins. The Hill End Trough had a longer depositional history, accounting for its much thicker sequence. In contrast to the Hill End Trough, both arcs flanking the Waitemata interarc basin were active during deposition, although the majority of volcanic input came from the west - the opposite of the Hill End Trough (Fig. 5.17).

Neither basin shows evidence of an oceanic basement, indicating sea-floor spreading had not commenced, and both basins were apparently floored by pre-rift sequences. Dimensions are similar, although the northern and southern extent of the Hill End Trough are poorly known. Magnetic interpretation suggests the Hill End Trough closes to the north of Goolma, west of Gulgong, although it is not known how this has been influenced by subsequent deformation. Glen and Watkins (1994) have suggested a basement high in the area of the Bathurst Granite which effectively provided a southern boundary for the Hill End Trough until the deposition of the Cunningham Formation. If these northern and southern constraints are accurate, the dimensions of the Hill End Trough to the north of the Bathurst Granite are broadly comparable to the Waitemata Basin.

5.6 SUMMARY AND CONCLUSIONS

Early Devonian sequences associated with the Capertee High can be broadly subdivided into two categories: (a) Platform sequences, which were deposited on the shallow marine to subaerial platform areas of the Capertee High and have been collectively defined as the c. 4000 m thick Kandos Group; and (b) Western Platform Margin Sequences, which were deposited in slope and base-of-slope settings along the western margin of the Capertee High platform and have been grouped as the
Figure 5.15 Location and regional tectonic setting of the Waitemata interarc basin. From Ricketts et al. (1989).

Figure 5.17 Sediment dispersal patterns in the Waitemata interarc basin. From Ballance (1974).

Figure 5.16 Block diagram of palaeogeography of the Waitemata interarc basin. From Ballance (1974).
Queens Pinch and Limekilns Groups. The latter sequences are transitional westwards into strata of the eastern Hill End Trough.

This thesis has focussed on: (a) elucidating the stratigraphy, depositional processes and environments, and palaeogeography of the Kandos Group (particularly in the Rylstone-Cudgegong area); (b) establishing a biochronological framework for the sequence by obtaining conodont faunas from the carbonates; and (c) attempting correlation of the Kandos Group with platform margin units and Hill End Trough sequences.

The Kandos Group spans virtually the entire Early Devonian and is separated from Late Silurian sequences by a widespread unconformity. The Kandos Group can be accurately correlated to the western platform margin sequences which are dated by conodonts; however, correlation with sequences of the Hill End Trough is imprecise and relies on a scatter of SHRIMP zircon dates (which are inaccurately aligned with conodont zones) and rare macrofaunas.

The Early Devonian evolution of the Capertee High has been divided into three timespans:

(1) **Lochkovian**

The Lochkovian portion of the Kandos Group is approximately 1900 m thick and represents a transgressive to regressive shallow marine sequence deposited on a west- to southwest-sloping shelf flanked to the east by eroding sedimentary basement and active silicic volcanoes, and passing to the west into deepwater facies along the western margin of the High (lower Mullamuddy Formation) and in the eastern Hill End Trough (upper Crudine Group). Sedimentation in the Kandos Group began in the early Lochkovian *woschmidti* Zone and comprised (in ascending order): a transgressive sequence of wave-dominated, fan-deltaic clastics (Warrah Conglomerate); a unit of transgressive biostromal to biohermal carbonates (Clandulla Limestone; *woschmidti* to *eurekaensis* Zones); a transgressive to regressive muddy, storm-influenced shelf deposit (Yellowmans Creek Formation; *eurekaensis* to *delta* Zones); and a thick regressive unit of storm-dominated siliciclastics deposited in middle shelf to near-shore settings (Roxburgh Formation; *delta* to *pesavis* or early *sulcatus* Zones). Facies prograded west during the late Lochkovian regression, culminating in shelf exposure and some erosion in the east. The early Lochkovian transgression may reflect a eustatic sea-level rise, whereas the late Lochkovian regression more likely indicates uplift of nearshore areas ahead of increasingly proximal volcanism.

Large conodont faunas were obtained from the Clandulla Limestone and Yellowmans Creek Formation. These early to middle Lochkovian faunas are unusual amongst Australian Early Devonian faunas in containing an abundance of icriodids. A small *pesavis-sulcatus* Zone conodont fauna was obtained from limestone lenses in the mainly siliciclastic Roxburgh Formation; it proved to be crucial in constraining the age of this unit.
(2) Pragian

Sequences of dacitic to rhyolitic pyroclastics and volcaniclastics up to 1 km thick (Riversdale Volcanics and Huntingdale Volcanics) indicate that voluminous subaerial silicic pyroclastic volcanism dominated the Capertee High during the early to mid Pragian *sulcatus* and *kindlei* Zones. Medium-scale calderas centred southeast of Cudgegong and in the eastern Capertee Valley were flanked by extensive, shallowly-dipping sheets of silicic ignimbrites and epiclastics which probably overlapped, forming a continuous sheet. Short-lived transgressions on the platform resulted in minor carbonate deposition during volcanically quiet periods. Vast volumes of volcanic detritus were transported to the west and deposited as lowstand fans and aprons along the margin of the Capertee High (Mullamuddy and Kingsford Formations) and in the Hill End Trough (Merrions Formation).

Important conodont faunas from the *sulcatus* and *kindlei* Zones were recovered from carbonate lenses in the Riversdale Volcanics and Kingsford Formation.

(3) Emsian

A eustatic transgression in the latest Pragian to earliest Emsian (*pireneae* to *dehiscens* Zones) initially reworked then drowned the existing volcanic topography, creating a broad southwest-sloping shelf with local eroding volcanic islands. Sedimentation in this setting was dominated by a laterally and temporally extensive regime of storm- and tidally-influenced nearshore to shelf siliciclastic sequences deposited in barrier island settings with local biostromal and biohermal carbonate buildups deposited in lagoonal and open-marine environments (Carwell Creek Formation, Myrtle Grove Formation and Mt Frome Limestone). Transgression cut off the supply of clastic detritus from slope and basin areas, and condensed mudstone sequences were deposited along the margin of the Capertee High (Warratra Mudstone and 'Rosedale Shale') and in the Hill End Trough (basal Cunningham Formation). A transgressive pulse in the *perbonus* Zone was followed by regression, local platform exposure and erosion of platform areas in the middle Emsian (late *perbonus* to *inversus* Zones), and a number of carbonate aprons developed along the western margin of the High (lower Sutchers Creek Formation and Ingleburn Formation).

A transgression (possibly eustatic) resulted in further deposition of carbonates on the Capertee High during the late Emsian *serotinus* Zone (Mt Frome Limestone and unreserved equivalents). Regression later in the *serotinus* Zone resulted in erosion of some of these platform carbonates and redeposition in fans along the western margin of the Capertee High (Jesse Limestone and upper Sutchers Creek Formation). This regression appears to continue into the earliest Eifelian *partitus* Zone, when carbonates of the Mt. Frome Limestone were replaced by wave-dominated shoreline and shelf clastics of the Boogledie Formation. Deposition of this clastic sequence continued for an unknown length of time into the early Eifelian before it was terminated by uplift and regression related to a
mildly manifested Middle Devonian deformation which effectively ended the depositional history of the Capertee High.

Biofacies control of conodonts is evident in Emsian sequences: polygnathids are scarce or absent in the shallow, frequently dolomitic carbonates deposited on the platform areas of the Capertee High, which are dominated by low diversity faunas of ozarkodinans, pandorinellids and simple cones. By contrast, polygnathids are relatively abundant in the deeper, western platform margin sequences.

A close modern analogue for the Early Devonian Capertee High-Hill End Trough-Molong High exists in the Sumisu Rift, a nascent intra-oceanic back-arc basin off the coast of Japan which displays marked similarities in facies, geography and scale. Similar facies also occur in the southern Havre Trough where it impinges on the North Island of New Zealand, and in the Sulu Basin between Borneo and The Philippines. An ancient analogue for the Hill End Trough and its margins is the Oligocene-Miocene Waitemata Interarc Basin of New Zealand, which displays close parallels in facies, geography, and scale.

Comparison with these analogues, together with recent gravity and magnetic data obtained during mapping the Dubbo 1:250 000 Sheet, suggest that the Hill End Trough was a small ensialic interarc rift basin which lacked oceanic basement and was floored by Late Ordovician mafic volcanics and volcaniclastics and Early Ordovician quartz turbidites and chert.
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Supporting Paper 1


EVOLUTION OF THE NORTHERN CAPERTEE HIGH


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In the Cudgegong district, Late Ordovician basalt and andesite were deposited as submarine lava and debris flows; geochemical data are typical of calc-alkaline lavas erupted in island arc settings. To the west and northwest of Rylstone, probable Late Ordovician chert and volcaniclastics represent pelagic deposits and turbidites.

No Llandoverian (Early Silurian) strata are known in the district. A thick, shallow marine to emergent, Wenlockian to Ludlovian (Middle to Late Silurian) sequence unconformably overlies Ordovician rocks in the Mudgee-Cudgegong area. The lowermost strata were deposited in a southeast-facing coastal zone and include fluvial channel, subtidal to supratidal flat, and marine shelf deposits. The subsequent eruptive episode produced dacite and rhyolite lava emplaced as a lava dome and a thick volcaniclastic ash and mud flow succession. Near Rylstone, lithologies and fauna are similar to the Cudgegong succession and represent deposition in a northwest-facing coastal zone.

Contacts between Late Silurian and Early Devonian strata are either faulted or unconformable, in the latter case representing the Bowning deformational event. In the Cudgegong-Rylstone district, the Lochkovian to Zlichovian Kandos Group (fine- to coarse-grained clastics, limestone, dacitic ignimbrite, volcaniclastics and lava) were deposited in shallow marine to subaerial environments. Equivalents of the Group can be traced as far as the Capertee Valley in the south and Mount Knowles in the north. In the Queens Pinch Belt, the bulk of the Early Devonian sequence is redeposited and the closely associated shelf and turbidite facies results from deposition adjacent to the western margin of the Capertee High. Middle Devonian strata are apparently developed only in the Mount Frome Limestone.

In the Mudgee-Cudgegong district, a thick Late Devonian sequence overlies older rocks mostly with low angle discordance (Lambian Unconformity). The strata result from a transgression producing initially fluviatile, then shallow marine conditions. The tectonism which deformed the Late Devonian strata predates intrusion of the Middle Carboniferous I-type Aarons Pass Granite (327 Ma), and extrusion of the latest Carboniferous (292 Ma) Rylstone Volcanics which consists of rhyolitic ignimbrite, lava and volcaniclastics.

Most of the region was affected by sub-Permian peneplanation; however, some Late Devonian erosional remnants protrude above the unconformity surface. Thin flat-lying remnants of the Early Permian Snapper Point Formation represent a sandy transgressive sequence near the western margin of the Sydney Basin.
Near the northeast margin of the exposed Lachlan Fold Belt, in the Cudgegong-Mudgee-Rylstone-Kandos district, the northern Capertee High was a shallow marine to subaerial palaeogeographic entity from the Late Silurian until the earliest Middle Devonian. Rocks of the former Capertee High are preserved in the Capertee Anticlinorium, together with Late Ordovician basement, Late Devonian fluvial to shallow marine sediments of the Lambie Group, Middle Carboniferous granitic intrusions and to earliest Permian silicic pyroclastics, lavas and epiclastics of the Rylstone Volcanics. The Permo-Triassic Sydney Basin succession conceals eastern portions of the Capertee High.

The Late Ordovician Cudgegong Volcanics are the products of basaltic and andesitic submarine volcaniclastic debris flows; geochemical data indicate an affinity with calc-alkaline lavas erupted in island arc settings. By comparison, coeval units to the east (Goomber Formation and Lue beds) represent deep water volcaniclastic and quartz-rich turbidites and debris flows, with radiolarian chert. Ordovician strata may represent rifted basement for the Capertee High which was established at the same time as the opening of the Hill End Trough.

Following an Early Silurian hiatus, voluminous shallow water silicic volcanics and associated epiclastics dominated the emerging Capertee High. In the Cudgegong-Mudgee district, a Wenlockian to Ludlovian, shallow marine to emergent succession of clastics and carbonates (Willow Glen Formation) is overlain by thick subaerial dacite lava and breccia (Windamere Volcanics) and laterally equivalent, shallow marine volcaniclastics (Toolamangang Formation), with an uppermost shallow marine shelf unit (Millville Formation). Coeval shallow marine sequences to the east and northeast (Moonbucca Formation and Dungeree Volcanics) formed in a siliciclastic/carbonate mass flow apron which interdigitated with a dacitic volcanic pile.

Extensive Early Devonian strata on the platform areas of the northern Capertee High were dominated by the early Lochkovian to middle Emsian Kandos Group, 4000 m of fine- to coarse-grained clastics, carbonates and silicic volcanics deposited in a variety of shallow marine and subaerial environments which were affected by a number of transgressions and regressions. The Kandos Group is best exposed in the Cudgegong-Rylstone-Kandos district; however, upper units of the Kandos Group identified from the Capertee Valley in the south to the Mount Knowles-Mount Frome area in the north, indicate the Capertee High was an elongate entity persistent during the Early Devonian. Conodont data suggest that sedimentation in the Mount Frome Limestone continued into the earliest Eifelian (Middle Devonian).

By comparison, the late Lochkovian to late Emsian sequences of the Queens Pinch Belt were deposited in fans and aprons adjacent to the western margin of the northern Capertee High and are characterised by mass flow volcaniclastics, siliciclastics and carbonates, with thick mudstone intervals. Vast volumes of detritus were transported to the west across the platform and were redeposited marginal to the Capertee High as well as providing fill for the Hill End Trough.

Regression in the earliest Middle Devonian effectively terminated deposition on the northern Capertee High, although major deformation of the strata did not occur until the Early Carboniferous.

In conclusion, Rocks of the Capertee High occur at the northeast margin of the exposed Lachlan Fold Belt (Fig. 1) and are now largely preserved in the Capertee Anticlinorium.
Packham (1960, 1968, 1969) originally defined the Capertee Geanticline as a discontinuous belt of Late Silurian to Early Devonian shallow marine strata from the Mudgee-Cudgegong district in the north (herein described as the northern Capertee High).
### Figure 2: Stratigraphic sequences on the northern Capertee High

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<th>Lue district</th>
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extending to the south through the Sofala district. Subsequent authors have referred to the Capertee High as a series of volcanic islands with fringing carbonate areas (Powell, 1984); a broad platform with a thin shallow water succession (Pickett, 1982; Powell, 1984); or generalised shallow marine deposition with local land areas (Cas, 1983). The aim of this paper is to provide stratigraphic evidence for the existence and subsequent development of depositional environments of the northern parts of the Capertee High. The result is a synthesis of mapping principally by University of Wollongong students and staff, together with theses from other N.S.W. universities.

The northern Capertee High (Pemberton et al., 1991) encompasses Late Ordovician basement, together with Late Silurian to earliest Middle Devonian sequences (Fig. 2) best known from Mudgee in the north to the Cudgegong-Rylstone-Kandos
district in the southeast (Fig. 3). The district is traversed by the Cudgegong River and
the village of Cudgegong now lies under Lake Windamere.

Capertee High strata continue to the south through the Sofala-Limekilns area and
to the southeast in the Capertee Valley (Fig. 1). Rocks of the northern Capertee High are
faulted against Hill End Trough strata to the west; overlain by the Late Devonian
Lambie Group; intruded by Carboniferous granite; and overlapped in the east by the
Permo-Triassic Sydney Basin succession.

LATE ORDOVICIAN

The oldest strata recognised in the region are Late Ordovician rocks assigned to the
Cudgegong Volcanics (Pemberton, 1989), the Coomber Formation (Colquhoun,
unpubl. data) and the Lue beds (Day, 1961).

The Cudgegong Volcanics (Fig. 4; Table 1) are Late Ordovician (Gisbornian)
basaltic and andesitic arenite, breccia and rare lava deposited by volcaniclastic debris
flows on the flanks of submarine volcanoes with minor fringing carbonate areas
[detailed descriptions in Pemberton (1989 and 1990)]. Basalt and andesite lavas of the
Cudgegong Volcanics have major, trace and rare earth element characteristics typical of
calc-alkaline lavas from island arc settings (Pemberton and Offler, 1985; Pemberton,
1990; Pemberton, in prep.) and their chemistry does not support the shoshonitic affinity

The Coomber Formation, by comparison, consists of thick- and thin-bedded
volcaniclastic, mudstone and radiolarian chert with minor andesite flows, allochthonous
limestone blocks and debris flow conglomerate, deposited as a deep marine volcanic-
lastic apron, with prolonged pelagic periods interrupted by turbidites and debris flows
(Colquhoun, unpubl. data). The extensive, yet poorly known, Lue beds include quartz-
rich clastics and volcanioclastics, chert and minor conglomerate, of probable deep water
origin. Ordovician (latest Darriwilian to earliest Gisbornian) conodonts have been
identified from chert collected from Bara Creek, just west of Lue (Fergusson and
Stewart, in prep.).

The Coomber Formation was a probable lateral equivalent of the Lue beds (Fig. 4)
and its tentative Ordovician age is based on lithological and stratigraphic similarities to
other Ordovician strata in the eastern Lachlan Fold Belt (Cas, 1983; Powell, 1984). The
Coomber Formation and the Lue beds represent deeper water conditions to the east of
the Cudgegong Volcanics. A similar configuration is reported to the south from the
Sofala-Palmers Oakey area (Packham, 1968; Bischoff and Fergusson, 1982; Fell, 1984).

There are no known Llandoveryan strata on the northern Capertee High. Contacts
with the overlying Wenlockian to Ludlovian rocks are probably unconformable,
although not observed. It has been proposed that rifting associated with the Benambran
Orogeny caused the opening of the Hill End Trough and the establishment of the Capertee
High (Gilligan and Scheibner, 1978). The Cudgegong Volcanics would, in this
model, represent rifted basement for the Capertee High.

LATE SILURIAN

Late Wenlockian to Ludlovian rocks crop out in the Carwell Creek area and in the
Cudgegong-Mudgee district. The sequences are recorded, for convenience, as Late
Silurian age based on the twofold division of the Silurian by the IUGS subcommission
on Silurian Stratigraphy (Holland, 1985).

Carwell Creek area

The Moonbucca Formation (Figs 4 and 5; Table 2) consists of mass flow- to
suspension-deposited clastics, including basal conglomerate (with clasts of the Coomber
### TABLE 1 LATE ORDOVICIAN BASEMENT ROCKS ON THE NORTHERN CAPERTEE HIGH

<table>
<thead>
<tr>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>CUDGEGONG VOLCANICS</strong> (McManus et al., 1965; Pemberton, 1989)</td>
<td>1600m - basalt lava (SiO₂ 47-51%), fine- to coarse-grained basaltic arenite and basaltic breccia; andesite lava (SiO₂ 55-56%) and fine- to medium-grained arenite; mudstone. Type section: GR587738 to GR570730</td>
<td>Thin autochthonous marl horizon contains coral (<em>Plasmoporella</em>), alga (<em>Vermiporella</em>).</td>
<td>Volcaniclastic debris flows on flanks of submarine volcanoes with minor fringing carbonate areas. Rare fine laminations, pillow lavas, reworked top of andesite lava provides facing.</td>
</tr>
<tr>
<td><strong>COOMBER FORMATION</strong> (new name) (Campbell, 1981; Colquhoun, unpubl. data)</td>
<td>1750m - volcaniclastic arenite and mudstone, with radiolarian chert and rare andesite lava, conglomerate and autochthonous limestone blocks. Rep. section: GR748652 to GR731642</td>
<td>Correlated with Late Ordovician deep water sequences.</td>
<td>Thick mass flow and suspension deposits in deep marine setting with pelagic periods interrupted by turbidites and debris flows. Slumping, flutes, Bouma AE and ABE.</td>
</tr>
<tr>
<td><strong>LUE BEDS</strong> (Day, 1961; Evans, 1968; O’Donnell, 1972; Southgate, 1975)</td>
<td>1000m - slate, quartz-rich arenite, chert, conglomerate, rare andesite lava</td>
<td>Ordovician (Darriwilian) conodonts in chert (<em>Fergusson and Stewart, 1993</em>).</td>
<td>Proposed deep water Bouma CDE, graded beds.</td>
</tr>
</tbody>
</table>

### TABLE 2 LATE SILURIAN ROCKS ON THE NORTHERN CAPERTEE HIGH

<table>
<thead>
<tr>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>MOONBUCCA FORMATION</strong> (new name) (Campbell, 1981; Colquhoun, unpubl. data)</td>
<td>610m - conglomerate (clasts of Coomber Formation), limestone breccia, mudstone, jasper; dacitic lava and arenite. Rep. section: GR732645 to GR720643</td>
<td>Late Silurian coral (<em>Phaulacis</em>, <em>Tryptops</em>, <em>Zenophila</em>, <em>Propora</em>), trilobite (<em>Pacificurus cf. mitchelli</em>) fauna</td>
<td>NW-facing shallow marine setting with terrigenous imbricated pebbles, carbonate slope and base of geopetals. Graded beds, slope deposits with subaqueous lava and volcaniclastics.</td>
</tr>
</tbody>
</table>
### TABLE 2 (Continued) LATE SILURIAN ROCKS ON THE NORTHERN CAPERTEE HIGH

<table>
<thead>
<tr>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
</tr>
</thead>
</table>
| **DUNGEREE VOLCANICS**
  (Offenberg et al., 1971; O'Donnell, 1972) | 1000m - dacite and rhyolite lava and volcanioclastics, mudstone, rare limestone | No distinctive fauna | Possibly shallow water to subaerial | None recorded |
| **MILLSVILLE FORMATION**
  (Pemberton, 1980b; Pemberton, 1989) | 250m - breccia (limestone, dacite and rhyolite clasts), limestone, dacitic conglomerate. Rep. section: GR501795 to GR497794 | Coral (*Phaulectis, Halysites*), trilobite (*Pacificurus mittelli*), brachiopod (*Kirkidium, Eospirifer, Leptaena, Rhizophyllum*) fauna of Wenlockian to Ludlovian age | Shallow marine shelf with bank, lagoon and beach deposits affected by limestone bank collapse | Geopetals |
| **TOOLAMANANG FORMATION**
  (Pemberton, 1980a; Pemberton, 1989) | 3000m - fine- to coarse-grained arenite (dacitic detritus) with black mudstone and fragmental basalt blocks | Lateral equivalent to, and fragmental version of, Windamere Volcanics | Dense, gravity-driven volcaniclastic ash and mudflows incorporating basement blocks | Slumping, loading |
| **WINDAMERE VOLCANICS**
  (Pemberton, 1980b; Pemberton, 1989) | 1500m - dacite lava (SiO$_2$ 65-69%), arenite and breccia; rhyolite lava. Rep. section: GR530793 to GR521766 | Conformably overlies Willow Glen Formation and overlain by Millsville Formation - both Wenlockian to Ludlovian age | Possible thick lava dome in very shallow marine to subaerial environment | Localised ripple marks, cross beds and scour and fill in arenite |
| **WILLOW GLEN FORMATION**
  (Pemberton, 1980a; Pemberton, 1989) | 1000m - conglomerate, pebbly litharenite and litharenite; fossiliferous mudstone and limestone. Type section: GR627676 to GR636720 | Coral (*Phaulectis, Halysites, Pycnostylus, Desmidiopora*), trilobite (*Pacificurus mittelli*), brachiopod (*Morinorrhynchus, Morristrophia, Kirkidium*) fauna of Wenlockian to Ludlovian age | SE-facing coastal zone includes fluvial channels; subtidal to supratidal flats, affected by transgressive-regressive cycles; and open marine conditions | Fining-up sequences, cross-beds, scour channels, imbricated pebbles, geopetals, evaporites |
Formation), litharenite, limestone breccia and calcarenite (with a late Wenlockian to Ludlovian fauna), mudstone and allochthonous jasper and limestone blocks. These lithologies interdigitate with a northward thickening unit of dacite lava, autoclastics and intraformational volcanoclastic (Colquhoun, unpubl. data). The depositional setting was a subaqueous mass flow apron with an associated silicic volcanic pile which accumulated at shallow depths on a northwest facing slope. Poorly known dacite lava and volcanioclastics, with minor mudstone, conglomerate and limestone, near Lue (the Dungeree Volcanics) are probable lateral equivalents of the Moonbucca Formation (Fig. 4).

Fig 4. Late Ordovician and Late Silurian units on the northern Capertree High.

Cudgegong-Mudgee district

A similar, yet much thicker, succession of conformable Wenlockian to Ludlovian shallow marine to probably emergent units occurs in the Cudgegong-Mudgee district
[Figs 4 and 5; Table 2; detailed descriptions in Pemberton (1989 and 1990)]. The Willow Glen Formation (conglomerate, pebbly arenite and arenite; mudstone and limestone with a late Wenlockian to early Ludlovian fauna) was deposited in a southeast-facing coastal zone which included fluvial channels, subtidal to supratidal flats (Jones et al., 1987) and more open marine shelf conditions (Fig. 6A).
Fig. 6. Palaeogeographic reconstructions for A - Willow Glen Formation; and B - Windamere Volcanics, Toolamanang Formation and Millsville Formation.

The overlying Windamere Volcanics are characterised by thick, undifferentiated dacite lava and breccia, with common rhyolite horizons, possibly emplaced as a subaerial to very shallow marine lava dome or perhaps cryptodome. The Toolamanang
Formation is a thick succession of fine- to coarse-grained lithic arkose to feldspathic litharenite with common very fine ash-sized horizons, fine-grained breccia and sporadic basaltic blocks. The unit is considered a lateral equivalent to, and a fragmental version of, the Windamere Volcanics produced by dense, gravity-driven, shallow marine volcaniclastic flows of ash-sized dacitic detritus, with associated mudflows of the finer detritus, both of which incorporated some eroded basement material (Fig. 6B).

This was followed by a period of localised volcanic quiescence during which breccia (limestone, dacite and rhyolite clasts), limestone, mudstone, and dacitic conglomerate and breccia of the Millville Formation were deposited. The limestone and mudstone contain a Wenlockian to Ludlovian fauna with similarities to that of the Willow Glen Formation. The unit was deposited on a shallow marine shelf with initially extensive bank and lagoon deposits subsequently affected by limestone bank collapse into slightly deeper water.

In the Sofala district, thinner and more localised equivalents of the Willow Glen Formation (the Tanwarrra Shale) and the Windamere Volcanics-Toolamanang Formation (the Bells Creek Volcanics) are recognised (Packham, 1968; Bischoff and Fergusson, 1982). There is no Millville Formation equivalent in the Sofala district as the Bells Creek Volcanics are conformably overlain by deep water Hill End Trough strata. Thick sequences of volcaniclastics, with silicic and intermediate lava, mudstone and limestone, crop out between Mudgee and Gulgong (the Eurundury Formation; Offenberg et al., 1971; Armstrong, 1983) and between Gulgong and Dunedoo (the Tucklan beds; Offenberg et al., 1971); these poorly known sequences remain undated and their relationship to the Late Silurian sequences in the Cudgegong-Mudgee district is unknown.

Comparison of the Ordovician and Silurian strata on the northern Capertee High shows clear differences in geochemistry and depositional environments. From the Wenlockian, voluminous shallow water silicic volcanics and associated epiclastics typify activity on the emerging Capertee High.

**EARLY DEVONIAN**

Early Devonian strata crop out extensively on the northern Capertee High in the Cudgegong-Rylstone-Kandos district and to the north of Ilford; in the Queens Pinch Belt; and in the Mount Frome-Mount Knowles district.

**Kandos Group in the Cudgegong-Rylstone-Kandos district**

An essentially conformable succession of Early Devonian (early Lochkovian to middle Emsian) shallow marine to subaerial units in the Cudgegong-Kandos district was redefined as the Kandos Group (Figs 7 and 8; Table 3), by Pemberton et al. (1991), named after Sussmilch's (1934) Kandos Series.

*Warrah Conglomerate* (Campbell, 1981)

The early Lochkovian Warrah Conglomerate is restricted to the southeast of the district (Fig. 7) where it overlies Late Silurian strata with a disconformable to low-angle unconformable contact. The unit is a fining-upwards sequence of pebble-granule conglomerate (containing clasts of various elements of the Moobucca Formation) and pebbly litharenite with subordinate litharenite, mudstone and limestone, the latter containing a sparse benthonic fauna (Table 3). The strata represent a transgressive, nearshore, shallow marine sequence in a wave-dominated, fan delta setting.

*Clandulla Limestone* (Pemberton et al., 1991)

The overlying early to middle Lochkovian Clandulla Limestone is best exposed in the Kandos quarries (Fig. 7) where conformable contacts with the underlying Warrah
Fig. 7. Geology of the Kandos Group between Cudgegong and Rylstone.

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Conglomerate and overlying Yellowmans Creek Formation, and the succession through the formation, are evident (Cook, 1988b). The unit comprises extensive shallow marine (depths of 0 to 24 m) carbonate deposition with biostromal and biothermal buildup dominated by stromatoporoid-Thamnopora communities and associated lagoon muds, forereef, interreef and backreef facies. The formation grades upwards into the Yellowmans Creek Formation reflecting deepening and a transition to more clastic sedimentation.

Yellowmans Creek Formation (Millstead, 1985)

The middle Lochkovian Yellowmans Creek Formation crops out extensively in the Carwell Creek valley and to the east of Cudgegong as mudstone, calcareous mudstone, limestone, sublitharenite and volcaniclastic arenite and breccia. The sequence indicates fairweather suspension deposition of fine-grained siliciclastic detritus, together with carbonate mounds, periodically interrupted by distal storm-generated incursions of quartz-rich sand and minor volcaniclastic debris flows on a low energy, subtidal muddy shelf. The formation represents the peak of the transgressive cycle which began during deposition of the Warrah Conglomerate and continued through the Clandulla Limestone. An increasing proportion of coarser grained clastics near the top of the unit indicates shallowing associated with initiation of regressive conditions which continued during deposition of the overlying sandstone-dominated Roxburgh Formation.

Roxburgh Formation (Pemberton, 1980a)

The late Lochkovian-early Pragian Roxburgh Formation occurs to the east of Cudgegong, along the Carwell Creek valley and to the north of Ilford (Colquhoun, in prep.). The formation thickens remarkably from 50 m in the east at an offlap contact with the Yellowmans Creek Formation near Carwell Creek (Fig. 8B) to a maximum of 750 m just east of Cudgegong (Fig. 8A). Lithologies include quartzarenite and sublitharenite with subordinate mudstone, litharenite, conglomerate, accretionary lapilli tuff and rare limestone. The unit yields, at several localities, an abundant shallow marine fauna dominated by brachiopods (Table 3).

Facies relationships indicate numerous cycles of small-scale (4 to 60 m thick) progradation and minor retrogradation superimposed on an overall upward-shallowing (regressive) trend (Colquhoun, in prep.). The depositional setting was a storm-dominated, southwest-sloping, shallow marine shelf, with outer to inner shelf, shoreface and coastal plain environments, flanked to the east by an eroding silicic volcanic terrain with some active volcanism, albeit distal to the setting. Palaeocurrent data indicate storm waves approached from the west-northwest producing dominantly southeast-directed longshore (on the shoreface) and southwest-directed offshore return currents over the shoreface and shelf.

Riversdale Volcanics (Wright, 1966)

The Pragian Riversdale Volcanics crop out extensively in the south of the district (Fig. 7) as dominantly welded, and lesser non-welded, purple dacitic ignimbrite (Fig. 8A). The unit thins very markedly to the west and north where it consists of dacite lava, immature coarse epiclastics and airfall tuff. Variation in both thickness and facies suggests emplacement of thick silicic intracaldera ignimbrites, with an explosive source to the south, which pass laterally to thin, coeval, shallow marine outflow equivalents. The base of the Volcanics is locally disconformable and implies a period of subaerial exposure prior to deposition of the thick subaerial ignimbrites. Towards the top of the sequence, reworking of dacitic detritus in a fan delta setting indicates the start of
### TABLE 3 EARLY DEVONIAN ROCKS OF THE KANDOS GROUP

<table>
<thead>
<tr>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>CARWELL CREEK</td>
<td></td>
<td></td>
<td>Low angle cross-beds</td>
</tr>
<tr>
<td>FORMATION (Offenberg et al., 1971; Pemberton, 1980a; Booth, 1990; Cook, 1988a, 1990; Millsteed, 1985, 1992; Colquhoun, unpubl. data)</td>
<td>Glendale Limestone Member - basal 180m - limestone, dolomitised limestone, conglomerate - overlain by 1050m litharenite, crinoidal litharenite, conglomerate, limestone, dacitic ignimbrite. Rep. section: GR681685 to GR688687</td>
<td>Non-diagnostic corals (favositids) and stromatoporoids.</td>
<td>Backreef to lagoon interrupted by volcaniclastic mass flows Shoreface to offshore storm-influenced setting with shoalwater limestone and minor volcanics</td>
</tr>
<tr>
<td>RIVERSDALE VOLCANICS (Wright, 1966; Pemberton, 1980a; Cook, 1988a, 1990; Millsteed, 1985, 1992; Colquhoun, unpubl. data)</td>
<td>1250m - purple dacitic ignimbrite with minor lava, litharenite, conglomerate, accretionary lapilli tuff. Type section: GR709607 to GR690603</td>
<td>Early Devonian brachiopods (Iridistophia) in litharenite near top</td>
<td>Thick subaerial intra-caldera ignimbrites passing laterally to thin outflow ignimbrites, epiclastics and lava. Marine incursion towards top</td>
</tr>
<tr>
<td>ROXBURGH FORMATION (Pemberton, 1980a; Booth, 1990; Cook, 1988a, 1990; Millsteed, 1985, 1992; Colquhoun, in prep.)</td>
<td>750m - quartzarenite, litharenite, mudstone. Rep. sections: GR657629 to GR709607</td>
<td>Early Devonian brachiopod-conodont-trilobite-mollusc fauna. Full faunal list in Colquhoun (in prep.)</td>
<td>Upward shallowing storm-dominated siliciclastics from middle-outer shelf to subaerial coastal plain setting</td>
</tr>
</tbody>
</table>

Localised erosional base; accretionary lapilli; fiamme; cross-bedded, bioturbated, laminated litharenite

Marine incursion towards top

Low angle tabular and trough cross-beds, hummicky cross stratification (HCS), wave ripples, bioturbation

Low angle cross-beds

Storm-influenced setting with shoalwater limestone and minor volcanics

Shoreface to offshore storm-influenced setting with shoalwater limestone and minor volcanics

Upward shallowing storm-dominated siliciclastics from middle-outer shelf to subaerial coastal plain setting

Low- to high-angle cross-beds, graded storm beds with rippled tops and scoured bases, intense bioturbation, abundant HCS
<table>
<thead>
<tr>
<th>ROCKS OF THE KANDOS GROUP</th>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
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<tbody>
<tr>
<td><strong>TABLE 3 (Continued)</strong></td>
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<tr>
<td><strong>YELLOWMANS CREEK</strong></td>
<td>500 m - calcareous mudstone, limestone, interbedded mudstone and litharenite, breccia. Type section: GR723616 to GR719613</td>
<td>Early Devonian brachiopods including (<em>Cyrtina, Anastrophia, Skenidiodes</em>) and trilobites (<em>Apocalyzene</em>) with ?middle Lochkovian (<em>eurekaenensis Zone</em>) conodonts (Colquhoun <em>et al.</em>, in prep.)</td>
<td>Quiet subtidal conditions with distal storm beds and carbonate deposition. Near top change to regressive cycle with storm deposits and very shallow conditions</td>
<td>Ripples, HCS in upper parts, bioturbation</td>
</tr>
<tr>
<td><strong>CLANDULLA LIMESTONE</strong></td>
<td>170 m - carbonate mudstone and variety of limestone types. Type section: GR735600 to GR736598</td>
<td>Early Devonian corals, stromatoporoids (<em>Amphipora</em>); brachiopods (<em>Machaeraria, Cyrtina, Salopina, Isorthis</em>). <em>woschmidtii</em> to <em>eurekaenensis</em> zone conodonts (Colquhoun <em>et al.</em>, in prep.)</td>
<td>Transgressive reef development from lagoon carbonate muds through colonisation stage of extensive bioherms and forereef, interreef and backreef facies</td>
<td>Biostromal to biotermal buildup flanked by well bedded limestones</td>
</tr>
<tr>
<td><strong>WARRAH CONGLOMERATE</strong></td>
<td>280 m - conglomerate (Late Silurian clasts), pebbly litharenite, litharenite, mudstone and limestone lenses. Rep. section: GR726610 to GR725609</td>
<td>Favositid corals and stromatoporoids (<em>Amphipora</em>)</td>
<td>Transgressive nearshore shallow marine sequence in a wave-dominated fan delta</td>
<td>Low angle cross-beds, normal grading, wave ripples, bioturbation</td>
</tr>
</tbody>
</table>
another transgressive cycle. The reworked horizon contains the brachiopod *Iridistrophia*, which is found throughout the late Lochkovian-early Pragian Roxburgh Formation, suggesting that any hiatus between the units was not prolonged. Furthermore, the Volcanics pass conformably into the Emsian Carwell Creek Formation, thereby constraining the age of the Volcanics to Pragian-early Emsian.
Carwell Creek Formation (Offenberg et al., 1971)

Stratigraphically the highest and most areally extensive unit of the Kandos Group, the early to middle Emsian Carwell Creek Formation crops out widely through the Cudgegong-Kandos district (Fig. 7) with probable equivalents to the south in the Capertee Valley and north as far as Mount Knowles. A discontinuous series of basal limestone and dolomite is defined herein as the Glendale Limestone Member (Table 3). The member and its lateral equivalents conformably overlie the Riversdale Volcanics, except in the east where the Carwell Creek Formation onlaps both the Clandulla Limestone and Late Silurian rocks (Fig. 8B). Maximum preserved thickness is 1230 m in the northeast; however, the top of the formation is nowhere exposed.

Lithologies include litharenite, crinoidal litharenite, pebbly litharenite and conglomerate as well as limestone and dolomite. The unit contains an abundant shallow marine fauna. The brachiopods Buchanathyris and Spinella suggest an earliest Emsian (dehiscens Zone) age (Garratt and Wright, 1988) and, in addition, dehiscens to perbonus Zone conodonts have been recovered from a number of localities (Table 3). The strata represent foreshore to inner shelf deposits influenced by fairweather- and storm-wave activity. The Glendale Limestone Member was mostly deposited in a restricted lagoon, suggesting a system of barrier islands and carbonate banks sheltering parts of the coastline. The setting was a broad, gently southwest-sloping shelf with rapid erosion of the Ordovician basement and the Riversdale Volcanics to the east and south. Some dacitic volcanoes remained active, possibly forming islands in the shallow shelf. The depositional phase was dominantly transgressive punctuated by brief regressive intervals and apparently long periods approaching aggradation.

Transgressive/regressive patterns recorded from the Kandos Group are similar to patterns recorded from various Australian basins, as well as on a world wide scale. Precise correlation of these trends, based on on-going conodont studies (Colquhoun), with eustatic sea level curves suggests a strong eustatic rather than tectono-sedimentary control on sedimentation patterns (Colquhoun et al., in prep.).

Kandos Group north of Ilford

Early Devonian strata containing elements of both the Kandos Group and marginal Capertee High successions crop out in limited exposures to the north of Ilford (Booth, 1990; Colquhoun, unpubl. data). The strata are faulted against the Toolamanang Formation in the west and unconformably overlain by Permian strata in the north, east and south (Fig. 9). The basal Warrah Conglomerate comprises up to 280 m of pebble-granule conglomerate, with litharenite horizons, deposited in a fluvial to shallow marine setting. The unit is ?conformably overlain by the Roxburgh Formation, 325 m of quartzarenite, sublitharenite and mudstone. Equivalents of the Clandulla Limestone and Yellowmans Creek Formation are missing from the Ilford area, 8 km south of the Carwell Creek area.

The Roxburgh Formation is apparently faulted against the Kingsford Formation (new name; Colquhoun, unpubl. data), at least 250 m of slope-deposited, mass flow volcaniclastic arenite and limestone breccia, with hemipelagic mudstone, and allochthonous and possibly autochthonous limestone. A thin clastic horizon towards the top of the formation contains a rich benthic fauna including the zonal brachiopod Boucotia australis (Booth, 1990) suggesting a late Lochkovian age (Garratt and Wright, 1988). Preliminary conodont studies (Colquhoun et al., in prep.) suggest a latest Lochkovian to early Pragian age (pesavis to sulcatus Zone). The Kingsford Formation is apparently a correlative of the lower parts of the Mullamuddy Formation of the Queens Pinch Belt, approximately 25 km to the northwest.
### TABLE 4 EARLY DEVONIAN ROCKS OF THE QUEENS PINCH BELT

<table>
<thead>
<tr>
<th>Formation</th>
<th>Thickness/Rock Types</th>
<th>Age</th>
<th>Environment</th>
<th>Structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ingleburn Formation</td>
<td>600m - dacite lava and breccia, mudstone, quartzarenite, arkose, litharenite, conglomerate (dacite and arenite clasts)</td>
<td>Coral and brachiopod fauna indicates late Pragian to early Emsian. Conodonts including <em>Polynathus inversus</em> indicate middle Emsian age (<em>inversus</em> Zone)</td>
<td>Shallow water slope setting with mass flow conglomerate and silicic volcanics</td>
<td>Finely laminated mudstone, cross bedded arenite at top of unit. Strongly cleaved mudstone</td>
</tr>
<tr>
<td>Sutchers Creek Formation</td>
<td>180m - mudstone to pebbly mudstone with allochthonous limestone blocks</td>
<td>Abundant tetracoral and brachiopod fauna, <em>perbonus</em> Zone (<em>Polynathus perbonus</em>) and <em>serotinus</em> Zone (<em>P. serotinus, P. inversus</em>) conodonts; at the top <em>Pandorinellina</em> cf. <em>steinhornensis</em> steinhornensis (McCracken, 1990)</td>
<td>Carbonate debris flow apron with extensive hemipelagic periods</td>
<td>Finely laminated mudstone with convoluted bedding. Graded arenite (Bouma AE and ABE)</td>
</tr>
<tr>
<td>Warratra Mudstone</td>
<td>120m - mudstone Abundant brachiopod fauna including <em>Spinella</em> indicates latest Pragian to earliest Emsian</td>
<td></td>
<td>Hemipelagic suspension deposits</td>
<td>Finely laminated, strongly cleaved mudstone</td>
</tr>
<tr>
<td>Taylors Hill Formation</td>
<td>240m - mudstone, volcanic, basaltic conglomerate, allochthonous limestone</td>
<td>Small coral, brachiopod and conodont fauna, including <em>Nadiastophia</em> and <em>Martinophyllum</em>, indicates middle to late Pragian age</td>
<td>Prolonged hemipelagic suspension periods interrupted by turbidity currents and mass flows</td>
<td></td>
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</tbody>
</table>
TABLE 4 (Continued) EARLY DEVONIAN ROCKS OF THE QUEENS PINCH BELT

<table>
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<th>structures</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>MULLAMUDDY</strong> FORMATION (Wright, 1966; McCracken, 1990)</td>
<td>780m - volcaniclastic arenite and conglomerate (clasts of dacite, limestone [some of which are Silurian] and basalt), mudstone, basaltic lava and breccia</td>
<td>Brachiopod, coral and trilobite fauna. Basal fauna of <em>Cyrtinopsis</em>, <em>Dolerorthis</em>, <em>Plectatrypa</em> (late Lochkovian). Higher beds include <em>Nadiastrophia</em>. Allochthonous limestone early Pragian (<em>sulcatus</em> Zone) conodonts (<em>Eognathodus</em>).</td>
<td>Submarine volcaniclastic apron, with mass flow deposits alternating with hemipelagic muds, and basalt lava and reworked material</td>
</tr>
</tbody>
</table>

TABLE 5 EARLY TO MIDDLE DEVONIAN ROCKS OF THE MOUNT FROME AREA

<table>
<thead>
<tr>
<th>thickness/rock types</th>
<th>age</th>
<th>environment</th>
<th>structures</th>
</tr>
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<tbody>
<tr>
<td><strong>BOOGLEDIE</strong> FORMATION (Wright, 1966; Offenberg et al., 1971; Hewitt, 1975; Lyons, 1976; Malnic, 1978)</td>
<td>950m - fine- to coarse-grained litharenite, quartzarenite, crinoidal litharenite, pebbly litharenite, conglomerate, mudstone, limestone</td>
<td>Coral and brachiopod fauna includes <em>Spinella</em> and <em>Maluropothia</em>, and <em>perbonus</em> Zone (late early Emsian) conodonts</td>
<td>Mixed siliciclastic and carbonate shelf and nearshore conditions</td>
</tr>
<tr>
<td><strong>MOUNT FROME LIMESTONE</strong> (Wright, 1966, 1969, 1981; Pickett, 1978; 1987)</td>
<td>178m - calcarenite with silty calcareous interbeds</td>
<td>Rich tabulate and rugose coral fauna. Conodonts (Colquhoun <em>et al.</em>, in prep.) demonstrate beds represent most of Emsian up to earliest Eifelian (<em>perbonus</em> to <em>partius</em> Zones)</td>
<td>Biostromal setting. Birdseye and fenestral textures suggest intertidal or supratidal conditions at times</td>
</tr>
</tbody>
</table>
Fig. 9. Geology of A — the Kandos Group north of Illford; B — the Mount Frome area; and C — the Queens Pinch Belt.

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Queens Pinch Belt
The Queens Pinch Belt of Early Devonian strata, centred 16 km southeast of Mudgee, is separated from the Hill End Trough sequence to the west by the Wiagdon Thrust, and to the east and south by faulted and locally disconformable contacts with Late Silurian rocks (Fig. 9). The sequence at Queens Pinch consists of sedimentary and volcanic rocks which were mostly deposited in slope and base of slope settings along the western margin of the northern Capertee High. The Devonian sequence is preserved in a number of fault-bounded, wedge-shaped blocks which are internally characterised by numerous, open to tight, gently plunging folds, with local overturning on the limbs (Wright, 1966). A number of stratigraphic, often fault-bounded and richly fossiliferous units have been recognised.

Mullamuddy Formation (Wright, 1966)
The formation consists of approximately 780 m of thickly bedded volcanioclastic arenite and conglomerate, lenses of conglomerate containing limestone blocks, mudstone, and a basalt lava horizon and fine- to very coarse-grained fragmental equivalents (Wright, 1966; Powis, 1975; McCracken, 1990). The basalts are petrographically and chemically similar to basalts from the Ordovician Cudgegong Volcanics.

Mainly brachiopod faunas have been recorded from two horizons (Wright, 1966, 1969). The lower fauna is indicative of a late Lochkovian age (*australis* Zone; Garratt and Wright, 1988). Conodonts obtained by McCracken from limestone clasts from stratigraphically higher conglomerates indicate a maximum age of *sulcatus* Zone (early Pragian). The higher of the two shelly faunas contains the stratigraphically important brachiopod *Nadiastrophia*, indicative of the Pragian *Nadiastrophia-loyolensis* zone (Garratt and Wright, 1988).

The depositional environment is envisaged as a volcanioclastic apron with mass flows of arenite, conglomerate, and limestone blocks alternating with quieter periods dominated by hemipelagic suspension deposition of mud. The unit was deposited during a period of major subaerial silicic volcanism on the platform of the Capertee High, accounting for the sporadic, high volume influxes of volcanioclastic sediment. The unit probably formed part of a low stand system tract apron or fan which formed along the western Capertee High margin during the early to middle Pragian as volcanic material was rapidly redeposited from the largely subaerial shelf areas.

Taylors Hill Formation (Wright, 1966)
The formation conformably overlies the Mullamuddy Formation and comprises up to 240 m of dark mudstone, thinly bedded volcanarenite, lenses of basaltic conglomerate and fossiliferous, massive to well-bedded, allochthonous and possibly also allodapic limestone (Wright, 1966, 1969). Good, although comparatively small, faunas were obtained by Wright (1966) from the base and near the faulted top of the unit, and may both be correlated (due to the presence of *Nadiastrophia* and the tetracoral *Martinophyllum*) with the *Nadiastrophia-loyolensis* assemblage of Garratt and Wright (1988), which spans most of the Pragian. In addition, a Pragian (possibly *sulcatus* to *kindlei* Zones: Garratt and Wright, 1988) conodont fauna is known from this unit. Depositional environments were similar to the Mullamuddy Formation with relatively prolonged periods of hemipelagic suspension and only minor turbidity currents of arenite. The decrease in volcanic material may indicate a cessation of volcanism, and possibly higher sea levels, on the platform areas of the Capertee High.
Warratra Mudstone (Wright, 1966)

This unit was originally defined by Wright (1966) as 120 m of grey mudstone underlying the Sutchers Creek Formation. McCracken (1990) altered the definition to include limestone conglomerate, lithic arenite and mudstone which Wright had placed in the overlying Sutchers Creek Formation. Here we recognise the two formations in their original sense. The Warratra Mudstone has a sparse macrofauna including rare Spinella; which is important in indicating a dehiscens Zone (earliest Emsian) age (Garratt and Wright, 1988). The unit was deposited during a major transgressive phase, which was responsible for deposition of part of the Carwell Creek Formation and its correlatives.

Sutchers Creek Formation (Wright, 1966)

The formation consists of up to 180 m of limestone conglomerate and blocks, usually commencing with an arenaceous turbidite unit; pebbly mudstone intervals and allodapic limestone are equally characteristic of the unit, which conformably overlies the Warratra Mudstone (Wright, 1966; McCracken, 1990). Wright (1966, 1969) listed extensive tetracoral and brachiopod faunas. Garratt and Wright (1988) noted perbonus Zone conodonts whereas McCracken recorded perbonus to early sentinus Zone conodonts from allodapic limestone, thus indicating a longer age span for the unit. Depositional environments were similar to those of the Warratra Mudstone yet with more and coarser lithic and limestone detritus; that is, a carbonate debris flow apron with extensive periods of hemipelagic sedimentation. The increase in carbonate detritus probably reflects regression and subsequent erosion of platform carbonate from the shelf areas of the Capertee High providing detritus for the Sutchers Creek Formation.

Ingleburn Formation (Wright, 1966)

The unit crops out as a fault-bounded block which contains 550 to 600 m of basal dacite lava, fossiliferous mudstone, conglomerate (with dacite clasts), andesite lava, cross-bedded quartzarenite and conglomerate (with dacite and quartzarenite clasts) at the top (Wright, 1966, 1969). Original age determination based on coral and brachiopod faunas (Wright, 1966, 1969) suggested a position in the sequence between the Taylors Hill Formation and the Warratra Mudstone (i.e., late Pragian-earliest Emsian). Garratt and Wright (1988) subsequently recorded inversus Zone (middle Emsian) conodonts, thus indicating partial equivalence with the Sutchers Creek Formation. The sampled limestone may be allochthonous and hence would give a maximum age for the unit. The depositional environment was at most an outer shelf/slope setting dominated by mass flows producing conglomerate and arenite, alternating with suspension sedimentation. This unit contains the youngest primary volcanics on the Capertee High as no correlatives from the platform areas are known; thus volcaniclastic deposition marginal to the Capertee High continued until very late in the Early Devonian.

Mount Frome-Mount Knowles area

In the vicinity of Mount Frome (Fig. 9), a distinctive Early to Middle Devonian sequence was recognised by Wright (1966), consisting of two formations: the lower Mount Frome Limestone, and the overlying Boogledie Formation (originally named the Melrose Formation).

Mount Frome Limestone (Wright, 1966)

The Mount Frome Limestone (6 km east of Mudgee) has been accorded either formation status (Wright, 1966) interdigitating with the Boogledie Formation or member status in the Boogledie Formation (Hewitt, 1975; Malnic, 1978; Powell and...
Edgecombe, 1978). Some authors (e.g., McCracken, 1990) suggested that the limestone may be allochthonous; however, conformable, autochthonous relations with surrounding sediments can be demonstrated. The Mount Frome Limestone is between 178 m (Wright, 1966, 1981) and 238 m (Pickett, 1978, 1987) thick, depending on whether separate outcrops near the Cudgegong River are included. The unit consists of beds of dark blue-grey calcarenite, from several cm to at least 6 m thick, separated by thick, sparsely fossiliferous, silty calcareous intervals. The Limestone contains rich tabulate and rugose coral and subordinate brachiopod faunas. A biostromal depositional environment, rich in food and oxygen, is envisaged (Wright, 1966). Birdseye textures are present, typically in the middle and lower parts, indicating ephemeral intertidal or supratidal conditions. Conodont studies (Philip, 1974; Pickett, 1978, 1987; Wright, 1981; Wright and Flory, 1981; Garratt and Wright, 1988; Strusz et al., 1988) collectively indicate the Mount Frome Limestone spans most of the Emsian with ages ranging from perbonus Zone (late early Emsian) for the basal beds, extending probably into the partitus Zone of the earliest Eifelian (Middle Devonian) in the uppermost 25 to 30 m.

Boogledie Formation (Offenberg et al., 1971)

A sequence largely consisting of clastic strata occurs to the east and northeast of Mudgee; this has been collectively assigned to the Boogledie Formation (Offenberg et al., 1971; and subsequent workers), a formation recognised initially on the basis of outcrops on the western side of the Mount Frome syncline and extended by later workers to include extensive outcrops around Eurundury, Budgee Budgee and Home Rule to the north. At Mount Frome, the unit has been interpreted as exhibiting a low angle unconformable contact with the Late Devonian Buckaroo Conglomerate (Powell and Edgecombe, 1978; Pickett, 1978). The base of the unit is poorly exposed, but locally developed unconformable contacts with poorly known Siluro-Devonian units have been reported with the Tingha Formation (Armstrong, 1983) and the Budgee and Eurundury Formations (Malnic, 1978). At Mount Frome, the formation interdigitates with the upper beds of the Mount Frome Limestone.

The Boogledie Formation consists of up to 950 m of fine- to coarse-grained, rarely fossiliferous, litharenite and quartzarenite (with small- and large-scale cross bedding and both symmetrical and asymmetrical ripple marks), lenses of pebbly arenite and pebble to cobble conglomerate, mudstone, and numerous small and large lenses of limestone, dolomitic limestone and dolomite (Wright, 1966; Hewitt, 1975; Lyons, 1976; Malnic, 1978).

Strata in the vicinity of Mount Knowles are included in the Boogledie Formation by Offenberg et al. (1971). The Mount Knowles Member (Wright, 1966; Hewitt, 1975) consists of lenses of thick or thin bedded limestone and dolomite (up to 120 m thick) and impure sandy limestone which crop out in the upper parts of the formation. Faunas vary from sparse (in the dolomite) to abundant (in the limestone) and are dominated by tabulate corals, stromatoporoids, receptaculitids, and minor tetracorals, with brachiopods locally abundant in thin beds; important brachiopods include Spinella and Malurostrophia. Pickett (1972) reported a probable perbonus Zone (late early Emsian) conodont fauna from the member, and further studies are in progress (Colquhoun). However, at Mount Knowles, the development of dolomite and quartz-rich crinoidal arenite are strongly reminiscent of the Carwell Creek Formation and the Mount Knowles member is probably a correlative of the Carwell Creek Formation which has been traced as far north as west of Lue (Rogis, 1974; Southgate, 1975; Lyons, 1976) and the 'Erang' property near Windamere Dam wall (Wright et al., in prep.).

The Boogledie Formation represents a mixed siliciclastic/carbonate shelf and nearshore sequence, probably an environment similar to that of the Carwell Creek
| **TABLE 6 LATE DEVONIAN ROCKS ASSOCIATED WITH THE NORTHERN CAPERTREE HIGH** |
|-----------------|-------|-----------------|-----------------|
| thickness/rock types | age | environment | structures |
| **DERALE SANDSTONE** (Wright, 1966; Killick, 1987) | 800m - quartzarenite and sublitharenite | Late Devonian brachiopods | Storm-dominated shallow marine shelf |
| **LAWSONS CREEK SHALE** (Wright, 1966; Killick, 1987; Millsteed, 1985, 1992) | 400m - thinly bedded mudstone with subordinate quartzarenite | Late Devonian brachiopods (cyrtospiriferids, *Microspirifer*) and *Lepiophloeum* flora | Upper offshore to shallow muddy shoreface with storm sand beds |
| **BUMBERRA FORMATION** (Wright, 1966; Killick, 1987; Cook, 1988a, 1990; Millsteed, 1985, 1992) | 400m - sublitharenite, quartzarenite and mudstone | Late Devonian (?Fammenian) fauna (cyrtospiriferid, chonetid and productid brachiopods) | Tidal flats (subtidal, intertidal and tidal channels) and wave-dominated foreshore to lower shoreface. Overall upward deepening trend |
| **BUCKAROO CONGLOMERATE** (Wright, 1966; Killick, 1987; Cook, 1988a, 1990; Millsteed, 1985, 1992) | 420m - purple conglomerate (chert, silicic volcanics, quartzarenite clasts) to pebbly litharenite and mudstone | ?Frasnian | Fluvial, braided river |

**STRATIGRAPHY AND ENVIRONMENTS OF CAPERTREE HIGH**

- *HCS*, low angle cross-beds with palaeocurrent directions to east
- Intensely bioturbated and well laminated with rare ripples, cross-beds, HCS. Palaeocurrent directions to east and northeast
- Cross-beds, ripples, parallel lamination, flaser bedding, bioturbation. Palaeocurrent directions to northeast
- Cross-beds, fining upwards cycles, red non-bioturbated mudstone with mudcracks. Palaeocurrent directions to northeast
Formation. Provenance is similarly a mixed volcanic/cratonic source (Wright, 1969; Hewitt, 1975; Lyons, 1976). Thick accumulations of richly fossiliferous, mainly biostromal carbonate indicate areas and times of limited clastic supply and suggest that, in general, more open marine, less restricted conditions prevailed than was the case in the Carwell Creek Formation carbonates.

**LATE DEVONIAN STRATA**

The Late Devonian Lambie Group is represented by a transgressive-regressive sequence at least 1700 m thick, of basal Buckaroo Conglomerate conformably overlain by a ?Famennian succession of Bumberra Formation, Lawsons Creek Shale and Derale Sandstone which crops out in a southeast-plunging syncline (the Pine Ridge syncline) from Mudgee to the west of Rylstone (Fig. 3; Wright, 1966; Killick, 1987).

The Buckaroo Conglomerate (Table 6) includes several fining-upwards cycles of red-purple conglomerate overlain by medium- to fine-grained litharenite and red mudstone (with mudcracks) and represents braided river deposits (Killick, 1987). The gradational contact with the overlying Bumberra Formation represents a transition to nearshore marine conditions. The Bumberra Formation consists of sublitharenite with thin mudstone interbeds and contains a possible Famennian fauna including cyrtospiriferid, and rare chonetid and productid brachiopods (Wright, 1966; Killick, 1987). The fauna, common bioturbation and abundant traction current structures (Table 6) indicate a shallow marine environment including tidal flats, foreshore and shoreface conditions. Thinly bedded mudstone with minor interbedded quartzarenite (the Lawsons Creek Shale) represent a rapid decrease in bed thickness, grain size and sand:silt ratio through the transgressive cycle. Abundant bioturbation and a brachiopod fauna including cyrtospiriferids and *Mucrospirifer* indicate deposition in a quiet, deepening environment of upper offshore to muddy lower shoreface conditions. The youngest unit of the Lambie Group in the Mudgee district, the regressive richly fossiliferous Derale Sandstone, represents thick sublittoral sheet sandstones deposited on a storm-dominated shallow marine shelf.

Palaeogeographic models for the Late Devonian (Killick, 1987) depict a northwest-southeast trending strandline, extending as far west as Parkes at the height of the transgression, with quartz-rich sediment derived from uplifted highlands to the west and southwest. Correlatives of the marine units in the Mudgee district are known from the Dulabree Syncline in the Ilford-Running Stream-Upper Turon area to the immediate south and elsewhere in the Lambie Group. However, the Buckaroo Conglomerate is apparently unique to the Mudgee-Cudgegong district. There are similar marine sequences in the Late Devonian Catombal Group on the Molong High (Killick, 1987) and it is apparent the ?Famennian marine transgression affected the majority of the northeast Lachlan Fold Belt (Webby, 1972).

A low-angle discordance (5 to 25°) between the Lambie Group and a variety of older rocks [the Lambian Unconformity of Powell and Edgecombe (1978)] has been recorded from numerous localities in the northeast Lachlan Fold Belt. The unconformity surface is exposed at several localities between Windamere Dam and the White Rock area (10 km northwest of Rylstone; Fig. 3). Recognition of the continuation of shallow marine sedimentation into the Middle Devonian at Mount Frome, together with the low-angle discordance of the Lambian Unconformity suggested by Powell and Edgecombe (1978), indicate that the Tabberabberan Orogeny had only mild effects on the northeast Lachlan Fold Belt.
## TABLE 7 CARBONIFEROUS AND PERMIAN ROCKS ASSOCIATED WITH THE NORTHERN CAPERTREE HIGH

<table>
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<tbody>
<tr>
<td><strong>AARONS PASS</strong></td>
<td>Massive, post-kinematic, minimum-melt, I-type biotite</td>
<td>Middle Carboniferous - 327Ma</td>
<td>Elliptical-shaped aureole, up to 1.5km wide, displays four prograde zones in pelites and psammites which are correlated with zones in calcareous hornfels (marble, calc-silicate, skarns) and metabasites</td>
<td>Intrudes Ordovician and Silurian units. Suspected faulted western contact. Nonconformably overlain by Permian rocks</td>
</tr>
<tr>
<td>GRANITE (Offenberg et al., 1971; Pemberton, 1990; Greenfield, 1992)</td>
<td>granite intruded at high crustal levels. Compositionally homogeneous yet considerable textural variation including pressure-quenched porphyritic granite. Associated with aplite dykes and veins, pegmatite and granophyre</td>
<td>cf. 320Ma from outlier immediately west (Vickary, 1983). Geochemically similar to Bathurst, Gulgong and Oberon plutons. Petrographically similar to Havilah and Pyangle Pass granites</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>RYLSTONE</strong></td>
<td>Massive, post-kinematic, minimum-melt, I-type biotite</td>
<td>Latest Carboniferous to</td>
<td>Proximal intracaldera facies</td>
<td>Shallow dipping sheets with broad, gently plunging folds. Fiamme, columnar jointing. Ripples, cross-beds, parallel laminations, flutes and accretionary lapilli in mudstone</td>
</tr>
<tr>
<td>VOLCANICS (Day, 1961; Campbell, 1980; Langworthy, 1986; Dicker, 1989; Shaw et al., 1989; Kuoni, 1991; Millsteed, 1992; Colquhoun, unpubl. data)</td>
<td>granite intruded at high crustal levels. Compositionally homogeneous yet considerable textural variation including pressure-quenched porphyritic granite. Associated with aplite dykes and veins, pegmatite and granophyre</td>
<td>earliest Permian - 292Ma</td>
<td>with subaerial ignimbrite sheets, co-ignimbrite breccias, lava dome growth and subordinate airfalls and ashflows. Shallow water (?lacustrine) reworking and fluvial channels drain basement</td>
<td></td>
</tr>
<tr>
<td><strong>SNAPPER POINT</strong></td>
<td>160m - basal polymictic conglomerate (clasts of immediately underlying basement) grading to pebbly arenite and litharenite</td>
<td>Late Early Permian (middle to late Artinskian)</td>
<td>Sandy transgressive shoreline with beach, nearshore and offshore environments. Fluvio-glacial character</td>
<td>Thin, flat-lying outliers throughout the Cudgegong-Mudgee district. Arenites contain bioturbation, cross-beds, ripple marks</td>
</tr>
<tr>
<td>FORMATION (Bembbrick, 1983; Bamherry, 1984; Pemberton, 1990)</td>
<td></td>
<td>brachiopods to north and south of Rylstone (Brown, 1974; Langworthy, 1986)</td>
<td></td>
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</table>
Deformation of the Pre-Carboniferous Rocks

The pre-Carboniferous rocks crop out in northwest-southeast trending, shallowly plunging, open, cylindrical, asymmetric folds, with well-developed axial plane cleavage; in the Late Ordovician strata, folding cannot be recognised. In particular, the Late Devonian strata are as intensely deformed as, and have similar structural characteristics to, the older rocks. This apparently indicates that the major folding and cleavage-forming deformation which affected strata on the northern Capertee High occurred in the Early to Middle Carboniferous, coinciding with the timing of the Kanimblian Orogeny (Powell and Edgecombe, 1978; Cas, 1983).

Aarons Pass Granite

The Aarons Pass Granite, to the southeast of Cudgegong, is an I-type, biotite granite stock, intruded at high crustal levels (Fig. 3; Table 7). The body is compositionally homogeneous yet the rocks show considerable textural variation including marginal phases which display probable pressure-quenched textures. The granite is associated with late stage aplite dykes, pegmatite and granophyre (Greenfield, 1992). A 328 Ma (Middle Carboniferous) age has been determined from Rb/Sr ratios of biotite separates (Pemberton, 1990). The intrusion produced an aureole up to 1.5 km wide, with four prograde zones identified in pelites and psammites. Other small I-type granite stocks occur at 'Havilah' and Pyangle Pass (Fig. 3).

Rylstone Volcanics

The Rylstone Volcanics are an extensive sequence of silicic pyroclastics, lavas and epiclastics (Table 7) which crop out for over 45 km along the western margin of the Sydney Basin (Fig. 3). They consist of thick sheets of subaerial dacitic and rhyolitic ignimbrite, with possible rhyolitic lava domes and interbedded volcanioclastics and basement-derived epiclastics (Langworthy, 1986). The facies association and volume of the ignimbrite sheets are consistent with the products of a small- to medium-volume ash flow caldera (Dicker, 1989; Kuoni, 1991). Radiometric data (Shaw et al., 1989) indicate an age of 292 Ma (earliest Permian; Roberts, et al., 1993). This evidence, together with nonconformable contacts with the Pyangle Pass Granite (Langworthy, 1986), refutes earlier proposals (Day, 1961; Ranocchia, 1981) that the Rylstone Volcanics and nearby Carboniferous stocks were synchronous and comagmatic.

Early Permian Strata

Permian rocks in the Cudgegong-Mudgee district are restricted to thin, flat-lying, conglomeratic veneers which overlie older rocks with angular unconformity. They represent the basal unit of the Snapper Point Formation, the lowermost member of the Shoalhaven Group at the extreme western margin of the Sydney Basin (Bembrick, 1983). The formation represents a sandy transgressive shoreline deposit with beach, nearshore and offshore environments under apparent fluvio-glacial conditions (Dulhunty and Packham, 1962; Bamberry, 1984). At Kandos and surrounding areas (Fig. 3) the sequence extends up into the Berry Siltstone and the Illawarra Coal Measures.

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Some of the University of Wollongong research documented here has been funded by grants to the Tasmanides Research Group from the University of Wollongong. We thank:— various University Geology Departments for access to Honours, Masters
and PhD theses; Des Strusz and John Pickett for faunal identifications, and John for his comments on the manuscript. Many farmers along the Cudgegong River valley and surrounds granted access to their properties as did the Department of Conservation and Land Management to their holdings associated with Cudgegong Waters State Recreation Park.

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Supporting Paper 3


GEOLOGICAL FIELD GUIDE

TO

THE CUDGEGONG-RYLSTONE AREA,

LACHLAN FOLD BELT

Saturday May 14-Sunday May 15, 1994

Gary Colquhoun, John Pemberton & Tony Wright
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Leaders

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Department of Geology, University of Wollongong, Wollongong, NSW 2522

This field conference has been organised with the support of the Association of Australasian Palaeontologists, the NSW Division of the Geological Society of Australia, and the Department of Geology, University of Wollongong.
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Cover diagram: Palaeogeographic reconstruction for the Silurian Windamere Volcanics-Toolamanang Formation eruptive episode (Pemberton et al. 1994).
Frontispiece 1: Photograph (*circa* 1934) of Philip Game (right) and Alan Voisey (left) in front of the old Cudgegong hotel, with the historic St James Church (1863) on the hill in the background.

Frontispiece 2: Philip Game and his supervisor, University of Sydney Lecturer in Geology Mr L. L. (Laurie) Waterhouse (*circa* 1934), standing by the governor's official car (a Crossley; also used to convey the Duke and Duchess of York, [later King George VI and Queen Elizabeth] during their visit to Australia in 1927), seen here as Game's field vehicle in the Cudgegong district.
GEOLOGICAL MAP
OF THE
CUDGEONG DISTRICT

LEGEND

POST-TERTIARY

TERTIARY

TRIASSIC

KAMILAROI

UPPER DEVONIAN

MIDDLE DEVONIAN

Limestone

Limestone Breccia and Limestone Conglomerate

Tuffs (acid)

Limestone Conglomerate

Conglomerates

Felsic Tuffs

Clay-Shales and Quartzites

Granite

Quartz-Porphyry, Quartz-Pebble and Quartz-Phryolite

ROAD

RAIL

FAULT

A \

Quartz-Rocks, Conglomerates, Acid Tuffs and Clay-Shales

Upper Marine Series

Upper Cool Measures

Basalts and Dolerites

Hawkesbury and Narrabeen Series

Alluvium

Limestone Breccia and Limestone Conglomerate

Tuffs (acid)

Limestone Conglomerate

Conglomerates

Felsic Tuffs

Clay-Shales and Quartzites

Granite

Quartz-Porphyry, Quartz-Pebble and Quartz-Phryolite

ROAD

RAIL

FAULT

Scale: MILES
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Fig. 1 Locality map of the Cudgegong-Mudgee district
1. HISTORICAL INTRODUCTION

The first white man to explore north from Bathurst into the Cudgegong River valley was James Blackman (Superintendent of Police at Bathurst) in 1821, accompanied by the legendary aboriginal guide Aaron (Harding 1984). Lieutenant William Lawson and John Blackman, looking for a route to the Liverpool Plains, reached the site of Mudgee town in November 1821 (Robinson et al. 1984). The region was settled very soon after by the Cox family in 1822, at 'Munna' ['Menah'] to the north of Mudgee and 'Burrundulla' on the southern outskirts of Mudgee, and the Rouse family at 'Guntawang' west of Gulgong. In the Cudgegong district (the name meaning 'red hill') early settlers included William Bowman who founded 'Tonabutta' by the 1830s, and James Jennings whose name was linked with 'Toolamanang' by 1830. In the Carwell Creek area, James Vincent founded 'Carwell' which passed to his son in law, John Nevell in 1848, the homestead dating from 1851.

The earliest geological investigations in the region were by Stutchbury (1852), Taylor & Thompson (1871), Taylor (1879), Clarke (1878) and Sussmilch (1933). The classic geological work was done by Philip Game (1935) who graduated with his BSc from the University of Sydney in 1934; his Cudgegong work was carried out as an honours project while he held the Deas-Thompson Postgraduate Scholarship in 1934, when he mapped a huge tract of rocks from Mudgee to Kandos (see copy of his 1935 map on p. iii).

Mudgee lies in the central tablelands district of NSW, and is 260 km NW of Sydney. Cudgegong lies 35 km SE of Mudgee at the junction of the picturesque Cudgegong River and Cudgegong Creek (Fig. 1). Windamere Dam lies 24 km SE of Mudgee, and was developed to provide irrigation for the Mudgee river flats and a regulated water supply for Mudgee, and for the Cudgegong Waters State Recreation Park.

Following undergraduate field mapping excursions to the Mudgee (especially Queens Pinch) district in the late 1950s, honours mapping by Sydney University students was
initiated there in 1960 (including Wright). Subsequently students from UNSW, Macquarie, ANU, Newcastle and Wollongong have been involved with the district. PhD studies in the geology of the district include those by Wright (1966), Killick (1987), Pemberton (1990) and Colquhoun (continuing).

2. THE CAPERTEE HIGH AND ASSOCIATED ROCKS - A SUMMARY OF THE GEOLOGICAL SETTING

From the Early Silurian until the Middle Devonian, the Lachlan Fold Belt probably occupied a backarc or intra-arc continental margin setting to the west of a west-dipping subduction zone located in the ancestral New England Orogen (Scheibner 1989; Fergusson & Coney 1992; Collins & Vernon 1992). Extension in the eastern Lachlan Fold Belt created a series of meridionally-trending deep water basins or 'troughs' flanked by shallow marine to terrestrial, often volcanioclastic, platforms or 'highs' (Packham 1960, 1969; Cas 1983; Powell 1984).

The Cudgegong area is located on the most easterly of these highs, the Capertee High, which was flanked to the west by the flyschoid Hill End Trough (Fig. 2). The Capertee High in the Mudgee-Cudgegong district was a shallow marine to subaerial palaeogeographic entity which existed in the extreme northeast of the exposed Lachlan Fold Belt from the Middle Silurian until the earliest Middle Devonian. The facies and palaeogeography of the Hill End Trough and its flanking volcanic margins (the Capertee and Molong Highs) during this time compare favourably with modern ensialic interarc or backarc basins, such as the southern Havre Trough in the Taupo Volcanic Zone of New Zealand (Cas & Jones 1979). Rocks of the former Capertee High are now largely preserved within the Capertee Anticlinorium (a subdivision of the Molong-South Coast Anticlinorial Zone: Scheibner 1976), along with Late Ordovician basement sequences, thick Late Devonian sedimentary sequences of the Lambie Group, Early Carboniferous granitic intrusions and latest Carboniferous-Earliest Permian silicic volcanics. Flat-lying Permian-Mesozoic sediments of
Figure 2. Regional geology of the Ilford-Mudgee-Rylstone district.
the Sydney Basin overlap the High in the east, whereas undeformed Mesozoic sediments of the Great Australian Basin cover the Capertee High in the north.

2.1.a. **Ordovician.** The Late Ordovician Cudgegong Volcanics (Fig. 3) consist of basaltic and andesitic arenite, rare lava and breccia deposited by volcanioclastic debris flows on the flanks of submarine volcanoes with fringing carbonate areas; geochemical data (from rare lavas) show an affinity with calcalkaline lavas erupted in island arc settings. By comparison, coeval units to the east (the Coomber Formation and the Lue beds) are deepwater sequences composed of volcanioclastic and quartz-rich turbidites, radiolarian chert, allochthonous limestone and rare andesite flows.

2.1.b. **Silurian.** Following an earliest Silurian hiatus, the Capertee High formed during rifting of the Ordovician basement in the Early Silurian. A thick, shallow marine to emergent succession of silicic volcanics and associated epiclastics was deposited in the Mudgee-Cudgegong district in the Wenlock-Ludlow interval. This consists of basal shallow marine to subaerial clastics and carbonates (the Willow Glen Formation), overlain by a thick unit of dacite and subordinate rhyolite lava and breccia (the Windamere Volcanics) and its lateral, fragmental equivalent, the Toolamanang Formation. The Millsville Formation, a shallow marine shelf unit of breccia, limestone, mudstone and dacitic conglomerate is the highest known unit in the sequence. In contrast, the approximately coeval Moonbucca Formation to the west of Rylstone represents deposition on a shallow siliciclastic/carbonate mass-flow apron which interdigitated with a silicic volcanic pile. The extensive, but poorly-known, Dungeree Volcanics of the Mudgee-Lue district contain similar lithologies to the Moonbucca Formation and are a likely correlative.

In the Sofala district, thinner and more localised equivalents of the Willow Glen Formation (the Tanwarra Shale) and the Windamere Volcanics-Toolamanang Formation (the Bells Creek Volcanics) are recognised. Thick sequences of Silurian (?) strata are known
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<tr>
<th>PERMIAN</th>
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<th>Artinskian</th>
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<td>TOOLAMANANG FORMATION</td>
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<td>COOMBER FORMATION</td>
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<td>LUE BEDS</td>
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Figure 3. Stratigraphic columns for the Mudgee-Rylstone-Lue district.
between Mudgee and Dunedoo; their relationship to sequences in the Cudgegong-Mudgee district is uncertain.

2.1.c. Early Devonian. After a period of erosion in the latest Silurian following a mildly manifested Bowning Orogeny, deposition of extensive Early Devonian sequences commenced. Between Mudgee and the Kandos-Rylstone-Cudgegong district, Early Devonian rocks occur in two belts separated from older rocks by faults or, more rarely, a disconformity to low angle unconformity. The eastern belt, which can be traced almost continuously from Mount Knowles east of Mudgee to southwest of Kandos contains shallow marine and subaerial sequences deposited on the platform areas of the Capertee High; by contrast, the western belt (the Queens Pinch Belt, centred 16 km southeast of Mudgee) contains mass-flow and hemipelagic sequences deposited in slope settings adjacent to the western margin of the Capertee High.

2.1.c.i Platform sequences: the Kandos Group

The platform areas of the northern Capertee High were dominated by the early Lochkovian to Emsian Kandos Group, an essentially conformable, dominantly shallow marine sequence up to 4000 m thick (Fig. 4). It consists of the (basal) Warrah Conglomerate (fan deltaic clastics), the Clandulla Limestone (reefal to biostromal limestone and dolomite), the Yellowmans Creek Formation (muddy shelf clastics and carbonates), the Roxburgh Formation (storm-dominated siliciclastics), the Riversdale Volcanics (dacitic ignimbrite with lavas and epiclastics) and, at the top, the Carwell Creek Formation (fine- to coarse-grained clastics and carbonates deposited in barrier island and wave- and tide-influenced shallow marine environments, the basal carbonate unit being known as the Glendale Limestone Member).

The constituent formations of the Kandos Group are characterised by rapid lateral changes in thickness and facies which locally alter superpositional relationships. All sedimentary units thicken markedly to the south and west and pinch out on or towards the
<table>
<thead>
<tr>
<th>Early Devonian</th>
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<tr>
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<td>* Carwell Creek Formation</td>
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<td></td>
<td>delta</td>
<td>* Yellowmans Creek Formation</td>
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<td></td>
<td>eurekaensis</td>
<td>* Clandulla Limestone</td>
</tr>
<tr>
<td></td>
<td>woschmidt</td>
<td>* Warrah Conglomerate</td>
</tr>
<tr>
<td>Late Silurian</td>
<td>woschmidt</td>
<td>Moonbucca Formation</td>
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Figure 4. Kandos Group formations and correlation with standard conodont zonation.
Silurian-Devonian unconformity in the northeast, which presumably marked the edge of the Early Devonian depositional basin. The Riversdale Volcanics thin rapidly to the north, east and west relative to a very thick central area.

Kandos Group units were affected by a number of major, partly eustatically-driven, transgressions and regressions; these include:

(a) an early- to mid-Lochkovian transgression peaking in the Yellowmans Creek Formation;
(b) a late Lochkovian-?early Pragian regression culminating in partial subaerial exposure and erosion at the top of the Roxburgh Formation;
(c) relative sea-level low stands during the Pragian, when the dominantly subaerial Riversdale Volcanics was erupted; and
(d) a transgression in the early Emsian, possibly punctuated by short-lived regressions and long periods approaching aggradation.

The Kandos Group is most fully exposed in the Cudgegong-Rylstone-Kandos district; however, probable equivalents of the Pragian and Emsian units of the Group (i.e., the Riversdale Volcanics and Carwell Creek Formation) have been identified from the Capertee Valley in the south, to the Mt. Knowles and Mt. Frome areas in the north. Equivalents of at least the two lower formations (Warrah Conglomerate and Roxburgh Formation) occur to the N of Ilford.

2.1.c.ii Western Platform Margin Sequences

The late Lochkovian to late Emsian sequences of the Queens Pinch Belt (Wright 1967, 1969, 1979) include mass-flow volcaniclastics, siliciclastics, and carbonates, with thick mudstone intervals, and rare volcanic horizons; these were deposited generally in slope and base-of-slope settings adjacent to the western margin of the northern Capertee High
and reflect the effects of transgression, regression, erosion and volcanism on the main platform areas of the High. Similar, coeval and older sequences also occur marginal to the Capertee High at Limekilns (Wright & Chatterton 1988; Wright & Haas 1990), the outskirts of Mudgee (Chatterton & Wright 1986), and possibly in the Ilford area and between Mudgee and Gulgong. Discussion of the formations present at Queens Pinch is given by Pemberton et al. (1994).

PALAEOGEOGRAPHY

By the Early Devonian, the Capertee High was a persistent shallow marine to emergent entity traceable from the southern Capertee Valley to the north of Mudgee (Fig. 5).

Lochkovian

Proven Lochkovian strata are largely restricted to the Cudgegong-Kandos district. Sedimentation began in the early to middle Lochkovian as a transgressive sequence from wave-dominated fan delta, eroding exposed basement to the east, through reefal-biostromal limestones to a muddy, storm-influenced shelf. Silicic volcanism, although distal to the depositional setting, commenced near the end of carbonate deposition and remained active through the Lochkovian. Storm-dominated siliciclastics deposited in a middle shelf to nearshore setting marks the start of regressive conditions in the late Lochkovian.

Pragian

Voluminous silicic pyroclastic volcanism dominated the largely subaerial Capertee High during the Pragian (Fig. 6a). Calderas, to the southeast of Cudgegong and in the Capertee Valley (Bembrick et al. 1969), were flanked by extensive ignimbrite sheets. Vast volumes of volcanic detritus were transported to the west and redepsoited in fans and aprons marginal to the Capertee High as well as providing fill for the Hill End Trough.
**MID DEV**

**Early Devonian**

Conodont Zones
- *australis*<br>- *costatus*<br>- *paritus*<br>- *patulus*<br>- *serotinus*<br>- *inversus*<br>- *perbonus*<br>- *dehiscens*<br>- *pereumae*<br>- *kindlei*<br>- *sulcatus*<br>- *pesavis*<br>- *delta*<br>- *eurekaensis*<br>- *woshmidtii*

**Platform**

- Rylstone-Cudgegong<br>- Ilford<br>- Capertee Valley<br>- Mt. Knowles-Buckaroo<br>- Mount Frone<br>- Boogledie<br>- Sutchers Creek<br>- Ingleburn Formation<br>- Warra <br>- Mudstone<br>- Taylors Hill Formation<br>- Mullamaddy Formation<br>- Merrians<br>- Tuff<br>- Waterbeach Formation<br>- Conodine Group<br>- Limestone Mudstone Breccia Facies

**Western Platform Margin Sequences**

- Queens Pinch<br>- Limekilns<br>- Queens Pinch<br>- Jesse Limestone<br>- Rosedale Shale<br>- Murrays Mudstone<br>- Tuff

**References**

4. Lyons (1976)

**Figure 5.** Stratigraphic columns and correlations of platform and western platform margin sequences of the Early-Middle Devonian of the Northern Capertee High.
Figure 6.a

Figure 6.b
Emsian

Volcanism waned at the end of the Pragian. Transgression in the earliest Emsian created a broad southwest sloping shelf with localised islands of eroding Pragian volcanics. Sedimentation was dominated by laterally and temporally extensive storm- and fairweather-wave influenced, nearshore to shelf siliciclastics with local carbonate buildups (both in lagoonal and more open marine settings) (Fig. 6b). The sequence gradually onlapped older strata to the east. During short-lived regressions, siliciclastic and carbonate detritus were transported westward to the shelf edge and redeposited as debris flows (Limekilns and Queens Pinch sequences).

Regression in the earliest Eifelian terminated marine carbonate deposition and effectively ended the history of the Capertee High as an active palaeogeographic element.

Middle and Late Devonian

Tectonism or regression in the earliest Middle Devonian effectively terminated deposition on the northern Capertee High, although major deformation of the sequences did not occur until the Early Carboniferous. A low-angle discordance (5° to 25°) between the Lambie Group and a variety of older rocks [the Lambian Unconformity of Powell & Edgecombe (1978)] has been reported from numerous localities in the northeast Lachlan Fold Belt. The unconformity is exposed at several localities between Windamere Dam and the White Rock area (10 km west of Rylstone) where it is overlapped by the Rylstone Volcanics. Recognition of the continuation of shallow marine sedimentation into the Middle Devonian at Mt Frome, together with the low-angle discordance of the Lambian Unconformity, suggest the Tabberabberan Orogeny had only mild effects on the northeast Lachlan Fold Belt.

The Late Devonian Lambie Group is represented by a transgressive, conformable
sequence at least 1700 m thick, commencing with the Buckaroo Conglomerate which is overlain by a ?Famennian succession of Bumberra Formation, Lawsons Creek Shale and Derale Sandstone in a southeast-plunging syncline (the Pine Ridge syncline) from Mudgee to the west of Rylstone (Wright 1966). Killick (1987) has interpreted these units as representing braided river, nearshore clastics, offshore to muddy shoreface and storm-dominated shelf environments respectively.

2.1.d. Post-Late Devonian rocks

Younger rocks of the Capertee Anticlinorium include the Early Carboniferous Aarons Pass (Pemberton et al. 1994), Havilah and Gulgong I-type granitic intrusions; and silicic pyroclastics, epiclastics and lava of the latest Carboniferous to earliest Permian (Shaw et al. 1989) Rylstone Volcanics which crop out along the western margin of the Sydney Basin from Botobolar to southwest of Kandos.

Sub-horizontal, marine to terrestrial strata of the Permian-Triassic Sydney Basin flank the High to the east and occur widely as outliers further west.

Tertiary alkaline intrusions (trachyte, teschenite, phonolite) and basaltic intrusions and flows (Wellman & McDougall 1974) crop out widely but are of only minor areal extent (Day 1961; Matson 1975).

Cainozoic deposits of unconsolidated clay, silt, sand and gravel occur along rivers and valley floors over much of the area. Large areas of such deposits between Mudgee and Gulgong have yielded economically significant quantities of gold, diamonds and clay (Taylor 1879; Matson 1975).
ITINERARY.

STOP 1. Top of Aarons Pass.

(GR 627602, Kandos 1:25,000 sheet).

**Turn east on to Perrams Road. Travel 400 m along this dirt farm track to farm gate on N side of track; walk 100 m NW to prominent lookout area.**

At this locality highly weathered Early Carboniferous (327 Ma: Pemberton 1990) Aarons Pass Granite (I-type biotite granite) is nonconformably overlain by up to 5 m of Permian polymict conglomerate (basal unit of the Snapper Point Formation: Bembrick 1983). Here the granite contains angular quartz and K-feldspar fragments from the granite together with chert and Late Devonian quartzite, and locally derived dacite and sandstone. The conglomerate forms a thin, flat-lying veneer near the extreme western edge of the Sydney Basin. Thicker outliers are seen at Mt Margaret (GR 542757), the Castle (GR 542705) and Mt Bocoble (GR 576665).

This is a useful vantage point. To the east and north, we see the exhumed sub-Permian peneplain. In the midground to the E the wooded hills are Early Devonian Kandos Group whereas in the distance the flat topped hill is Tongbong Mountain, and the large rounded hill is one of the Tertiary alkaline intrusions. To the northwest, Late Devonian rocks protrude well above the peneplain and represent exposed land masses not eroded prior to Permian deposition. **The relief on the sub-Permian surface can vary up to 200 m between Mudgee and Kandos.** In the midground to the north and northwest are undulating wooded hills of Late Silurian Toolamanang Formation, and in the foreground large slabs and tors of the Aarons Pass Granite and associated aplitic dykes are seen. To the west is the prominent Mt Bocoble.

*Drive carefully back on to the highway and head north to the intersection of the Mudgee road and the Cudgegong-Rylstone road (5.9 km). All stops on the Mudgee-*
Lithgow road are located in terms of distance from the Cudgegong-Rylstone turnoff. Considerable care is needed to park carefully on the verges, and in crossing the road.

En route to Stop 2, we drive for 7 km through dark-coloured Toolamanang Formation on both sides of the road; several cuttings show a massive succession of fine- to coarse-grained dacitic arenite with black dacitic ash horizons and clasts, and sporadic basaltic blocks. The road then passes through a thin conglomerate of the Willow Glen Formation, and into andesitic arenite of the Cudgegong Volcanics near Francis Skinner Bridge at 7.6 km. Immediately west of the bridge, basaltic arenite and breccia outcrop and a prominent basalt lava is exposed in the bridge spillway.

Stop 2. Cudgegong Volcanics.

(GR 577738, Broombee 1:25,000 sheet).

11.8 km. Large road cutting on E side just over brow of sweeping right bend, before steep descent, 1.6 km before Windamere Dam turnoff. Best exposures of pillow lavas next to first white post with red reflector.

The Cudgegong Volcanics consist of Late Ordovician (Gisbornian) basaltic and andesitic arenite, breccia and rare lava deposited by volcaniclastic debris flows on the flanks of submarine volcanoes with minor fringing carbonate areas. Basalt and andesite lavas have major, trace and rare earth element characteristics typical of calc-alkaline lavas from island arc settings and their chemistry does not support the shoshonitic affinity proposed by Wyborn (1992). The rocks show the imprint of prehnite-pumpellyite to greenschist facies metamorphism. Assemblages include the metamorphic phases albite, calcite, chlorite, actinolite, epidote, sphene, sericite, prehnite and pumpellyite with relict clinopyroxene and amphibole. Detailed descriptions are given in the accompanying copy of Pemberton
In this cutting, the rocks are mainly structureless fine- to medium-grained andesitic and basaltic arenite with some coarse-grained basaltic breccia. Calcite veining and small scale faulting are common. Numerous bulbous masses are seen in the cutting; however, near the white post rounded, pillow-shaped basaltic lava bodies with fine-grained dark rims up to several cm thick are concentrated. 50 m to the NW in this cutting, basaltic breccias are well exposed.

The best exposures of the Cudgegong Volcanics crop out to the south of stop 2. Pemberton (1989) included a map noting basaltic boulder breccia, fine basaltic ash, andesite lava, several syenite sills and thin fossiliferous marl. Access is limited by locked CALM gates (permission required from Paul Gibb, Mudgee office, ph. 063-724044). The most accessible gate is about 500 m to the southeast of the cutting, the track being four wheel drive only.

Stop 3. Willow Glen Formation.

(GR 575746, Broombee 1:25,000 sheet).

12.6 km. Second road cutting north of stop 2. Top of steep long descent. Southeast cutting.

The Willow Glen Formation (conglomerate, pebbly arenite and arenite; fossiliferous mudstone and limestone with a distinctive Wenlock to Ludlow brachiopod-coral-trilobite fauna) was deposited in a southeast-facing coastal environment which included a fluvial channel zone; subtidal to supratidal flats, affected by a series of transgressive-regressive cycles including common incision by fluvial channels; and more open marine conditions. The best and most typical exposures occur to the east; however, the majority, including several important faunal localities, are now under the waters of Lake Windamere.
Pemberton (1989) included maps and stratigraphic sections for these areas together with detailed faunal lists.

At this locality 220 m of conglomerate and pebbly arenite represent the fluvial channel zone and are atypical of the majority of the Willow Glen Formation. The conglomerate is dominated by white and black silicic volcanic clasts in a quartz- and plagioclase-rich matrix. The cuttings show cross bedding, aligned and imbricated clasts, graded beds and clear southwest dips. Low angle asymptotic cross beds together with draped sandy material over imbricated pebbles indicate local overturning. The Willow Glen Formation at this locality lies on the overturned northeast limb of a large scale northwest-trending anticline, with the Cudgegong Volcanics in the core.

Stop 4. Windamere Volcanics/Willow Glen Formation contact.

(GR576745, Broombee 1:25,000 sheet).

12.7 km. Same road cutting, about 50 m downhill from Stop 3.

A conformable contact between massive, strongly jointed, typical dacite lava and breccia of the Late Silurian Windamere Volcanics and strongly cleaved and weathered pebbly arenite of the Willow Glen formation is clearly exposed in the cutting. At first appearance a prominent shallowly south-dipping shear zone appears to represent the contact; however, detailed examination reveals a sharp contact (steeply south dipping) between a cleaved pebbly arenite and massive weathered dacite.

Stop 5. Windamere Volcanics.

(GR 587754, Lue 1:25,000 sheet).

Proceed downhill and, at 13.4 km, turn right to Windamere Dam. Cross bridge over spillway and park about 50 m further towards dam. 14.2 km.

The Late Silurian Windamere Volcanics are characterised by thick, undifferentiated dacite lava and breccia, with common rhyolite horizons, possibly emplaced as a subaerial
to very shallow marine lava dome. The rocks exhibit low-grade metamorphic assemblages similar to those in the Cudgegong Volcanics together with relict quartz and amphibole. Detailed descriptions together with maps and representative sections were presented by Pemberton (1989).

Typical massive strongly jointed, quartz-veined, cleaved and sheared dacite lava and breccia are best exposed in the road cuttings leading up to the bridge and in the dam spillway.

At this locality north of the spillway, exposures of the Windamere Volcanics are considered atypical. On the northwest side of the cutting, large prominent rounded and angular dacite boulders occur in a fragmental dacite matrix. Finely laminated, northeast dipping arenite often drapes over the blocks. On the southeast side of the cutting, dacite lava blocks have slumped into steeply dipping cleaved mudstone.

A more complete Windamere Volcanics sequence crops out to the northwest of the district in the Millsville area (access via a farm track east of 'Woodgrove', owned by the Rayner family). Exposures include both underlying (Willow Glen Formation) and overlying (Millsville Formation) Wenlock to Ludlow formations and less common lithologies such as dacitic conglomerate, dacite breccia with limestone clasts, flow-layered dacite lava, mudstone. Here the rocks crop out on the southern upright limb of the anticline.


(GR 590754, Lue 1:25,000 sheet).

14.4 km. About 150 m past stop 5 in the same cutting on the spillway road.

The shallowly east-dipping unconformity surface is now poorly exposed. Intensely weathered pale brown porphyritic dacite lava of the Windamere Volcanics is in sharp contact with the basal conglomerate (containing quartz pebbles and chert clasts) of the Late Devonian Buckaroo Conglomerate. An excellent photograph of the fresh surface taken just
after the cutting was exposed is shown in Pemberton (1990).

STOP 7. Buckaroo Conglomerate.

(GR 592756, Lue 1:25,000 sheet).

14.9 km. 500 m east of STOP 6 on the spillway road.

In road cuttings opposite the roundabout are good exposures of typical lithofacies of the Late Devonian Buckaroo Conglomerate dipping towards 050° at about 55°, representing parts of fining-up, probably fluvial sequences (Killick 1987). These are the basal reddish, coarse-grained, quartz-rich conglomerate, passing up through sandstone into red shale which probably represents overbank deposits. The formation, the lowest unit of the Late Devonian sequence in the Mudgee-Rylstone district, has a maximum thickness of about 150 m, and can be recognised over a strike distance of some 30 km from the northern end of the Pine Ridge Syncline near 'Buckaroo' to near 'Glendale' (Cook 1988a).

At this point members of the group may wish to examine the dam, vital statistics for which are given in Appendix 2. This will also be the lunch spot, as good facilities are located on the downstream side of the dam wall (but safely away from its threat). The walls of the discharge system for the dam are cut into very highly deformed Bumberra Formation which overlies the Buckaroo Conglomerate. About 1 km to the NW along strike, the Bumberra Formation has yielded poorly preserved productid and chonetid brachiopods which led to the suggestion that the Derale Sandstone (the highest formation in the sequence in the Pine Ridge Syncline) may have been Early Carboniferous; restudy of similar but much better preserved material from the Dulabree Syncline does not support the Carboniferous age determination.

After lunch we will drive back to Cudgegong, turn east on to the Rylstone road, and proceed 1.1 km to the entrance area to Cudgegong Waters Park, parking well off the
entrance road. Across the Dam it is possible to see reddish soils from Late Silurian limestone just S of the location from which Jones et al. (1987) described Silurian evaporitic limestones. The contact, immediately to the E of these soils between the Silurian and Early Devonian limestone of the Yellowmans Creek Formation, is the Cudgegong Fault.

Walk carefully along road to east for 100 m to Stop 8. From the Rylstone turnoff, the road traverses Late Silurian Toolamanang Formation, lateral equivalent of the Windamere Volcanics, for 2 km.

Stop 8. Toolamanang Formation.

(GR 643658, Kandos 1:25,000 sheet).

1.2 km. Road cutting, which has not been examined in great detail at this stage, is on south side of Rylstone road immediately east of Cudgegong Waters Park entrance.

The Toolamanang Formation is a thick undifferentiated succession of fine- to coarse-grained lithic arkose to feldspathic litharenite with common black very fine ash-sized horizons, fine-grained breccia and sporadic basaltic blocks. The blocks are petrographically and chemically similar to basalts in the Cudgegong Volcanics; blocks vary from tens of cm to 3 m in size, and were deposited as megaclasts in the sandy and muddy detritus. Loaded and slumped features are present on sand bases.

The unit is considered to be laterally equivalent to, and a fragmental version of the Windamere Volcanics, produced by dense, gravity-driven, shallow marine volcaniclastic flows of ash-sized dacitic detritus, with associated mudflows of finer detritus, both of which incorporated some eroded basement material.

At Stop 8, typical arenite and fine ash-sized rocks are exposed in the cutting. Atypical, finely laminated east-dipping arenite is clearly observed. To the east several basaltic blocks are present; the first is several m across at the top of the cutting whereas the
second and perhaps third block is a thicker basaltic horizon near the eastern end of the cutting.

**Roxburgh Formation Road Cuttings**

The Early Devonian (late Lochkovian) Roxburgh Formation is a >750 m thick, siliciclastic (and lesser volcaniclastic) sequence in the middle of the Kandos Group; it crops out extensively around the shores of Lake Windamere, in the Carwell Creek valley (further east) and in a small area north of Ilford. Facies analysis (Fig. 7) reveals 10 lithofacies, ranging from storm-dominated middle to outer shelf (Facies A and B), through inner shelf (Facies D) to amalgamated hummocky shoreface (Facies E), along with shoaling wave-dominated shoreface conditions (Facies F), tide-dominated shoreface to tidal inlet complex (Facies G), and foreshore (Facies H). Contemporaneous silicic volcanism is indicated by sporadic occurrences of subaqueously-deposited accretionary lapilli and ashfall tuff (Facies C), together with gravelly volcaniclastic deltas (Facies J) which prograded into the nearshore zone, probably following eruptions. Rare bodies of biostromal limestone also occur (Facies I). Facies A, D, and I contain rich shallow marine faunas which taphonomic studies show to have been strongly influenced by storm events.

The Roxburgh Formation displays an overall regressive trend, which may be linked to a 3rd order eustatic fall of sea level. Ubiquitous high frequency (?)5th order) cyclic alternations of facies (parasequences) are superimposed upon this upward shallowing trend. These typically comprise stacked, highly asymmetric, partial shelf to shoreface progradational sequences which are 5 to 15 m thick and dominantly regressive in character, although local transgressive portions are preserved as transition zones between cycles. The characteristics and durations of these cycles suggest allocyclic forcing mechanisms. Cyclicity is also present on larger (?)4th order) and smaller scales, and some sections, particularly the inner shelf and lower shoreface, display essentially random, chaotic sequences.
Reconstruction of shelf hydraulic regimes suggests storm waves approached the north-northeast-south-southwest trending shoreline from the northwest, generating a coastal set-up...
which produced offshore-directed bottom return currents which deflected consistently to
the left as they flowed across the shelf (flowing southwest to west) due to geostrophic
veering caused by Coriolis Force. Tidal inlets (high in the unit) and wave-dominated
shoreface conditions became dominant during fairweather periods with low fairweather
wave energy. Palaeowave estimates derived from the inner shelf (Facies D) indicate
conditions typical of large waning storm waves with moderate to large wave periods and
heights. The study area had a low palaeolatitude and was positioned propitiously to
receive frequent long period, hurricane-generated waves approaching, down the Hill End
Trough, from the north.

The Roxburgh Formation is well-exposed along a 3.5 km stretch of road cuttings on
the Rylstone road to the south of Lake Windamere. Sequences here crop out in a major
anticline bounded to the east and west by faults. The cuttings display locally intense
folding and faulting, especially adjacent to the bounding faults and in the core of the
anticline.

STOP 9. Roxburgh Formation.

Road cutting (RC1) at GR 651654, Kandos 1:25,000 sheet, 2 km.

This road cutting is located within the Roxburgh Formation immediately east of the
Cudgegong Fault, a major fault separating the Ordovician-Silurian and Devonian strata, and
the Roxburgh Formation here is intensely deformed. Well-developed tectonic melange and
broken formation are centred around a major listric fault at the east end of the cutting.
Excellent pinch and swell structures are developed in the sandstone beds. The rest of the
cutting is intensely fractured and faulted with extensive development of secondary
cleavages and 'scaly' mudstone fabrics. Although largely useless for sedimentology due to
the deformation, some mudstones at RC1 yield Zoophycos and Chondrites.
STOP 10.  Roxburgh Formation.

Road cutting (RC2) at GR 660652, Kandos 1:25,000 sheet, 3.2 km.

RC3 displays uniformly west-dipping Roxburgh Formation disrupted by five or six faults, some with small displacements. Extensive development of Facies A (massive fossiliferous mudstone to very fine sandstone) can be seen at the western end, adjacent to the Marble Crusher; these beds are typical (although not notably brachiopod-rich) fossiliferous Facies A. Further east, excellent exposures of three high frequency coarsening-upward cycles (parasequences) display rapid alternation of outer shelf to shoreface facies. Good exposure of proximal to medial storm beds (Facies D) and amalgamated hummocky cross-stratified sandstone (Facies E) occur towards the eastern end of the cutting.

STOP 11.  Roxburgh Formation.

Road cutting (RC4) at GR 663652, Kandos 1:25,000 sheet, 3.6 km.

The sequence exposed here is also uniformly west-dipping with gentle warping to kinking of strata. The western end contains excellent exposures of alternating distal and proximal storm beds of Facies B and Facies D displaying numerous small scale cycles of bed thickness change. Good exposures of the sedimentary structures (including HCS) which comprise the typical Facies D storm bed occur. Bioturbation is locally apparent in the interbedded shales. A 5 m thick horizon of Facies F (bimodal trough cross-bedded sandstone with minor HCS deposited in a wave dominated middle to upper shoreface setting) is well exposed at the eastern end. Note that Facies F occurs at the top of a small coarsening-upward parasequence and is separated from the Facies B and D storm beds by an undulating erosion/flooding surface.
STOP 12. Roxburgh Formation.

Road cutting (RC5), at GR 666652, Kandos 1:25,000 sheet, is a small section disrupted by two major faults and folded at its eastern end, 3.9 km.

A good coarsening-upward parasequence (Facies A, B, D, E, ?F) is exposed at the western end. Fossiliferous horizons are reasonably exposed in Facies A mudstone-fine sandstone at the base of this cycle. Fossils present include sparse brachiopods, well-preserved fenestellid bryozoans, and favositid and solitary cystimorph tetracorals, the latter locally in growth positions.

STOP 13. Roxburgh Formation.

Road cutting (RC6), at GR 666652, Kandos 1:25,000 sheet, 4 km.

This is a large cutting close to the centre of the major anticline. A beautifully-exposed box anticline, faulted in its hinge area, dominates the centre of the cutting. Strata to the west of the hinge fault are extensively disrupted by numerous faults and fractures. Strata to the east are uniformly east-dipping and contain several faults, some with small displacements. Facies E, D and B storm beds are all well exposed along with 2 distinct calcareous horizons (Facies I) of smothered and slumped limestone and fossils (brachiopods, crinoids, corals, etc.); a small, non age-diagnostic conodont fauna has also been recovered from these limestone fragments.

STOP 14. Roxburgh Formation.

Road cutting (RC10), at GR 675653, Kandos 1:25,000 sheet, 5 km.

Approximately 500 m long, containing around 200 m of section, this is an entirely east-dipping section occurs on the eastern limb of the major anticline. Several normal and reverse faults disrupt the section and become more common towards the eastern end. Interesting exposures of medium- to large-scale boudinage occur approximately 200 m from the western end. Several coarsening upward parasequences (5 to 20 m thick) are well exposed, as are Facies D, B, and E; also prominent are numerous horizons of Facies G
(bimodal-bipolar cross-bedded sandstone [?tidal shoreface to tidal inlet]) and foreshore deposits of Facies H. Several fossiliferous horizons are exposed in Facies D towards the eastern end. Easternmost outcrops are very poor, being extensively fractured and iron stained.

From the Roxburgh Formation outcrops, we drive to the east through Devonian strata, passing across a poorly exposed contact with Late Devonian beds marked by outcrops of ironstone. These surficial deposits commonly mark faults in this district (e.g., Ironstone Creek contact between Silurian and Devonian; near the Kandos quarries between the Clandulla Limestone and Silurian strata; and at 'Glendale', between Early Devonian beds and probable Late Devonian beds. The rugged hills on the west side of the road largely consist of Riversdale Volcanics.

STOP 15. Yellowmans Creek Formation, Roxburgh Formation and Riversdale Volcanics.

(GR 707642, Kandos 1:25,000 sheet, 9.1 km).

At the northern end of the Carwell Creek valley, a large part of the Kandos Group is well-exposed on the E side of Carwell Creek; cross creek to examine the gently folded and condensed sequence on the hill opposite the gates to 'Blackberry Gully' where one of the last members of the early settler family (Nevell) still lives. The original (1851) 'Carwell' homestead still stands further to the S along the valley.

Formations exposed low on the slopes include the upper part of the Yellowmans Creek Formation, a very thin wedge of Roxburgh Formation, thin bouldery Riversdale Volcanics and, at the top of the hill, perhaps 50 m of Carwell Creek Formation. Of these we plan to examine only the contact between the Roxburgh Formation and the Riversdale Volcanics.
We will see that there is a prominent erosional contact where the Riversdale cuts down through about a metre of Roxburgh Formation. On the western side of the creek are good exposures of the Yellowmans Creek Formation.

To examine the remainder of the Kandos Group, we will drive through 'Lucknow' (property owner Ross Lloyd), crossing Carwell Creek in the process and drive around to the eastern side of the folds exposed in the two low hills to the E of Stop 16.

**STOP 16. Kandos Group.**

(GR 719647 to 717647, Kandos 1:25,000 sheet).

The section, exposed on the east-facing slope of a south-draining tributary of Carwell Creek reached by driving through 'Lucknow', includes the following formations in a complete but condensed sequence through the Kandos Group:

- Carwell Creek Formation
- Riversdale Volcanics
- Roxburgh Formation (?)
- Yellowmans Creek Formation
- Clandulla Limestone
- Warrah Conglomerate (lowest)

The original plan was to examine the Warrah, Clandulla and Yellowmans Creek units at the Kandos Quarries, but access to the quarry area has been suspended at short notice, pending clarification of the policy of access by the operating company.

The lowest formation, the Warrah Conglomerate, is poorly exposed and highly sheared; preserved thickness may reach 50 m. Bedding is poorly defined, and tectonic rotation of (? imbricated) clasts (dominantly volcanic, chert and rare limestone) is suspected. Sandy
beds seen at the Quarries are absent here. The usually gradational contact with overlying strata is not seen.

The Clandulla Limestone dips gently to the west, is extensively dolomitised, and has a thickness estimated at 100 m compared with 150 m in the quarries. Fauna is limited to *Amphipora* and *Cladopora*-like twigs, rare snails and some silicified brachiopods such as *Machaeraria* and *Isorthis*.

The contact with the overlying formation is poorly exposed. The Yellowmans Creek Formation in this section consists of yellowish thin platey limestones with some volcanic breccia units, whereas it normally consists of grey thin-bedded silts and muds with thin stringers of sand, closely resembling the Late Devonian Lawsons Creek Shale. The topmost 70 m approximately are exposed here.

The overlying Roxburgh Formation and its lower contact are poorly exposed and, from outcrops on the hillside to the west, the Roxburgh Formation is known to be no more than 50 m in this area, in contrast with the thickness of 750 m several km to the west. At least some of this may be explained by the erosional upper contact. The unit is represented here by very restricted outcrops of cross-bedded quartz arenite which are quite similar to some beds in the Yellowmans Creek Formation.

The Riversdale Volcanics in this area is represented by a continuous series of low outcrops of boulder beds and, as for the underlying horizon, the thickness can only be estimated at about 10 m. Better outcrops of the latter 2 units and their contact will be seen on the western side of the hills, at STOP 15.

The Carwell Creek Formation (maximum known thickness 1230 m) is typically crinoidal quartz-rich dolomitic limestones and arenites; rare shelly fossils can be seen in sandy beds. Spectacular cross bedding is developed in places in the sandy horizons. Depending on how far we proceed up this section, we may only see the lower calcareous sands; however, sandy, crinoidal float indicates that other lithologies are developed close above the base of the formation. The preserved thickness here in the N-plunging synclinal core
is very small, and higher beds are seen in the creek to the north.

Stop 17.  Rylstone Volcanics and Related Mineralisation.

The Rylstone Volcanics form an extensive sequence of silicic pyroclastics, lavas, and epiclastics which crop out along the western margin of the Sydney Basin from Upper Botobolar in the north to south of Kandos (Offenberg et al. 1971) - a distance of over 50 km. Much of the unit remains poorly-known, although three recent studies (Ranocchia 1981; Langworthy 1986; Kuoni 1991) have dealt at length with the stratigraphy, volcanology, petrography and geochemistry of the Volcanics in the Rylstone-Kandos district.

Outcrops to the west and northwest of Rylstone illustrate some of the diverse nature of the Rylstone Volcanics. The Rylstone Volcanics to the northwest of the Rylstone in the vicinity of 'White Rock' are composed almost entirely of rhyolitic ignimbrites belonging to the White Rock Ignimbrite Member of Langworthy (1986). These rocks are white- to buff-coloured and generally form low, rounded outcrops, but locally are silicified and crop out as prominent, elongate, NW-trending ridges (e.g. 'White Rock'; GR 705684). A pumice lenticle foliation defines bedding in a number of outcrops and rare imbrication of these pumice lenticles was used by Langworthy (1986) to indicate a northerly palaeoflow direction. Based on matrix and biotite contents, Langworthy (1986) has inferred 4 compound flow units in this area, ranging from 5 m to 37 m thick.

Further south, near 'Moonbucca' homestead (GR 744642), a diverse range of lithologies is exposed, the most prominent being a flow-layered rhyolite lava which crops out around Cumber Melon Creek. Named the Coomber Flow-banded Rhyolite Member by Langworthy (1986), it is characterised by a prominent flow-layering (3 mm to 15 cm separation) defined by colour contrasts and recessive weathering. The layering is generally planar in the thicker layers, gently undulating or harmonically to disharmonically folded in the intermediate layers, and irregularly distorted in the thinner layers. The harmonic
flow-layered folds at GR 745642 are consistently overturned towards the NE, indicating a
general palaeoflow of lava from SW to NE. Langworthy (1986) records a consistent
NE-SW strike and subvertical dips for the flow-layering and proposed a domal
emplacement origin along a major fracture oriented in this direction. Columnar jointing,
localised areas of massive, aphyric lava and autobrecciated lava also occur and are all well
exposed in Cumber Melon Creek (GR 742638).

Overlying the rhyolite lava to the south and west are a variety of lithologies grouped
under the name Quarry Road Siltstone Member by Langworthy (1986). The most common
lithology is a very fine-grained white tuff which is best exposed at GR 738630. Outcrops
are generally massive or, more rarely, well-bedded (10-30 cm) and parallel laminated.
Additionally, both Ranocchia (1981) and Langworthy (1986), record cross-beds, ripple
laminations, and flute marks, indicating redeposition of the ash as a tuffaceous siltstone.
However, several 10-25 cm thick horizons rich in accretionary lapilli were noted in this
study, suggesting at least these areas may be primary airfall deposits which have undergone
little or no reworking; in addition, their presence implies a source vent within 10s of
kilometres (Moore & Peck 1962).

In the basal 5-18 m of the Rylstone Volcanics, the white tuff/tuffaceous siltstone
passes rapidly into a prominent clast- to matrix-supported breccia or conglomerate. This
lithology displays angular to subrounded, granule- to boulder-sized clasts of local basement
- typically chert, lithic sandstone, and shale from the Coomber Formation and Moonbucca
Formation - in a matrix of either coarse poorly-sorted volcanolithic sandstone or white
tuffaceous siltstone. Best exposures of this lithology occur at GR 734631 and GR 734632.
The breccia/conglomerate occurs in sharp-based beds which lack bedding, stratification,
imbrication or other internal sedimentary structures, although several horizons at GR
735637 display abundant plant fossils in the tuffaceous matrix. These features suggest
deposition by subaerial debris flows (Nemec & Steel 1984). Interspersed throughout the
breccia are rare beds of ripple-laminated tuffaceous sandstone, and prominent thin horizons
of well-bedded airfall tuff. The framework grains occur in evenly-spaced laminations which display both normal and reverse grading.

In this area is the 'Rockwell' Antimony Mine, located approximately 5 km SW of Rylstone at GR 737643. Access to the mine is via a dirt track 1.5 km SW from the 'Moonbucca' homestead. The deposit was discovered in 1881 and in 1883 several shafts were sunk, encountering 3 narrow lodes; however, for reasons unexplained, the area was abandoned in 1884 after only 20.3 tonnes of ore had been mined and 800 pounds expended on exploration and development. The ore was apparently not sold, and has been reported "lost at sea". The mine site has since received desultory attention from government mine inspectors (Whiting 1943; Nicholson & Suppell 1967) and, more recently, from a number of exploration companies (BHP 1982; Sunshine Gold Search 1984; Alkane Exploration 1988; CRA 1989 to present). Results of the geochemical and geophysical programmes undertaken by these companies have so far been uniformly negative, with all concluding that only subeconomic grades and tonnages exist.

The mine is located in the ?Ordovician Coomber Formation on the lower of two andesite sills or flows and surrounding sediments. It was reported, erroneously, by Matson (1975) that the geological host was the Late Carboniferous Rylstone Volcanics, possibly in deference to Offenberg et al. (1971) who mapped the area as belonging to this unit. Although the contact of the Rylstone Volcanics passes close to the mine, and the deposit itself may be related genetically to it, the mineralisation itself is clearly hosted by the steeply-dipping Lower Palaeozoic sediments and volcanics - a conclusion also reached by numerous earlier workers.

The mineralised quartz veins are localised in a systematic joint set which occurs widely in the surrounding units and is probably related to the regional Early Carboniferous deformational period. The hydrothermal fluids may also have been related to this deformation, or to some post-orogenic event - most likely the extrusion, in the latest Carboniferous-earliest Permian (Langworthy 1986), of a thick dome of rhyolitic lava and
breccia in the Rylstone Volcanics immediately adjacent to the mineralisation site. This theory is supported by the fact that mineralisation and quartz veining only occur contiguous to the Rylstone Volcanics contact and do not extend along strike. Epigenetic deposits in the basement around rhyolite domes are an important exploration target, particularly for precious metals (Cas & Wright 1987) and exploration in and around the Rylstone Volcanics in recent years has focussed on finding such deposits.

Stop 18. Moonbucca Formation.

(GR 725625, Kandos 1:25,000 sheet. 11.7 km).

The Silurian Moonbucca Formation overlies the Ordovician Coomber Formation with disconformable to faulted contact, and is, in turn, disconformably overlain by sediments of the Early Devonian Kandos Group. The unit crops out over 8 km² just east of Carwell Creek and comprises 250 m to 610 m of upward-fining mass-flow- to suspension-deposited clastics (including polymict conglomerate, litharenite, limestone breccia, calcarenite, shale, and allochthonous jasper blocks), interbedded with the Oakborough Volcanic Member - a northward-thickening series of massive and flow-banded dacite lavas, associated autoclastics, and intraformational volcarenites. The depositional setting was a subaqueous, terrigenous/carbonate, mass-flow apron and a closely associated silicic volcanic pile, both of which accumulated at moderate to shallow depths on a NW-facing slope. The limestone breccia clasts and calcarenite beds contain a late Wenlockian to Ludlovian coral, trilobite, and brachiopod fauna. The undifferentiated Moonbucca Formation lacks firm correlatives, although it is temporally and lithologically, but not palaeoenvironmentally, similar to the Willow Glen Formation and Tanwarra Shale.

Good exposures of the Moonbucca Formation occur in several creeks to the north of the Cudgegong Road just east of Carwell Creek (GR 725624, 725629, and 725634; Kandos 1:25,000 Sheet). Features of interest in this area include: (a) the type section for the unit
(GR 730625 to 723624) with good exposures of conglomerate, limestone breccia, and shale. Of particular interest in this section are several large allochthonous blocks of limestone near the base of the conglomerate (GR 730625); (b) particularly good exposures of the limestone breccia occur in a small quarry in a creek at GR 725629; (c) there are a number of interesting outcrops in a around a creek at GR 725624. Rhyolitic to dacitic lavas of the Oakborough Volcanic Member are exposed in western sections of the creek. The irregular shape of this rhyolite body and evidence of mixing and quenching at contacts with the shale suggest it may be a shallow intrusive cryptodome, in contrast with lavas elsewhere in the unit. At GR 723634 a couple of small allochthonous blocks of jasper can be seen. At GR 725624 a channel of conglomerate appears to be incised into the other lithologies. The conglomerate here displays normal and reverse grading, scouring, and well-developed pebble imbrication indicating currents from the south and southeast. Further up the creek, white volcanioclastic mass-flow arenites of the Oakborough Volcanic Member crop out in thick massive beds. Good exposures of this lithology also occur on the Cudgegong Road at GR 724624 where the volcanioclastics are thin bedded, alternate with shales, and display partial Bouma sequences.

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APPENDIX 1

The following material was prepared by Gary Colquhoun in the expectation that the field trip would examine exposures in the Kandos Quarries. A little more than one week before the trip, access to the quarries was withdrawn because of potential insurance problems. We have, nevertheless, included this material for the sake of completeness.

Kandos Quarries

Kandos limestone quarries are located approximately 5 km west of the town of Kandos and can be approached via Quarry Road either from Cudgegong Road (turn off 5.8 km west of Rylstone) or the Ilford Road (6 km southwest of Kandos). They are operated by Kandos Quarries Australia Ltd. (formerly Australian and Portland Cement Ltd.) who have quarried the limestone here for around 100 years. The area contains three main quarries: the old Charbon Quarry, which ceased production many years ago, is located approximately 1 km southwest of the Kandos Quarry; the No. 1 Quarry which ceased production a number of years ago and now houses processing facilities including the start of a 6 km aerial skipway with the town of Kandos; and the presently worked and larger No. 2 Quarry. Current production from the Kandos No. 2 quarry has an average grade of about 60% CaCO₃, reaching 97.5% CaCO₃ in very pure samples (Lishmund et al. 1986). MgCO₃ values are generally low; however, problems are occasionally encountered with selective dolomitisation along faults and fractures, which has caused MgCO₃ in excess of 5%, rendering the limestone unsuitable for cement production (Thomson 1988). More pervasive dolomitisation is encountered in the Clandulla Limestone several kilometres to the north near the 'Lucknow' property.

The Quarries are located on the Early Devonian (early-mid Lochkovian: ?woschmidtii to eurekaensis Zones) Clandulla Limestone of the lower Kandos Group. Strata in the Kandos No. 2 Quarry and the old Charbon Quarry are mostly uniformly west-dipping at moderate angles, although the unit is disrupted by minor faulting and mesoscopic folding. The No. 1 Quarry is centred on a major anticline and is bounded to the south and east by complex folded and faulted zones marking a contact with the Riversdale Volcanics.

The Clandulla Limestone attains a thickness of 170 m in the No. 1 Quarry and 110 m in the No. 2 Quarry (Cook 1988b); conformable contacts with underlying Warrah Conglomerate and overlying Yellowmans Creek Formation are well displayed in these quarries, as is the succession through the unit. The unit represents extensive shallow marine (depths of 0 to 24 m) carbonate deposition with biostromal and biothermal buildup dominated by stromatoporoid-Thamnopora communities with associated lagoon muds, forereef, interreef and backreef facies. The limestone grades up into the Yellowmans Creek Formation where brachiopod communities near the base of the formation indicate deepening and a transition to more elastic
Kandos No. 1 Quarry. (GR 735598 Kandos 1:25,000 Sheet)

The No. 1 quarry is notable mainly for the spectacular folding displayed in its north wall. In the core of the main anticline the contact between the Warrah Conglomerate and the Clandulla Limestone is well displayed.

The Warrah Conglomerate consists, in its upper levels, of pebble to granule conglomerate and fine sandstone in moderate to thick, sharp-based beds. Cross-bedding is visible in coarse sandstone horizons indicating sediment transport towards the southwest. Conglomerate clasts are dominantly black calcareous siltstone and quartz. The conglomerates are interpreted as fan deltaic deposits which prograded into a sheltered dysaerobic lagoon; continued transgression eventually cut off the terrigenous clastic supply and lead to carbonate deposition dominated by stromatoporoid-\textit{Thamnopora} communities (Clandulla Limestone).

The other notable feature of the No. 1 Quarry occurs on the top bench of the eastern wall. Here, a unit of calcareous shale with thin nodular limestone interbeds marks the transition zone to the overlying Yellowmans Creek Formation. The calcareous shale displays an undescribed, prolific though low diversity, silicified fauna of: brachiopods (\textit{Isorthis} sp., \textit{Eoschuchertella} sp., \textit{Cyrtilina} sp.), gastropods, rostroconchs, trilobites (\textit{Apocalymene} sp.), tabulate corals (\textit{Cladopora} sp.), bryozoans, fish and ostracodes (Cook 1988). Conodonts from the more limey horizons in this area include \textit{Icriodus postwoschmidtii}, \textit{Ozarkodina remschiedensis remschiedensis}, and \textit{?Amydrotaxis sexidentata}, indicating the \textit{eurekaensis} Zone of middle Lochkovian age.

Other facies of the Clandulla Limestone, well displayed in the No. 1 Quarry, are often inaccessible due to the dangerous nature of the walls.

Section between Carwell Creek and Charbon Quarry. (GR 729594 Kandos 1:25,000 Sheet)

This section, located just south of Carwell Creek, was proposed as the type section for the Clandulla Limestone (then the Kandos Limestone) by Campbell (1981). The underlying Warrah Conglomerate and the overlying Yellowmans Creek Formation in this section provide an excellent picture of the lower Kandos Group transgression.

The section begins in a small creek where the upper parts of Warrah Conglomerate are displayed - mainly massive pebble-granule conglomerate deposited in fan deltaic environments. The base of the Warrah Conglomerate is, at this locality, apparently faulted against the silicic volcanics of the Riversdale Volcanics approximately half way up the large hill to the immediate east. The Warrah Conglomerate grades into the
Clandulla Limestone, the transition zone being represented by numerous thin beds of granule conglomerate which gradually decrease upwards. The Clandulla Limestone is only around 50 m thick in this section in contrast to its much greater thickness in the quarries, and appears to represent mainly biostromal deposition without development of reefs. The limestone consists of mainly thick-bedded biomicrite at the base dominated by high density but low diversity stromatoporoid (mainly Amphipora) and coral (mainly favositid) faunas, suggesting a shallow, somewhat restricted environment. Silicification is typically well-developed throughout the section. Towards the top of the Clandulla Limestone, the limestones become more thin-bedded and brachiopods appear in increasing numbers, reflecting deepening of the environment and more open marine conditions. The upper 2 m of the Clandulla Limestone, containing thin beds of densely packed brachiopods (Cyrina, Eoschuchertella, Machaeraria and Schizophoria), possibly represents condensed faunal sections deposited during times of low carbonate accumulation. Conodonts from these brachiopod beds are similar to the above-mentioned fauna from the top bench of the No. 1 Quarry.

Above the brachiopod beds the Yellowmans Creek Formation begins and consists of interbedded calcareous, often fossiliferous, shales and thin 10 to 50 cm thick beds of flaggy allodapic limestone. The latter are occasionally fossiliferous, though rarely silicified. Such a facies is typical of deeper fore-reef settings and deeper carbonate ramps. Interbedded with these shales and limestones are several beds of volcaniclastic breccia to volcaniclastic sandstone which probably represent subaqueous debris flows which may have developed directly from subaerial pyroclastic eruptions or from slumping of rapidly deposited pyroclastic and epiclastic detritus. These are the oldest suggestions of active volcanism in the Kandos Group. Volcanic detritus in overlying units suggests volcanism remained sporadically active throughout the rest of the history of the Kandos Group.

Approximately 100-150 m above the base of the Yellowmans Creek Formation, the calcareous shales and allodapic limestones give way to barren grey-green shales, indicating eventual drowning of the carbonate platform and cutting off of the supply of carbonate detritus due to continued transgression.

Kandos No. 2 Quarry. (GR 725615, Kandos 1:25,000 sheet)

The west wall of the quarry reveals a section through the Clandulla Limestone which contrasts with the previous section by showing evidence of reefal buildups. Cook (1988) has divided the limestone here into six biofacies.

The sequence begins with the Black Shale Biofacies which overlies the Warrah Conglomerate with a contact similar to that seen in Kandos No. 1 Quarry. The unfossiliferous Black Shale Biofacies exhibits extensive pyritisation, suggesting quiet, dysaerobic lagoonal conditions at the base of the limestone.
This facies is overlain by a Stromatoporoid Biofacies which represents development of stromatoporoid biostromes and bioherms, dominated by hemispherical stromatoporoids which were able to overcome the oxygen-poor benthic conditions caused by a slight increase in water depth.

The overlying *Thamnopora* Biofacies is characterised by biostromal development of *Thamnopora* thickets, with *Thamnopora* acting as sediment bafflers in the quiet water conditions.

The Reef Biofacies is dominated by the ramose, branching stromatoporoid *Amphipora*. Such a reef-building role for *Amphipora* is only possible in quiet, restricted waters.

The Flank Biofacies represents carbonate mud and fine sand deposition on the reef margin. Some degree of reef orientation is suggested by lithological zonation with this biofacies in Quarry No. 2, with fine carbonaceous muds to the south and slightly coarser carbonate sands to the north of the reef structure indicating back- and fore-reef deposits respectively.

The Flank Biofacies is overlain by calcareous shale, allodapic limestone and volcaniclastic breccia/arenite of the basal Yellowmans Creek Formation, well-exposed on the top two benches of the western side and similar to the previous section.
## Windamere Dam in facts and figures

| **Location** | Cudgegong River about 23 kilometres by road south-east of Mudgee, NSW |
| **Storage capacity** | 353,000 megalitres. (1 megalitre = 1,000 cubic metres or approx. 200,000 gallons) |
| **Area submerged** | 2,000 hectares |
| **Catchment area** | 4,400 kilometres |
| **Maximum water depth** | 53 metres |
| **Main embankment** | Earth and rockfill |
| Height | 65 metres |
| Length of crest | 800 metres |
| Width of crest | 8 metres |
| Maximum width of base | 215 metres |
| **Embankment quantities:** | |
| Impervious core | 440,000 cubic metres |
| Filters | 140,000 cubic metres |
| Rockfill | 1,070,000 cubic metres |
| Total fill | 1,650,000 cubic metres |
| Foundation excavation | 200,000 cubic metres |
| **Spillway** | Open unlined spillway and discharge channel of 1,220,000 cubic metres excavation located in a saddle formation about one kilometre west of the damsite. A bridge to be constructed over the spillway to carry the access road to the dam. |
| **Maximum spillway discharge** | 4,800 cubic metres per second. |
| **Diversion tunnel** | Length | 188.5 metres |
| Diameter (concrete lined) | 3.6 metres |
| **Outlet works** | Outlet tower height | 57.0 metres |
| Diameter of penstock | 1.60 metres |
| Diameter of by-pass pipe | 610 mm |
| Outlet valves | One 1,200 mm diameter dispersion cone valve |
| One 610 mm diameter fixed dispersion cone valve. |
| **Total concrete** | Outlet tower | 1,630 cubic metres |
| Diversion tower | 1,340 cubic metres |
| Dissipator structure | 1,000 cubic metres |
| including valve block |
| **Maximum outlet works discharge** | 1,800 megalitres per day. |

**Trunk Road No. 55 deviation**

A section of Trunk Road No. 55 (Mudgee-Wallerawang Road) passed through these proposed spillway site and storage area and required to be located around these areas by a road deviation of length approximately 15.5 kilometres.
APPENDIX 3

Sunday afternoon option: Aarons Pass Granite and associated hornfels field guide

The Aarons Pass Granite is an I-type, biotite granite to adamellite stock, 10 km in diameter and covering an area of 45 km², both north and south of Aarons Pass. The body is compositionally homogeneous yet the rocks show considerable textural variation from typical coarse-grained porphyritic granite (quartz, plagioclase, microcline, microperthite and brown biotite) to marginal phases (coarse-grained phenocrysts with an aphanitic groundmass) suggesting probable pressure-quenched textures at the pluton carapace (Greenfield 1992). The granite is associated with late stage aplite masses and dykes, pegmatite (quartz-microcline) and granophyre (probable sill 2 km long by 500 m wide at southern margin).

A 328 Ma (Middle Carboniferous) age has been determined from Rb/Sr ratios of biotite separates (Pemberton 1990) and Vickary (1983) reported a 320 Ma age from a small outlier of the granite 3 km from the western margin.

The intrusion produced a low pressure elliptical-shaped aureole up to 1.5 km wide (best seen along track to old marble quarry between GR 607634 to GR 607646). The aureole is widest in the north (1.5 km) and south (1.2 km), and may reach 800 m in the east; there is no evidence for contact metamorphism along the poorly exposed western margin where possible thrust faulting is inferred. Four prograde contact metamorphic zones are identified in pelites and psammites, ranging from an outer zone 1 (chlorite + muscovite + biotite; up to 500 m wide), zone 2 (cordierite + muscovite + biotite; up to 500 m wide), zone 3 (cordierite + K-feldspar + muscovite + biotite; up to 150 m wide) and an inner very restricted zone 4 (andalusite + cordierite + muscovite + biotite). Other hornfels types include marble, metabasite and skarns (calcite + quartz + diopside + wollastonite + garnet). The contact metamorphism is superimposed on low-grade regional metamorphism. Other small I-type granite stocks occur at "Havilah" and Pyangle Pass.

Possible Itinerary

Stop 1

*From Tabrabucca, turn north onto the Sydney road. Travel for 3.0 km past Aarons Pass to the base of the Pass, with heavily wooded prominent hill slope to your east. Park vehicle by side of road, climb fence at gate and walk east along creek bank by the edge of the wooded hill (rhyolite lava of Toolamanang Formation marks contact with Aarons Pass Granite). After 1.1 km, head north up slope for 50 m to old shallow workings. GR 634622, Kandos 1:25,000 sheet.*

Old workings in garnet-rich skarn. Abundant coarse-grained garnet (+ calcite, quartz and diopside).
Stop 2

*From Tabrabucca, turn north onto the Sydney road. Travel to Aarons Pass (500 m), turn west onto the Pyramul road. Follow this dirt road for 2.7 km to locked gate near prominent cattleyards (CALM holdings; access required from Paul Gibb; agistment held by Trevor Harding at 'Tabrabucca'). Enter gate, travel north along 4 wheel drive track down slope into granite country (tors and slabs abundant yet strongly weathered) to Y junction, take right fork through gate to base of hill where granite outcrop ends and rolling hills mark hornfels outcrops. Continue around this hill as track swings to the north at the edge of a small dam. By the edge of the track a variety of hornfels types are present including calc-silicate, skarns and spotted pelite. Travel further north along track, cross creek noticing abandoned and overgrown marble (wollastonite abundant) quarry at GR 608646 (access very difficult). Better and more accessible exposures of wollastonite marble in disused quarry GR 605648 over hill to NW (take west track back at creek crossing, instead of east to overgrown quarry).*

Stop 3

*Best bedded skarn crops out at the NW extremity of the granite contact. However, access is difficult, involving old tracks and some cross country driving. Consult the Kandos 1:25,000 sheet GR 583643.*

Stop 4

*From Tabrabucca, turn south onto the Sydney road and travel 1.8 km to the Crudine road. Turn onto the road and travel over deeply weathered granite 1.7 km to 'Donnasville' farmhouse (Ian Moore, owner). Need to drive SE of farmhouse, past numerous sheds and E along farm track to prominent granite outcrop about 300 m E of farmhouse. GR 622560, Ilford 1:25,000 sheet. Here a 3 m high and 10 m wide exposure of interlayered sheets of aplite and pegmatite (cut by veins of pegmatite, aplite and quartz; and underlain by aphanitic granite) probably represents the roof of the pluton.*
Stop 5

Leave 'Donnasville', turn south onto the Crudine road for 0.9 km. Road cuttings on both sides of road and loose boulders on east side of road, just before creek crossing. GR 613552, Ilford 1:25,000 sheet.

Cuttings include a variety of hornfels types (spotted pelite, calc-silicate and silicic volcanics) associated with a narrow exposure of the Willow Glen Formation very close to the granite margin. Immediately southeast on the hill slopes, the granophyre sill clearly crops out.

Stop 6

Continue south, crossing creek and turn west to 'Nolac' property (Moss family, owners). Past farmhouse about 100 m, double farm gates. Take southern gate, travelling west along farm track over hill (through another farm gate), around hill slope to the north. So far 1.5 km on farm track. Steep gully to the south. GR 600555, Ilford 1:25,000 sheet.

On the upper western slopes of gully, the contact between spotted pelite of the Late Silurian Chesleigh Formation and the granite is well exposed. In several localities, pelite blocks up to 5m across, crop out within the granite which contains common pelite xenoliths.
Supporting Paper 4

Supporting Paper 5
The Capertee High in the Mudgee-Capertee region was a Late Silurian to earliest Middle Devonian shallow marine to subaerial palaeogeographic unit to the E of the Hill End Trough. Rocks of the Capertee High largely occur as structurally complex sequences along with Ordovician basement, thick Late Devonian siliciclastic sequences of the Lambie Group, and Middle Carboniferous granitic intrusions. Flat-lying Permian-Mesozoic sequences overlap the High in the E and N.

Broadly, a two-fold subdivision of the Early Devonian sequences associated with the Capertee High is apparent:

Platform Sequences were deposited on the main shelf areas of the High and are dominated by the ~4000 m thick Kandos Group. Recent conodont dating of this sequence of fine- to coarse-grained siliciclastics, carbonates, and silicic volcanics indicates it spanned almost all of the Early Devonian. The Kandos Group is most fully exposed in the Rylstone-Cudgewong district where it progressively onlaps the Silurian sequence; however, a correlative of the Pragian Riversdale Volcanics (silicic pyroclastics and volcaniclastics) occurs in the Capertee Valley (Huntingdale Volcanics), and the early Emsian Carwell Creek Formation (shallow marine siliciclastics and carbonates) outcrops almost continuously from the southern Capertee Valley (Myrtle Grove Formation) to northeast of Mudgee. Western Platform Margin Sequences consist of siliciclastic and carbonate mass-flow deposits, hemipelagites and rare silicic to intermediate volcanics deposited in slope and base-of-slope settings adjacent to the western margin of the High. These sequences crop out in the Queens Pinch Group south and east of Mudgee, in the Limekilns Formations at Limekilns, and in the Kingsford Formation near Ilford. Most units are well dated by conodonts, allowing correlation with platform sequences; correlation with the units of the eastern Hill End Trough is more difficult as their ages are comparatively poorly constrained. Sedimentation in marginal High sequences is linked to episodes of transgression, regression, erosion and volcanism on the platform areas of the High.

Palaeogeography and Sedimentation: Lochkovian platform sequences are known only from the Rylstone-Cudgewong area, although very similar facies were deposited 150 km to the S in the Lochkovian Tangerang Formation. The Lochkovian portion of the Kandos Group is a transgressive to regressive shallow marine sequence deposited on a W- to SW-sloping shelf flanked to the E by eroding sedimentary basement and active silicic volcanoes, and passing to the W into deepwater facies along the western margin of the High (Mullamuddy Formation) and in the eastern Hill End Trough (upper Crudine Group). Basal fan deltaic clastics (Warrah Conglomerate) are overlain by mainly biostromal carbonates (Clandulla Limestone) which, in turn, are overlain by the muddy shelf deposits of the Yellowmans Creek Formation. The thick regressive siliciclastic sequences of the storm-dominated Roxburgh Formation overlie the Yellowmans Creek Formation and possibly extend into the earliest Pragian. Facies prograded W during the late Lochkovian regression, culminating in shelf exposure and some erosion in the E.

During the Pragian, relative low stands of sea level dominated and shelf areas of the High were largely subaerial, apart from a narrow shallow marine shelf margin. Voluminous silicic pyroclastic volcanism dominated the shelf area and calderas, probably centred southeast of Cudgewong and in the eastern Capertee Valley, were blanked by extensive ignimbrite sheets. Short-lived transgressions on the platform resulted in minor carbonate deposition during volcanically quiet periods. Vast volumes of volcanic detritus were transported to the W and deposited as low-stand fans and aprons along the margin of the Capertee High and in the Hill End Trough. Transgression in the latest Pragian-earliest Emsian partly eroded, then drowned the existing volcanic topography, a retrograding complex of shallow marine shelf and shoreline siliciclastics and carbonates was then deposited in barrier island to open marine settings in the platform areas. Some silicic volcanoes remained active, perhaps forming islands on the shallow shelf. Concurrent sedimentation in the marginal High sequences was dominated by fossiliferous condensed mudstone sequences. During the mid to late Emsian, at least two major regressions caused erosion of early Emsian platform carbonates and siliciclastics; these were redeposited in mass-flow submarine fans and aprons adjacent to the W margin. Minor silicic volcanism occurred in these marginal High areas at this time. Carbonate deposition on the platform continued into the earliest Eifelian (at Mt Frame) before being replaced by shallow marine siliciclastics of the Boogledie Formation. Deposition was probably terminated in the early Eifelian by uplift associated with the weak Middle Devonian Tabberabberan deformation.
The Mudgee 1:100,000 Geological Sheet forms the most southeasterly of the six sheets comprising the Dubbo 1:250,000 Geological Sheet and encompasses the northeastern margin of the exposed Lachlan Fold Belt and the western margin of the Sydney Basin. Compilation of existing geological data and field mapping of the Mudgee 1:100,000 Sheet were undertaken in January to June 1995 by the Geological Survey of New South Wales and AGSO as part of the National Geoscience Mapping Accord. Mapping was aided by airborne magnetic and radiometric data flown in 1991 at a 400 m line-spacing. The stratigraphy and structure differ significantly from those embodied in the original Dubbo sheet (1st ed., 1971); salient features of the new map include:

Ordovician: major changes to the extent and structure of the Ordovician units, particularly the volcanic units of the Cabonne Group which are currently significant exploration targets for Cu-Au mineralisation. The Lue beds - previously considered Silurian - have been recognised as Ordovician (see Fergusson and Colquhoun, this volume) and divided into the Early Ordovician Adaminaby Group (quartz turbidites) and the Late Ordovician-?Early Silurian Coomber Formation (mafic volcanioclastics, lavas, intrusives and mudstone). The mafic-volcaniclastic Burrah Formation, previously regarded as Early Devonian, contains a Late Ordovician coral fauna in allochthonous limestones and has a similar radiometric signature to the Coomber Formation; it is suspected this unit may have been the source of much of the alluvial gold in the nearby Gulgong gold fields. North of Aarons Pass, the mapped outcrop and structure of the Sofala Volcanics has altered considerably.

Silurian: inclusion and definition of the Tannabutta Group, a Capertee High unit of Late Silurian shallow marine clastics, carbonates and silicic volcanics outcropping between Mudgee and Cudgewang in the west, and Botobolar and Kandos in the east. The silicic volcanics of the Group contain a number of massive sulphide and vein gold prospects. A small area of the Wenlockian Tanwarra Shale (Mumbil Group) has been mapped to the east of the Sofala Volcanics, west of Ilford. In the Hill End Trough strata, the turbiditic Chesleigh Formation (Mumbil Group) was subdivided into a lower unit of state and lithic sandstone and an upper unit of predominantly volcanioclastic sandstone which includes two concordant rhyolite to dacite and coarse volcanioclastic horizons, the upper formerly known as the Nulling Formation.

Devonian: the complex Early Devonian shallow marine to subaerial sequences of the Kandos Group (see Colquhoun, this volume) - mostly undifferentiated on the original sheet - have been mapped in detail, largely following the work of University of Wollongong students and staff. The fault-bounded Early Devonian sequences south of Mudgee (the Queens Pinch Group) remain little changed from the original map; however, some constituent units of Mudgee have been traced northwards to the southern outskirts of Mudgee, and new units are recognised north of Mudgee (Tinja Formation) and near Ilford (Kingsford Formation). The stratigraphy and structure of the Crudine Group of the Hill End Trough have been modified: the Lana Formation was found to be a junior synonym of the Turondale Formation, and the Dunnoogin and Guroba Formations are lateral facies equivalents of the Waterbeach Formation. The four-fold subdivision of the Late Devonian Lambie Group remains unchanged from the original sheet; however, recent mapping has increased knowledge of the complex structure to the west of Lue, where an en-échelon fault system cuts the sequence into several openly-folded blocks.

Carboniferous-Permian: Mid Carboniferous granites of the Mudgee Sheet lack the concentric radiometric and magnetic zonation discovered in many plutons of the Bathurst Batholith. Instead the radiometrics highlighted anomalous zones in the Aarons Pass (328 Ma) and Camboon Granites of relative thorium enrichment and depletion separated by sharp, linear contacts; these features are yet to be explained. In addition, magnetic data indicate the Aarons Pass and Havilah Granites have considerable shallow sub-surface extent and an associated network of mostly unexposed dykes. Outcrop of the earliest Permian (292 Ma) Rylstone Volcanics along the edge of the Sydney Basin has been refined and extended. This unit now constitutes an important exploration target following the discovery of a major epithermal vein and disseminated Ag-Pb-Zn deposit in the unit near Lue.

Hill End Trough strata are characterised by tight, gently-plunging folds with plunge reversals common over several kilometres. Capertee Zone rocks vary from open to isoclinal folded and contain several generations of reverse and normal faults. One major deformation (D2) affected the sheet area in the Early Carboniferous, resulting in folding, thrusting and cleavage development. Evidence for earlier deformation (D1) is meagre and restricted to steeply plunging folds and lineations in the Ordovician Adaminaby Group. Areas of kinking and folding of S2 cleavage in the Lue-Havilah area indicate significant D3 deformation (probably also Early Carboniferous). Major changes in the regional bedding and cleavage trends occur across lineations in the Bocoble and Havilah areas, suggesting significant north-south compression postdating D3 but un-timed with respect to D2. Disconformities between the Ordovician-Silurian and Silurian-Devonian sequences suggest erosion and exposure at these times, due either to gentle uplift or sea-level fall.
Supporting Paper 8

Stratigraphy, Structure and Mineralisation of the Mudgee 1:100 000 Geological Map Sheet

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ABSTRACT

The Mudgee 1:100 000 geological map sheet forms the south-eastern sixth of the Dubbo 1:250 000 geological map sheet and encompasses the exposed north-eastern margin of the Lachlan Fold Belt and the western margin of the Sydney Basin. Recent remapping of the map sheet area by the Geological Survey of New South Wales and the Australian Geological Survey Organisation (AGSO) resulted in a major reinterpretation of the stratigraphy and structure, particularly of the Late Ordovician volcanic units of the Cabonne Group — currently significant exploration targets for Cu–Au mineralisation. The Lue beds — regarded as Silurian on the original Dubbo geological map sheet — have been divided into the Early Ordovician Adaminaby Group (quartz turbidites) and the Late Ordovician–Early Silurian Coomber Formation (mafic volcaniclastics, lavas, intrusives and mudstones). The mafic volcaniclastic Burranah Formation, previously regarded as Early Devonian, contains a Late Ordovician coral and conodont fauna in allochthonous limestones and has a similar radiometric signature to that of the Coomber Formation.

The Silurian sequence is dominated by the Tannabutta Group (new name), a Capertee High unit of Late Silurian shallow marine to subaerial clastics, carbonates and silicic volcanics outcropping between Mudgee and Cudgegong in the west, and Botobolar and Kandos in the east. The silicic volcanics of the group contain numerous massive sulphide and vein gold prospects. A small area of the Wenlock Tanwarra Shale (Mumbil Group) has been mapped to the east of the Sofala Volcanics, west of Ilford. In the Hill End Trough sequence, the turbiditic Chesleigh Group has been subdivided into a lower unit of slate and lithic sandstone (Biraganbil Formation) and an upper unit of predominantly rhyolitic volcaniclastics, latite lava and tuffaceous mudstone (Piambong Formation); the latter has been intruded by thick Early Devonian mafic sills.

The complex Early Devonian shallow marine to subaerial sequences of the Kandos Group have been mapped in detail. The Queens Pinch Group is proposed for the fault-bounded, trough margin sequences south of Mudgee, which have been modified slightly from the mapping of Wright (1966). Constituent units of the group have been traced northwards to the outskirts of Mudgee; in addition, a new formation of the group has been recognised near Ilford (Kingsford Formation). The stratigraphy of the Crudine Group sequences of the northern Hill End Trough has been modified: the ‘Lana Formation’ and ‘Dummoogin Formation’ are considered junior synonyms of the Turondale and Waterbeach Formations respectively, and the Guroba Formation was recognised as a widespread sandy facies between the Waterbeach and Merrions Formations. The four-fold subdivision of the Late Devonian Mount Knowles Group (new name) remains unchanged from the original map sheet; however, recent mapping has increased knowledge of the complex structure west of Lue where several en echelon faults cut the Late Devonian sequence into openly-folded blocks.

Carboniferous granites of the Mudgee map sheet area lack the concentric radiometric and magnetic zonation observed in many plutons of the Bathurst Batholith. In contrast, the radiometric results highlighted anomalous zones in the Aarons Pass and Pyangle Pass Granites, of relative thorium enrichment and depletion separated by sharp, linear contacts; these features are yet to be explained. In addition, magnetic data indicate the Aarons Pass and Havilah Granites have considerable shallow subsurface extent and an associated network of mostly unexposed dykes.
INTRODUCTION

The Mudgee 1:100 000 map sheet (8832) occupies the south-east corner of the six 1:100 000 map sheets which comprise the Dubbo 1:250 000 geological map sheet (figure 1). Since the original mapping and compilation of the Dubbo geological map sheet by Offenberg et al (1971) — and a minor update by Matson (1973) — the area has been the subject of considerable university thesis work, most notably by students of the University of New South Wales in the 1970s and the University of Wollongong in the 1970s to 1990s. The map sheet area has a range of mineral occurrences and has attracted considerable exploration activity, yielding additional geological data. Mapping and compilation of the Mudgee 1:100 000 map sheet area took place between January and July 1995 and was a joint undertaking by the Geological Survey of New South Wales and the Australian Geological Survey Organisation (AGSO) as part of the National Geoscience Mapping Accord. The project also involved significant collaboration with the staff of the School of Geosciences at the University of Wollongong who have considerable experience in the Mudgee-Cudgegong district. Mapping was greatly assisted by airborne radiometric and magnetic data collected at 400 m flight line spacing across the area by AGSO in 1991. A programme of SHRIMP zircon dating of the rocks was undertaken in an effort to establish a better chronological framework for the stratigraphy (Fanning in prep). The new map (Colquhoun et al 1996b) represents the first thorough synthesis of this comparatively poorly known area of the Lachlan Fold Belt and differs markedly from the Mudgee portion of the original Dubbo geological map sheet. This paper expands on findings which were summarised in Colquhoun et al (1996) and aims to detail the stratigraphy and structure of the Mudgee 1:100 000 map sheet area and the implications for mineralisation.
The Mudgee map sheet area is located at the north-east margin of the exposed Lachlan Fold Belt (figure 1A). The area has three main structural subdivisions (after Scheibner 1993):

1. the Hill End Zone, which is predominantly composed of Late Silurian–Early Devonian deepwater sedimentary sequences representing the deformed fill of the Hill End Trough (Packham 1960).

2. the Capertee Zone, containing shallow marine to subaerial sequences (and their slope-deposited equivalents) of the Late Silurian to early Middle Devonian Capertee High in addition to Ordovician basement, thick Late Devonian siliciclastic sequences and Carboniferous granites.

3. the flat-lying Permian–Triassic sequences of the Sydney Basin (figure 1B).

STRATIGRAPHY

A summary of the revised stratigraphy of the Mudgee 1:100 000 map sheet area is presented in figure 2.

Ordovician

The oldest rocks are those of the Early Ordovician Adaminaby Group, which consists of submarine fan deposits of turbiditic quartz-rich sandstone, slate and bedded radiolarian chert occurring in two belts between Botobolar and Lue, and to the south of Lue (figure 3) (Fergusson & Colquhoun 1996, Colquhoun et al in review); these rocks were formerly mapped as part of the Silurian Lue beds by Offenberg et al (1971). Conodonts of probable late Darriwilian age have been obtained from cherts near Bara Creek (Stewart & Fergusson 1995).

Three distinct Late Ordovician mafic volcanic successions have been placed in the Cabonne Group, defined in the legend of the recent Bathurst 1:250 000 geological map sheet.

The first is the Sofala Volcanics, which occurs in four distinct areas and extends south into the Bathurst 1:100 000 map sheet area (figure 3). The most southerly outcrop is a deep marine succession of black mudstone and black radiolarian chert, with rare lithic sandstone, basaltic boulder conglomerate and basalt sills or flows. The three more northerly areas of outcrop — formerly the Cudgegong Volcanics of Pemberton (1989, 1990) — represent a more proximal volcanic succession consisting of basalt to andesite lava, fine to coarse grained basaltic sandstone and breccia, mudstone and rare limestone. The unit contains a Gisbornian coral and alga fauna (Pickett 1982; Pemberton 1989, 1990) and poorly preserved graptolites of probable Bolindian age (Bo3: A.H.M. VandenBerg, written comm to Pemberton 1995) — very close to the Ordovician–Silurian boundary of Jones (1996). The unit appears to have spanned the Late Ordovician and, as the graptolites occur well below formation top, possibly extended into the earliest Silurian.

The second unit, the Coomber Formation (Pemberton et al 1994), is a deep marine apron deposit consisting of volcanioclastic sandstone, mudstone, chert, basaltic sills and flows, and intermediate intrusives (Fergusson & Colquhoun 1996, Colquhoun et al in review). The Coomber Formation concordantly overlies the Adaminaby Group, or is faulted against it (figures 2 & 3) and shows a distinctive, relatively potassic, response in the radiometric data. The formation, which was formerly mapped as part of the Lue beds (Offenberg et al 1971), has not been dated directly but is constrained on stratigraphic grounds to Late Ordovician–Early Silurian (Fergusson & Colquhoun 1996).

The Burranah Formation is a unit of mafic volcanioclastic sandstone, basalt, intermediate to mafic intrusives and rare limestone which occurs between Mudgee and Gulgong (Exon 1962, Armstrong 1983, Watkins 1996). This unit was shown as Early Devonian on the original Dubbo geological map sheet (after Exon 1962); however, the discovery during recent mapping of a Late Ordovician coral–alga fauna (Plasmoporella sp and Vermiporella sp.,...
Figure 2  Time–space diagram for the Mudgee 1:100 000 map sheet area. Ages are by U/Pb SHRIMP techniques except where otherwise indicated.
Figure 3 Simplified geology of the Mudgee 1:100 000 map sheet area. See figure 2 for key to unit names and the text for unit descriptions.
The Tanawarra Shale (Mumbil Group) is the most northerly outcrops of the Lachlan Fold Belt and a similar allochthonous limestone of the Toolamanang Formation at GR 772200 6366100 and a dacite lava from GR 753800 6393800 in the Botobolar area gave an age of 419.2 ± 4.8 Ma (Fanning in prep). A rhyolite lava from newly-recognised outcrops of the unit east of Dunedoo (GR 753240 6452750) in the Cobbora 1:100 000 map sheet area, gave an age of 415 ± 7 Ma (Fanning in prep). The fauna displays similarities with faunas of the late Wenlockian to early Ludlovian, Willow Glen Formation in the Cudgegong District (Pemberton 1989). The SHRIMP ages confirm that the Dungeree Volcanics is partially equivalent to the Windamere Volcanics and, presumably, the Toolamanang Formation.

Depositional environment for the Dungeree Volcanics was apparently somewhat varied: southern outcrops suggest a carbonate/terrigenous slope environment influenced by a growing volcanic pile whereas farther north the unit appears to represent a thick shallow marine to emergent volcanic pile with fringing clastics and carbonates, and adjacent slope areas.

Devonian

Early Devonian

Early Devonian sequences occur in three groups:
Kandos Group, a ∼4 km thick shallow marine to subaerial succession deposited on the main platform areas of the Capertee High

- Queens Pinch Group (new name), consisting mostly of siliciclastic and carbonate mass flow sequences deposited in slope settings along the western margin of the Capertee High

- Crudine Group, composed of mainly deepwater volcaniclastic successions representing the fill of the Hill End Trough.

Units of the Kandos Group crop out extensively in the Ryllstone-Cudgegong district and have been described in detail by Pemberton (1980), Cook (1990), Milstead (1992), Pemberton et al (1994), Colquhoun (1995a, 1996, and in prep.) and Wright et al (1994); the conodont biostratigraphy of the sequence has been discussed by Colquhoun (1995b). The sequence spans almost the entire Early Devonian and represents an early to middle Lochkovian transgressive sequence of basalt fan deltaic clastics (Warrah Conglomerate) overlain by reeval to biotromal carbonates (Clarendon Limestone) which in turn are overlain by a muddy storm-influenced shelf sequence (Yellowmans Creek Formation). The thick regressive siliciclastic sequences of the late Lochkovian storm- and tide-influenced Roxburgh Formation overlie the Yellowmans Creek Formation and complete the Lochkovian transgressive-regressive cycle. The Roxburgh Formation is overlain with partial disconformity by the Pragian Riversdale Volcanics, a sequence of subaerial to shallow marine dacitic pyroclastics, volcaniclastics, epiclastics and rare limestone.

Three samples from the Riversdale Volcanics were submitted for U/Pb dating by SHRIMP. Dacitic ignimbrite from close to the base of the unit (GR 76870 6363000) gave an age of 410±8 Ma, whereas purple dacitic ignimbrite close to the top of the unit (GR 770280 6362280) gave 389±8 Ma (Fanning in prep). Due to the large error bars caused by multiple zircon populations, these dates only confirm an Early Devonian age for the Riversdale Volcanics, which is constrained by biostratigraphic evidence to the early to middle Pragian (Colquhoun 1995b). The third sample from the unit gave an anomalous age of 418±5.5 Ma (Fanning in prep), suggesting inheritance of zircons from Late Silurian volcanics.

The latest Pragian to Emsian Carwell Creek Formation conformably overlies the Riversdale Volcanics and comprises over 1 000 m of siliciclastics and carbonates deposited in barrier island and shoreline settings. As the Kandos Group progressively onlaps the Late Silurian Dunageree Volcanics near Carwell Creek (GR 772015 6364800), only the uppermost unit of the Group (the Carwell Creek Formation) is present to the north. This unit has been mapped almost continuously to the vicinity of Havilah property, 30 km north, where it is offset by the Havilah Fault and continues north of the Lue Road in the Mount Knowles–Buckaroo–Eurundy district (figure 3). Equivalents of the Riversdale Volcanics and Carwell Creek Formation also occur extensively in the Capertee Valley to the south (Pemberton et al 1994, Colquhoun 1995b).

The Mount Frome Limestone crops out 5 km south-east of Mudgee and consists of ∼200 m of richly to sparsely fossiliferous limestone deposited in shallow subtidal to intertidal environments (Wright 1966, 1969; Pemberton et al 1994). Deposition of the unit extended from the early Emsian (perbonus Zone) to the earliest Eifelian (partitus Zone). The Mount Frome Limestone is overlain by the Boogledie Formation, a unit of sandstone (rich in silicic volcanic detritus and frequently crinoidal), shale, pebbly sandstone and conglomerate deposited in wave-dominated shallow marine conditions. The Boogledie Formation is the youngest unit of the Kandos Group and is presumably early Middle Devonian (Eifelian) in age (figure 2). Mapping has redefined the Boogledie Formation to include only the clastics which overly the Mount Frome Limestone in the Mount Frome area, and not the older (early Emsian) clastics and carbonates of the Mount Knowles–Buckaroo area which are now thought to be part of the Carwell Creek Formation (see above).

The Queens Pinch Group (new name) consists of 1 700 m of siliciclastic and carbonate mass flow deposits, hemipelagites and silicic to basic volcanics deposited in slope and base-of-slope settings along the western margin of the Capertee High. The type area of the Group is in the Queens Pinch district, 16 km to the south of Mudgee, where it occurs in a series of fault-bounded, wedge shaped blocks between the Mudgee Thrust in the west and the Silurian Tannabutta Group in the east. Stratigraphy and structure of the area was elucidated by Wright (1966, 1967, 1969) and remains largely unchanged. Most units are accurately dated by conodonts and macrofossils (Garratt & Wright 1988, McCracken 1990, Pemberton et al 1994) allowing correlation with the Kandos Group units. The sequence at Queens Pinch consists of the late Lochkovian to Pragian Mullamuddy Formation (mass flow volcaniclastic conglomerate, mostly allochthonous limestone, shale and basalt) overlain by the late Pragian Taylors Hill Formation (dark shale, volcaniclastic turbidite and alloapidic limestone). The presence of basalt lava in the Mullamuddy Formation, together with the recent SHRIMP age determinations demonstrating widespread Early Devonian mafic to intermediate intrusive in Ordovician and Silurian sequences (see above), indicates Early Devonian magmatic activity on the Capertee High was bimodal (silicic/mafic).

In a separate fault block, the earliest Emsian Warratra Mudstone (mudstone and shale) is conformably overlain by the middle to late Emsian Sutchers Creek Formation (limestone conglomerate and allochthonous blocks, alloapidic limestone and mudstone intervals). The middle Emsian Ingleburn Formation crops out in another fault-bounded block and comprises basal dacite lava followed by fossiliferous shale, dacite, mass flow conglomerate and cross-bedded sandstone. Dacite lava from GR 754222 6371751 in the Ingleburn Formation yielded a U/Pb SHRIMP age of 403±8 Ma (Fanning in prep). During recent mapping, the Sutchers Creek Formation and the Warratra Mudstone were traced north from the type areas originally mapped by Wright (1968) to the southern outskirts of Mudgee. Early Devonian outcrops directly north of Mudgee (Tinja Formation of Offenberg et al 1971) have also been placed in the Sutchers Creek Formation. The unit at this locality unconformably overlies or is faulted against the Ordovician Burrarah Formation and comprises shale, mudstone, volcaniclastic sandstone, pebbly mudstone, rare limestone blocks and diorite — the latter of Early Devonian age, similar to the diorite intruding the Burrarah Formation.
Macrofaunas similar to those of the Sutchers Creek Formation in its type area were recovered from these outcrops (Wright 1995).

The Kingsford Formation occurs in a small area to the north of Ilford where it is faulted against the Roxburgh Formation. It is composed of volcanoclastic turbidite, dark mudstone and allochthonous limestone blocks and breccia. The Kingsford Formation was considered to be a proximal slope equivalent of the Riversdale Volcanics by Colquhoun (1995) and has been placed in the Queens Pinch Group.

The Crudine Group of the Hill End Trough comprises ~2,600 m of turbiditic sandstone (of mainly silicic volcanic provenance) and shale, with rare concordant volcanic horizons. Stratigraphy is broadly similar to the well-described Trough sequences farther south (Packham 1968, 1969; Glen & Watkins 1994; Pogson & Wyborn 1994). The Cookman Formation consists of up to 400 m of slate and thinly interbedded quartz sandstone and vitric tuffs. The unit thins steadily towards the north of the map sheet area; north of GR 741250 6369850 the Cookman Formation is absent and the overlying Turondale Formation rests directly on the Late Silurian Pambong Formation. The Turondale Formation contains up to 900 m of thick-beded volcanoclastic and lithic turbidites of fine sand to pebble conglomerate, along with tuffaceous mudstone and concordant rhyolitic porphyries (probably thick sills). A U/Pb SHRIMP age of 417.3 ± 3.4 Ma was obtained from these porphyries at the Turon River in the Bathurst map sheet area to the south (Fanning in prep). Where zircons from sandstones in the upper part of the unit yielded a SHRIMP age of 409.5 ± 4.2 Ma (Jagodzinski & Black in prep), these ages suggest that the porphyry sills must have been approximately synsedimentary, given the overlap in the error bars.

The Turondale Formation passes conformably into the Waterbeach Formation. This ~600 m thick unit consists of 120 m of shale to siltstone at the base passing up into a thick sequence of shale interbedded with lithic-rich 30 cm to 3 m thick turbiditic sandstone beds. The Dunmoogin Formation, a unit mapped widely in the northern Hill End Trough by Offenberg et al. (1971), was shown by the recent mapping to be a junior synonym of the Waterbeach Formation and this name has been suppressed. The Guroba Formation — a unit mapped extensively in the northern Hill End Trough by Offenberg et al (1971) and defined to the north of Burrendong Dam — was found to occur extensively in the Mudgee map sheet area north of GR 747100 6355510 as a distinctive sandy facies between the Waterbeach Formation and the Merriens Formation; the unit has also been mapped widely in the Euchareena 1:100,000 map sheet area to the immediate west. The Guroba Formation consists of massive to thin-beded turbiditic sandstone of volcanic and basement lithic provenance, and mudstone.

The Merriens Formation (formerly the Merriens Tuff) is composed of thick-beded volcanioclastic sandstone, shale and concordant silicic porphyries, interpreted by Cas (1977, 1978) and Cas and Jones (1979) as cold, subaqueous mass-flows, background hemipelagic sedimentation and subaqueous lavas respectively. Several recent U/Pb SHRIMP dates place the Merriens Formation between 408 Ma and 405 Ma (Jagodzinski & Black in prep). The similarity of these ages to that of the Riversdale Volcanics of the Capertee High to the east, together with palaeocurrent and petrographic data of Cas (1977), suggest that thick dacitic ignimbrites of the Riversdale Volcanics and its lateral equivalent the Huntingdale Volcanics in the Capertee Valley provided the source for the very thick-beded volcanioclastic turbidites in the upper Merriens Formation.

The ungrouped Cunningham Formation is the youngest unit of the Hill End Trough and consists of over 3,000 m of slate (predominant) and thin-bedded volcanioclastic sandstone, with horizons of pebble-granule conglomerate. It occurs extensively in tightly folded sequences in the south-west corner of the Mudgee map sheet area. The unit has yielded a poorly preserved shelly fauna near its base (Wright 1994) which may correlate with the late Pragian Taylors Hill Formation fauna of the Queens Pinch Group. The Limekilns Formation, believed to correlate with the upper Cunningham Formation (Packham 1968), extends up into the late Emsian serotonus Zone (Mawson & Talent 1992). These data suggest deposition of the Cunningham Formation extended from about the late Pragian until at least the late Emsian.

Late Devonian

Siliciclastic rocks occupy the core of the Pine Ridge Syncline (Wright 1966), extending south-east from Mudgee to the west of Rylstone, where they occur in complexly folded and faulted sequences which overlie the Kandos Group with a low angle discordance (Powell & Edgewcombe 1978, Pemberton et al. 1994).

The Mount Knowles Group (new name) in the Mudgee district is at least 1,700 m thick and consists of several formations. To avoid confusion, it is recommended that the name Mount Knowles Member (Pemberton et al 1994) be abandoned, as this unrelated unit is now considered to be part of the Carwell Creek Formation.

The basal Buckaroo Conglomerate, a unit composed of quartz-rich purple conglomerate, quartz sandstone and red mudstone representing braided stream and overbank deposits (Wright 1966, Killick 1987), is conformably overlain by quartz and silicic porphyry of the Bumberra Formation, deposited in an open, tide- and storm wave-influenced, shallow marine setting. The muddy, storm-influenced shelf deposits of the Lawsons Creek Shale conformably overlie the Bumberra Formation and are, in turn, overlain by the regressive, richly fossiliferous quartz sandstone of the Derale Sandstone, deposited in a storm- and wave-influenced shallow marine environment. These units were previously placed in the Lambie Group (Connolly 1969, Killick 1987), however the sequence is very different to the Lambie Group of the Mount Horrible and Mount Dulabree Synclines farther south and is therefore placed in a new group.

In the south of the Mudgee map sheet area, thick-beded richly fossiliferous quartz sandstone of the Gibbons Creek Sandstone represents the most northerly occurrence of the basal unit of the Mount Dulabree Syncline sequence. Brachiopod faunas (Wright & Yan in prep) suggest correlation with the Bumberra Formation in the Mount Knowles Group.

Carboniferous

Discordant, unfoliated I-type, biotite granite occurs at Aarons Pass, Havilah, the Camboon area (east of Tongbong...
the boundaries of these intrusions and as (figure 3). Of these, only the Aarons Pass Granite has been isotopically dated as Middle Carboniferous (328 Ma; Pemberton 1990). Mapping has refined the boundaries of these intrusions and the aeromagnetic survey results suggest an associated network of largely exposed dykes, some of which may be up to 300 m wide. One of these is well exposed in the Pyramul area, adjacent to the Crudine road. The magnetic data also indicate that the Aarons Pass Granite extends considerably to the west of its exposed margin in shallow subsurface. This is confirmed to some extent by the intersection of granite in a drillhole at Mount Pleasant, 3 km west of the mapped boundary. The Havilah and Botobolar Granites, in particular, are in the process of being unroofed; both display numerous roof pendants and areas of greisen, giving rise locally to very complex outcrop patterns. The Aarons Pass and Pyangle Pass Granites have unusual radiometric signatures — zones relatively enriched and depleted in thorium are separated by linear contacts, creating a part-yellow, part-red radiometric colour pattern. No difference in granite petrography or geochemistry has been noted between the zones and these radiometric anomalies are unexplained at present.

**Permian—Mesozoic**

Permian and Triassic strata of the western Sydney Basin occur mainly in the east of the map sheet area, with a thin veneer (usually less than 10 m thick) present as outcrops and hillcappings above a partially exhumed palaeoenplain which rises steadily from less than 600 m above sea level (asl) in the east to over 900 m asl in the west of the map sheet area. The Rylstone Volcanics occur discontinuously at the base of the sequence and are restricted to the east of the map sheet area. The dominant lithologies are dactitic to rhyolitic ignimbrite, ashfall deposits, flow-banded rhyolitic lava, epiclastics and basement-derived conglomerate. The unit has been dated by Rb/Sr at 292 Ma (Shaw et al 1989), placing it in the Early Permian (Roberts et al 1996).

The Shoalhaven Group unconformably overlies the Rylstone Volcanics and thins towards the north of the map sheet area. It comprises a mainly marine sequence of basal conglomerate and sandstone (Snapper Point Formation) overlain by siltstone of the Berry Formation. The Late Permian Illawarra Coal Measures conformably overlie the Shoalhaven Group and consist of a basal alluvial fan conglomerate (Marrangaroo Conglomerate) overlain by shale, sandstone, conglomerate and coal seams deposited in a fluviatile to deltaic environment (Bembrick 1983). The coal measures thin steadily towards the north of the map sheet area. The Narrabeen Group conformably overlies the Illawarra Coal Measures and is composed mainly of thick units of quartz and lithic sandstone, with shale, which form prominent cliff lines along the eastern edge of the Mudgee map sheet area.

Low-lying, unnamed deposits (Pu) of Permian sandstone, siltstone and conglomerate occur in valleys to the north and west of Mudgee (figure 3). Varves and other glacial features have been noted in these deposits just north of Mudgee by Dulhunty and Packham (1962). At Macdonalds Creek, McMinn (1983) obtained polymorromorphs from these sediments which suggest correlation with Permian Stage 2. A *Gangamopteris* flora has also been documented by Exon (1962). Dulhunty and Packham (1962) and during recent mapping. Two samples from this unit were submitted for palynomorph extraction but both were barren. Mesozoic alkaline intrusives occur widely in the east of the map sheet area, in most places intruding the Illawarra Coal Measures (Day 1961). Only three of these have been dated isotopically: Tongbong Mountain is an exhumed sill of coarse olivine basalt or teshenite dated at 185 ± 8 Ma (Early Jurassic) by Dulhunty (1976). A possible flow of olivine trachyte at Jimmy Jimmy Mountain (13 km north-west of Lue) was dated at 216 ± 8 Ma (Late Triassic) by Dulhunty (1976), and a laccolith of aegerine augite phonolite at Pinnacle Mountain (10 km north-west of Lue) was dated at 178 ± 7 Ma (Middle Jurassic) (Dulhunty 1976).

**Tertiary**

Tertiary olivine basalt flows occur widely in the Mudgee map sheet area (figure 3). Most were extruded onto fairly planar erosional surfaces; however, diatremes have been described at Mount Fitzgerald, east of Ilford (Carne 1904), and just south-east of Kandos (Dove & Lee 1994), and are common to the east of Rylstone. Few of the basalt flows have been dated isotopically but Dulhunty (1971) carried out five K/Ar date determinations on basalt from the map sheet area and found two distinct levels and ages of extrusion:

(a) high-level Upper Eocene—Lower Oligocene flows at Mount Carcalong, Mount Vernon, Boiga Mountain, Mount Bocoble and Cherry Tree Hill forming cappings to hills mainly in the south of the Mudgee map sheet area.

(b) low-level middle Miocene flows filling deep lead valleys around Mudgee and Gulgong.

**Quaternary**

Large areas of alluvial sand, silt and gravel (Qa) occur around the Cudgegong River and its tributaries to the east, south and north of Mudgee and as elevated gravels to the south of Mudgee (Cza) (figure 3). Water bore drilling indicates these deposits are up to 50 m thick. Other significant areas of alluvium occur in the Lawson Creek valley around Havilah.

**STRUCTURE**

The sheet area spans three main structural domains which correspond to the structural subdivisions outlined above. The Sydney Basin strata in the east are largely undeformed, although local dips of up to 12° to the east have been reported and the Early Permian Rylstone volcanics has dips in airfall tuffs — interpreted as mainly depositional — of up to 30° near the edge of major palaeovalleys (Langworthy 1986). Strata of the Hill End Zone in the west of the map sheet area are characterised by north-west to north-east trending folds and few faults, whereas the Capertee Zone contains north-west to north-east trending structures and numerous major reverse and normal faults. The latter two areas are separated by the Wiagdon Thrust and Mudgee Thrust to the south and north of the Aarons Pass Granite respectively.

Evidence for significant pre-cleavage deformation is meagre and refold patterns observed in the Ordovician strata to the south (eg Palmers Oakey area, Ferguson 1979; Sofala Volcanics, Powell et al 1978) are not found in the...
Mudgee map sheet area. Steeply plunging lineations (defined by bedding and cleavage intersections) occur in the Ordovician Adaminaby Group in the Bara Creek area (Fergusson & Colquhoun 1996) and are the main evidence for an earlier (D₁) folding event. In addition, unconformities occur between the Late Ordovician–Early Silurian Coomber Formation and the Late Silurian Dungeree Volcanics west of Rylstone, and between the Sofala Volcanics and Willow Glen Formation (Pemberton 1989). However, a lack of structural data close to the contact makes it difficult to gauge the degree of angular discordance, although available data suggest it is minimal. A tight east-west trending anticline in the Sofala Volcanics and Willow Glen Formation occurs in the Apple Tree Flat area 12 km south-south-east of Mudgee (figure 3); this fold is possibly attributable to D₁. D₂ is therefore poorly time-constrained but appears to be post Late Silurian–pre Early Carboniferous and lacks an associated cleavage. A similar weak D₁ has been reported from the Orange 1:100 000 map sheet area to the south-west (Glen & Watkins 1994).

Unconformities are also present between the Late Silurian and Early Devonian sequences to the west of Rylstone and in the Queens Pinch area (figure 2). However, the degree of angular discordance is low (<5°) and this break may be best attributed to erosion due to sea-level fall or very gentle uplift and tilting. An unconformity has been widely reported between the Early and Late Devonian sequences. Powell and Edgecombe (1978) found a low angle discordance of 5 to 25° between the Early and Late Devonian sequences in a structural study of the Pine Ridge Syncline. They concluded that erosion, uplift and broad tilting had occurred during the Middle Devonian.

Measurements during recent mapping of the angular discordance farther south in the Kandos to Lue area are within the range given by Powell and Edgecombe (1978) — normally less than 10° — and are in accord with their conclusion that the Middle Devonian Tabberabberan Orogeny had only mild effects in this part of the Lachlan Fold Belt (figure 2).

The major Early Carboniferous D₃ deformation (Powell & Edgecombe 1978, Fergusson & Colquhoun 1996) was responsible for the majority of folding, thrusting and cleavage formation in the Mudgee map sheet area. The Hill End Trough strata are characterised by tight, upright to steeply inclined F₁ folds with long planar limbs and angular hinges, normally with well developed axial plane cleavage (S₁). F₂ folds typically plunge gently to the south, although several instances of plunge reversal over several kilometres were noted.

D₂ structures in the Capertee Zone are considerably more varied and complex. The major bounding structures, the Wiagdon and Mudgee Thrusts, are mainly west-dipping thrust faults with Hill End Zone strata thrust eastwards over Capertee Zone rocks. F₂ folding changes considerably throughout the Capertee Zone with interlimb angles from open to tight. Map scale structures such as the Pine Ridge Syncline and numerous folds in the Early Devonian sequence to the west of Rylstone are commonly fairly open and north-west-south-east trending with broad rounded hinge areas. Tight map scale folds occur in the Adaminaby Group and the Coomber Formation in the Bara-Botobolar area east of Mudgee, and an overturned anticline with S₂ in the core occurs in the Windamere Dam area. Fold plunges are usually gentle (<20°) and highly variable in direction with double plunging folds common at map scale (figure 3).

Cleavage (S₂) is commonly well developed and axial-planar to the major folds; it ranges from a planar slaty cleavage in the finer grained lithologies to anastomosing closely spaced cleavage in coarser sandstones and volcanics. D₃ was responsible for major faulting in the Capertee Zone, creating reverse faults and possibly reactivating older normal faults. Thrust faults commonly occur parallel to strike, eg the Riversdale Faults (which have wedged a block of Late Devonian strata into the Early Devonian Kandos Group), the east bounding fault of the Queens Pinch Group, the Mount Bara Thrust (where the Adaminaby Group has been thrust over the Coomber Formation) and the Walkers Lane Fault (figure 3). Other faults cut across the prevailing strike and fold axes at moderate to high angles. These are best displayed in a series of steeply dipping, meridionally trending en echelon faults (Havliah Fault, Erang Fault, Cudgegong Fault and others) which divide the Pine Ridge Syncline south of Havilah into a series of openly folded blocks. Some of these faults may be older normal faults reactivated during Carboniferous compression. Regional metamorphic grade is in most places greenschist to subgreenschist facies and has not been observed to vary systematically throughout the map sheet area.

In the Lue–Havliah–Bara area, S₂ has been folded into an F₃ antiform to the east and south of the Havilah Granite, suggesting significant D₃ deformation (Fergusson & Colquhoun 1996). D₃ is also probably responsible for folding of the Mount Bara Thrust, the trend of which changes by 90° before being overlain by the Rylstone Volcanics (figure 3). Mesoscopic kink folding of S₂ is also widely developed in slaty lithologies of the Coomber Formation and Dungeree Volcanics in this area — and at Queens Pinch (Warratra Mudstone) and in the Cudgegong district (Yellowmans Creek Formation) — and is probably attributable to D₃ which field relationships indicate is also probably Early Carboniferous in age.

One of the striking features of the Mudgee map sheet area is the marked change in the regional structural grain which occurs across east-north-east trending photolinesations in the Bocble and Havilah areas. Across the Pyramul Lineament (Scheibner & Stevens 1974) near Mount Bocble, the structural grain changes gradually from north-east trending to north-west trending, with the effects decreasing to the east and west of Mount Bocble. Termined the Bocble Orocline by Scheibner and Stevens (1974) and the Pyramul Megakink by Powell et al (1985), the change in orientation appears to be too gradual for a megakink and lacks an associated crenulation cleavage. A more gradual change in the structural grain was indicated in the radiometric data in the Havilah area across the Havilah Lineament (new name) which appears to extend well into the Hill End Zone and causes kinklike folding of the Merrins Formation evident in the radiometric data. Significantly, these lineaments both pass through granites (the Aarons Pass and Havilah Granites respectively), suggesting that extensional forces during folding/kinking may have created a preferential conduit for rising granitic magma. These features affect bedding, cleavage and fold axes, suggesting a significant north–south compression event postdating D₃ folding and thrusting (possibly reactivating many D₂ faults), but untimed with respect to D₃ structures of the Lue–Havliah area. Similarly, Kosaka (1994) has recorded brittle deformation in the Bathurst Batholith to the south. He attributed this...
information to a north–south shortening event which postdated the regional ($D_2$) deformation. Alternatively, these features may represent buttressing or hinging around some anomaly or weakness in the deeper basement during the latter parts of $D_2$ deformation.

**MINERALISATION**

The Mudgee map sheet area is characterised by a wide variety of mineralisation types and styles which to date have yielded only comparatively small tonnages.

The most important vein gold mineralisation occurs in the Windeyer area and is hosted by the turbiditic sediments of the Crudine Group. The mineralisation is similar to that at Hill End to the south and occurs in association with quartz veins and minor stockworks in the Merrions and Waterbeach Formations along the hinge zone of a large, gently south-east plunging symmetrical anticline. Gold is disseminated in quartz reefs, usually with minor pyrite, arsenopyrite and galena.

Numerous vein gold occurrences exist in dacite and rhyolite lava of the Late Silurian Windamere Volcanics to the west of Apple Tree Flat (figure 3). These were mapped as the Ordovician Sofala Volcanics by Offenberg et al (1971) but recognised as Silurian by Pemberton (1989, 1990). Gold occurs with sulphides (pyrite, arsenopyrite, minor galena and chalcopyrite) and is commonly associated with chlorite, silica and albite alteration. Gold distribution appears to be controlled by a north–south trending fault system in the Windamere Volcanics, and this unit should now be regarded as an important exploration target.

A significant epithermal Pb–Zn–Ag vein deposit was discovered by Golden Shamrock Mines and CRA Exploration Pty Limited at Bowdens Gift, 4 km north-east of Lue (figure 3). Mineralisation consists of vein and disseminated sphalerite, galena and silver-bearing sulphides and is hosted by hydrothermally brecciated tuff and massive crystal tuff of the Early Permian Rylstone Volcanics. In late 1995, the deposit was estimated to contain an indicated and inferred resource of 18.8 million tonnes at 99 g/t Ag. This discovery has highlighted the potential of the Rylstone Volcanics as a target for precious and base metal mineralisation. Similar silver/base metal mineralisation occurs at the Coomber Prospect, 5 km north-west of Rylstone, which is also hosted by the Rylstone Volcanics.

Porphyry Cu–Au deposits associated with Ordovician volcanics and intrusives — an important exploration target in the Lachlan Fold Belt — are largely absent from the Mudgee map sheet area, reflecting the paucity of primary volcanics and intrusives of this age. The Cheshire copper mine and the Milfor prospect in the Sofala Volcanics near Aarons Pass (figure 3) consist of disseminated and layered chalcopyrite and minor gold associated with quartz in basaltic volcanics and appear to be the only known occurrences of this type in the Mudgee map sheet area.

Molybdenum and tungsten mineralisation is associated with Carboniferous granites in the Mudgee map sheet area. Most prominent among these occurrences is the Mount Pleasant prospect to the west of Aarons Pass. Mineralisation is hosted by silicic to intermediate volcanics and sediments of the Late Silurian Chesleigh Formation and occurs within a fracture-controlled network of veins over an area of 3 to 4 km². Other smaller molybdenum prospects occur adjacent to the Botobolar and Havilah Granites.

Exploration in the 1970s located subeconomic massive sulphide type mineralisation in rugged country to the south of Havilah hosted by dacites and slates of the Late Silurian Dungeree Volcanics (Dead Horse prospect, figure 3). Mineralisation appears to be concentrated along the unconformity between the Early Devonian Carwell Creek Formation and the Dungeree Volcanics and is expressed at the surface as a series of gossans which assayed high in Pb, Zn and Cu. Four kilometres south of Lue, also hosted by dacites of the Dungeree Volcanics, is a series of vein type barite occurrences in irregular anastomosing quartz veins, many following cleavage planes; two of these have been mined.

The Rockwell antimony mine to the north-west of Rylstone is hosted by andesites and basalt’s of the Late Ordovician Coomber Formation and is the only such occurrence in the Mudgee map sheet area (figure 3). Mineralisation at Rockwell comprises stibnite and minor sulphides and occurs in quartz veins which are localised in a joint set. The mineralisation may be related to the extrusion nearby of a large dome of subaerial rhyolite in the Early Permian Rylstone Volcanics. A cinnabar occurrence is located near the Cudgegong River, 11 km west of Rylstone, just above the top water level of Lake Windamere. This prospect apparently occurs in alluvial gravels and attempts by various companies to locate the primary source have failed.

A large number of residual ironstone occurrences in the region are localised close to major faults and unconformities (figure 3). The Lue ironstone mine, 1 km south of Lue, is located close to a major fault between the Adaminaby Group and the Dungeree Volcanics. The Lucknow and Glendale prospects in the Kandos district occur on the Silurian–Devonian unconformity and in a fault separating the Early and Late Devonian sequences respectively. The major Ironstone Creek occurrences lie close to an unconformity between the Late Silurian Windamere Volcanics and the Late Devonian Buckaro Conglomerate and adjacent to the Cudgegong Fault. The Eurundry ironstone occurrences are located along an unconformity between the Carwell Creek Formation and the Dungeree Volcanics.

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Responsibility for mapping and compiling the Mudgee geological map sheet was as follows:

- **Geological Survey of New South Wales**: Simone Meakin (Mudgee and Botobolar 1:25 000 map sheets), Jan Krynen (Broombee 1:25 000 map sheet), Gary Colquhoun (Lue, Kandos, Illford, and NE corner of Windeyer 1:25 000 map sheets)

- **Australian Geological Survey Organisation (AGSO)**: Tony Henderson and Elizabeth Jagodzinski (Tunnabidgee and Windeyer (remainder) 1:25 000 map sheets).

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Figure 2 and 3 were drawn for publication by Cheryl Hormann and Li Li respectively.

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Wright A.J. & Yan J-X. in prep. Late Devonian productoid and chonetoid brachiopods from New South Wales.
This publication, written by Erwin Scheibner and edited by Helena Basden, defines a structural framework of NSW as a reference for compiling earth science data of the State.

Modern geophysical data have enabled the definition of 32 zones (including one subzone) and 169 blocks; definitions of the structural units are given and contentious issues are discussed. Each structural block is described in Appendix 1.

The status of terrane analysis in the Tasman Fold Belt System of eastern Australia is also assessed, and sutures and terrane boundaries defined.

The text of volume 1 is divided into three parts plus the appendix. A map of the structural framework of New South Wales at 1:1 500 000 scale is included in a pocket at the back of the book.

PART 1
Introduction
• Introductory notes
• History

PART 2
Structural framework of NSW
• Crustal and lithospheric structure of New South Wales
• Australian Proterozoic Craton and paratectonic fold belts
• Tasman Fold Belt System
• Late Carboniferous to Holocene cratonic cover

PART 3
The status of terrane analysis in the Tasman Fold Belt System
• Sutures in the Tasman Fold Belt System
• Terranes in the Tasman Fold Belt System

APPENDIX 1.
Summary of structural units in NSW.

This work is a quality publication of xvi + 295 pages. It is presented as a hard cover case bound book with numerous full colour text figures and photographs. Priced at $90.00

Volume 2: Geological Evolution, currently in preparation, will contain descriptions of selected time slices which represent significant episodes in the tectonic development of New South Wales, and each tectonic stage will be illustrated by palaeogeographic maps. Again, definitions will be given and contentious issues discussed.

It is expected that Volume 2 will be available in the third quarter of 1997.
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NEXT ISSUE
Mineral Deposits of the Glen Innes 1:100 000 map sheet area
R.E. Brown
Articles below removed for copyright reasons, please refer to the citation:

Supporting Paper 9
CRATONIC TO ARC PROVENANCE SWITCHING: AN EXAMPLE FROM THE EARLY PALEOZOIC OF SOUTHEASTERN AUSTRALIA

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ABSTRACT: Early Paleozoic strata east of Mudgee, northeastern Lachlan Fold Belt, Australia, consists of an Early Ordovician sequence (Adaminaby Group) of quartz-rich turbidites and chert, and the conformably overlying Late Ordovician-?Early Silurian Coomber Formation, a mafic volcaniclastic sequence of lithic sandstone and mudstone with mafic volcanic and intrusive rocks. Analysis of detrital modes from the Adaminaby Group indicate a uniform quartzose petrofacies (Q$_{91}$F$_{2}$L$_{7}$) composed of deep-water quartz and sublithic sandstones considered to have undergone long-distance transport by turbidity currents from a source in the Ross-Delamerian mountain chain of South Australia and Antarctica. By contrast, the lithic sandstones of the Coomber Formation represent a volcanolithic petrofacies (Q$^{i}$F$_{4}$L$_{46}$) of mainly intermediate volcanic detritus with a tendency to become more silicic up-section. These sandstones are considered to be derived from a calc-alkaline andesitic-basaltic volcanic center in the eastern Lachlan Fold Belt. The change from a quartz-rich petrofacies to a volcanolithic petrofacies was sharp with no evidence of mixing. This change primarily reflects a rapid outpouring of mafic volcanic material in the Late Ordovician, together with an approximately concurrent eustatic rise in sea-level (thereby cutting off the supply of clastic detritus to the quartz turbidite fan), and a lowering of relief in the source terrain of the quartz sandstones.

INTRODUCTION

The petrography of modern sands and ancient sandstones and its relationship to tectonic setting has been the subject of intense research in recent years and a large body of literature now exists (e.g., Dickinson and Suczek 1979; Dickinson and Valloni 1980; Valloni and Maynard 1981; Dickinson et al. 1983; Dickinson 1985; Marsaglia and Ingersoll 1992) graphically presented their data in a series of ternary diagrams outlining characteristic detrital modes for various tectonic provenances such as magmatic arc, continental block and recycled orogen.

Ordovician strata of the Lachlan Fold Belt of southeastern Australia comprise two petrographically distinct units: a lower unit of monotonous turbiditic quartz-rich sandstones and chert, and an upper unit of mainly graptolitic black shale, except in the northeast of the fold belt where it consists of a mafic volcaniclastic succession including turbiditic sandstones (Fergusson and Coney 1992; VandenBerg and Stewart 1992; Wybom 1992a, 1992b). However, few detailed petrographic analyses or provenance studies of the sandstones of these units have been published (Crook 1955; Fell 1984; Fenton et al. 1982; Pound et al. 1994; Gray and Webb 1995), and their characteristics are not well known. Sandstone petrography provides a valuable tool for interpreting the paleogeography and tectonic setting of the Ordovician rocks of southeastern Australia, both of which are not well understood. Furthermore, the contact between the lower quartz-rich sandstone unit and upper lithic sandstone unit in the northeast of the fold belt is apparently sharp and conformable, with only very rare evidence of interbedding of the two facies; the mechanisms for this rapid switching of provenance have not been addressed by previous authors.

This paper discusses the results of petrographic analyses carried out on quartz-rich and lithic-rich Ordovician sandstones to the east and southeast of Mudgee, central New South Wales, Australia. The aim is to document the sandstone petrography, discuss the provenance of the contrasting petrofacies, suggest possible reasons for the rapid switching of provenance, and discuss the implications for the Ordovician paleogeography and tectonic setting of the Lachlan Fold Belt.
REGIONAL GEOLOGY AND TECTONIC SETTING

The Lachlan Fold Belt of southeastern Australia is a Cambrian to Carboniferous orogenic belt which dominates the basement outcrop of southeastern Australia (Fig. 1). Ordovician to Early Silurian strata of the eastern Lachlan Fold Belt are characterised by two fundamentally different lithofacies associations (Fergusson and Coney 1992).

The first is a monotonous Early Ordovician succession of quartz-rich turbidites and chert at least several kilometers thick which occurs widely throughout much of Victoria and southern to central New South Wales, and is known as the Adaminaby Group (VandenBerg and Stewart 1992). The Adaminaby Group occurs extensively in the Capertee Zone at the exposed northeastern margin of the fold belt (Pogson and Wybom 1994) and has been recently documented farther north, to the east of Mudgee (Fig. 1). These latter exposures represent the oldest rocks in the region and the most northeasterly outcrop of the quartz turbidite succession in the Lachlan Fold Belt (Fergusson and Colquhoun 1996). The quartz turbidite succession has been interpreted to represent the deposits of a huge submarine fan of comparable dimensions to the present Bengal-Nicobar submarine fan complex (Cas et al. 1980; Powell 1984; Fergusson and Coney 1992) which we interpret as a foreland sedimentary prism infilling an oceanic realm. Strontium isotopic ratios of the quartz sandstone suggest probable derivation from Cambrian sedimentary rocks of the Kanmantoo Group in South Australia or Cambrian igneous rocks in western Tasmania (Gray and Webb 1995).

The second lithofacies association consists of a Late Ordovician sequence of mafic volcanic rocks, volcanioclastic strata, and limestone, which outcrop in several north-south trending belts in the northeastern Lachlan Fold Belt (Packham 1969, 1987; Powell 1984; Wyborn 1992a, 1992b). This association has recently been documented to the east and southeast of Mudgee (Pemberton et al. 1994; Fergusson and Colquhoun 1996), where it is known as the Coomber Formation. Data from these petrographic analyses were compiled to discriminate tectonic provenance, using the tectonic provenance diagrams of Dickinson and Colquhoun (1996) interpreted the Coomber Formation as consisting of mass-flow and suspension deposited sediments lacking any features indicative of shallow marine environments; the presence of radiolarian cherts suggest a setting seaward of the paleoshelf-slope break, probably in relatively deep water (Jones and Murcchey 1986). Sedimentation was dominated by variable, but often prolonged, periods of hemipelagic and pelagic settling from suspension which were interrupted episodically by volcanioclastic turbidites - of highly variable volume and sediment concentration - and rare debris flows of conglomerate and limestone blocks. Syn-sedimentary basalt sills and flows were emplaced into/onto the unconsolidated sedimentary pile accompanied by subvolcanic mafic to intermediate intrusions. Between Mudgee and Sofala (Fig. 1B) occur the Sofala Volcanics, a Late Ordovician succession dominated by primary mafic volcanic rocks with abundant coarse to fine mafic to intermediate volcanioclastic strata that are lateral and coarse-grained equivalents of the Coomber Formation.

Contacts between the two lithofacies associations are commonly faulted in the Mudgee area; however, where conformable, the contact is sharp with no sign of interstratification of detritus. Elsewhere in the northeast Lachlan Fold Belt, such as north of Palmers Oakey (Fig. 1B), contacts are similarly faulted or conformable and sharp; however, rare instances of interstratification have been recorded southeast of Palmers Oakey (Fergusson 1979; see also Powell 1984, p. 294) where lithic sandstones and quartz sandstones are interbedded in intervals up to hundreds of meters thick.

The majority of paleogeographic interpretations depict the volcanic/volcanioclastic sequence as a chain of island volcanoes that developed adjacent to, and finally overwhelmed, the quartz turbidite fan (Webby 1976; Cas et al. 1980; Powell 1984). This paleogeography has been widely interpreted as representing a single island arc related to west-dipping subduction (Scheibner 1974; Cas et al. 1980; Powell 1984; Fergusson and Colquhoun 1996), although Wyborn (1992a) considered it to represent discrete centers of mantle-derived volcanism that were unrelated to subduction.

METHODS

Thin sections were prepared for 98 outcrop samples (Fig. 1). Representative samples (9 from the Adaminaby Group and 12 from the Coomber Formation) were chosen for modal analysis using the Gazzadi Dickinson method with 500 points counted per thin section. Framework grain parameters are shown in Table 1 and modal analyses in Table 2. Framework quartz grains were classified using the criteria of Basu et al. (1975). Data from these petrographic analyses were compiled to discriminate tectonic provenance, using the tectonic provenance diagrams of Dickinson and Suczek (1979) and Dickinson et al. (1983). No significant grain size effect was determined for any of the framework parameters. Representative sandstones from the Coomber Formation were analysed for pyroxene and feldspar compositions using a Cameca Camebax electron microprobe at Macquarie University.

SANDSTONE PETROGRAPHY

Adaminaby Group

Sandstones of the Adaminaby Group are poorly to moderately sorted with subrounded to rounded grains. Most samples have a grain size in the very
fine to fine sand (approximately 3 phi) range and are framework supported (matrix of 2% to 23%). The sandstones are mainly quartzarenite and sublitharenite with minor subarkose and litharenite compositions (after Folk, 1974) (Figs 2, 3A, Table 2). Quartz is dominant (61-93%, mean 80%) and shows strong to strongly undulose extinction. Polycrystalline quartz commonly shows stretched metamorphic textures, strong undulose extinction and a polygonal fabric, typical of a metamorphic source terrain. Feldspar is strongly undulose extinction and a polygonal fabric, multiply twinned and untwinned albite, the latter checked by microprobe. Lithic fragments (0.6-20.6%, mean 6%) are mainly quartz-mica schist with minor micaceous crenulated phyllite, fine-grained sedimentary grains and rare quartz-rich plutonic fragments. Matrix consists solely of sericite. Rare accessory tourmaline, zircon and opaques are also present.

**Coomber Formation**

Sandstones generally consist of angular to subrounded grains of very fine to coarse sand (up to 4 mm) with poor to very poor sorting and low sphericity values. Sandstones are matrix- and framework-supported, texturally and mineralogically immature, and are classified as lithic arkose, feldspathic litharenite and litharenite (after Folk 1974) (Figs 2, 3B, Table 2). Intermediate and mafic volcanic rock fragments are dominant (31-83.4%, mean 51%) and have microlitic and porphyritic textures with subhedral phenocrysts of plagioclase (mainly albite) and highly altered mafic minerals (?Clinopyroxene) in a felted to trachytic groundmass of plagioclase microlites, sericite and opaques. Plagioclase grains are less abundant (3.2-38.6%, mean 27%) than volcanic fragments and are typically more sodic than Ab90 (determined from microprobe). Other framework grains consist of minor mudstone-shale fragments (1-2%) and rare quartz (< 2%).

The sandstones are subdivided into two distinct suites, based on the presence or absence of clinopyroxene. Where present, clinopyroxene (augite-diopside, compositions determined from microprobe) is a major component (up to 18%) and is subrounded to subhedral (Fig. 3B). The matrix is difficult to distinguish from framework grains and is composed of a mixture of fine-grained volcanic material, chlorite and opaque grains. Calcite is occasionally present as cement. The Coomber Formation sandstones are extensively altered with assemblages containing sericite, chlorite, prehnite, epidote, actinolite and calcite. Assemblages indicate sub-greenschist to greenschist metamorphic facies conditions. Biotite is also present in zones of contact metamorphism adjacent to granites.

**PROVENANCE**

Sandstones from the Early Paleozoic strata of the Mudgee district contain two distinct petrofacies (Figs 2, 3, Table 2) which display no signs of transition which might indicate mixing of the source terrains. Sandstones from the Adaminaby Group fall within the craton interior and recycled orogen field whereas those from the Coomber Formation straddle the transitional arc and undissected arc fields (Fig. 2).

Quartz-rich rocks of the Adaminaby Group contain detritus indicative of a low- to moderate-grade metamorphic source terrain (quartz-mica schist, phyllite) with rare silicic intrusions (quartz-rich, plutonic fragments). This is in agreement with previous studies of provenance of the Ordovician turbidites of the Lachlan Fold Belt, which have suggested that the unit was derived from a mountain chain in the Delamerian Fold Belt in South Australia (Fig. 1C; Cas et al. 1980; Gray and Webb 1995) and is also consistent with paleocurrent trends that indicate a westerly source (Fig. 4; Cas and Vandenberg 1988; Fergusson et al. 1989; Jones et al. 1993).

Sandstones of the Coomber Formation were derived from an intermediate to mafic volcanic terrain (Figs 4, 5). Volcanic rock fragments lack vitriclastic textures and are dominated by microlitic and porphyritic textures, suggesting derivation mainly from coherent volcanic rocks rather than pyroclastic rocks. Volcanic detritus is predominantly intermediate; however, the presence of a distinct suite of pyroxene-rich sandstones indicates local derivation from a more mafic source. The pyroxene-rich sandstones occur in the lower part of the succession, indicating that the source volcanic pile became more silicic with time, possibly in accord with eruption from a differentiating magma chamber. The immaturity of the sandstones together with the presence of mafic intrusions and mafic lava flows (Fergusson and Colquhoun 1996) indicates a fairly proximal position on the volcanic apron. The Sofala Volcanics to the southwest contain abundant breccias and primary volcanic rocks of basaltic to andesitic composition (Pemberton 1989) and probably are a more proximal succession derived from the same source as the Coomber Formation.

**DISCUSSION**

The modal data from the Adaminaby Group and the Coomber Formation reveal the existence of two discrete petrofacies in the Ordovician strata that lack any evidence of interstratification or mixing (Fig. 5). With the exception of the Palmers Oakey district, northeast of Bathurst (Crook 1955; Fergusson 1979), this is a characteristic feature of the Ordovician rocks of the northeastern Lachlan Fold Belt (Fig. 5). Any paleogeographic or tectonic model must therefore account for the rapid switching from craton-derived quartz-dominated detritus in the Early Ordovician to arc-derived mafic volcanioclastic detritus in the Late
Ordovician (Figs 4, 5). Several factors may have contributed to this change.

The first is a decrease in relief in the source area of the Adaminaby Group. Petrographic, isotopic (Gray and Webb 1995), zircon age (Williams et al. 1991) and paleocurrent (Cas and VandenBerg 1988; Fergusson et al. 1989) data are consistent with the Ordovician turbidite fan having been derived from the Delamerian Fold Belt in South Australia, western New South Wales and western Victoria (Figs 1C, 4). This source region was deformed, metamorphosed and uplifted during the Late Cambrian and Early Ordovician Delamerian Orogeny (Preiss 1990; Gibson and Nihill 1992; Mills 1992) and provided a long-lived mountain range during this interval. Relief was probably considerably reduced by the late Early Ordovician, and should be reflected by an upward-fining succession in the upper Adaminaby Group. In southeastern New South Wales, a transition has been recorded between the Sunlight Creek Formation of the Bendoc Group and the Warbisco Shale, where progressively thinner bedded quartz turbidites grade into mudrocks (Glen et al. 1990). Similarly, the transition to the southeast of Goulburn (Fig. 1B) is marked by an upward-fining sequence of thin, ripple-bedded sandstones interpreted as contoursites (Jones et al. 1993). However, a similar thin-bedded transition zone was not recorded at the Adaminaby Group- Coomber Formation contact. This could be due to a lack of exposure near the contact (these rock types tend to weather recessively) or may indicate that the volcaniclastic deposition commenced slightly before the inception of black shale deposition elsewhere.

Second, a rise in sea-level cutting off supply of quartzose detritus to the fan. Jones et al. (1993) attributed the deposition of the black shale, which dominates Late Ordovician strata in the southeastern Lachlan Fold Belt, to a eustatic sea-level rise. VandenBerg and Stewart (1992) regarded this transition from the Early Ordovician turbidites to the Late Ordovician black shales as very slow (lasting over perhaps two graptolite zones in the latest Darriwilian and earliest Gisbornian) and also favoured an extra-basinal mechanism such as global sea-level rise. Ross and Ross (1992) found in studies of six sequences in Europe and North America that a major transgressive event occurred at or near the Nemagraptus gracilis zone, the first zone of the Australian Gisbornian stage. Similarly, Struckmeyer and Brown (1990) recorded a major third order transgression close to the Darriwilian-Gisbornian boundary. It appears, therefore, that deposition of the black shale/volcaniclastic strata was synchronous with a major eustatic rise in sea-level, around the Early-Late Ordovician boundary.

Third, a major pulse of mafic volcanism replaced the supply of quartz-rich detritus. Volcanism in the Late Ordovician occurred over a wide area of the Lachlan Fold Belt in contrast to the more limited outpourings in the Early Ordovician which were localised in the Blayney-Wellington and Parkes regions (Figs 1B, 5B; Webby 1976; Wyborn 1992a). Furthermore, in these latter regions, the Early Ordovician mafic volcanism did not contribute significant detritus to the turbidite fan, as is reflected by the quartz-rich sandstones of the Adaminaby Group in surrounding areas (Figs 2, 5B). In contrast to the Late Ordovician volcanic succession, the Early Ordovician volcanic strata lack any evidence of shallow marine deposition, indicating that the volcanic centers never rose close to or above sea-level and consequently were not significant contributors of detritus to the adjacent basin. Furthermore, these volcanic centers did not significantly affect regional paleocurrent patterns in the Adaminaby Group since, in the Mudgee region, the fan was derived from the west-southwest (Fergusson 1979; Fergusson and Colquhoun 1996), which is close to a direct path from the localised volcanic centers (Fig. 4A).

Mafic volcanism in the Late Ordovician rapidly built numerous volcanic edifices, which were subsequently subjected to erosion by subaerial processes and wave action. Extensive volcanic aprons formed and rapidly overlapped and smothered any waning deposition on the quartz turbidite fan. Thick sequences of shallow platform carbonates were deposited during volcanically quiet periods adjacent to the volcanic edifices, whereas further offshore, deep water sedimentation prevailed (Webby 1976; Cas et al. 1980). The paucity of interbedding between the quartz-rich and volcanolithic petrofacies suggests a period of non-deposition on the quartz turbidite fan, due to sea-level rise and lowering of relief in the source terrain, prior to the major volcanic pulse. The widespread nature of this sharp contact between the two petrofacies suggests this was probably the case throughout most of the northeast Lachlan Fold Belt. Rare areas of interbedding, such as southeast of Palmers Oakey, were probably formed around volcanic centers that built to sea-level slightly earlier than elsewhere, and hence were contributing volcaniclastic sediment to the adjacent basin before deposition on the quartz turbidite fan ceased.

CONCLUSIONS

(1) Modal petrographic analysis of an area of Ordovician strata in the northeastern Lachlan Fold Belt, southeastern Australia, reveals two distinct petrofacies: a quartz-rich Early Ordovician sequence derived from distant cratonic areas to the west of the Lachlan Fold Belt, and a mafic volcaniclastic sequence derived from erosion of a more local volcanic center.

(2) The contact between the two petrofacies is conformable and sharp with no sign of interbedding or mixing of detritus, suggesting rapid switching between source terrains. This rapid switch was probably caused by a lowering of relief in the source
terrain of the Adaminaby Group and a rise in sea-level in the early Late Ordovician, which combined to terminate deposition on the quartz turbidite fan just prior to a major pulse of mafic volcanism in the Late Ordovician.

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Manuscript received ; revised

FIGURE CAPTIONS

Fig. 1.—A) Geology of the district east and southeast of Mudgee with locations of sandstone samples. B) Structural zones in the northeastern Lachlan Fold Belt and location of sample map A). Small circles indicate the extent of Permian-Mesozoic cover rocks. C) Location of the Lachlan Fold Belt and neighbouring fold belts in southeastern Australia. D) Location of C) in Australia.

Fig. 2.— QFL, QpLvmLsm, LmLvLs and, QmFLt plots of sandstones from the Adaminaby Group (Early Ordovician) and Coomber Formation (Late Ordovician-Early Silurian) from the Mudgee district. Sandstone classification after Folk (1974). Key to provenance fields on QFL and QmFLt plots: CI = craton interior, TC = transitional continental, BU =
basement uplift, RO = recycled orogen, DA = dissected arc, TA = transitional arc, UA = undissected arc, QR = quartzose recycled, M = mixed, TR = transitional recycled, LR = lithic recycled. Provenience fields after Dickinson and Suczek (1979) and Dickinson et al. (1983).

Fig. 3.—Photomicrographs showing the petrographic contrast between sandstones from A) the quartz-rich petrofacies (crossed polars, Adaminaby Group), and B) a pyroxene-volcanolithic sandstone from the volcanolithic petrofacies (crossed polars, Coomber Formation). Symbols: p = plagioclase, px = pyroxene, v = volcanic rock fragment. Scale bars = 0.10 mm.

Fig. 4.—Paleogeographic reconstructions (based on present-day exposure) of A) Early and B) Late Ordovician paleogeography of the Lachlan Fold Belt (modified from Powell, 1984, Fig. 189). Paleocurrent data in the Adaminaby Group reported in (1) Fergusson and Colquhoun (1996), flute casts; (2) Powell (1984, fig. 189, after Fergusson 1979), flute casts; (3) Fergusson et al. (1989), flute casts; and (4) Powell (1984), mainly Bouma C micro-crosslaminations with some flute casts.

Fig. 5.—Cross sectional models of A) Late and B) Early Ordovician strata in the Lachlan Fold Belt; note that the effects of Silurian and younger folding have been removed (updated and modified from Fergusson and Coney 1992).

TABLE CAPTIONS

Table 1.—Framework grain parameters (for Gazzi-Dickinson method)

Table 2.—Modal analyses (Gazzi-Dickinson point-counting method) and recalculated data (see Table 1 for component abbreviations) for the Adaminaby Group (AG) and Coomber Formation (CF)
Exposed Precambrian Delamerian basement

Terrestrial to shallow marine Deep marine strata

Mafic-intermediate volcanic and volcanioclastic rocks (including limestone)

Limestone

Fig. 4

Combined Legend

Exposed Precambrian
Terrestrial to shallow marine
Mafic-intermediate volcanic and volcanioclastic rocks (including limestone)
Limestone

Delamerian basement
Deep marine strata (sandstone, mudstone)

Fig. 5
Table 1. Framework grain parameters (for Gazzi-Dickinson method).

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(Recalculated data for Table 1 component abbreviations.)
EARLY DEVONIAN FOSSILS FROM THE CARWELL CREEK FORMATION, WINDAMERE DAM, MUDGEE DISTRICT, NEW SOUTH WALES, AUSTRALIA: TAXONOMY, AGE AND PALAEOECOLOGY

by A. J. WRIGHT1, B. D. E. CHATTERTON2 and G. P. COLQUHOUN1

1. School of Geosciences, University of Wollongong, Northfields Ave. Wollongong NSW 2522
2. Department of Geology, University of Alberta, Edmonton, Alberta T6G 2E3, Canada

ABSTRACT. Early Devonian macrofossil and microfossil species are described from limestones assigned to the Glendale Limestone Member of the Carwell Creek Formation, from 'Erang' near Mudgee, New South Wales, Australia. Silicified taxa include the trilobites Cyphaspis windamerensis sp. nov., Maurotarion erangensis sp. nov., Apocalympene sp. and Dudleyaspis sp. and the brachiopods Skendioides johnsoni sp. nov., Aulacella boucoti sp. nov., Resserella rhodesi sp. nov., and M. (Mesodouvillina) savagei sp. nov.; the tetracoral Sociophyllum wilsoni sp. nov. is also described.

The occurrence of the conodonts Pandorinellina steinhornensis miae (Bultynck), Ozarkodina pseudomiae (Mawson et al.), O. buchanensis (Philip) and O. excava (Branson and Mehl) indicates correlation of the fauna with the dehiscens Zone (late Pragian to early Emsian: medial Early Devonian). A summary of Early Devonian macrofossil biostratigraphy is presented.

The high proportion of new macrofossil taxa in this fauna emphasises the paucity of documentation of Australian faunas of this age from both limestone and clastic facies.

INTRODUCTION

EARLY Devonian limestones on 'Erang' property near Windamere Dam, 25 km southeast of Mudgee, New South Wales, Australia (Text-fig. 1) have yielded variably silicified and frequently deformed brachiopods, trilobites, conodonts and ostracodes, as well as rare corals.

Aspects of the stratigraphy of the region, which lies in the northeastern part of the Lachlan Fold Belt of eastern Australia, were discussed by Pemberton et al. (1994), and a geological map of the area is given in Text-fig. 2. Briefly, the highly deformed limestone, possibly 100 m thick, occurs as an inlier faulted on both sides against Late Devonian strata of the Bumberra Formation (and possibly Lawsons Creek Shale) on the southwest, and the Buckaroo Conglomerate on the northeast. The folded limestone beds are in contact with crinoidal sandstone typical of the Carwell Creek Formation and are, in places, highly cleaved; the sequence within the Early Devonian beds at this locality is uncertain. Highly altered silicic volcanics to the south of the Early Devonian inlier (Text-fig. 2) are interpreted as the Late Silurian Windamere Volcanics (Pemberton 1980b) unconformably underlying the Carwell Creek Formation.

The only other limestones known from the immediate vicinity of Windamere Dam were intersected during exploratory drilling for the construction of Windamere Dam (Text-fig. 1). These limestones documented by Pickett (1974, 1976) occur on the SW side of the main outcrop of the Buckaroo Conglomerate, immediately upstream (SW) of the dam (see Text-fig. 1). From borehole W37 (GR 597750 Lue 8832-1-S 25,000 sheet) Pickett (1974) recorded mainly conodont faunas from the intervals 42.4 m - 59.4 m and 85.3 m - 85.9 m. In particular, the lower level yielded Spathognathodus rem scheidensis? Ziegler as well as silicified ostracodes, and the conodont occurrence was used by Pickett to indicate provisionally a Gedinnian age. Pickett (1976) noted the occurrence of corals (indeterminate to species) in cores from borehole DDH 102 (GR 595752, Lue 8832-1-S 25,000 sheet) from several levels from 9.5 m - 13.4 m. On the basis of lithological similarities between the 'Erang' limestone and samples described by Pickett (1974, 1976), it is possible that these are all occurrences of the same limestone unit.

LOCAL LITHOSTRATIGRAPHY AND CORRELATION

The best understood development of the local Early Devonian sequence is developed in the Kandos Group (about 4000 m thick) in the Cudgegong-Carwell Creek district, about 12 km to the SSE of Windamere Dam, as documented by Pemberton (1989), Millsteed (1992), Cook (1990), Pemberton et al. (1994) and Colquhoun (1995a, 1995b, and in prep.). The mostly shallow marine sequence has been subdivided into several
formations yielding benthic faunas which have been, for the most part, dated precisely by diagnostic conodont faunas identified from the sequence by Colquhoun (1995b). Provisional ages for the stratigraphic units currently recognised are as follows:

(highest) Carwell Creek Formation (Offenberg et al. 1971); Zlichovian; Glendale Limestone Member (Pemberton et al. 1991); late Pragian-Emian; Riversdale Volcanics (Wright 1966); rare, non-diagnostic fossils; Roxburgh Formation (Pemberton 1980a); Pragian; Yellowman's Creek beds (Millsteed 1985); Lochkovian; Clandulla Limestone (Pemberton et al. 1991); Lochkovian; (lowest) Warrah Conglomerate (Campbell 1981); Lochkovian; unfossiliferous.

Various levels in the Clandulla Limestone, the Yellowman's Creek beds, and the Carwell Creek Formation yield varied and distinctive Devonian shelly faunas in the district. Affinity at the generic level allows tentative correlation of the Yellowman's Creek fauna with delta Zone faunas of the Garra Limestone (Savage 1968a, 1968b, 1969, 1973; Farrell 1992). The Roxburgh Formation contains a variety of faunas, almost entirely from clastic strata (Pemberton 1980a). The Carwell Creek Formation is predominantly clastic but limestone (typically pink or grey) developed just above the base of the Carwell Creek Formation in the Cudgegong-Kandos region was named the Glendale Limestone Member by Pemberton et al. (1991, 1994). Most of these limestones are stromatoporoid-dominated, but some occurrences of benthic shelly fossils and conodonts are known; rare Spinella and Buchananthrys have been reported from clastic beds of this formation (Millsteed 1992).

Reference to this stratigraphic framework permits assignment of distinctive lithologies at 'Erang' to a particular Early Devonian formation recognised elsewhere in the district. Crinoidal sandstones associated with the 'Erang' limestones are assigned to the Carwell Creek Formation, and the pink-yellow-grey 'Erang' limestone is, therefore, correlated with the Glendale Limestone Member.

SIGNIFICANCE OF THE FAUNA
The 'Erang' fauna is one of the few diverse faunas from the Carwell Creek Formation in the Cudgegong area; rich but very different and younger faunas are known from limestones at Mount Knowles and Brogans Creek which have also been assigned to the Carwell Creek Formation (Pemberton et al. 1994). Common species in the 'Erang' fauna include the trilobites Cyphaspis windamerensis sp. nov., Maurotarion erangensis sp. nov., Apocalymene sp. and Dudleyaspis sp. and the brachiopods Skenidioides johnsoni sp. nov., Aulacella boucoti sp. nov., Resserella rhodesi sp. nov., and M. (Mesodouvillellina) savagei sp. nov. A unique specimen of one species of tetracoral, Sociophyllum wilsoni sp. nov., is described. Less common and generally poorly preserved taxa left in open nomenclature include: Salopina sp., Douvillina sp., Apocalyptum sp. and Dudleyaspis sp. Occasional poorly preserved leptaenid, chonetid, orthotetid, pentamerid, atrypid, spiriferid (?Cyrtina) and rhynchonellid brachiopods also occur, as well as stromatoporoids, the tetracoral Rhizophyllum and tabulate corals.

The striking aspect of this fauna is the high proportion of new brachiopod and trilobite taxa. This is presumably a reflection of our limited knowledge of east Australian shelly faunas, especially those of this biofacies and age.

AGE
Despite the deformation, a varied and reasonably preserved conodont fauna discussed by Colquhoun (1995b), includes Panderodus unicostatus (Branson and Mehl, 1933; P. recurvatus (Rhodes, 1953); P. cf. valgus (Philip, 1965); P. sp.; Pandorinellina steinhornensis miae (Bultynck, 1989) [Pa elements], Amydrotaxis sp., Ozarkodina pseudomiae Mawson et al., 1992 [Pa elements], O. cf. buchanensis (Philip, 1966) [Pa elements], O. excavata excavata (Branson and Mehl, 1933) [Pa, Pb and M elements] and Ozarkodina sp. O. pseudomiae ranges from the pireneae to dehiscons zones (Mawson et al. 1992) and P. s. miae ranges from dehiscons to lower inversus zones (rarely occurs in pre-dehiscons strata: Klapper and Johnson 1980). Thus, the conodont fauna most probably should be assigned to the early Emian dehiscons Zone (medial Early Devonian), with some slight chance of assignment to the pireneae zone.

This age assessment is consistent with the age indicated by the brachiopods Spinella and Buchananthrys in the Carwell Creek Formation (Millsteed 1992; Pemberton et al. 1994). These brachiopods were considered by Garratt and Wright (1988) to be indicative of the dehiscons Zone (of that time) which has now been subdivided into the underlying pireneae and the overlying dehiscons zones.

The absence in the limestone fauna of Spinella, Nadiastrophia, Malurostrophia and Buchananthrys (which are present in beds at this level elsewhere in Eastern Australia) seems best attributable to ecological control, although the 'Erang' limestone may predate this zonal assemblage.

Conodonts from the 'Erang' limestone are similar to those extracted from the Glendale Limestone Member on the former property of Riversdale on the old road to 'The Dolomites', and from the 'Brogans Creek limestone'.

The brachiopod and trilobite species are not age-diagnostic, as a result of our limited knowledge of the stratigraphic ranges of Australian species of this age. Genera such as the brachiopods Resserella, Aulacella, and M. (Mesodouvillellina), and the trilobites Cyphaspis, Dudleyaspis, Harpella and Apocalyptum are long-ranging Silurian to Devonian forms, as is the operculate tetracoral Rhizophyllum.

Savage (1974) described shelly fossils from clastic strata of the underlying Maradana Shale. Some long-ranging genera from both the silicified and 'clastic' faunas (e.g. Resserella, Skenidioides) occur in the 'Erang' fauna.
The most abundant species are the brachiopods *Skenidioides johnsoni* n. sp. and *Aulacella boucoti* n. sp., and the trilobites *Apocalymene* sp. and *Cyphaspis windamerensis* n. sp. Other, less abundant taxa include the brachiopods *Resserella rhodesi* n. sp., and *M. (Mesodouvillina) savagei* n. sp. and the trilobites *Maurotarion erangensis* n. sp. and *Dudleyaspis* sp., as well as conodonts and ostracodes. The operculate coral *Rhizophyllum* sp. indet. and rare fragmental and indeterminate rhyynchonellid, chonetid, orthotetid, leptaenid, pentamerid, atrypid and spiriferid brachiopods also occur in the fauna. Although the indeterminate rhynchonellid, chonetid, orthotetid, leptaenid, pentamerid, atrypid and spiriferid brachiopods are present, with a high abundance of articulated brachiopods shells. Juvenile stages of less common taxa like *Dudleyaspis* and *Apocalymene* are not rare; well-preserved ostracod material occurs commonly. Less common taxa do not display such an obvious growth series, and relatively few number of slipper-shaped corallites.

These 'Erang' fossils are preserved in situ or at worst were para-autochthonous, and were not subjected to storm waves, as very little current sorting has occurred. This fauna appears to correspond to the subtidal BA 3 of Boucot (1975). Associated crinoidal quartz-lithic arenites represent a higher energy environment brought about by delta switching or sea level change.

**CONODONT BIOFACIES**

The rarity or absence of polygnathids in this and approximately coeval faunas (The Basin', Brogans Creek, Glendale, Capertee Valley) from Emsian strata in this part of the Capertee High is striking, as they are well developed in limestones in the Queens Pinch area (e.g., Garratt and Wright 1988), Mount Frome (Pickett 1978; Pickett and Wright in Strusz et al. 1988), Flirtation Hill (Chatterton and Wright 1986) and Mount Knowles (Pickett 1978; Colquhoun 1995b). Mawson and Talent (1989), Mawson et al. (1988) and Mawson et al. (1992) noted that nearshore Early Devonian environments commonly are dominated by ozarkodinan elements and simple cones. For this situation they proposed the Ozarkodinan Biofacies (analogous to the icriodid Biofacies of the northern hemisphere), which passes offshore into quieter, deeper conditions dominated by polygnathids and pandorinellinids. It appears that some Emsian platform carbonates of the Capertee High were dominated by the Ozarkodinan Biofacies in contrast to the Polygnathid Biofacies which dominated deeper areas marginal to the High (Queens Pinch, Limekilns); the presence of polygnathids in carbonate sequences (Mt Knowles, Mt Frome) on the platform cannot be explained as the result of the progressive deepening of the platform area.

**EARLY DEVONIAN FAUNAL SUCCESSION**

Recognition of a sequence of Early Devonian (Lochkovian, Pragian and Emsian) macrofaunas in the Lachlan Fold Belt has been hampered by the lack of uninterrupted and continuously fossiliferous sections, and the lack of good biostratigraphic control; however, this situation is being progressively ameliorated by the ongoing flow of conodont data.

The aim of this section is to review Early Devonian macrofossil biostratigraphy for the Lachlan Fold Belt by integrating new biostratigraphic data, especially those on silicified faunas and conodonts from the Garra Limestone and from several units in the Cudgegong-Carwell Creek area.


Sherwin (1992) recorded a number of faunal assemblages from the Lochkovian-Pragian of central N.S.W.; these faunas are poorly constrained by conodont data (Pickett and Ingpen 1990; Pickett and McClatchie 1991). Being dominated in part by *Howellella*, they are difficult to relate to the zones recognised by Garratt and Wright. Young (1995, p. 18) commented that "Previous assemblages now dated on associated conodonts give ages about half a stage older than before (cf. 1989 chart)".

**Boola-Maradana-Mullamuddy faunas.** One particular group of shelly faunas of similar generic composition, central to this discussion, was considered late Lochkovian by Garratt and Wright (1988). They include those described by Philip (1962) from the Boola Siltstone in Victoria and Savage (1974) from the Maradana Shale, Manildra district, N.S.W. as well as the lowest fauna from the Mullamuddy Formation, at Queens Pinch near Mudgee. The faunas are dominated by long-ranging brachiopod genera such as *Spirigerina, Eospirifer, Cyrtinopsis* and *Lissatrypa*, and lack hystero lidit brachiopods. This very generalised generic brachiopod assemblage appears to be strongly facies-controlled (as suggested by Savage (1974, fig. 4) and, at the same time, long-ranging.

Age control is now available for all 3 occurrences. The Garra Limestone with *eurekaensis* Zone conodonts (Mawson et al. 1988) overlies the Maradana Shale; in Victoria, the Coopers Creek Limestone with a *sulcatus* zone conodont fauna overlies the Boola Siltstone (Wall et al. 1995); and at Mudgee a *sulcatus* (to kindlei) has been suggested by Professor Ruth Mawson for conodonts including *P. exigua philipi* extracted from limey nodules occurring
in the fossiliferous strata of the Mullamuddy Formation. Species-level analysis of this brachiopod fauna is apparently crucial, as the genera are long-ranging.

Mawson and Talen (1994) advocated a correlation for the Boola Formation slightly above the Lochkovian-Pragian boundary. The presence in the Boola Siltstone of Boucotia australis had led Garratt and Wright (1988) to assign it to their 'late Lochkovian' australis zone. At the specific level, the occurrence at Manildra of taxa such as Notanoplia pherista Gill, Maoiristrophia banksi Gill and Notoleptaena, all absent from the Mullamuddy Formation, support the pre-eurekaensis age discussed below for the Maradana Shale.

Summary of Devonian faunal succession
Described brachiopod-coral faunas correlate approximately with 10 conodont-defined intervals. These are:-

i. woschmidtii Zone. Zonal conodont faunas occur in several units with largely undescribed macrofaunas: in the upper Elmside Formation ("Upper Trilobite Bed" of Brown 1956) in the Yass district of N.S.W. (Link and Druce 1972), in the Windellama Limestone (Mawson 1986); and in the basal Clandulla Limestone (Colquhoun 1995b).

Jones et al. (1981) identified Spathognathodus sp. cf. S. remscheidiensis from concretions in the Efflux Siltstone Member of the Freme Hill Formation (Bauer 1994) at Bungonia, N.S.W. Mawson (1986) suggested that the conodont illustrated by them might be S. r. eostehomensis and that the age might be Pridolian rather than Lochkovian. Wright and Bauer (1994) reported the occurrence of the brachiopod Cyrina from the Sawtooth Ridge Limestone Member of the Freme Hill Formation, noting that this genus is usually taken as a Devonian indicator.

ii. eurekaensis Zone. Faunas from the eurekaensis zone are relatively uncommon. The best known occurrence is in the Manildra (N.S.W.) sequence described by Savage (1968a), where the Garra Limestone (formerly placed by Savage in his Mangadery Creek Formation, but see Mawson et al. 1988 for reassignment) was assigned to the eurekaensis zone (Mawson et al. 1988). Savage's silicified faunas, characterised by the Reticulatrypa-Ogilviella brachiopod fauna (Savage 1968a, 1968b, 1969, 1971), were from this limestone.

A non-silicified fauna including the latter characteristic brachiopods occurs in the uppermost beds of the Waterbeach Formation at Paling Yards in the Hill End Trough sequence; no conodont data are available for this occurrence, but Garratt and Wright (1988) assigned it to their Lochkovian B. australis zone.

The Clandulla Limestone (Carwell Creek, Kandos district, N.S.W.), which ranges from the woschmidtii zone to the eurekaensis zone (Colquhoun 1995b), has yielded undescribed macrofaunas dominated by the apparently long-ranging brachiopod Machearia catomalensis Strusz, 1970 (woschmidtii-eurekaensis age at Carwell Creek; delta-pesavis age at The Gap [Farrell 1992]).

iii. delta Zone. At 'The Gap' near Wellington (Farrell 1992) listed the Ogilviella-Reticulatrypa fauna, ranging through at least 140 m of section consisting at least 40 m of delta zone and about 100 m of pesavis zone. The same brachiopod fauna is known in delta age strata of the Yellowmans Creek Formation (Colquhoun 1995a) in the Cudgegong-Kandos district, which has yielded conodonts assigned to the eurekaensis (in the basal 65 m approximately), delta and possibly the pesavis zones. Thus the fauna originally described by Savage from Manildra ranges from the eurekaensis zone through the delta zone into the pesavis zone.

iv. pesavis Zone. Some macrofauna from this zone, including Cyrtinopsis, were described by Farrell (1992, pl. 6, figs 1-3; as Megakozlowsiella) and mentioned by Sorentino (1989).

v. sulcatus Zone. Macrofaunas of this zone were believed by Garratt and Wright (1988) to be characterised by Nadiastrophia, Cyrtinopsis, and the appearance of Martinophyllum. The Nadiastrophia-Boucotia loyolensis (Gill) assemblage of Garratt and Wright (1988) is first seen in this zone, but its upper limit now seems to lie in the kindlei zone. This macrofaunal zone occurs widely; in the Garra Limestone (Lenz and Johnson 1985a, b); and in the Queens Pinch sequence near Mudgee (Wright 1966; M'Cracken 1990) where the Mullamuddy Formation contains both Cyrtinopsis and Nadiastrophia, and where the Taylors Hill Formation yields the first Martinophyllum. Murtifera? appears in this sequence just above the base of the Taylors Hill Formation, over 400 m above the appearance of Nadiastrophia. Neither of these clastic faunas (Mullamuddy and Taylors Hill formations) contains brachiopods similar to those of the Erang fauna.

The basal fauna of the Mullamuddy Formation at Queens Pinch, near Mudgee (Garratt and Wright 1988) includes Cyrtinopsis, Dolerorthis and Plectatrypa, but lacks Nadiastrophia, and was correlated by them with Garratt's (1983) B. australis zone. Conodont studies have demonstrated the presence of the Pragian sulcatus Zone extending from the lowest Mullamuddy Formation fauna (Professor Ruth Mawson, pers. comm. 1996) to about 150 m (M'Cracken 1990) below the earliest occurrence of the Nadiastrophia fauna in the Mullamuddy Formation: the range is at least 150 m. The upper limits of this zone are not constrained in this section until the kindlei zone is provisionally recognised in the overlying Taylors Hill Formation, this may provide some constraints on the australis zone, assuming it is correctly recognised here.

In the Wellington district of N.S.W., Nadiastrophia occurs in the Garra Limestone; according to Wilson (1989), the silicified trilobites described by Chatterton et al. (1979) and brachiopods described by Lenz and Johnson (1985a, b) from this formation are of sulcatus age. Wilson
suggested that some parts of the Garra range above *sulcatus*, as previously indicated by Druce (1971); Philip and Pedder (1967) noted possible *dehiscens* age conodonts and corals, although not with certainty suggested that some parts of the Garra range above the Coopers Creek Limestone in Victoria contains a good *sulcatus* fauna, and overlies the Boola Formation (with its *B. australis* brachiopod assemblage); they infer from these conodont data that the Boola Formation probably falls within the *sulcatus* zone. The Boola Formation was assigned a pre-*sulcatus* age by Garratt and Wright (1988). Depending on correlation of shelly faunas and the Lilydale Limestone in the Melbourne district, the *B. australis* assemblage zone may be correctly interpreted as laterally equivalent to (part of) the early to late Pragian Lilydale Limestone, which was shown by Wall et al. (1995) to contain horizons referable to possibly the *sulcatus* zone, and definitely the *kindlei* and *pirenae* zones.

The well-known but poorly understood Baton River fauna from New Zealand may correlate with this zone, although it does appear to be an atypical *dehiscens* zone fauna.

The type and only locality is on Erang' property, at Mount Frome below the appearance of Hipparionyx. Frome, and Limekilns. The brachiopods *Megastrophia* and *Zdimir* occur at Mount Frome below the appearance of *Bipennatus palethorpei* (Telford) - *Phacellophyllum furcatum* Pedder assemblage and the Malurostrophia-Taemostrophia-Howittia assemblage, which also includes the tetracorals *Xystriphyllum pavimentum* (Pedder) and *Xystriphyllum michellii* (Etheridge). Occurrences of the * inversus* zone reported have by Garratt and Wright (1988) not been confirmed.

vi. *kindlei* Zone. No characteristic macrofaunas of this age have been defined (Talent and Yoklin 1987; Mawson et al. 1988; Wilson 1989) from N.S.W. Wall et al. (1995) established that the Lilydale Limestone, Victoria, can be correlated in part with the *kindlei* zone (see above). Colquhoun (1995b) recorded and illustrated *Eognathodus sulcatus* cf. *kindlei* which occurs with *Boucota australis* in the Kingsford Formation in the Cudgegong district, NSW.

Conodonts of probable *kindlei* age have been identified from low in the Taylors Hill Formation at Queens Pinch by Professor Ruth Mawson; the macrofauna at this level includes *Nadiastrophia* superba Talent and eospiriferids. Combined with the new data from the Lilydale area discussed above, this suggests that hysterolitid and acrospiriferid brachiopods appeared in Eastern Australia no earlier than the *kindlei* zone, as forms like *Hysterolites lilydalensis* (Chapman) are well known from the Humevale Formation underlying the Lilydale Limestone (Garratt 1983). Garratt (1983) placed the incoming of *Hysterolites lilydalensis* and other species at the base of his *australis* zone to which he assigned a 'late Lochkovian' age. The new conodont data thus suggest that Garratt's 'late' Lochkovian age is too old, although Carter et al. (1994) indicated that both the Hysterolitinae and Acrospiriferinae appeared in the upper Lochkovian.

vii. *pirenae* Zone. Until macrofaunas are well described from this zone, only taxa described in this paper can be considered as even potential zonal fossils. Currently, the *P. steinhormensis miace - O. pseudomiae - linearis* fauna is known from the Glenale Limestone Member of the Carwell Creek Formation ('Erang' limestone and in the Carwell Creek district). The Erang fauna might correlate with this zone, although it does appear to be an atypical *dehiscens* zone fauna.

viii. The *Spinella-Buchanathyris* assemblage was the only *dehiscens* zone fauna recognised by Garratt and Wright (1988). In addition to the name-giving taxa, this assemblage includes coral faunas characterised by *Zelolasma, Chalcedophyllum* and *Tipheophyllum*. The Reefson sequence in New Zealand probably correlates with this assemblage.

Pickett and M'CIntochie (1991) and Sherwin (1992) concluded that occurrences of *Spinella*? in the Jerula Formation of central N.S.W. may be as old as *sulcatus* Zone, and Mawson et al. (1992) suggested that the genus extends down to at least the *pirenae* Zone.

Range data for constituent conodont taxa are given by Mawson et al. (1992), and indicate that this fauna correlates best with the *dehiscens* Zone and, therefore, is probably of earliest Emsian age.

Wright (1994) discussed a probably Pragian shelly fauna from the basal beds of the Cunningham Formation in NSW. The occurrence there of *Nadiastrophia?* and *Spinella?* apparently indicates a correlation with the *Spinella-Buchanathyris* assemblage, although *Spinella* and *Nadiastrophia* are normally exclusive. The *Spinella* in this fauna may be thus of *kindlei* age.

ix. *perbonus* Zone faunas are characterised by the *Malurostrophia-Taemostrophia-Howittia* assemblage, which also includes the tetracorals *Trapezophyllum pavimentum* (Pedder) and *Xystriphyllum michellii* (Etheridge). Occurrences of the * inversus* zone reported have by Garratt and Wright (1988) not been confirmed.


The brachiopods *Megastrophia* and *Zdimir* occur at Mount Frome below the appearance of *Bipennatus palethorpei* (Telford) - (late Emsian to early Eifelian according to Mawson 1993) in the *serotinus* zone (Wright 1981; in Strusz et al. 1988). The two above brachiopods occur in a silicified fauna immediately below the level at which *B. palethorpei* appears in the Mount Frome Limestone (Strusz et al. 1988).

**SYSTEMATIC PALAEONTOLOGY**

The type and only locality is on 'Erang' property, at
grid reference 616747 (Lue 1:25,000 topographic sheet 8832-1-S, 1st edition). Authorship of brachiopods and trilobites is attributed to Wright and Chatterton, and that of the coral to Wright; conodont identifications are by Colquhoun.

Because most specimens have been tectonically deformed, accurate measurements are of dubious taxonomic value, and so have been omitted. Furthermore, the outlines of the specimens, and internal features such as brachiopod muscle fields have, in some cases, been altered by this deformation (see Aulacella herein). We have tried, where possible, to eliminate the effect of such distortion from our diagnoses and descriptions.

Institutional abbreviations. AMF, Australian Museum, College Street, Sydney, New South Wales.

**PHYLUM CNIDARIA** Hatschek, 1888
**ORDER TETRACORALLIA** Haeckel, 1866
**FAMILY STRINGOPHYLLIDAE** Wedekind, 1922
**Genus Sociophyllum** Birenheide, 1962

Type Species: *Spongophyllum elongatum* Schlüter, 1881, from the Givetian (late Middle Devonian) Rech-Horizont, Loogher Schichten, Bemdorf, Hillesheimer Mulde, Eifel region, Germany; by original designation.

**Sociophyllum wilsoni** sp. nov.
Plate 1, figures 1-5

**Material.** Holotype AMF 88794.

**Derivation of name.** The species is named for the late Bill Wilson, former proprietor of 'Erang'.

**Diagnosis.** *Sociophyllum* with maximum corallite diameter of 11 mm; 20-24 long major septa with abundant monocanths; minor septa mostly confined to septal stereozone; dissepiments rare; tabulae mostly complete and strongly concave in mature corallites, but flattish in juveniles.

**Description.** Colony broadly conical with a diameter of 20 cm, with phaceloid corallites closely packed and weakly diverging; increase is parricidal, with up to four offsets being produced. Average corallite diameter about 6 mm, maximum 11 mm. From 20-24 major septa extend from half to at least two thirds distance to axis, axial space occupied by discrete monocanths; minor septa generally confined to septal stereozone up to 0.5 mm thick, but may reach one sixth of distance to axis, and are rarely seen as monocanths. Septa weakly dilated close to stereozone, with wedge-shaped ends. Dissepiments seen as rare, vertical, flattened cysts. Tabulae always concave and complete in mature corallites, often deeply concave, with an axial steepening; distinctly flattish in immature corallites.

**Remarks.** The dissepiments, concave tabulae, and monocanthine septa argue for assignment to *Sociophyllum*. No closely comparable species are known, although two European species warrant comparison.

*S. (S.) semiseptatum* (Schlüter; Birenheide 1962, pl. 9, fig. 6a, b) has an incomplete series of preseppiments and flattish to concave tabulae; it differs from *S. (S.) wilsoni* in corallite diameter (7-8 mm), in lacking minor septa and having 24-30 major septa, and in its tabulae, which are irregularly flattish.

*Crista varia* Tsyganko, 1971 is a fully to loosely cerioid species from the lower Eifelian of the northern Urals with monocanthine septa, sparse dissepiments and flattish to weakly concave tabulae; this species differs from *S. wilsoni* mainly in the nature of the tabulae. *Crista* has been assigned to the Stringophyllidae by Pedder and Wright (in prep.).

**Phylum BRACHIOPODA** Dumeril, 1806
**Class ARTICULATA** Huxley, 1869
**Order ORTHIDA** Schuchert and Cooper, 1932
**Family SKENIDIOIDES** Schuchert and Cooper, 1931

**Type species.** *Skenidioides billingsi* Schuchert and Cooper, 1931, from the Middle Ordovician Black River Limestone, Paquette Rapids, Ontario, Canada; designated by Schuchert and Cooper, (1931).

**Remarks.** In addition to the type species, well-known species of *Skenidioides* are from Oklahoma (Amsden 1958), Poland (Biernat 1959), Nevada (Johnson et al. 1973), Utah (Sheehan 1976), the Yukon (Lenz 1977), Czechoslovakia (Havliček 1977), Canadian Arctic (Smith 1980), District of Mackenzie (Perry 1984) and New South Wales (Lenz and Johnson 1985a, b); these taxa are discussed in detail below where appropriate. *Skenidioides henryhousensis* Amsden, 1958 (Early Devonian, Oklahoma), is referred to *Skenidioides* with doubt as no internal structures are known. Boucot et al. (1966) discussed the generic assignment of several other species.

*Skenidioides johnsoni* sp. nov.
Plate 2, figures 1-17; Plate 3, figures 1-9

?1966 *Skenidioides* sp.; Boucot et al., p. 365, pl. 40, figs 6-9 only.


**Derivation of name.** This species is named for the late Jess Granville Johnson, formerly of Oregon State University.

**Material.** Holotype AMF 65440 (brachial valve). Paratypes AMF 65441-65453, AMF 88767-88772, AMF 88793.

**Diagnosis.** *Skenidioides* with semi-circular shape in plan view, apsacline pedicle interarea, very low brachial interarea; fold and sulcus wide and shallow; brachiophores ventrally directed.

**Description.** Ventrri-biconvex *Skenidioides* with almost semi-circular outline, and a distinct brachial sulcus, a very high brachial median septa and a correspondingly high pedicle valve. Pedicle interarea is subplanar and apsacine (rarely catacline), with an
apical angle of about 75° to 90°, delthyrium has an apical angle of about 30°, with apical 1/3 to 1/2 filled internally by spoon-shaped spondylum supported by short median septum. Low brachial interarea is anacrine, approaching apsacline. About 30 costellae at the margin of a mature valve, mostly present at 1/3 valve length; increase is by bifurcation. Maximum width of about 6 mm is close to midlength.

Pedicle musclefield has distinct low, broad median ridge, with shallow lobate adductor scars extending to 2/3 length of valve; adductor scars are weak, and extend beyond midlength. Stubby hinge teeth are transversely semi-oval, and are elongate and flattened parallel to hingeline at margin of delthyrium.

Cardinal process is rarely swollen, and is continuous (in plan view and lateral profile) with median septum; median septum is highest at midlength of valve, and almost reaches anterior margin of valve; brachiophores are very long and slender, ventrally directed; cruralium is low and sessile in juveniles, elevated in adults, being formed by deeply sagging crural plates, and reaches forward to almost 1/3 of valve length. Long lobate adductor scars flank median septum, reaching at least 2/3 valve length, flanked by low muscle bounding ridges for most of this length; sockets small, floored by fulcral plates and covered distally by interarea.

Remarks. The type species, S. billingsi Schuchert and Cooper, 1931, differs from S. johnsoni in having a distinct brachial interarea and has (Schuchert and Cooper 1932, p. 72) a rather deeply sulcate brachial valve; S. billingsi is poorly known internally.

The species known in most detail are based, almost without exception, on silicified material. S. robertseni Johnson et al., 1973 is transverse and has a relatively high brachial interarea, a steeply apsacline to catacline pedicle interarea, more costellae and a relatively short cruralium compared with S. johnsoni. S. variabilis Lenz, 1977 differs from S. johnsoni in: being more mucronate, having fewer and more irregular costellae (up to 26), a longer cruralium, anteriorly directed brachiophores and a weakly convex brachial valve with a longer interarea. S. johnsoni n. sp. differs from S. cf. S. pyramidalis (Hall, 1852) from the Silurian of Utah (see Sheehan 1976), in the following: the hinge line is relatively shorter; the maximum width is farther forward, being closer to the midlength; costellae are mostly added earlier in the ontogeny; and the spondyline does not underlie as much of the delthyrium (1/3 to 1/2 as opposed to 2/3). S. blussoni Perry, 1984 differs from S. johnsoni in being more transverse and in having fewer (16-26) costellae, long sockets and highly divergent brachiophores. This species is poorly known internally. Bassett and Cocks (1974) discussed but did not illustrate "Skenedioides acutum" (Lindstrom, 1861) and "Skenedioides lewisi" (Davidson) from the Silurian of Gotland. Boucot et al. (1966) listed "Scenidium baylei" (Roault in Oehlert 1888), "Scenidium moelleri" Tscherneychev, 1887 and S. uralicum Tscherneychev, 1887 as taxa not assignable to genus as the internal structures are not known.

Havlicek (1977) described 6 species of "Scenidium" from Czechoslovakia including S. suburbanus (Havlicek, 1956) and the new species moranus, famulus, boucoiti, fascianatus and cingulatus. Of these, fascianatus is only known from the pedicle valve and has up to 17 costellae; boucoiti and cingulatus lack a fold or sulcus and have few ribs, both moranus and famulus have an erect beak, and famulus is further distinguished by its narrow deep brachial sulcus.

Most other Australian representatives of "Skenedioides" are poorly known. Boucot et al. (1966) reported Early Devonian occurrences of "Skenedioides" from Victoria, and listed Devonian occurrences of the genus. Some other Australian occurrences of "Skenedioides" from Early Devonian rocks of New South Wales are listed by Wright (in Packham 1969) and Savage (1974) described one form; these are further discussed below.

"Skenedioides" sp. (Savage 1974, p. 12, pl. 8, figs 24-26) from the Lochkovian (see Klapper and Johnson 1980; Mawson et al. 1988) Maradana Shale of N.S.W. differs mainly in having fewer costellae (25) and may be conspecific with the Lochkovian (Garratt 1983) form reported from the Crudine Group at Paling Yards, N.S.W. by Wright (in Packham 1969). Comparison of moulds (as is the case for these two occurrences) with silicified material (as in S. johnsoni, S. robertseni and S. variabilis) is not simple. However, the "Skenedioides" sp. described by Boucot et al. (1966, p. 365, pl. 40, figs 6-9) from the Lochkovian Bell Shale of Tasmania has a similar number of costellae to, and may be conspecific with, S. johnsoni. In addition, S. aff. variabilis Lenz described by Lenz and Johnson (1985a) has some morphological features in common with S. johnsoni.

Strusz (1982) described a new species, "Skenedioides thronax", from Wenlock strata near Canberra; S. johnsoni differs in: being less transverse and usually less alate; having a longer median septum in both valves; and having a cardinal process which is not as distinctly swollen.

Family DALMANELLIDAE Schuchert, 1913
Subfamily RESSERELLINAE Lazarev, 1970
Genus RESSESELLA Bancroft, 1928

Type species. Orthis canalis J. de C. Sowerby in Murchison, 1839, from the Wenlock of Coalbrookdale Formation, Wooton Hope, Herefordshire, England; by original designation.

Remarks. Havlicek (1977, p. 169) stated that there is a transition from the multicostellate (and parvicostellate) Resserella to the coarsely fascicostellate Parmormithina and thence to the fascicostellate/plicate Fascizetina. Multicostellate Resserella occur in both Silurian and Early Devonian strata, but parvicostellate forms apparently occur only in Early Devonian rocks.
Further studies may warrant formalising and resurrection of the currently invalid (as no type species has been designated) genus Curranella (see Chatterton 1973, pl. 35, figs 1-3) for this species group as indicated by Havlichek (1977, p. 153).

**Resserella rhodesi** sp. nov.
Plate 2, figures 18-27

**Derivation of name.** This species is named for the late Valentine Rhodes, former owner of 'Windamere' and 'Stoney Pinch' properties, both of which are now largely submerged beneath the waters of Windamere Dam.

**Type material.** Holotype AMF 65454 (incomplete bivalve shell). Paratypes AMF 65455-65458, AMF 65464-65467, AMF 65469, AMF 88773, AMF 88775.

**Diagnosis.** Generally parvicostellate, transverse *Resserella* with up to 10 major costellae and a wide, shallow brachial sulcus at maturity; maximum width posterior of midlength. Juvenile brachial valves with pronounced narrow, deep sulcus; sharp median brachial ridge bisecting long, narrow, lobate adductor field and extending to anterior margin; and costellate sculpture.

**Description.** Ventriloculated *Resserella* with wide, shallow brachial sulcus and parvicostellate sculpture, with 17 costellae per 5 mm at distances of 5 and 20 mm from beak; up to 60 costellae on a mature shell, including about 8-10 major costellae. Shells up to 12 mm wide and 10 mm long; valves transversely or oblong semi-oval (depending on orientation of strain), with maximum width located posterior of midlength. Juvenile shells costellate.

Pedicle interarea curved and apsacline; teeth of moderate size, surmounting curved, rounded dental ridges continuous with short dental plates ventrally, pedicle valve with very short, shallow and triangular muscle field, with broad, short, low median platform anteriorly, and a well-defined transverse ridge forming the anterior edge of muscle field; a convex plate closes apical part of delthyrium which has an apical angle of 50°-60°.

Brachial interarea low and anacline; cardinal process short, stout, bilobed posteriorly, with abrupt anterior face at posterior end of very low, broad median ridge; brachial adductor scars shallow and extend to about mid-length, bounded by very low muscle ridges posterolaterally, muscle bounding ridges disappear and converge anteriorly; brachiotheca bases stout and triangular in section; sockets narrow, with axes highly divergent posteriorly at about 160°, falcular plates well-developed; setal grooves shallow for genus.

**Remarks.** Walmsley and Boucot (1971) illustrated a number of species of *Resserella*, they recognised only one parvicostellate species, *R. elegansulcatus* (Kozlowski, 1929) from the Gedinnian of Podolia (see also Johnson et al. 1973). *R. elegansulcatus* is best known from Kozlowski (1929, fig. 10) and Nikiforova et al. (1985); Podolian specimens are less transverse than *R. rhodesi* and have relatively robust brachiotheca and more primary ribs. Johnson et al. (1973), Savage (1974), Lenz (1977, with some uncertainty) and Perry (1984) all referred material to *elegantuloides*, these materials form a compact group which are, however, almost certainly not conspecific and need further study. A genus that could be a suitable receptacle for these species is *Curranella* Chatterton (see above).

A number of other parvicostellate species are important for comparison and discussion. Probably the best known of these is *Resserella careyi* Chatterton, 1973, from Zlichovian limestones near Good Hope, New South Wales; it possesses a distinctive character which it shares with *R. rhodesi*, this being the deltidial plate that closes the apex of the delthyrium. *R. rhodesi* can be distinguished from *R. careyi* by having: a broader brachial sulcus; less ponderous brachiotheca; less distinct muscle bounding ridges and median ridge in the brachial valve; less impressed muscle scars in the pedicle valve; coarser ribbing (17-18 per 5 mm at 10 mm from beak as opposed to 21-24 per 5 mm); and less distinct setal grooves.

*Salopina? herkion* Farrell (1992) was based on material from late Lochkovian *pesavis* Zone beds low in the Garra Limestone at 'The Gap'. It differs from *R. rhodesi* in lacking the pronounced, narrow brachial sulcus. Farrell's distribution charts show that *S.? herkion* occurs with an undescribed species of *Resserella*.

No closely similar species to *R. rhodesi* were described by Lenz and Johnson (1985a) from *sulcatus* Zone beds of the Garra Limestone. They described *Prokopia hilliae* (Chatterton) (which has a brachial adductor field like that of *R. rhodesi*) and *S.? herkion*, but differs in having a high brachial median septum. Species of *Salopina* described by them include *S. submurifer* Johnson et al., which differs in having a relatively wide brachial median ridge and fascicostellate sculpture; the authors note a distinct brachial sulcus. A second species they described is *Salopina cf. kemezyasi* Chatterton which has a totally different brachial adductor field, but does have subfascicostellate sculpture.

*Resserella brownsportensis* (Amsden, 1949) from the Ludlow of North America is parvicostellate and may share with the Australian taxa the small apical plate in the delthyrium, as described by Amsden (1949). However, Walmsley and Boucot (1971) stated that the delthyrium of that species is open, and the plate is not visible in specimens illustrated by those authors (1971, pl. 98, fig. 5b) or by Sheehan (1976, pl. 5, figs 4-7). In other characters, this species is not particularly similar to *R. rhodesi*.

Another species which appears to belong to this small parvicostellate group is *R. strussi* Savage, 1974, as it also shows the apical development of deltoidal plates; however, it is distinct in various features of the brachial valve, including the anteriorly convergent muscle bounding ridges, more prominent falcular plates, the lack of a long, prominent median ridge in juveniles, its plano-convex form and triangular outline. *R. strussi* further differs from *R. rhodesi* in having: fewer (up to 50) costellae, maximum width located more anteriorly, and pedicle muscle-bounding ridge located more anteriorly. Other Australian species, although not parvicostellate, include: *R. impensa* Philip, 1962 from...
the Early Devonian Boola Beds of Victoria is distinct in having coarse fascicostellate ribbing; *Parmorthis vandieni* Gill, 1948 (which, according to Philip 1962, may not be a *Resserella*), and *Parmorthis aff. allani* (Shirley) of Gill (1950, p. 253), which is again poorly known.

*R. walmsleyi* Havlicek, 1977 (p. 166, pl. 41, figs 3-7) is a parvicostellate species known from mostly incomplete material. It differs from *R. rhodesi* in the broad, high median ridge in the brachial muscle field and the massive brachiophores, and in lacking the deltiodial plate. Another species differing from *R. rhodesi* in sculpture is *R. amsendeni* Walmsley and Boucot, 1975, a Ludlow species characterized by a brachial median ridge and fascicostellate ribbing.

Subfamily Isothinae

Genus **SALOPINA** Boucot, 1960

Type species. *Orthis lunata* J. de C. Sowerby, 1839; from the Early Devonian of northwest France; by original designation, although Nikiforova *et al.* (1985) cite 1827 as the date of authorship of this species.

Salopina sp. indet.

Plate 3, figures 10-11; Plate 4, figures 13-17

Material. AMF 79021-4, 88774, 88776.

Remarks. This material is assigned to this taxon with some reservation, as juveniles of *Resserella rhodesi* are not known. The brachial valves are small, transverse, gently sulcate and simply ribbed, and exhibit a long slender median ridge and a long narrow muscle field constrained by a pair of bounding ridges. Pedicle valves are cap-shaped, strongly apsacline, and have a short muscle field terminated anteriorly by a transverse ridge. This form strongly resembles forms like *S. submurifer* Johnson *et al.* (1973) from the Pridolian to early Gedinnian of central Nevada.

Family RHIPIDOMELLIDAE Schuchert, 1913

Genus **AULACELLA** Schuchert and Cooper, 1931

Type species. *Orthis Eifliensis* Schnur (1853, p. 213, plate 37, fig. 6 a-c) from the "Kalk" of the Eifel; by original designation.

Remarks. The name of the type species has been bedevilled by a confused history. Anderson *et al.* (1969, p. 121) correctly drew attention to the fact that the taxon mentioned but not formally erected by de Verneuil (1850, p. 161: "Orthis Eifliensis, Verm. [manuscr.]") for material from Veneros (Spain) and the Eifel is a *nomen nudum*, and attributed the species name *eifliensis* to Schnur (1853), the first describer. Schuchert and Cooper (1931) cited *O. eifliensis* (sic.) Schnur as type species of their new genus *Aulacella* (as did Chatterton, 1973), and in their 1932 monograph they cited *O. eifliensis* de Verneuil as the type species.

Biemat (1959) cited *O. eifliensis* Schnur as type species. Schnur (1853) and Kayser (1871a, 1871b) termed this species *Orthis Eifliensis* de Vern.

Chatterton (1973) regarded *Dalejina* as a junior subjective synonym of *Aulacella*. Perry (1984, p. 25) stated that "Chatterton's discussion appears partially valid", but followed the common western North American practice of using *Dalejina* for dalmanellids of this type. Those *"Dalejina"* with subtriangular shape are, we agree, distinctive. The species has been described from the Eifel, Poland and Burma.

*Aulacella? boucoti* sp. nov.

Plate 4, figures 1-12

Derivation of name. This species is named for Arthur J. Boucot of Oregon State University.

Type material. Holotype AMF 88782 (originally bivalved shell, subsequently broken). Paratypes AMF 65468, 88783-88789.

Diagnosis. Biconvex, *Aulacella?* with gentle pedicle fold and brachial sulcus, subcircular to transversely oval, up to 44 ribs. Brachial muscle field very shallow, with very low median and muscle bounding ridges, and indistinct transmuscle ridge; pedicle muscle field very shallow and weakly bilobed anteriorly, with low median ridge of moderate width and low, outwardly convex muscle bounding ridges that disappear anteriorly.

Description. Subequally, moderately biconvex *Aulacella?* with gentle pedicle fold and brachial sulcus becoming weaker anteriorly; subcircular to transversely oval in outline, maximum width only slightly in front of midlength. Pedicle interarea steeply apsacline, enclosing an apical angle of about 90°, brachial interarea anacrine. Delthyrium and notothyrium open. Sculpture of up to 44 ribs increasing by bifurcation and intercalation in a valve 10 mm wide and 7 mm long.

Brachial interior. Cardinal process bilobate or trilobate. Stout brachiophore bases convex anteromedially, sockets stout. Brachial muscle field extends to about mid-length, very shallow and subquadrate, with very low median and muscle bounding ridges, and indistinct transmuscle ridge.

Pedicle interior. Pedicle muscle field very shallow and weakly bilobed anteriorly, not reaching mid-length; with low median ridge of moderate width and low, outwardly convex muscle bounding ridges that disappear anteriorly. Pedicle callist and crural fossettes present. Short dental plates diverge at just less than 90°.

Remarks. Chatterton (1973) described *A. philipi* and *A. stoermerti* from Zlichovian strata near Good Hope, New South Wales. Of the two, *A.? boucoti* is more similar to *A. stoermerti*, but differs in the following: the muscle fields are shallower, the ribs appear to be more rounded, with more numerous distinct growth lines (almost lamellae); and the cardinal process is shorter. The Pragian species *Dalejina ovata* Lenz and Johnson (1985) from the Garra Limestone differs in having a subpentagonal outline and a strong myophore in front of the stout cardinal process.
Order STROPHOMENIDA Opik, 1934
Family STROPHEOdontidae Caster, 1939
Genus MESODOUVILLINA Williams, 1950

Type species. Stropheodonta (Brachyprion) subinterstrialis seretensis Kozlowski (1929, p. 87, pl. 4, fig. 1-7), from the Early Devonian upper Borschov al. seretensis, Rem arks.

in the subspecies Type species. Stropheodonta (Brachyprion) designation.

concavo-convex forms are assigned to convexity of the valves and nature of sculpture; weakly genus into three subgenera, two of which are separated Derivation of name.

from the nominate subgenus on the basis of relative strongly rugose. The 'Erang' material therefore belongs M. Savage of the University of Oregon at Eugene.

sp. nov.

Plate 5, figures 1-15, Text-figures zz-yy

Derivation of name. This species is named for Norman M. (Protocystrophia) savagei from M. subinterstrialis Kozlowski (Nikiforova et al. 1985, pl. 4) and M. stecki Lenz, 1977. Species known to possess similar ridges include M. cf. varistrata (Conrad): (see Johnson 1970, pi. 23), M. equicosta Smith, 1980; and M. delormei Perry, 1984 differ in other important features.

Several previously-described Australian species are assigned or assignable to Mesodouvilla, including Stropheodonta limbimura Talent, 1965 (see also Savage 1974, p. 23-24), Cymostrophia dickinsi Chatterton, 1973 and C. multicosstella Chatterton, 1973, the latter two species were assigned to their new subgenus Mesodouvilla (Protocystrophia) by Harper and Boucot (1978). All three species have more primary costellae than M. savagei, although the arrangement of sockets, teeth, muscle-bounding ridges and brachial ridges in C. dickinsi especially is similar to that of M. savagei.

C. dickinsi (see Chatterton 1973, pl. 6, figs 5, 6) and M. savagei (see camera lucida sketches of the internals of brachial valves: Text-fig. 8) exhibit considerable variation in the development of ridges in the brachial valve; one specimen (AMF 65459) also resembles a specimen of Talaeoshaleria padaukinensis (Reed, 1908) figured by Anderson et al. (1969, pl. 4, fig. 9). The same general arrangement of muscle-bounding ridges and median ridge is seen in the possibly descendant plexus of Nadiastrophia, Malirostrophia and Taemostrophia.

Comparison of this Windamere material with Mesodouvilla limbimura (Talent, 1965) is limited by the available illustrations of Lochkovian material from Heathcote, Victoria (Talent 1965, pl. 5, fig. 11; pl. 14, figs 1-3) and from Manildra, N.S.W. (Savage 1974, pl. 5, figs 1-5). Savage's diminutive material shows about 16 primary costellae; Talent illustrated a large specimen (1965, pl. 14, fig. 1) with a much greater number of primary costellae. Without more definitive material, adequate comparison is difficult.

Genus DOUVILLINA Oehlert, 1887

Type species. Orthis dutertrei nobis Murchison, 1840,
Material. AMF 88781, AMF 65409-10.

Remarks. This stropheodontid is represented by only a few fragmentary pedicle valves, which may not be conspecific. They differ from Mesodouvillella savagei n. sp. in having a lobate muscle-bounding ridge and a strong median ridge. They cannot be assigned with certainty to Douvillella but strongly resemble Douvillella arcuata (Harper and Boucot, 1978, pl. 26, figs 8-10). Fragments of Leptaena occur rarely in the fauna.

Class TRILOBITA Walch, 1771
Family AULACOPLEURIDAE Angelin, 1854
Genus MAUROTARION Alberti, 1969

Type species. Harpidella maura Alberti, 1967 from the Ludlow of Morocco; by original designation.

Remarks. This genus and its constituent species have been revised by Adrain and Chatterton (1995), who assigned a number of species previously assigned to Harpidella to Maurotarion. Australian species of the latter genus are M. struszi (Chatterton, 1971) from the Emsian of N.S.W.; M ? munroei (Strusz, 1964) from the Pragian of N.S.W. and M. mystax Holloway and Sandford, 1993 from the Llandovery of Tasmania. Specimens illustrated by Chatterton et al. (1979, e.g., pl. 105, figs 4, 21-22), and described as a free cheek of Otarion listron and cranidia of Otarion sp. B, should be assigned to Maurotarion.

Maurotarion erangensis sp. nov. Plate 6, figures 1-13

Derivation of name. This species is named after 'Erang' property, where the type locality is located.

Type material. Holotype AMF 65404 (cranidium). Paratypes AMF 65405-65408, AMF 65784-65788, AMF 88792.

Diagnosis. Maurotarion with sculpture of low tubercles on back of almost cylindrical glabella; small, isolated L1 lobes; anterior branches of facial suture strongly divergent forward; fixigenae field narrow opposite eyes; occipital node low; librigenae smooth except for terrace lines near margin and on doublure; elongated axial spine on one thoracic segment.

Description. Maurotarion with a sculpture of low tubercles towards back of cylindrical glabella, on occipital ring and on palpebral lobes; length of preglabellar field (sag.) about equal to that of anterior border; fixigenae narrow, with strongly convex palpebral lobes close to axial furrows; anterior branch of facial suture strongly divergent forward; L1 small and little inflated; S1 and occipital furrow comparatively shallow, S2 apparently absent; eye ridges very low; palpebral lobes opposite midlength of glabella in front of occipital ring, and not quite as high as glabella; free cheek smooth, with comparatively short, sharp, furrowed genal spine (length about equal to that of rest of cheek); lateral border furrow shallows slightly and constricted close to genal angle.

Pygidium with 3 distinct segments, with a sculpture of a distinct row of prominent granules on each segment, an indentation of the margin behind first segment, and with comparatively shallow furrows on the axis and pleurae.

Remarks. There is no certainty that any pygidia recovered belong to this species, most probably belonging to the more abundant Cyphaspis windamensis n. sp.

M. erangensis n. sp. differs from M. maura in having a less convex margin to the cranidium, a more densely tuberculate sculpture on the glabella, usually lacking sculpture on the preglabellar field, a more cylindrical glabella, and a less prominent occipital node. Although clearly congeneric, these species are not very similar.

The most similar Australian taxa to M. erangensis n. sp. are M. munroei (Strusz, 1964) and Harpidella sp. B (as Otarion sp. B of Chatterton et al., 1979); these two species are from shales and limestones respectively of the Pragian Garra Limestone. M. erangensis differs from M. munroei in that the occipital furrow and lateral border furrows are shallower. However, as the type material of M. munroei is missing and consists of a few rather poorly preserved cephalas (see Strusz 1964; Chatterton et al. 1979), the species may be best discarded.

The material of H. sp. B of Chatterton et al. (1979) consists of only a few silicified cranidia. M. erangensis differs from this form in having: a sculpture of tubercles on the back of the glabella and of granules on the anterior border; a more semicylindrical and less conical glabella; and palpebral lobes set closer to the axial furrows.

M. erangensis resembles M. struszi (Chatterton, 1971), from Zlichovian strata near Good Hope, New South Wales, in having small, little inflated L1 lobes, a narrow fixigena, a distinctly divergent anterior branch to the facial suture, an anterior border which is similar in length to the preglabellar field, and the extension of the posterior border furrow on to the preglabellar field. It differs from that species, however, in having rarely a row of fine granules on the anterior border, in lacking an obvious occipital spine in small holaspis cranidia; in having a palpebral lobe that is farther forward, and a genal spine that is much shorter.

Some similarities exist between M. erangensis and Maurotarion plautum (Whittington and Campbell, 1967) from the Silurian of Maine, including: the relative lengths of the preglabellar field and anterior border; the angle of divergence of the anterior branches of the facial sutures; the position and size of the palpebral lobe; and the short, sharp nature of the genal spines. M. erangensis differs from H. plautum in: lacking a sculpture on the preglabellar field and the front of the glabella; having smaller L1 lobes; having...
shallower occipital and S1 furrows; and having a more distinctly furrowed genal spine.

Genus CYPHASPIS Burmeister, 1843

Type species. Phacops ceratophthalmus Goldfuss, 1843, Eifelian from Gees, near Gerolstein, Germany; by original designation.

Remarks. Adrian and Chatterton (1994, 1996) considered Otarion and Cyphaspis sister taxa. They provided (1994, p. 312) a differential diagnosis for these genera, and pointed out that Wenlock species from northwestern Canada which belong to these two clades are so similar that, without knowledge of later lineages with distinct apomorphies, they would almost certainly be included in the same genus. Some of the distinctive features of these genera require well-preserved and undistorted material to allow discrimination of the diagnostic character states (not the case for our material). We have included our new species in Cyphaspis because of:- the comparatively inflated glabella (which appears to overhang a short preglabellar field); the narrow interocular fixigenae; the wide and short free cheeks with comparatively long genal spines; the small pygidia (apparently with only 3 axial rings and a terminal piece) and the comparatively broad pygidial doublure.

The two best-known Australian species of Cyphaspis are C. dabrowni (Chatterton, 1971) (see Adrian and Chatterton, 1996) and C. listron (Chatterton et al., 1979), Adrian and Chatterton (1995, p. 309) reassessed the type material of C. listron, and suggested that a small cranium included among the types of listron (Chatterton et al., 1979, pl. 105, figs 1-3) may belong to Otarion, and that a free cheek (Chatterton et al., 1979, pl. 105, fig. 4) should be assigned to Maurotarion.

Cyphaspis windamereensis sp. nov.

Plate 7, figures 1-17; Plate 8, figures 1-22

Derivation of name. This species is named after the property of 'Windamere' (now largely submerged under Lake Windamere, a reservoir), close to the type locality.

Type material. Holotype AMF 65412 (cranium). Paratypes AMF 65077, AMF 65411, AMF 65413-65420, AMF 65771-65783, AMF 88750-88757.

Diagnosis. Cyphaspis with a long juvenile occipital spine, reducing to a stout node in larger specimens; sculpture of large and small tubercles on the glabella, fixigenae and preglabellar field; tubercles on anterior and lateral margins of cephalon develop from short spines in early growth stages. Inflated L1 lobes.

Description. Cyphaspis with moderately sized L1; anterior border approximately same length as or shorter than preglabellar field; anterior margin occasionally has (tri)angular form. Sculpture of variably sized tubercles on elongate glabella, cheeks and preglabellar field, especially on anterior and lateral margins of cephalic border, long, slender, sub-vertical occipital spine in juveniles becomes a distinct, more posteriorly directed spine to occipital node in adults. Most of palpebral lobe located in front of L1; eye prominent, opposite cephalic mid-length, and palpebral lobe raised above fixigenae. Posterior border furrow short and deep, swinging abruptly anteriorly at genal angle to join lateral border furrow which is narrow and deep on posterior part of cheeks, becoming shallower and wider towards anterior margin of cephalon.

Free cheeks with prominent, long genal spine diverging abruptly from curve of border opposite posterior border furrow, so that gently curved genal spine is more posteriorly directed; sculpture of fine radially directed spines becomes coarser in mature stage.

Pygidial sculpture consists of granules on first three segments (in a single row mediadly and two rows distally on each segment); comparatively shallow furrows on short, anteriorly wide axis and pleural lobes of transverse pygidium.

Remarks. Pygidia of the two otarionid taxa represented at 'Erang' are not distinguished with certainty, possibly because of the deformation disguising any differences in shape or sculpture. However, the abundance of the pygidia assigned to C. windamereensis parallels that of cranidia and librigenae.

C. windamereensis n. sp. differs from C. dabrowni (Chatterton, 1971), from Zlichovian limestones near Good Hope, New South Wales, in the following: L1 are larger and more inflated; there are fewer tubercles in the sculpture on the cranium, and there is a distinct row of granules along the anterior margin; the eyes are slightly farther forward; the genal spine is slightly more laterally directed relative to the lateral margin of the cheek; and the pygidium is more transverse, with less distinct furrows and more numerous granules on the pleural lobes. C. windamereensis also lacks or has a much less distinct ridge that interrupts the border furrow just in front of the genal angle, a feature that is very distinct in C. dabrowni (called the genal trunk by Adrian and Chatterton, 1994).

C. windamereensis differs from C. listron (Chatterton et al., 1979), from Pragian limestones of the Garra Limestone near Wellington, New South Wales, in the following: L1 are larger and more inflated; the sculpture is not as reduced in size and amount anteriorly (note that this feature is not as prominent in specimens considered by Adrian and Chatterton to belong to listron); the preglabellar area is more steeply inclined, and the genal spine is finer. The cranidia of Cyphaspis sp. A of Chatterton and Wright (1986), from late Zlichovian to early Dalejan limestones at Mudgee, N.S.W., are very similar to those of C. windamereensis but that form has a much more distinct genal trunk interrupting the border furrow and appears to lack genal caecae on the free cheek.

Another similar Australian species of Cyphaspis is C. horani (Etheridge and Mitchell, 1894) which was redescribed by Chatterton (1971, pp. 95-96), who placed the species in Otarion (Otarion). Following a review of the Aulacopleuridae by Thomas...
and Owens (1978), Chatterton and Wright (1986) placed the species in Cyphaspis, a decision supported by Adrain and Chatterton (1996) in their review of the genus. C. horani is very similar to C. windamerenesis but has slightly more divergent anterior branches to the facial sutures, more densely arranged tubercles on the cranidium, and lacks the variation in size of the sculpture. The sculpture of fine tubercles below the eye on the librigena of C. horani is often arranged in orderly rows, as also seen in C. horani on AMF 36117 (not a type).

AMF 28354 (selected as lectotype of C. horani by Chatterton 1971) was not figured by Etheridge and Mitchell (1894, plate 7), and probably is not a syntype; for this reason (article 74(b).1 of the International Code of Zoological Nomenclature) and because of the intent of Recommendation 74(a) of the ICZN, it may not be a validly designated lectotype. However, it is a toptype collected by Mitchell, and certainly one of the best and most complete specimens of horani in the Australian Museum. It is an external mould, and far more complete than any of the specimens illustrated by Etheridge and Mitchell (1894), which were all internal moulds. Chatterton and Campbell (1980) figured some well-preserved material of C. horani as aff. horani, from Wenlock limestones and shales from near Canberra.

Of the other species of 'aulacopleurid' trilobites described from the Silurian-Devonian sequence of the Yass Basin, Cyphaspis bowingensis Mitchell, 1887 differs in the nature of the librigena and should be assigned to Otarion (see Adrain and Chatterton, 1994). Cyphaspis yassensis Etheridge and Mitchell, 1894 was reillustrated and reassigned to Prantia by Chatterton and Campbell (1980). Cyphaspis filmeri Mitchell, 1919 is based on a single inadequate type specimen (highly squashed, incomplete cranidial internal mould) and should probably be considered a nomen nudum. Cyphaspis rotunda Etheridge and Mitchell, 1894 is based on what appears to be a meraspid stage; until far more is known about the ontogeny of proetide trilobites from the Yass region, it appears impossible to identify adult specimens for this taxon which may be a small specimen of Prantia yassensis.

Family CALYMNIDAE Burmeister, 1843
Genus APOLYMENE Chatterton and Campbell, 1979

Type species. Apolympene hippocrepis Chatterton and Campbell, 1979, from Wenlock strata near Canberra, Australia; by original designation.

Apolympene sp.
Plate 9, figures 1-15; Plate 10, figures 1-9

?1979 Apolympene sp. A; Chatterton et al., p. 813, pl. 104, figs 28, 29; pl. 106, figs 10, 16-27.

Material. AMF 65421-65430, AMF 65789-65797, ?AMF 65798, AMF 88758.

Remarks. Characteristic features of this form include:

a dense, bimodal sculpture of tubercles; interpleural furrows that run almost uninterrupted from the margin to the axial furrow of the pygidium; the distal band of the pygidium is only very slightly depressed where the pygidium was overlapped by the cephalon during enrolment; and pleural furrows that are slightly deeper than the interpleural furrows. Juvenile pygidia have a long, subrectangular glabella and rather spinose pygidia.

Windamere specimens are very similar to material from the Garra Limestone near Wellington Caves described by Chatterton et al. (1979) as Apolympene sp. A. 'Erang' specimens only appear to differ in having a slightly less depressed distal band to the pygidium; however, the significance of this feature is uncertain as its strength increases ontogenetically in our material, and the illustrated Wellington material includes larger specimens.

We follow Chatterton et al. (1979) in not formally proposing a name for this species since we consider that most of the Australian Silurian and Devonian species of Apolympene are in need of revision; the proposal of new taxa might make that task more difficult. Holloway (1980) considered this genus a junior subjective synonym of Sthenarocalymene Siveter, 1976. Chatterton and Campbell (1980, p. 96) stated that Sthenarocalymene is best regarded as a subgenus of Gravicalymene as the anterior border is a variable character and the form of the glabella is like that of Gravicalymene.

Family ODONTOPLEURIDAE Burmeister, 1843
Subfamily ODONTOPLEURINAE Burmeister, 1843
Genus DUDLEYASPIS Prantl and Pribyl, 1949

Type species. Dudleyaspis quinquespinosa Lake, 1896, from the Wenlock of England; by original designation.

Remarks. Revisions and remarks relating to this genus were given by Thomas (1981) and Chatterton and Perry (1983).

Dudleyaspis sp. indet.
Plate 9, figures 16-27; Plate 10, figures 10-13

Material. AMF 65022, AMF 65038, AMF 65431-65439, AMF 65470-65473, AMF 65770, AMF 65798.

Remarks. This taxon is, in several respects, intermediate between Dudleyaspis bowingensis (Etheridge and Mitchell, 1897) from the Ludlow "Lower Trilobite Bed" (= Yarwood Siltstone Member of the Black Bog Shale; Link 1970) at Bowing, New South Wales, and Dudleyaspis campbelli (Chatterton, 1971) from Zlichovian limestones near Good Hope, New South Wales. Intermediate characters include: the sculpture on the cephalon (size and number of tubercles, and distinctness of caecae on cheeks); the number of pairs of border spines distal to the major pair on the pygidium; and the length/width ratios of the pygidia (lower in D. bowingensis and higher in D. campbelli). Juvenile pygidia in the two silicified taxa are very different, as are the hypostomes.
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EXPLANATIONS OF FIGURES

TEXT-FIG. 1. Fossil locality, present and former properties, and toposgraphic features.

TEXT-FIG. 2. Geological sketch map of the 'Erang' area.

PLATE 1. Sociophyllum wilsoni new species, holotype AMF 88794. 1, 2, transverse views, X4. 3-5, longitudinal views, all x4.

PLATE 2. 1-17, Skenidioides johnsoni new species. 1-2, holotype AMF 65440, brachial internal and lateral views, both x6. Paratypes: 3, AMF 65441, x6; 4-5, AMF 65443, external x6, and lateral x4, brachial views, ; 6, AMF 65444, posterior view of conjoined valves, x4; 7-8, AMF 65445, posterior and anterior views of conjoined valves, both x6; 9, AMF 65446, lateral view of conjoined valves, x6; 10, AMF 65447, pedicle exterior, x6, 11, AMF 65448, broken shell viewed from anterior, x6; 12-13, AMF 65449, internal views of pedicle valve, both x5; 14-15, AMF 65450,
same, both x4; 16, AMF 65451, pedicle interior, x5; 17, AMF 65452, pedicle interior, x5. 18-27, *Resserella rhodesi* new species. 18, AMF 65454, holotype, view of brachial valve of conjoined shell, x3. Paratypes: 19, AMF 65455, brachial interior, x2; 20, AMF 65456, brachial interior, x2; 21, AMF 65457, brachial interior, x2; 22, AMF 65458, brachial interior, x2; 23, AMF 65465, brachial interior, x2; 24, AMF 65464, broken shell, posterolateral view, x3; 25, AMF 65466, pedicle interior, x2; 26, AMF 65467, pedicle interior, x3; 27, AMF 65469, pedicle interior, x3.

**Plate 3.** 1-9, *Skendioides johnsoni* new species, scanning electron micrographs of paratypes; scale bar = 0.5 mm: 1-2, AMF 65442, brachial interior and lateral view showing septum and fulcrum plates; 3-6, AMF 86769, internal brachial views, showing crura, septum and fulcrum plates; 7, AMF 88767, pedicle interior, 8-9, AMF 88772, lateral views of broken, articulated shell showing high septum and long crura. 10-11, *Salopina* sp. indet., scanning electron micrographs: 10, AMF 88774, brachial interior; 11, AMF 88776, brachial interior.

**Plate 4.** 1-12, *Aulacella? boucoti* new species. 1-3, holotype AMF 88782, brachial external view of conjoined shell, pedicle interior, brachial interior, all x4. Paratypes: 4-5, AMF 88783, pedicle and posterior views of conjoined shell, both x3; 6, AMF 88784, brachial interior, x4; 7, AMF 88785, brachial interior, x6; 8, AMF 88786, brachial interior, 1.5; 9, AMF 88787, pedicle interior, x2; 10, AMF 88788, pedicle interior, x2; 11, AMF 88789, pedicle interior, x2; 12, AMF 88788, pedicle interior, x3. 13-17, *Salopina* sp. indet. 13, AMF 79021, brachial interior, x; 14, AMF 79022, brachial exterior, x; 15, AMF 79023, pedicle interior, x; 16, AMF 79024, pedicle interior, x; 17, AMF 79025, pedicle interior. 18-19, *Douvilleina?* sp. 18, AMF 88781, pedicle interior, x4; 19, AMF 65409, pedicle interior, x2.

**Plate 5.** 1-15, *M. (Mesodouvillina) savagei* new species. 1, holotype AMF 65459, brachial interior, x3. Paratypes: 2, AMF 65460, brachial interior, x5; 3-4, AMF 65461, conjoined valves showing pedicle exterior x3, and lateral profile, x2; 5, AMF 65462, pedicle interior, x2.5; 6, AMF 65463, brachial interior, x4; 7, AMF 88777, conjoined valves in ventral view, x2.5; 8, AMF 88776, juvenile pedicle exterior showing protegulum; 9, AMF 88778, brachial valve interior x2; 10-11, AMF 88759, pedicle exterior and enlargement of sculpture; 12, AMF 88763, pedicle exterior; 13, AMF 88764, pedicle exterior; 14, AMF 88765, pedicle exterior; 15, AMF 88766, brachial interior.

**Plate 6.** 1-13, *Maurotarion erangensis* new species. 1, holotype AMF 65404, craniumid, x8. Paratypes: 2, AMF 65405, craniumid in dorsal and lateral view, x15; 3-4, AMF 65784, dorsal and posterolateral view, 5-6, AMF 65785, dorsal and lateral view, 7, AMF 65786, dorsal view, 8-9, 65787, dorsal and anterolateral view, 10, AMF 65788, pygidium; 11, AMF 65406, mature librigena, x10; 11-12, AMF 65407, librigena in dorsal and ventral view, both x8; 13, AMF 65408, librigena in ventral view, x6.

**Plate 7.** 1-17, *Cyphaspis windamensis* new species. 1-5, holotype AMF 65412 in dorsal, anterior, lateral, anterodorsal and anterolateral view, all x8. Paratypes: 6, AMF 65411, x10; 7, AMF 65413, x10; 8, AMF 65414, x6; 9-12, checks AMF 65415 x6, AMF 65416 x5, AMF 65417 x10, AMF 65419 x5, all in dorsal view; 13, AMF 65418, pygidium, x10; 14-15, AMF 65077, pygidium in dorsal and ventral view, both x12; 16, AMF 65420, pygidium, x8; 17, AMF 88750, x20.

**Plate 8.** 1-22, *Cyphaspis windamensis* new species, scanning electron micrographs of paratypes: cranidia 1-2, AMF 65774, lateral views; 3-4, AMF 65775, anterolateral view and dorsal view; 5-6, AMF 65776, anterolateral and lateral views; 7 AMF 65772; 8, AMF 65771; 9, AMF 65778; 10, AMF 65773; 11, AMF 65779, librigena; 12, AMF 65780, librigena; 13, AMF 65781, librigena; 14, AMF 65782, thoracic segment; 15, AMF 65783, librigena; pygidia 16, AMF 88753; 17, AMF 88756; 18, AMF 88754; 19, AMF 88755; 20, AMF 88757; 21, AMF 88751; 22, AMF 88752.

**Plate 9.** 1-15, *Apocalymene* sp. 1-3, AMF 65421, dorsal view x5, posterodorsal view x4, and lateral view x4; 4-5, AMF 65422, dorsal view x3, and anterior view x4; 6-7, AMF 65424, dorsal and ventral views of librigena, both x7; 8, AMF 65423, librigena, x5; 9, AMF 65425, hypostoma, x15; 10, AMF 65427, rostral plate, x12; 11, AMF 65426, rostral plate showing outer surface, x9; 12, AMF 65428, tip of thoracic segment, x8; 13, AMF 65429, pygidium, x8; 14, 15(?), AMF 65430, dorsal view x15, and ventral view of juvenile pygidium x12. 16-27, *Dudleyaspis* sp. indet. 16, AMF 65432, cephalon x15; 17, AMF 65433, cephalon x7; 18, AMF 65434, cephalon x10; 19, AMF 65435, cephalon x12; 20, AMF 65436, librigena x7; 21, AMF 65437, free cheek x15; 22, AMF 65437, tip of thoracic segment x10; 23, AMF 65438, pygidium x10; 24, AMF 65439, pygidium x6; 25, AMF 65472, pygidium x15; 26, AMF 65471, pygidium x12; 27, AMF 65470, pygidium x6; 29, AMF 65431, pygidium.

**Plate 10.** Scanning electron micrographs. 1-9, *Apocalymene* sp. 1, AMF 65790; 2, AMF 65798; 3, AMF 65791; 4-5, AMF 65795, rostral plate, external and internal views; 6, AMF 65793; 7, AMF 65794; 8, AMF 65797; 9, AMF 88758. 10-13, *Dudleyaspis* sp. indet. 10, AMF 65770; 11, AMF 65798; 12, AMF 65038; 13, AMF 65022. 14-15, 2 x. *aulacopleurid.* 14, hypostome, AMF 65792; 15, juvenile thorax plus pygidium, AMF 65796.