Tecton relationship between the Ordovician sedimentary and volcanic successions of the Lachlan fold belt, central

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TECTONIC RELATIONSHIP BETWEEN THE ORDOVICIAN SEDIMENTARY AND VOLCANIC SUCCESSIONS OF THE LACHLAN FOLD BELT, CENTRAL/SOUTHERN TABLELANDS, NEW SOUTH WALES.

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

DOCTOR OF PHILOSOPHY

from

UNIVERSITY OF WOLLONGONG

by

SUZANNE ISOBEL MURRAY BSC (HONS)

School of Geosciences
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View looking south over Lake Oberon from the Goulburn Road west of Oberon. The Vulcan Fault separating the two Ordovician lithological provinces occurs to the east and traverses the hill to the left. Tors of Oberon Granite can be seen outcropping around Lake Oberon with the hornfels ridge capped by Tertiary Basalt in the centre right of the photograph. Racecourse Hill on the left of the photograph is formed by a resistant Tertiary Basalt flow overlying the Oberon granite.
ABSTRACT

Ordovician rocks of the Oberon-Taralga region were deposited in a deep marine environment in the forearc basin outboard of a long-lived Ordovician oceanic island arc and now form basement rocks to the Hill End Zone. The succession consists of two contrasting provinces: the Molong Volcanic Province characterised by mafic-intermediate volcanic and volcaniclastic rocks with associated deep-marine chert and the Quartzose Sedimentary Province composed of monotonous quartz-rich turbidites and chert overlain by graptolitic black shale.

The Quartzose Sedimentary Province is represented in the region by rocks of the Adaminaby and Bendoc Groups. The Adaminaby Group is subdivided into three conformable formations: the undifferentiated Adaminaby Group, the Numeralla Chert and the Bumballa Formation. The undifferentiated Adaminaby Group comprises mainly thick-bedded quartz rich turbidites of Middle Ordovician age, has a minimum thickness of 1000 m, with beds displaying Bouma sequences and ranging from 60 to 100 cm thick. The Numeralla Chert is composed of thin-bedded (2-10 cm) chert bands commonly separated by thin mudstone interbeds, and has a maximum thickness of 40 m in the Oberon-Taralga region. Based on conodonts, it is of late Darriwilian to earliest Gisbornian age and contemporaneous with the Sunlight Creek Formation mapped farther south in Victoria (VandenBerg et al. 1992). The Bumballa Formation conformably overlies both the Numeralla Chert and the Adaminaby Group, where it is composed of thin-bedded (5-30 cm) unfossiliferous quartz-rich turbidites. The Bendoc Group conformably overlies the Adaminaby Group, and in the Oberon-Taralga region is composed of only one formation. The Warbisco Shale comprises siliceous black shales and slates with minor interbedded quartz sandstone. The units of the Adaminaby and Bendoc Groups represent a fining upward sequence of craton-derived turbidites, together with pelagic chert and black shale, deposited on a terrigenous deep-marine fan outboard from Gondwana. The regional stratigraphy recognised from the northern part of the Eastern Belt is similar to the regional stratigraphy of the Quartzose Sedimentary Province previously described from farther south.
The Triangle Formation and the Rockley Volcanics represent rocks of the Molong Volcanic Province within the Oberon-Taralga region. They were deposited in a deep-marine basin adjacent to the volcanic arc and received pulses of volcaniclastic detritus from the arc interspersed with pelagic sedimentation. The Triangle Formation is thought to span the entire Ordovician to Early Silurian and is subdivided into four members including: the Budhang Chert Member, Gidyen Volcaniclastic Member, Mozart Chert Member and undifferentiated Late Ordovician volcaniclastics. The basal Budhang Chert Member consists of tightly folded, dark coloured chert and mudstone of early Bendigonian age. A large stratigraphic hiatus representing approximately 25 Ma occurs between the Budhang Chert Member and the overlying Gidyen Volcaniclastic Member. The Gidyen Volcaniclastic Member consists predominantly of crystal-rich volcaniclastic sandstone with minor conglomerate, and was deposited as either turbidites or debris flows. The Mozart Chert Member conformably overlies the Gidyen Volcaniclastic Member, comprises moderately bedded chert bands with interbedded mudstone and sandstone beds and contains late Darriwilian to earliest Gisbornian conodonts making it contiguous with the Numeralla Chert recognised from the Quartzose Sedimentary Province. Conformably above the Mozart Chert Member a second pulse of undifferentiated Late Ordovician crystal-rich mafic volcaniclastic detritus occurs. Within the Oberon area, the Rockley Volcanics overlie the undifferentiated Late Ordovician Triangle Formation, however, in some places relationships are not clear and the two units may interdigitate. The Rockley Volcanics are composed of brecciated basaltic pyroxene porphyritic lava and volcaniclastic conglomerate, sandstone and dark mudstone/shale. No lavas were noted from the Oberon area although they are known to occur in the Rockley district to the west. The major difference between the Rockley Volcanics and the volcaniclastics of the Triangle Formation is the presence of primary hornblende in the Triangle Formation and the more proximal nature of the Rockley Volcanics.

During the Middle Ordovician, rocks of the two lithological provinces were deposited distant from each other. The Quartzose Sedimentary Province is considered to have been located further outboard of the arc than the deep-marine Molong Volcanic Province succession. However, the two provinces
were juxtaposed by the middle Late Ordovician, when a rapid increase in volcanlastic sedimentation due to collision of the subduction zone with an oceanic plateau resulted in the expansion of the volcanlastic apron over units of the Quartzose Sedimentary Province.

Petrographic, geochemical and isotopic analyses of the volcanic arc succession within the Molong Volcanic Province show that these rocks have oceanic island arc characteristics and were derived from mantle sources with little or no crustal contamination. Volcanism is interpreted as having occurred synchronously with subduction, with the arc region floored by oceanic crust; however, based on the shoshonitic characteristics of the magmatism, the arc is considered to have undergone a major perturbation during the Darriwilian. Based on the occurrence of a large Bendigonian to Darriwilian stratigraphic hiatus across the entire arc, followed by significant uplift with discontinuous limestone deposition and the shoshonitic character of the volcanic rocks during the early Late Ordovician, the arc is interpreted to have been rifted and broken up during the Late Ordovician. This has been related to the onset of the Benambran deformation during the Darriwilian, possibly following the collision of an oceanic plateau or ridge with the Ordovician subduction trench. Onset of the Benambran Orogeny also resulted in uplift and is interpreted as the factor responsible for cutting off the supply of quartzose sediment to the turbidite fan during the late Middle Ordovician resulting in widespread deposition of chert and black shale, and providing the source of renewed quartz turbidite sedimentation during the latest Ordovician and Early Silurian synchronous with the cessation of arc volcanism.

Within the Oberon-Taralga region the Ordovician rocks of both provinces have been affected by three deformation events. D1 was a weak deformation resulting in early pre-cleavage folds. The timing of this event is poorly constrained; however, it may be related to Late Ordovician “type-Benambran” deformation, corresponding to the formation of an angular unconformity in the Oberon and Thompson Creek areas following deposition of the ?Lower Silurian Bald Ridge Formation. The second deformation event (D2) is the strongest event recognised in the study area. It is characterised by north
to northeast trending folds, an associated axial planar cleavage (S2) and related faults. Timing constraints on the age of this deformation are poor, and evidence from the different study areas may indicate it is diachronous across the Oberon-Taralga region. D3 was also a weak deformation event, caused by north-south shortening and resulted in gentle folding of D1 and D2 structures throughout the region. It is post-dated by intrusion of the Carboniferous granites in the Oberon area. The intrusion of granites associated with the Bathurst Batholith during the Middle Carboniferous marks the termination of tectonic development within the northeastern Lachlan Fold Belt.
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CHAPTER 1

INTRODUCTION

1.1 INTRODUCTION
This study concentrates on Ordovician rocks of the Lachlan Fold Belt occurring between Bathurst and Goulburn in the Central and Southern Tablelands of New South Wales (Figures 1.1, 1.2). The Ordovician rocks of the Lachlan Fold Belt are divided into a quartz turbidite succession (Quartzose Sedimentary Province) and a mafic volcanic belt (Molong Volcanic Province, VandenBerg & Stewart 1992). Variably deformed rocks of both lithological provinces form basement in the area of investigation, where they have undergone regional metamorphism up to greenschist facies. Rocks of the Molong Volcanic Province occur in the northern half of the area mapped, whereas rocks belonging to the Quartzose Sedimentary Province occur throughout the whole area (Figures 1.3, 1.4).

1.1.2 Objectives
The major objectives of this thesis are listed below.

- To undertake detailed mapping of the stratigraphy and structure of Ordovician units at 1:12 500 scale or less, in selected areas of the Oberon-Taralga region in the Central to Southern Tablelands of New South Wales (Figure 1.4).
- To determine the nature of the contact and/or transition between the Ordovician quartz turbidite succession (Quartzose Sedimentary Province) and the Ordovician mafic volcanic belt (Molong Volcanic Province) in the Oberon-Taralga region.
- To geochemically characterise the Ordovician mafic volcanic rocks of the Molong Volcanic Province in an attempt to resolve the island arc or intraplate tectonic affinity of this succession.
- To synthesise all the collected structural data and correlate it with data already documented from the area to construct a structural history for the Ordovician rocks of the Oberon-Taralga region.
- To synthesise all collected data for reconstruction and analysis of the original basin of deposition of the eastern Lachlan Fold Belt.
Figure 1.1 Location of the Lachlan Fold Belt and other Palaeozoic fold belts making up the "Tasmanides" in eastern Australia. The Delamerian Fold Belt and Georgetown and Coen Inliers are also shown.
1.2 LACHLAN FOLD BELT - GEOLOGY AND TECTONIC SETTING

The Lachlan Fold Belt has an areal extent of approximately 300 000 km\(^2\) and lies in the southeastern part of the Australian continent (Figure 1.1). It forms the southern part of the north-south trending Palaeozoic to early Mesozoic Tasman Orogenic Belt, which extends along eastern Australia and includes part of the Delamerian Fold Belt, the Lachlan, Thomson and New England Fold Belts, and part of Tasmania. The Lachlan Fold Belt is bordered to the east by the Tasman Sea and to the northeast by the Late Palaeozoic/Mesozoic Sydney-Bowen Basin. To the north, it is unconformably overlain by the Mesozoic Great Australian Basin and to the west by the Mesozoic-Cainozoic Murray Basin. In Victoria, the southern part of the Lachlan Fold Belt is buried beneath the Otway, Gippsland and Bass Basins adjacent to Bass Strait (Figure 1.2).

Rocks within the Lachlan Fold Belt comprise weakly to intensely deformed, Cambrian to Early Carboniferous deep-marine to terrestrial sedimentary and volcanic rocks, which are intruded by voluminous granitic plutons and exhibit low to moderate metamorphic grades (Gilligan & Scheibner 1979; Cas 1983). These successions record the gradual conversion of an Early Palaeozoic deep-marine continental margin formed outboard of eastern Gondwana, to part of the Gondwanan continental landmass by the end of the Palaeozoic. These rocks are divided into three main lithological associations: (1) Cambrian mafic volcanic and marine sedimentary rocks of oceanic affinity; (2) an Ordovician to Early Silurian deep-marine succession with interspersed mafic volcanic centres, and (3) a complex Late Silurian to early Carboniferous association consisting of clastic and limestone deposits, volcanics and intrusions (Cas 1983; Coney et al. 1990). Based on variations in the Ordovician stratigraphy and differences in the orientation and timing of deformation, the Lachlan Fold Belt has been divided into two belts: the Eastern and Western Belts (Figure 1.2, Powell et al. 1990). The structural zone subdivisions used in Figure 1.2, and referred to in the text are based on Fergusson et al. (1986), VandenBerg and Stewart (1992), Glen (1992), Scheibner and Basden (1998) and Fergusson (1998a).
Figure 1.2 Location and extent of the Eastern and Western Belts of the Lachlan Fold Belt. The main structural zones and boundary faults within these belts are also shown. Structural zones modified after Glen. (1992).
1.2.1 Cambrian

Cambrian rocks are the oldest rocks in the Lachlan Fold Belt. Their exposure is confined to four major north-south trending elongate fault zones (from west to east, the Stavely, Avoca, Heathcote and Mount Useful Fault Zones), with other small outcrops on Phillip Island, at Barrabool Hills, and at Cape Liptrap (Figures 1.2, 1.3; Cas 1983; Crawford et al. 1984; Crawford 1988; Henry & Birch 1992; Morand et al. 1995). Mafic volcanic breccia associated with Middle to Late Cambrian limestone are also known from the south coast of New South Wales, where they have been interpreted as an ancient seamount (Bishoff & Prendergast 1987).

A stratigraphy has been recognised in the greenstones of the Heathcote and Mount Wellington greenstone belts. It comprises basal andesitic arc detritus occurring below boninitic lavas, which are in turn overlain and intruded by slighter younger low-K tholeiites and co-magmatic shallow intrusive dolerite and gabbro (Crawford & Keays 1987). Although unfossiliferous, the greenstone belts have been assigned an Early Cambrian age based on stratigraphic relationships with overlying Middle to Late Cambrian and Early Ordovician fossiliferous units (Kennedy 1971; Jell 1985; Crawford 1988; VandenBerg 1992; Englebrechtson & Brock 1996; Fergusson 1998a).

The Late Cambrian to earliest Ordovician Delamerian Orogeny caused compressional deformation, regional metamorphism and igneous intrusion in western Victoria, western New South Wales (Wonominta Block) and eastern South Australia. This event has been related to the closure of a marginal sea between an outboard volcanic arc and the Gondwanan continent resulting in an island arc - continental margin collision (Gray & Webb 1995). Convective removal of lithospheric mantle followed the Delamerian orogenic thickening and resulted in both the rapid exhumation of some 10-15 km of thickened orogenic crust and termination of convergent deformation (Turner et al. 1996). A transcontinental mountain belt was produced as a result of the Delamerian Orogeny. Erosion of the Delamerian mountains during the Ordovician fed quartz-rich sediment offshore as turbidites, to the
Figure 1.3 Diagram shows the location of the two Ordovician-Early Silurian lithological provinces in the Lachlan Fold Belt and localities mentioned in the text.
1.2.2 Ordovician

Ordovician rocks dominate much of the exposure across the Lachlan Fold Belt, and these rocks are characterised by two fundamentally different lithological associations: the Quartzose Sedimentary Province and the Molong Volcanic Province (Packham 1969; VandenBerg & Stewart 1992; Figure 1.3).

QUARTZOSE SEDIMENTARY PROVINCE

The Quartzose Sedimentary Province (Quartzose Flysch Province of VandenBerg & Stewart 1992) is an extensive deep-marine sedimentary province. It is characterised by a basal Lower to Middle Ordovician quartz-turbidite succession of at least 3000 m of quartz-rich turbidites with interbedded chert and black shale horizons (Figure 1.2, VandenBerg & Stewart 1992). In the Eastern Belt, this turbidite succession is overlain by an extensive black shale horizon previously thought to have been interbedded with the turbidites (VandenBerg & Stewart 1992).

The basal quartz turbidite succession extends across the width of the Lachlan Fold Belt for 600 km from the Avoca Fault to the east coast of New South Wales, and for a maximum of 800 km in a north-south direction. Undeformed, these strata would have extended at least 1200 km in an east-west direction, dimensions comparable to the present-day Bengal-Nicobar submarine fan complex (Cas et al. 1980; Powell 1984; Fergusson & Coney 1992a). Strontium isotopic ratios of the quartz sandstone suggests probable derivation from either the Cambrian sedimentary rocks of the Kanmantoo Group in South Australia or Cambrian igneous rocks in western Tasmania (Gray & Webb 1995). Derivation from the Kanmantoo Group is also supported by consistent palaeocurrent directions from the west (Powell 1984; Fergusson et al. 1989; VandenBerg & Stewart 1992; Jones et al. 1993), variations in sandstone composition (Pound et al. 1994; Fergusson & Tye 1999), Sm-Nd isotopic evidence (Turner...
Figure 1.4 Map with the distribution of Molong Volcanic Province and Quartzose Sedimentary Province rocks in the Oberon-Taralga region, and the location of the five areas discussed in the text (see enclosures 1-14): (1) Oberon area, (2) Thompson Creek area, (3) Silent-Oakey Creeks area, (4) Golspie area, (5) Crookwell River area. Also shown are the locations of Enclosures 2 and 3 in the Oberon Area.
et al. 1993), $^{40}\text{Ar}/^{39}\text{Ar}$ detrital mica ages (Turner et al. 1996) and detrital zircon ages (Williams et al. 1991; Ireland et al. 1998).

East of the Mount Useful Fault Zone, the Lachlan Fold Belt is divided into eight zones (Figure 1.2). The lowermost contact of the turbidite succession has only been observed in the Tabberabbera Zone, where it conformably overlies the Lancefieldian Howqua Chert (Fergusson 1998a). The base of the Adaminaby Group here is La2, slightly younger than that occurring further west in the Bendigo-Ballarat Zone. Apart from minor variations in chert and black shale facies occurring during the Bendigonian and Chewtonian, there are no other major lateral facies variations within the turbidite succession across the Zones of the Eastern Belt.

A major lateral difference does exist between the Western and Eastern Belts. This is the cessation of most quartz turbidite deposition during late Middle Ordovician (Darriwilian-Da2) in the Eastern Belt, whereas turbidite deposition continued through to the late Late Ordovician (Bolindian) in the Western Belt (VandenBerg et al. 2000). Provenance studies from both the Eastern and Western Belts indicate there is a ‘maturing’ of the turbidite sandstone composition up-section throughout the Early and Middle Ordovician, with lithic fragments decreasing and an increase in the relative proportion of quartz between the Lancefieldian and the Darriwilian (Pound et al. 1994; Fergusson & Tye 1999).

In the Eastern Belt, the Adaminaby Group is overlain by a widespread Late Ordovician graptolitic black shale (Warbisco Shale of VandenBerg et al. 1992). In some places; however, latest Darriwilian chert (Numeralla Chert of Lewis et al. 1994) and early Gisbornian thin-bedded grey shale (Sunlight Creek Formation of Orth et al. 1995) and thin-bedded turbidites (stripy facies of Fergusson & VandenBerg 1990; Chakola Formation of Lewis et al. 1994) occur between the Adaminaby Group and the Warbisco Shale.
MOLONG VOLCANIC PROVINCE

The Molong Volcanic Province is restricted to the Cowra-Canberra-Buchan, Molong and Capertee Zones of the Eastern Belt and comprises a northerly trending suite of ultramafic to intermediate volcanic and volcaniclastic rocks with fringing carbonate platforms and associated mafic intrusions (Figures 1.2, 1.3; Packham 1969, 1987; Powell 1984; Wyborn 1992a, 1992b; Percival 1999a; Percival et al. 1999). Based on the present-day spatial distribution of the rocks making up the Molong Volcanic Province, it has been subdivided into smaller volcanic belts (see Webby 1976; Wyborn 1992a; Glen et al. 1998). From west to east, the main subdivisions are Junee-Narromine, Molong, Rockley-Gulgong and Kiandra Volcanic Belts (Figure 1.3; Glen et al. 1998).

Early Ordovician rocks are only known from the Molong and Junee-Narromine Volcanic Belts within the Molong Volcanic Province. They indicate active volcanism occurred in the Eastern Belt of the Lachlan Fold Belt during this time and that in the Molong Volcanic Belt at least, the submarine volcanoes grew to shallow water depths (Packham 1969; Krynen et al. 1990; Percival et al. 1999; Butera et al. 2001). Castlemainian-Yapeenian rocks are not known from the Molong Volcanic Province. This hiatus has been attributed to either a period of non-deposition or the later removal of shallow water deposits by erosion (Percival et al. 1999). There was a major increase in volcanic activity in all four of the volcanic belts during the late Darriwilian, and shallow-marine conditions prevailed in some areas (Pickett 1984; Percival et al. 1999). By the early to middle Late Ordovician (Gisbornian to Eastonian) widespread accumulation of shallow marine limestone occurred in the Molong and northern part of the Rockley-Gulgong Volcanic Belts (Zhen & Webby 1995; Percival et al. 1999; Percival 1999a).

RELATIONSHIP BETWEEN THE PROVINCES

The relationship between the two Ordovician lithological provinces is problematical. Contacts between the two successions are commonly faulted; some conformable contacts have been documented but the contacts tend to be sharp with rare evidence of interdigation (Fergusson &
Colquhoun 1996). Recent mapping in the Bathurst area indicates that Molong Volcanic Province rocks in the Rockley-Gulgong Volcanic Belt conformably overlie the Adaminaby Group, but in most areas these relationships are only inferred (Pogson & Watkins 1998). More work is required to document and establish the nature of this transition in detail because no ash or volcanic detritus has been found in the quartz-turbidite units.

PALAEOGEOGRAPHY AND TECTONIC SETTING THROUGHOUT THE ORDOVICIAN

The palaeogeography of Ordovician rocks in the Wonominta Block of western New South Wales consisted of a continental shoreline and shelf (Gnalta Shelf of Webby 1976), which continued into western Victoria and Tasmania (Tasmanian Shelf of Webby 1976). Ordovician mafic volcanics of the Molong Volcanic Province represented a north-northwest trending, eastward facing volcanic island chain floored by oceanic crust and separated from the Gondwanan continent by a marginal sea (Wagga Trough of Webby 1976; Wagga Marginal Sea of Powell 1983, 1984). This arrangement is similar to the present day Andaman-Nicobar system of the northeast Indian Ocean (Cas et al. 1980; Powell 1983, 1984; Fergusson & Coney 1992a).

This palaeogeography is interpreted tectonically as a convergent margin setting with the following features: a west-dipping Mariana B-type subduction zone causing the development of a volcanic arc (Ordovician mafic volcanic rocks) with a marginal sea formed by backarc spreading, separating the arc from the Gondwanan continent (Powell 1983, 1984; Packham 1987; Scheibner 1987, Scheibner & Basden 1998). Based on the regional stratigraphy identified within the Quartzose Sedimentary Province, VandenBerg and Stewart (1992) argued for a passive margin tectonic setting especially during the Late Ordovician black shale deposition, with the mafic volcanic rocks generated by plume related hot spot volcanism. Wyborn (1992a) emphasised the shoshonitic composition of the mafic volcanics and related the source of the mafic magmatism to melting of a metasomatized lithosphere, triggered by overturning within an intraplate tectonic setting and underlain by continental crust. More recently, further geochemical analysis of the mafic volcanics have indicated they are subduction
related with isotope values indicating derivation from mantle-derived melts with little sediment input (Carr et al. 1995; Glen et al. 1998).

Previous attempts at palaeogeographic and tectonic models have been frustrated by controversy over the nature of the basement to the fold belt. A Cambrian oceanic substrate similar to the Cambrian greenstones in Victoria has been advocated (Crawford & Keays 1978; Crook 1980; Cas 1983; Powell 1983, 1984). Conversely, a Proterozoic (and more recently, Cambrian) continental substrate has been inferred based on granite geochemistry and the restite model (Chappell et al. 1988; Wybom 1992a; Chappell 1994). However, new granite geochemical models are now emerging, which indicate that the voluminous Silurian and Devonian granites have been sourced by the mixing of two or three melt components (Rossiter & Gray 1996; Collins 1996; Keay et al. 1997). These components are derived from a basaltic oceanic substrate, similar to the Victorian greenstones, lower crustal melts of Ordovician quartzose turbidites and the mantle (Collins 1998).

Based on these models, there now seems to be a general consensus that oceanic crust forms basement to much of the Lachlan Fold Belt. Some investigators argued that continental margin areas of the Gondwanan craton were rifted away during the Late Proterozoic and these form microcontinents under parts of the Lachlan Fold Belt (Scheibner & Basden 1998). One area that appears likely is the Melbourne Zone in Victoria, where magnetic lineations extending from Proterozoic rocks in Tasmania can be traced across Bass Strait to Victoria (VandenBerg et al. 2000).

1.2.3 Late Ordovician – late Early Silurian (Benambran Orogeny)
The Benambran Orogeny began during the Late Ordovician and was concentrated in the 200 km wide, north-south trending Wagga-Omeo Metamorphic Zone and the Girilambone Group to the north (Fergusson & Coney 1992b). The Wagga Marginal Basin was intruded by granites and locally metamorphosed to greenschist-amphibolite facies to form the Wagga-Omeo Metamorphic Zone. It remained as an uplifted landmass for most of the Silurian.
The Benambran Orogeny has been linked to the facies change from Late Ordovician black shale to overlying Early Silurian quartzose clastics (VandenBerg et al. 2000). Early Silurian deep-marine clastics and limestone olistoliths were probably derived from the uplifted Wagga-Omeo Metamorphic Zone to the west (Crook et al. 1973; Powell 1983; VandenBerg 1999).

Within the Wagga-Omeo Metamorphic Zone and in the Eastern Belt, the Benambran Orogeny is characterised by east-west trending, tight, upright folds, devoid of cleavage (Crook & Powell 1976; Powell et al. 1978; Powell 1983, 1984; Fergusson 1987; Morand 1988; Stuart-Smith et al. 1992; Glen 1992). Because these folds are strongly oblique to the main north-south structural trends, they have been related to dextral transpression (Fergusson & Coney 1992b). The Benambran Orogeny caused widespread and complex deformation, and has been attributed to a change in plate convergence from decoupled Marianas-type to coupled Chilean-type subduction (Powell 1983, 1984). More recently, $^{40}$Ar/$^{39}$Ar data from fault zones across the Lachlan Fold Belt have been used to provide constraints on the specific areas and timing of deformation within the Lachlan Fold Belt, (Gray 1997; Gray & Foster 1998; Foster et al. 1999). The model of Gray and others, based on the apparent complexity of deformation patterns, is interpreted as diachronous deformation occurring within multiple subduction complexes.

Deformation events attributed to the Benambran Orogeny continued episodically on a local scale until the late Early Silurian (Collins & Hobbs 2001). Based on unconformities between Early and Middle to Late Silurian units in the Yass, Canberra and Tantangara areas, Crook et al. (1973) proposed the Quidongan Orogeny. This is now considered to be the latter part of the Benambran Orogeny (VandenBerg 1999). By the early Late Silurian (Wenlock) the Tumut Trough had begun to open (Powell 1983) and VandenBerg (1999) reported rhyolitic volcanics accumulating in rift-like basins of Wenlock-Ludlow age, heralding the onset of a transtensional regime.
1.2.4 Late Silurian – Middle Devonian
The Late Silurian to late Devonian is characterised by the development of horsts and grabens and dominated by felsic volcanism with associated plutonism. The oldest silicic volcanics are known from the Tumut Trough and have a SHRIMP age of 428 Ma (Stuart-Smith et al. 1992), as are the earliest granites dated from north of Omeo (Keay et al. 1999). The presence of granitic clasts in basins in eastern Victoria indicate unroofing of Early Silurian granitoids of the Wagga-Omeo Metamorphic Complex by the late Early Silurian (VandenBerg 1999). The timing and style of deformation varied across the Lachlan Fold Belt, but was widespread with the exception of several areas (see detailed summary by Fergusson & Coney 1992b).

A number of variably complex extensional to compressional tectonic settings have been proposed for the Late-Silurian to Middle-Devonian including:

- 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 & Coney 1992b),
- a dextral transform plate margin setting similar to the Basin and Range Province of western North America (Cas 1983; Powell 1983, 1984), and
- double subduction zones, one on the eastern boundary of the Lachlan Fold Belt, the other between the Eastern and Western Belts. The one between the Eastern and Western Belts became a dextral transform fault in the Early Devonian and resulted in the juxtaposition of the Eastern and Western Belts by the Middle Devonian (Gray 1997; Foster et al. 1999).

1.2.5 Middle – Late Devonian (Tabberabberan Orogeny)
By the Middle to Late Devonian, much of the Lachlan Fold Belt had been uplifted and intensely deformed, with prevailing fluviatile and shallow-marine conditions (Powell 1984). The Tabberabberan Orogeny occurred during the Middle Devonian, with deformation most evident in the Melbourne Zone, where east-west compression formed upright northerly trending folds (Powell 1984;
Fergusson & Coney 1992b). In the Eastern Belt (including the Omeo Metamorphic Complex),
deformation occurred along a conjugate strike-slip fault system (Fergusson & Coney 1992b).

1.2.6 Carboniferous (Kanimblan Orogeny)
The fluviatile and shallow marine conditions continued into the Early Carboniferous, when the
orogenic history of the Lachlan Fold Belt was concluded by widespread deformation resulting from
the Kanimblan Orogeny (Powell 1984). Deformation was strongest in the northeastern Lachlan Fold
Belt, where it resulted in abundant upright folds with easterly tectonic transport (Powell 1984;
Fergusson & Coney 1992b). Throughout other parts of the Lachlan Fold Belt, Late Devonian to Early
Carboniferous rocks were folded, forming close to gentle upright synclines with bounding thrust
faults (Fergusson & Coney 1992b). 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 & Coney 1992b).

Post-deformational, I-type granitoids were subsequently emplaced along the eastern margin of the
Lachlan Fold Belt during the Middle Carboniferous (Gray 1997). A mild north-south compressive
deformation immediately post-dating intrusion of granites of the Bathurst Batholith has been
documented near Bathurst (Kosaka 1994).

1.3 MAPS, ORIENTATION DATA AND SAMPLES
The main geological and structural maps (Enclosures 1-14) are appended to the thesis. The maps were
digitally produced using SPHERISTAT 2.2, ARCVIEW 4.0 and CORELDRAW 8 using a geographic
base compiled from 1:25 000 and 1:50 000 topographic maps that cover the study area. Original
mapping was undertaken using 1:25 000 scale air photographs and topographic maps.

A list of samples collected for geochemistry, palaeontological work, and petrology (thin sections) are
listed along with grid references and lithological details in Appendix A. Samples are all held in the
collection of the School of Geosciences at the University of Wollongong.
CHAPTER 2
ORDOVICIAN TO EARLY SILURIAN STRATIGRAPHY
OF THE OBERON-TARALGA REGION

2.1 INTRODUCTION
Detailed geological mapping of Ordovician and Early Silurian rocks in several study areas in the Oberon-Taralga region has been undertaken (Figure 1.3). The study areas investigated include (1) the Oberon area, (2) the Thompson Creek area west of Burraga, (3) the Silent-Oakey Creeks area in the Abercrombie National Park, (4) the Golspie area including Burra Burra and Monkey Creeks northwest of Taralga, and (5) the Crookwell River area north to northwest of Crookwell (Figure 1.4). The Ordovician lithostratigraphic framework outlined in Chapter 1, consisting of the Quartzose Sedimentary Province and the Molong Volcanic Province (Figure 1.3), is used to describe the rocks. Firstly, descriptions of the stratigraphic units comprising each lithological province are given, and secondly, the stratigraphic relationship between the two provinces is addressed. Stratigraphic correlation between the different areas is based on both lithology and age control provided by sparse graptolite and conodont occurrences (see Chapter 3 and Appendix B). The regional stratigraphy developed for the Oberon-Taralga region is placed in a palaeogeographic context within the Eastern Belt of the Lachlan Fold Belt (Figure 1.2; see Glen 1992).

The stratigraphy of Ordovician rocks between Goulburn and Bathurst has been complicated by mapping stratigraphic units that occur in both lithological provinces. For example, the Triangle Group described by both Packham (1968, 1969) and Scheibner (1970, 1973) includes both Ordovician volcanioclastic and quartz turbidite units. The various units defined and used by previous investigators are summarised in Table 2.1. Many of local stratigraphic names used in the past are now abandoned in favour of a regionally defined nomenclature partly following Glen (1994) and Pogson and Watkins (1998).
2.2 ORDOVICIAN TO EARLY SILURIAN STRATIGRAPHY
QUARTZOSE SEDIMENTARY PROVINCE

The Adaminaby Group, Bendoc Group and the Bald Ridge Formation (new name; Figure 2.1) represent the Quartzose Sedimentary Province in the study area. The bulk of the Adaminaby Group is not further subdivided due to the monotonous nature of the quartz turbidites. The Numeralla Chert and Bumballa Formation occur in the upper Adaminaby Group and are overlain by the Warbisco Shale of the Bendoc Group. In the Oberon-Taralga region the units of the Quartzose Sedimentary Province span the Middle to late Late Ordovician and probably continue into the Early Silurian (Llandovery/Wenlock), although supporting fossil evidence for the latter is lacking. These units comprise siliceous, deep marine sedimentary rocks that outcrop in all the study areas mapped, although their distribution is limited to the eastern and southern parts of the Oberon area (see Enclosures 1 to 7).

2.2.1 Adaminaby Group

The name "Adaminaby beds" was first used informally by Adamson (1951) to describe the metasedimentary rocks around the old town of Adaminaby (Figure 1.3), which was partially flooded by Lake Eucumbene. Fairbridge (1953) formally published the name and used it to describe a turbiditic sandstone-slate unit in the Snowy Mountains of New South Wales. The Adaminaby beds were redefined and upgraded to group status following recognition of imbricate thrusting within the El Paso reference section (Figure 1.3; see Glen et al. 1990). The redefined Adaminaby Group comprises an Early to Middle Ordovician quartz-rich turbidite-chert submarine fan deposit that has since been described from many areas of the Lachlan Fold Belt (VandenBerg & Stewart 1992; Glen 1994; Fergusson & Colquhoun 1996; Colquhoun et al. 1997; Raymond et al. 1997; Fergusson 1998a).

In the Bega region (Figure 1.3), Glen (1994) divided the Adaminaby Group into three unequal parts comprising the basal “undifferentiated quartz turbidites”, overlain by the distinctive and fossiliferous late Middle Ordovician Numeralla Chert. The Chakola Formation, which also consists of quartz-rich turbidites, occurs above the Numeralla Chert east of the Murrumbidgee Fault (Figure 1.3). Pogson and
Table 2.1 Ordovician to Middle Silurian stratigraphic nomenclature used on the Bathurst and Goulburn 1:250 000 sheets in the Oberon-Taralga region, as well as that outlined by Glen (1994) from the Bega/Mallacoota 1:250 000 sheet.
Figure 2.1 Stratigraphic columns showing the stratigraphy recognised from each of the five study areas within the Oberon-Taralga region. All age data is based on fossil ages. ◆ Indicates a Bendigonian (Be1/2) age based on conodonts, * indicates a Darriwilian to Gisbornian (Da3/Gi1) age based on conodonts, ■ indicates an Eastonian to Bolindian graptolite age (Ea4/Bol1). (a) Early Silurian Quartzose Sedimentary Province stratigraphy from the Thompson Creek area. (b) - (c) Molong Volcanic Province stratigraphy. (b) Stratigraphic column from the Triangle Formation type section in Triangle Creek (from Fowler & Iwata 1995). Units in this column have been patterned based on their lithology for comparison with (c). (c) The stratigraphy recognised from Oberon West. (d) Stratigraphic column from Oberon East shows presence of the units from the two lithological provinces during the Late Ordovician. (e) - (g) Columns show stratigraphy of the Quartzose Sedimentary Province units in the (e) Crookwell River area, (f) Golspie area, and (g) Silent-Oakey Creeks area.
Watkins (1998, p. 19) proposed that the Adaminaby Group "should include all Ordovician turbidites in the eastern part of the Lachlan Fold Belt". In this study, the Adaminaby Group includes the basal undifferentiated quartz-rich turbidites of Middle Ordovician age, the overlying late Middle Ordovician Numeralla Chert, and the early Late Ordovician Bumballa Formation (new name).

2.2.2 Undifferentiated Adaminaby Group

**DISTRIBUTION**
Undifferentiated quartz turbidites of the Adaminaby Group outcrop in all the areas mapped in detail within the Oberon-Taralga region except the Thompson Creek area. Exposures exhibiting the best preservation of Ordovician rocks occur in the Golspie area along Monkey Creek upstream of Burra Burra Creek (Enclosure 6A) and along Silent Creek in the Oakey-Silent Creeks area (Enclosure 5). The rocks in these study areas were previously mapped as the Burra Burra Creek Formation (Scheibner 1970, 1973).

**LITHOLOGY**
The undifferentiated Adaminaby Group is lithologically similar across the entire Oberon-Taralga region, comprising alternating beds of quartz-rich sandstone, siltstone and shale. Layers have a suite of sedimentary structures characteristic of the Bouma sequence and are interpreted as having formed by turbidite deposition. Sandstone beds are commonly light- to blue-grey in colour, weathering to an orange-brown with iron oxide staining. These beds are normally between 0.5 and 0.8 m thick, but occasional massive sandstone beds of up to 1.2 m occur. The sandstone beds have sharp erosional basal contacts and are overlain by either massive or planar-bedded medium-grained sandstone (Bouma divisions A and B), that commonly fines upward into convolute-bedded fine-grained sandstone (Bouma division C), plane-laminated fine sandstone to siltstone (Bouma division D) and these, in turn, grade into laminated purple, chocolate and/or blue-grey coloured mudstone or shale (Bouma division E). Common Bouma sequences displayed along Monkey and Burra Burra Creeks include ABCDE, BCDE, ACD, ABC and AC (Figures 2.2a, 2.3).
Sandstone from this unit is predominantly composed of poorly sorted subangular to subrounded quartz grains with subordinate muscovite commonly altered to phengite, with minor albitised plagioclase and metamorphic biotite. Elongate intraformational mudstone rock fragments and cryptocrystalline chert fragments are also common in minor amounts. Quartz grains commonly show undulose extinction but rarely show flattening and elongation in the S2 cleavage, whereas micas are commonly aligned in S2.

Along the Crookwell River, much of the undifferentiated Adaminaby Group has undergone contact metamorphism within the aureole of the Wyangala Batholith. (Figure 2.2c, d). Grain characteristics are difficult to determine due to the recrystallisation in hornfelses; however, close to quartz-feldspar porphyry dykes associated with the batholith, contact metamorphism has enhanced the preservation of sedimentary features and younging directions are easily determined (Figure 2.2).

Within the Oberon area, the undifferentiated Adaminaby Group displays varying degrees of contact metamorphism, particularly within the aureoles of the Carboniferous Oberon, Sloggetts, Rossdhu and Duckmaloi granite plutons (Enclosure 1). Spotted hornfels are common, and comprise recrystallised granoblastic quartz grains and dispersed clots of fine-grained metamorphic biotite. Both metamorphic biotite and muscovite are developed in close proximity to the granite plutons, but gradually decrease in size and abundance away from the pluton. The development of fine-grained metamorphic biotite has obscured sedimentary features.

**RELATIONSHIPS**
The lower contact of the undifferentiated Adaminaby Group was not observed anywhere in the Oberon-Taralga region, whereas the upper contact was mapped in the Oberon, Silent-Oakey Creeks, Golspie and Crookwell River areas. The upper contact is commonly faulted; however, where it is conformable it is marked by a gradual decrease in bed thicknesses up-section, accompanied by a reduction in the grain size of sandstone layers. The predominant thick-bedded turbidites of the
Figure 2.2 Representative photographs of the Adaminaby Group. The length of the geological hammer used for scale is 33 cm. (a) Thick-bedded quartz turbidites, recumbently folded and younging towards the north (left of photograph), Monkey Creek, Golspie at GR748548 6206180 (Golspie 1:25 000 topographic sheet 8829-III-N). Backpack is approximately 30 cm wide at the base. (b) Upward younging thick-bedded quartz turbidites in Silent Creek at GR747255 6217422 (Fullerton 1:25 000 topographic sheet, 8829-IV-S). Bed thickness ranges from 50 cm to 120 cm. (c) Quartz turbidites within the contact aureole of the Wyangala Batholith north of Crookwell at GR718518 6189996 (Crookwell 1:50 000 topographic sheet, 8729-II and III). The bedding direction is oblique to both cleavages. (d) Thick-bedded quartz turbidites younging towards the north (left of picture) on the Crookwell-Binda Road at GR719320 6190123 (Crookwell 1:50 000 topographic sheet, 8729-II and III). Darriwilian conodonts have been identified from a thin (2 cm) chert horizon at the top of the dark Bouma D/E horizon in the centre of the picture. (e) Folded and hornfelsed Adaminaby Group thick-bedded quartz turbidites forming the wall of the Oberon Dam at GR765521 6264750 (Oberon 1:25 000 topographic sheet, 8830 1-S). The fence above the dam is approximately 1.2 m in height. (f) Vertically dipping turbidite bed beside the Duckmaloi River exhibiting Bouma B, C and D/E layers (GR769388 6258171, Edith 1:25 000 topographic sheet, 8830-II-N). The compass is 8 cm wide.
Figure 2.3 Logged section through the uppermost thick-bedded quartz turbidites of the undifferentiated Adaminaby Group as they thin up-section into thin-bedded turbidites of the Bumballa Formation. The outcrop is located in Monkey Creek at GR748548 6206180 (Golspie 1:25 000 topographic sheet, 8829-III-N, Figure 2.2a).
undifferentiated Adaminaby Group thus become mainly thin-bedded turbidites of the Numeralla Chert or the Bumballa Formation where the Numeralla Chert is absent.

In the Oberon area, the upper contact of the undifferentiated Adaminaby Group is recognised where the thick-bedded turbidites grade into thin-bedded siliceous turbidites 5-10 cm thick, then into a thin (2-3 cm) bed of unfossiliferous chert east of the Oberon Dam and on Goulburn Road (GR765131 6265515, Oberon 1:25 000 topographic sheet, 8830-I-S and GR759993 6257995, Edith 1:25 000 topographic sheet, 8830-II-N, Enclosure 1). Along Hampton Road in the east of the Oberon area, the sandstone component of the turbidites disappears up the section, and the Numeralla Chert is absent (Enclosure 1, 2, Figure 2.1). Over some tens of metres, sandstones of the undifferentiated quartz-turbidites become fine-grained, and decrease in abundance until they disappear leaving massive, fine-grained cream-grey siltstone. This changes up the section into massive blue-grey mudstone overlain by carbonaceous black shale of the Warbisco Shale. Northeast of Black Springs, the Adaminaby Group and Triangle Formation are juxtaposed along the north-south to northwest-southeast trending Vulcan Fault (see Chapter 4).

Further south in the Silent-Oakey Creeks area most of the contacts between the undifferentiated Adaminaby Group and the overlying units of the Adaminaby Group are faulted; however, the contact is particularly well exposed at one outcrop. Thick-bedded turbidites outcrop at the junction of Oakey and Silent Creeks (GR749361 6218457, Fullerton 1:25 000 topographic sheet, 8829-IV-S), where they are overturned, dipping gently towards the west but young eastwards upstream into Oakey Creek (Enclosure 5). In Oakey Creek, approximately 20 m upstream from the junction with Silent Creek, beds within the thick-bedded turbidites thin to 15-30 cm, losing the Bouma A divisions and sandstones decrease in grain size. These thin-bedded turbidites are also interbedded with occasional 10-15 cm thick cream-grey bands of chert and cherty mudstone (Enclosure 5, Figure 2.4a). The first chert bed is taken as the base of the Numeralla Chert. The Numeralla Chert is absent at another contact exposed further downstream in Silent Creek near the junction with the Abercrombie River.
The Numeralla Chert is also absent in the Golspie and Crookwell River areas except for a thin fault-bounded slice occurring at GR747904 6209255 (Fullerton 1:25 000 topographic sheet, 8829-IV-S) east of Burra Burra Creek (Enclosure 6A). Elsewhere in these two areas, the bed thicknesses of the undifferentiated Adaminaby Group turbidites gradually thin over 5 to 10 m from 60-120 cm to thicknesses of 15-30 cm, in thin-bedded turbidites of the Bumballa Formation (Enclosures 6A, 7).

**THICKNESS**
The thickness of this unit is difficult to determine because it contains no marker horizons and is structurally complex. However, the greatest thickness determined in the Oberon-Taralga region is a minimum of 1500 m estimated for the undifferentiated Adaminaby Group of the Crookwell River area (Enclosure 7, Figure 2.1). Only 600 m is exposed in the Oberon area, and minimum thicknesses of 450 m and 300 m occur in the Golspie and Silent-Oakey Creeks areas respectively (Enclosures 1, 3, 4, 5, 6A, Figure 2.1).

**AGE**
A single 2-cm thick chert horizon at the top of a turbidite layer in the Crookwell River area (Enclosure 7, Figure 2.2d) contains conodont elements indicating a late Darriwilian age (Da3-4; see Chapter 3). Apart from this single fossil locality, the Adaminaby Group in this area is unfossiliferous. In the Silent-Oakey Creeks area the undifferentiated Adaminaby Group conformably underlies Numeralla Chert of Darriwilian age. Based on these data, the Adaminaby Group is Darriwilian or older across the entire area.

**ENVIRONMENT OF DEPOSITION**
The undifferentiated Adaminaby Group is mainly composed of thick-bedded and less abundant thin-
bedded turbidites (facies C to D of Mutti & Ricci Lucchi 1975). This is indicated by the common Bouma sequences ABCDE, BCDE, ACD, ABC and AC recognised in turbidite sandstone-mudstone couplets. The continuity and lateral regularity of beds is also consistent with turbidite deposition on the margins of the turbidite fan rather than in channels. Therefore, the succession is interpreted to have formed in a deep-marine setting, on the middle to outer part of a large submarine fan.

**CORRELATIVES**
Within the Eastern Belt of the Lachlan Fold Belt (Figure 1.2) usage of the name "Adaminaby Group" has gradually been supplanting a plethora of names previously used to describe Early-Middle Ordovician monotonous quartz-rich turbidites (e.g. Mallacoota beds, Boltons beds, Abercrombie beds and Nungar beds). This unit has been recognised from other areas of the Bathurst and Goulburn 1:250 000 sheets (Raymond et al. 1987; Fergusson 1998b), as well as further north, where it is mapped east of Mudgee along Bara Creek (Figure 1.3; Fergusson & Colquhoun 1996).

The Canberra region to the southwest of the Oberon-Taralga region is also within the Eastern Belt (Figures 1.2, 1.3), and here Ordovician quartz-rich sandstone, siltstone and shale with minor occurrences of black shale, chert and calcareous sandstone have been mapped as the Pittman Formation (Abell 1991). Conodons identified from the Pittman Formation indicate a Darriwilian (Da3/4) or late Middle Ordovician age. Graptolites from the Acton Shale Member at the top of the Pittman Formation extend to the early Bolindian stage (Nicoll 1980; Abell 1991). Therefore, the Pittman Formation is correlated with both the Adaminaby and Bendoc Groups in the Oberon-Taralga region.

Ordovician rocks occurring further west in New South Wales are known from the Wagga-Omeo Zone and the Girilambone Sub-zone (Figure 1.2). Low-grade, but locally high-grade, centimetre to metre thick, well-bedded quartzose-metasandstone, phyllite and schist are included within the Girilambone Group in the Girilambone Sub-zone (Figure 1.2, Gilligan et al. 1994). Bouma sequences have been
recognised from these rocks indicating they were also deposited as turbidites. Although the age of these turbidites is unknown, Darriwilian (Da3/4) conodonts have been identified from the Ballast beds, a chert and sandstone unit occurring at the top of the Girilambone Group (Stewart & Glen 1986). Thus, at least the top of the Girilambone Group is probably a temporal and lithological correlative of the Adaminaby Group in the Oberon-Taralga region.

On the Cootamundra 1:250 000 sheet, the Wagga Group occurs in the Wagga-Omeo Zone (Figure 1.2), where it is mapped as a single unit consisting of slate, phyllite, siltstone, sandstone and minor chert (Warren et al. 1995). Quartzose sandstone and massive to structureless quartzite, commonly with fine-grained cross-laminated sandstone tops (Warren et al. 1995), are correlated with the undifferentiated Adaminaby Group in the Oberon-Taralga region. In eastern Victoria, the Adaminaby Group consists of a single formation, the Pinnak Sandstone, which consists of quartz turbidites of Early to Middle Ordovician age (VandenBerg & Stewart 1992; Fergusson 1998a).

In the Western Belt of the Lachlan Fold Belt (Figure 1.2), the Lower and Middle Ordovician succession is represented by the Castlemaine Group (VandenBerg & Stewart 1992). This is a widespread unit composed of quartz turbidites and black shale and ranges in age from earliest Ordovician (Lancefieldian), up to the top of the Darriwilian.

2.2.3 Numeralla Chert

DERIVATION, NOMENCLATURE AND TYPE SECTION
The Numeralla Chert was named after the Numeralla River, approximately 15 km northeast of Cooma (Figure 1.3), and was first described as a mappable Darriwilian chert-rich interval occurring near the top of the Adaminaby Group around Cooma (Figure 1.3; Glen 1994). The type section is located on the Dry Plains Road 15 km northwest of Cooma (Figure 1.3; Glen 1994, p. 21). In the type section, the Numeralla Chert consists of one or more mappable chert bands, which vary in thickness from 20 m up to 100 m and commonly contain late Darriwilian conodonts (Glen 1994).
Figure 2.4 (a) - (c) Representative photographs of the Numeralla Chert. (a) Base of the Numeralla Chert consisting of downward younging siliceous mudstone interbedded with thin (2-5cm) chert bands in Oakey Creek at GR747274 6216920 (Fullerton 1:25 000 topographic sheet, 8829-IV-S). At the top of the photograph are thick-bedded quartz turbidite beds of the Adaminaby Group, also overturned and younging into the Numeralla Chert. (b) - (c) Ribbon-bedded chert with siliceous mudstone interbeds also occurring in Oakey Creek (GR747302 6216936; Fullerton 1:25 000 topographic sheet, 8829-IV-S), but stratigraphically higher in the Numeralla Chert than shown in photograph 2.4a. Hammer used for scale in photographs (a) and (b) is 33 cm long, and half the handle width (approximately 1.5 cm) is also visible on the left hand side of photograph (c). (d) - (g) Representative photographs of the Bumballa Formation. (d) Interbedded siltstone and mudstone in a road outcrop south of the Crookwell River at GR722232 6188711 (Crookwell 1:50 000 topographic sheet, 8729-II and III). Swiss army knife is 9 cm in length. (e) Thin-bedded quartz turbidites just below the contact with the Warbisco Shale (centre left of photograph) in Monkey Creek, Golspie at GR748024 6205262 (Golspie 1:25 000 topographic sheet 8829-III-N). Person in bottom right of photograph for scale. (f) Thin-bedded quartz turbidites further down-section than photograph (e). Monkey Creek, Golspie at GR748014 6205232 (Golspie 1:25 000 topographic sheet 8829-III-N). Person in bottom right of photograph for scale. (g) Thin-bedded quartz turbidites younging northwest (right of picture) in a road outcrop beside Jenolan Street (GR765275 6265515, Oberon 1:25 000 topographic sheet 8830-1-S).
**DISTRIBUTION**

Within the study area the Numeralla Chert occurs in Oakey Creek approximately 20 m upstream of the junction with Silent Creek and as thin, up to 60 cm thick fault slivers in Silent Creek (Enclosure 5). Two thin (2-3 cm) chert horizons outcrop in the Oberon area, the first located south of Oberon at GR765131 6265515 (Oberon 1:25 000 topographic sheet, 8830-1-S) and the second located on Goulburn Road between Oberon and Black Springs (GR759993 6257995, Edith 1:25 000 topographic sheet, 8830-II-N, Enclosure 1). A fault-bounded slice of Numeralla Chert was also mapped on the eastern bank of Burra Burra Creek in the Golspie area at GR747904 6209255 (Fullerton 1:25 000 topographic sheet 8829-IV-S, Enclosure 6A).

**LITHOLOGY**

The most complete section is exposed in Oakey Creek in the Silent-Oakey Creeks area (Enclosure 5, Figure 2.4a, b, c). Here, the base of the Numeralla Chert is composed of alternating thinly bedded (5-15 cm) siliceous sandstone and mudstone (thin-bedded turbidite facies) that are interbedded with occasional thin (~5-15 cm) chert bands (Figure 2.4a). Approximately 5 m stratigraphically above this interval, sandstone layers in the thin-bedded turbidites gradually become finer-grained and more siliceous. Light blue-grey chert bands approximately 5-10 cm thick, separated by thin (1-2 cm) pale creamy coloured mudstone interbeds become more dominant up the section (Figure 2.4b, c). The chert and mudstone beds commonly display internal plane lamination.

The 0.6 m thick horizon of chert on the eastern bank of Burra Burra Creek in the Golspie area (Enclosure 6A) is composed of alternating chert and thin pale mudstone layers, similar in appearance to the bedded chert in Oakey Creek. Here, the Numeralla Chert is faulted between quartz turbidites of the undifferentiated Adaminaby Group and the Warbisco Shale.

At Oberon, sandstone layers of the undifferentiated Adaminaby Group become thinner and the grain size decreases up-section. Gradually they change to siliceous 5-10 cm thin-bedded turbidites and are
overlain by a thin horizon of creamy coloured, fractured and unfossiliferous chert. Although still recognisable as chert bands, they have undergone hydrothermal alteration and recrystallisation within the aureole of the Oberon Granite (Enclosure 1).

In thin section, the Numeralla Chert is very fine-grained, composed of cryptocrystalline quartz grains with subordinate amounts of pyrite and opaque iron oxides. Recrystallised relict radiolarians are common.

RELATIONSHIPS AND THICKNESS
The Numeralla Chert conformably overlies the Adaminaby Group in both the Silent-Oakey Creeks and the Oberon areas (Figure 2.1). Conformably overlying the Numeralla Chert in the Silent-Oakey Creeks area are thick-bedded and thin-bedded turbidites of the Bumballa Formation (Enclosure 5). The total thickness of the Numeralla Chert in the Silent-Oakey Creeks area is approximately 40 m (Figure 2.1). At Oberon, however, the Numeralla Chert is less than 10 cm in thickness. Both of these occurrences are considerably thinner than in the type section where it is estimated at up to 100 m thick (Glen 1994).

AGE
Poorly preserved conodonts and radiolaria occur in the chert beds. Recrystallisation of radiolarian tests precludes their usefulness as a dating tool, and subsequent deformation has extended and broken the conodont elements precluding the use of acid digestion recovery methods. An alternative thick sectioning technique (described in Chapter 3) has resulted in the identification of Darriwilian conodonts. *Histiodella* sp., *Cordylodus horridus* and *Periodon* sp. elements were identified in chert samples from Oakey Creek (samples R16475 and R16476) (Enclosure 5) indicating a probable mid Darriwilian (Da2/Da3) age. *Pygodus serra* and *Periodon* sp. elements indicate a slightly younger Darriwilian to Gisbornian (Da3/early Gi1) age for the fault-bounded chert (sample R16486) in Burra Burra Creek in the Golpsie area (Enclosure 6A). The Numeralla Chert is therefore considered to range
in age from middle to latest Darriwilian/earliest Gisbornian (Da2-4/Gil) across the study area.

ENVIRONMENT OF DEPOSITION
The fine grainsize and lamination in both chert and mudstone beds are consistent with deposition of the Numeralla Chert in a quiet, deep-marine environment with gentle bottom currents and aerated conditions. The conodonts identified in the chert are predominantly from the Cold Faunal Realm (Miller 1984) and are consistent with a deep-marine environment of deposition (see Chapter 3). The chert has probably formed from pelagic settling and accumulation of mainly radiolarians with minimal terrigenous input.

The presence of Da3/Gil conodonts identified from a thin 2 cm chert horizon occurring at the top of a turbidite flow in the Crookwell River area indicates the undifferentiated Adaminaby Group was still being deposited in this area during the early Darriwilian. Therefore the undifferentiated Adaminaby Group in the Crookwell River area was deposited concomitantly with the Numeralla Chert in the Silent-Oakey Creeks area. During the latest Darriwilian/earliest Gisbornian (Da3/4-Gil) a single thin chert lens was deposited in the Golspie area. In the Oberon, Crookwell River and other parts of the Golspie areas however, both thick- and thin-bedded quartz turbidites were being deposited during this time. Therefore, vertical and lateral facies changes existed across the submarine fan throughout the Darriwilian with synchronous deposition of the Numeralla Chert, undifferentiated Adaminaby Group, and thin-bedded quartz turbidites of the Bumballa Formation.

CORRELATIVES
Chert beds of Darriwilian age have been recognised and described from both the Eastern and Western Belts of the Lachlan Fold Belt (Figure 1.2). Within the Eastern Belt, most of these occurrences have been reported from the Bega/Mallacoota 1:250 000 sheet area and along the south coast of New South Wales at Batemans Bay, Burrewarra Point and Narooma and are included within the Numeralla Chert (Stewart & Glen 1991; Glen 1994). East of the Great Dividing Range 24 km northeast of Taralga (Figure 1.3), Whalan (1986) mapped two different chert horizons near and at the top of the Jocks
Creek Formation (a synonym of the Adaminaby Group). Conodonts from these horizons indicated a mid Darriwilian (Da2/Da3) age for the lower unit and a late Darriwilian to early Gisbornian (Da3/Gi1) age for the higher unit. The lower unit correlates with the Numeralla Chert in the Silent-Oakey Creeks area and the higher unit corresponds in age to chert in the Golspie area and from the type area northwest of Cooma. In the Jocks Creek area, Whalan (1986) mapped quartz turbidites similar to those of the undifferentiated Adaminaby Group between the two chert horizons.

Other occurrences of Darriwilian chert have been reported from the Mudgee and Goulburn areas where they were associated with thin-bedded turbidites of the upper Adaminaby Group (Stewart & Fergusson 1995; Fergusson & Colquhoun 1996; Colquhoun et al. 1997).

The Ballast Formation occurs at the top of the Girilambone Group in the Girilambone Sub-zone of western New South Wales, where it is composed of quartz turbidites with less common chert and mafic-intermediate igneous rocks (Figure 1.2, Iwata et al. 1995). Conodonts and radiolarians from chert in the Ballast Formation indicate that this unit is also of Darriwilian age (Stewart & Glen 1986; Iwata et al. 1995) and, therefore is correlated with the Numeralla Chert.

Although the chert units form lenses, the widespread occurrence of these lenses possibly indicates major changes in sedimentation patterns on the turbidite fan during the Darriwilian.

### 2.2.4 Bumballa Formation

**Derivation and Type Section**

The name "Bumballa Formation" is here used for a unit comprising thick- to thin-bedded quartz turbidites conformably overlying the Numeralla Chert and the undifferentiated Adaminaby Group where the Numeralla Chert is absent (C. L. Fergusson, personal communication 2000). The unit is named after Bumballa Parish, in which the type section is located along the southern bank of the Shoalhaven River (30 km east of Goulburn, Figure 1.3). This unit was previously informally referred
to as the “stripy facies” (Fergusson & VandenBerg 1990; Jones et al. 1993) and the Sunlight Creek Formation (Stewart & Fergusson 1995).

**DISTRIBUTION**
The Bumballa Formation occurs in the Golspie area (Enclosure 6A, Figures 2.1, 2.4e, f), where it was previously mapped as the Bubalahla Formation of Scheibner (1973). In the Silent-Oakey Creeks area, the Bumballa Formation is also well exposed along Oakey Creek (GR747034 6216050 to 747725 6216500, Fullerton 1:25 000 topographic sheet 8829-IV-S), but occurs less commonly in Silent Creek (Enclosure 5). In the Crookwell River area, the Bumballa Formation outcrops along the Crookwell River, and is well exposed on a track north of a black shale quarry at GR722413 6188668 (Crookwell 1:50 000 topographic sheet, 8729-II and III, Enclosure 7, Figure 2.4d). It also occurs to the south and east of Oberon (Enclosures 1, 2, Figure 2.4g).

**LITHOLOGY**
The Bumballa Formation consists of quartz-rich, thin, rhythmically-bedded fine sandstone and mudstone/shale. Sandstone/mudstone couplets are generally between 5 and 15 cm in thickness (Figures 2.4d, e, g), but at the base of the unit, thin-bedded turbidites up to 30 cm thick occur (Figures 2.4f). The fine sandstone is orange-brown to cream in colour, moderately to well-sorted and exhibits ripple cross-bedding. Bouma A and B divisions not normally thicker than 20 cm occur, and the total turbidite thickness is between 15 and 30 cm. The mudstone/shale beds are dark blue-grey, purple or chocolate in colour, and normally display planar-lamination (Bouma D). Thin-bedded turbidites at the base of the unit are similar to and easily mistaken for thin-bedded quartz turbidites that occur amongst the dominant thick-bedded turbidites of the undifferentiated Adaminaby Group, especially where the Numeralla Chert is absent.

Although the Bumballa Formation occurs as rhythmically-bedded sandstone and mudstone south of Oberon (Enclosure 1, Figure 2.4g), along Hampton Road to the east (Enclosure 2), the Bumballa
Formation is composed of massive cream siltstone, which fines up-section, through blue-grey mudstone into the Warbisco Shale.

In thin section, sandstone of the Bumballa Formation is very similar to that of the undifferentiated Adaminaby Group. It is predominantly composed of subangular to subrounded quartz grains with subordinate plagioclase and muscovite/phengite. Lensoidal rock fragments composed of chert and low-grade slate occur.

RELATIONSHIPS AND THICKNESS
The Bumballa Formation conformably overlies either the Numeralla Chert or the undifferentiated Adaminaby Group where the Numeralla Chert is absent. The conformable contact between the Bumballa Formation and the underlying Numeralla Chert is exposed in Oakey Creek (Enclosure 5). The contact is marked by a change from chert to an overlying 5 m section of thick-bedded alternating quartz-rich sandstone and mudstone beds, which is in turn overlain by thin-bedded sandstone and mudstone beds.

Where the Numeralla Chert is only locally developed or absent, as in the Crookwell River and Oberon areas, the transition from thick-bedded quartz turbidites of the undifferentiated Adaminaby Group to thin-bedded turbidites of the Bumballa Formation is more gradual. It occurs over a few metres, as is exposed along the Crookwell River between GR723380 6189920 and 724540 6189900 (Crookwell 1:50 000 topographic sheet, 8729-II and III, Enclosure 7) and in Monkey Creek in the Golspie area (GR748522 6206232, Golspie 1:25 000 topographic sheet, 8829-III-N; Enclosure 6A).

The Bumballa Formation is conformably overlain by the Warbisco Shale in the Crookwell River and Golspie areas (Enclosures 6A, 7, Figure 2.1). It is a sharp contact, marked by the disappearance of sandstone and the onset of the dark blue-black siliceous Warbisco Shale.
The thickness of the Bumballa Formation is variable both across the Oberon-Taralga region as a whole and within the Golspie area. The thickest section outcrops in the Crookwell River area, where up to 300 m is estimated (Enclosure 7, Figure 2.1). A minimum thickness of 180 m is exposed both in Oakey Creek in the Silent-Oakey Creeks area (Enclosure 5) and on the western bank of Burra Burra Creek at Golspie (Enclosure 6A). However, in other exposures where both contacts are preserved in the eastern Golspie area, the Bumballa Formation ranges between 30 m and 90 m in thickness. Thin-bedded turbidites of the Bumballa Formation south of Oberon are only 60 m thick (Enclosure 1), whereas to the east along Hampton Road, the transitional massive siltstone and mudstone changes in thickness from 20 m to 50 m (Enclosure 2).

**AGE**
No fossils have been recovered from the Bumballa Formation in the Oberon-Taralga region; however, its age is constrained by conformable relationships with the underlying and contemporaneous Numeralla Chert and overlying Warbisco Shale. These constraints indicate that the Bumballa Formation spans a maximum age range from latest Darriwilan (Da3/4) up to the top of the Gisbornian.

**ENVIRONMENT OF DEPOSITION**
The Bumballa Formation comprises thick- to thin-bedded and very thin-bedded turbidites, commonly with incomplete Bouma successions, either lacking or only having thin Bouma A and B divisions and lacking bioturbation. These turbidites are representative of facies D and E of Mutti and Ricci Lucchi (1975) and are interpreted as having formed in a deep marine, middle to outer submarine fan deposit.

On the basis of ubiquitous ripple cross-bedding, basal scours, moderate to good sorting and heavy mineral laminae, Jones et al. (1993) interpreted the thin-bedded, turbidites of the Bumballa Formation in the Shoalhaven River Gorge as partly having formed by deposition from contour currents (i.e. contourites). Although no contourites have been recognised in the Oberon-Taralga region, turbidites of the Bumballa Formation were deposited concomitantly with thick-bedded turbidites at the top of
the undifferentiated Adaminaby Group, chert of the Numeralla Chert and black shale of the lowermost portions of the Warbisco Shale. The facies variations recognised across the submarine fan during the late Darriwilian/Gisbomian are indicative of a waning sediment supply to the middle and outer margins of the submarine fan during this time.

CORRELATES
Probable lateral correlatives of the Bumballa Formation in the Eastern Belt of the Lachlan Fold Belt (Figure 1.2) include the Chakola Formation of Glen (1994) and the Sunlight Creek Formation of VandenBerg et al. (1996). The Chakola Formation occurs above the Numeralla Chert east of the Murrumbidgee Fault in the Cooma area (Figure 1.3) and is composed of 300-400 m of thick-bedded quartz-turbidites similar to the Adaminaby Group (Glen 1994). The Glen Fergus Mudstone Member occurs in the top 50-100 m and comprises interbedded quartz sandstone and mudstone/slate. Glen (1994) described the sandstone beds as generally less than 10 cm thick, but occasionally up to 1 m thick, and characterised by sets of low-angle cross-laminations.

The Sunlight Creek Formation was originally described as a thin and discontinuous basal Member of the Warbisco Shale (VandenBerg et al. 1992). It was upgraded to formation status by Orth et al. (1995). The formation occurs in eastern Victoria, where it is composed of rhythmically alternating green to black siltstone and white fine-grained quartz-sandstone of latest Darriwilian to early Gisbornian (Da4-Gi1) age (VandenBerg et al. 1996). It has also been recognised in the Bega/Mallacoota region, where it overlies the Numeralla Chert west of the Murrumbidgee Fault (Glen 1994).

A possible correlative to the Bumballa Formation has also been reported from the Cootamundra region (Figure 1.3). At the top of the Wagga Group, a generally thin-bedded sequence of interbedded sandstone, siltstone and minor chert displaying graded bedding and cross-lamination has been recognised (Warren et al. 1995). Graptolites indicating a late Gisbornian age have been identified from this unit (Sherwin 1986).
2.2.5 Bendoc Group
The name "Bendoc Group" was used by VandenBerg et al. (1992) to encompass three fine-grained Upper Ordovician units in east Gippsland and southeast New South Wales. The three units included the basal Sunlight Creek Formation, overlain by the Warbisco Shale and Akuna Mudstone. The name is derived from the Bendoc region in East Gippsland, Victoria, where all three formations are exposed in the type section located in Mountain Creek 33 km west of Bendoc (Figure 1.3, VandenBerg et al. 1992). Within the Oberon-Taralga region only one fine-grained Upper Ordovician unit, the Warbisco Shale, is included in the Bendoc Group.

2.2.6 Warbisco Shale

NOMENCLATURE, DERIVATION AND TYPE SECTION
The name "Warbisco Shale" was formalised by VandenBerg et al. (1992), but had been used informally prior to this time (VandenBerg 1981; Cas & VandenBerg 1988; VandenBerg & Webby 1988). The name is derived from the Warbisco Track in the Yalmy region of eastern Victoria (Figure 1.3) and is used for a unit of black mudstone and shale. The type section, located along Mountain Creek in the Yalmy region, is the only known section where the formation is complete (VandenBerg et al. 1992). Faunal gaps in other sections indicate faults either within or at the contacts of the formation.

DISTRIBUTION
Between Bathurst and Goulburn, the Warbisco Shale is widespread and in many places usually occurs as fault-bounded slices between the more competent interbedded quartz sandstone and mudstone of the undifferentiated Adaminaby Group, Bumballa Formation and Bald Ridge Formation. The unit has been mapped in the Oberon, Silent-Oakey Creeks, Golspie and Crookwell River areas (Enclosures 1, 2, 5, 6A, 7, Figure 2.5a-e). It also occurs east of the Silent-Oakey Creeks area on the Oberon-Goulburn Road, north and south of the Abercrombie River and northeast of the Thompson Creek area south of Sophia Creek adjacent to the Rockley-Burraga Road (Figure 1.4).
LITHOLOGY
The Warbisco Shale is uniform throughout the Oberon-Taralga region, consisting of siliceous black shale at the base of the formation with more fissile carbonaceous black shale higher in the succession. Where fresh, it is dark grey to black and weathers to pale grey/white or pale orange if iron-stained. In thin section, the shale consists almost entirely of very fine-grained quartz, relict radiolarians and carbonaceous material with small amounts of pyrite. It either contains thin (2-4 cm) siliceous beds or is laminated. Higher in the unit the shale is more fissile and splits easily into 1 - 2 mm thick laminae, especially where bedding and cleavage are coplanar, whereas the stratigraphically lower more siliceous shale splits into 5-10 mm laminae. Where cleavage is oblique to bedding, splitting the shale is very difficult.

A single thin sandstone bed was observed interbedded with the shale in Monkey Creek in the Golspie area (GR748132 6204991, Golspie 1:25 000 topographic sheet 8829-III-N). This black sandstone bed was only 20 cm thick and composed of fine-grained quartz, sedimentary rock fragments and carbonaceous material (sample R16029).

AGE
Sections through the Warbisco Shale are commonly fault-bounded and structurally complex. Recovery of graptolite faunas is needed to precisely constrain the stratigraphic level within the shale. Graptolite ages from the Eastern Belt of the Lachlan Fold Belt indicate that the Warbisco Shale extends from the late Gisbornian (Gi2) through to the middle Bolindian (Bo3) stages of the Late Ordovician (VandenBerg & Stewart 1992).

In the Oberon-Taralga region graptolites have only been collected in this study from black shale in the Golspie area, and were identified by A. H. M. VandenBerg (personal communication 1996). Poorly preserved graptolites were collected from weathered grey-white shales beside Leihwood Road at GR747087 6203050 (Golspie 1:25 000 topographic sheet, 8829-III-N, Enclosure 6A). Although
Figure 2.5 Representative photographs of the Warbisco Shale. (a) Siliceous cleaved and folded Warbisco shale outcropping beside Hampton Road at GR768751 6265096 (Oberon 1:25 000 topographic sheet, 8830-1-S). The fence at the top of the outcrop is approximately 1 m in height. (b) Warbisco Shale outcropping in Burra Burra Creek at GR749586 6205994 (Golspie 1:25 000 topographic map, 8829-III-N). The shale in this outcrop is very siliceous and forms open folds with axial planar to the northwest trending S2 cleavage. The hammer is 31 cm long. (c) Chevron-folded siliceous Warbisco Shale beside Crookwell River at GR721639 6188986 (Crookwell 1:50 000 topographic sheet, 8729-II and III). The knife used as a scale is 9 cm in length. (d) Carbonaceous Warbisco Shale with numerous silica partings and quartz veins in a quarry south of Crookwell River at GR722355 6188563 (Crookwell 1:50 000 topographic sheet, 8729-II and III). The hammer is 33 cm in length. (e) Folded and weathered carbonaceous Warbisco Shale beside Monkey Creek, GR748249 6205341 (Golspie 1:25 000 topographic sheet, 8829-III-N). Geological hammer is 33 cm in length. Sulphides weathering out of the black shale are contaminating Monkey Creek, the water beside this outcrop was tested and had a pH of 4.0.
dominated by indeterminate biserial graptolites, a single *Dicellograptus* sp. indicative of the Eastonian was identified from this site. Fresh black shale outcropping above Burra Burra Creek at GR749075 6206350 (Golspie 1:25 000 topographic sheet, 8829-III-N, Enclosure 6A) yielded a single specimen of *Climacograptus uncinatus* showing the characteristic split virgella. This is the index species for the Bolindian *C. uncinatus* graptolite zone. The sample also contained a fully grown specimen of *Dicellograptus ornatus*, also diagnostic for the Bolindian, and a large stipe thought to belong to *Dicellograptus gravis*, index for Ea4 but also known to extend into Bo1. Therefore, the Warbisco Shale above Burra Burra Creek is earliest Bolindian (*Bo1- C. uncinatus* zone) in age.

During previous mapping of the Goulburn 1:250 000 geological sheet, graptolites have been recovered from black shale in the Crookwell River, Silent-Oakey Creeks and the Golspie areas. A complete list of graptolite faunas from the areas covered within this study is included within Appendix B. These data indicate that the Warbisco Shale ranges in age from Gisbomian to early Bolindian across the study area; however no complete sections without faunal breaks have been identified.

**THICKNESS**

Sections through the Warbisco Shale are discontinuous and structurally complex, consequently it is very difficult to estimate even a minimum thickness for this unit. Based on cross-sections in the Oberon-Taralga area, the thickness of the Warbisco Shale ranges from 60-90 m in the Oberon area to between 35 m and 145 m in the Golspie area, with approximately 100 m occurring in the Crookwell River area. This is significantly less than in the type section in eastern Victoria, where the thickness has been estimated at 400 to 500 m (VandenBerg *et al.* 1992).

**RELATIONSHIPS**

In the Oberon-Taralga region, most of the upper and lower contacts of the Warbisco Shale are faulted or exhibit polished surfaces indicating some slip has taken place parallel to the contact. The lower boundary is preserved along the Crookwell River at GR722280 6188700 (Crookwell 1:50 000
boundary is preserved along the Crookwell River at GR722280 6188700 (Crookwell 1:50 000 topographic sheet 8729-II and III, Enclosure 7). Here, 2-3 m of shaly mudstone occurs at the top of the Bumballa Formation and gradually changes colour from khaki-grey to pale weathered black shale. The lower boundary is also preserved along Hampton Road, in the Oberon area where at the top of the Bumballa Formation massive mudstone changes from blue-grey to black and becomes more indurated (Enclosures 1, 2).

The upper contact of the Warbisco Shale is exposed at the junction of Monkey and Burra Burra Creeks (GR748517 6206364, Golspie 1:25 000 topographic sheet 9928-III-N, Enclosure 6A) where the Warbisco Shale is overlain by thin-bedded and thick-bedded turbidites of the Bald Ridge Formation. At a second locality in Monkey Creek (GR748156 6205007, Golspie 1:25 000 topographic sheet 9928-III-N, Enclosure 6A), the Warbisco Shale is sharply overlain by carbonaceous siltstone and thin-bedded turbidites at the base of the Bald Ridge Formation. In the Oberon area, including the Hampton Road Section, the Warbisco Shale is overlain at a sharp contact by undifferentiated volcaniclastics of the Triangle Formation (Enclosures 1, 2).

**ENVIRONMENT OF DEPOSITION**

The Warbisco Shale is commonly siliceous, laminated and contains a sparse graptolite fauna, partially dissolved radiolarian tests and pyrite nodules. These characteristics are indicative of the "Black Shale Facies" outlined by Berry and Wilde (1978). It is interpreted as having been deposited in deep-marine, oxygen-reduced environment.

**CORRELATIVES**

Late Ordovician fissile, siliceous black shale is widespread across the Lachlan Fold Belt (Figure 1.2, see Figure 3 in VandenBerg & Stewart 1992). The Warbisco Shale is mapped from eastern Victoria through southeastern New South Wales and into the Oberon-Taralga region (VandenBerg et al. 1992; Glen 1994, Orth et al. 1995; VandenBerg et al. 1996; Simpson et al. 1997; Fergusson 1998a, b). The
Acton Shale Member at the top of the Pittman Formation in the Canberra district comprises a pale black, thinly laminated siliceous shale which, based on graptolites recovered from the shale, ranges in age from Gisbornian to Bolindian and is a correlative of the Warbisco Shale (Abell 1991).

Correlatives of the Warbisco Shale have been recognised from the Cowra-Canberra-Buchan zone and Girilambone sub-zone west of the Oberon-Taralga region in New South Wales (Figure 1.2). In the Parkes and Forbes areas of these zones, a succession of Late Ordovician to Early Silurian chert, black siltstone and shale with calcareous bands has been mapped as the Cotton Formation (Sherwin 2000). To the south in the Cootamundra and Forbes areas several Upper Ordovician units have been mapped and are all composed predominantly of siliceous siltstone, mudstone, chert and claystone with minor sandstone (Warren et al. 1995).

Black siliceous shales of Darriwilian to Bolindian age have also been recognised from central Victoria. The Mount Easton Shale occurs in the Melbourne Zone (Figure 1.2), where it is composed of a basal portion of cherty shale and bioturbated mudstone of Darriwilian to early Gisbornian age, overlain by black carbonaceous shale containing late Gisbornian to mid-Bolindian graptolites (VandenBerg & Stewart 1992).

2.2.7 Bald Ridge Formation (new name)

**Derivation and Type Section**
The Bald Ridge Formation is named after the property of “Bald Ridge” located 14 km west of Burraga (Figure 1.4). It is introduced to refer to Early Silurian quartz-rich sandstone and mudstone outcropping in the Oberon-Taralga region. The type section is located along Thompson Creek (Enclosure 4) between GR723000 6241900 and 729850 6243125 (Abercrombie 1:50 000 topographic sheet, 8730 II and III).
**DISTRIBUTION**

Within the Oberon-Taralga region, the Bald Ridge Formation occurs in all of the areas mapped (Enclosures 1, 2, 4, 5, 6A, 6B, 7). In the Thompson Creek area it forms the core of the Thompson Creek Anticline (Enclosure 4). In this area, these rocks were previously mapped as part of both the undifferentiated Adaminaby Group and the Triangle Formation (Raymond *et al.* 1997; Pogson & Watkins 1998). The Bald Ridge Formation also forms north-south trending fault-bounded tracts in the eastern part of the Oberon area (Enclosures 1, 2). It occurs as poorly exposed hornfels and quartz sandstone in the east of the Crookwell River area (Enclosure 7), and in the Golspie and Silent-Oakey Creeks areas (Enclosures 5, 6A, 6B). In all of these areas, it has previously been placed in either the Burra Burra Creek Formation or the Adaminaby Group (Scheibner 1970, 1973; Raymond *et al.* 1997; Pogson & Watkins 1998).

**LITHOLOGY**

In the Oberon-Taralga region, the Bald Ridge Formation is composed of thick-bedded (0.6 to 4.0 m) buff to grey, quartz-rich, coarse to fine-grained sandstone and mudstone. In the Thompson Creek area, the unit has been contact metamorphosed, and is light to dark grey because of the presence of fine biotite. Medium- to fine-grained sandstone beds are commonly massive, forming Bouma A and B horizons of up to 1.0 m thick with little obvious change in grainsize throughout (Figure 2.6a, c, f, g, h). These are overlain by laminated fine sandstone and mudstone beds between 15-50 cm thick. Where graded bedding occurs, Bouma divisions A, B and D/E are usually thin (10-15 cm) compared to thick (0.8-1.5 m) Bouma C horizons (Figure 2.6a, c), with the Bouma A and B divisions comprising coarse to medium-grained sandstone and rare conglomerate composed of subrounded, intraformational quartz sandstone clasts.

Quartz sandstones are typically bimodal and composed of larger well-rounded, spherical poly- and monocrysaline quartz grains and sedimentary rock fragments up to 2-3 mm contained amongst smaller medium to fine sand grains. The latter are dominated by well sorted, subangular to
Figure 2.6 Representative photographs of the Bald Ridge Formation. (a) Thick-bedded, fine-grained quartz turbidites in Thompson Creek at GR723927 6241088 (Abercrombie 1:50 000 topographic sheet, 8730-II and III). Individual bed thicknesses are commonly between 0.8 m and 3.0 m. (b) Rare thin-bedded fine sandstone beds showing bioturbation in Thompson Creek at GR724075 6241699 (Abercrombie 1:50 000 topographic sheet, 8730-II and III). Swiss army knife is 9 cm in length. (c) Thick Bouma A/B horizons composed of fine quartz sandstone with much thinner Bouma C/D/E mudstone horizons in Bald Ridge Formation in Thompson Creek at GR728069 6242084 (Abercrombie 1:50 000 topographic sheet, 8730-II and III). Geological hammer is 33 cm in length and lying along the strike direction. (d) The contact between the Bald Ridge Formation and the overlying Upper Silurian Kangaloolah Volcanics on Bald Ridges Road, GR720943 6244685 (Abercrombie 1:50 000 topographic sheet, 8730-II and III). The contact is an angular unconformity, which predates and has subsequently been folded by F2 folding. The top 30 to 40 cm of the Bald Ridge Formation immediately below the contact has been contact metamorphosed and contains large biotite flakes. The measuring tape case at the base of the outcrop is 35 cm from the base to the top of the handle. (e) Rounded granitic boulder within a quartz sandstone turbidite of the Bald Ridge Formation, Bald Ridge Creek (GR723713 6243192, Abercrombie 1:50 000 topographic sheet, 8730-II and III). The Swiss army knife is 9 cm in length. (f) Thick-bedded coarse-grained quartz sandstone of the Bald Ridge Formation forming thick Bouma B division at GR748210 6204989, Monkey Creek (Golspie 1:25 000 topographic sheet 8829-III-N). Hammer head is 18 cm in length. (g) Thick-bedded, upward-younging quartz-rich turbidites in “Blue Hill Quarry” south of Oberon (GR766739 6256692, Edith 1:25 000 topographic sheet, 8830-II-N). (h) Same outcrop as in 2.6f, photograph shows thick-bedded quartz turbidites in Monkey Creek, Golspie. Hammer length is 33 cm and is sitting in the middle of a single turbidite deposit. Strike of bedding is across the photograph.
subrounded quartz grains of up to 1 mm with subordinate sedimentary rock fragments, accessory zircon and tourmaline and minor amounts of plagioclase. Rock fragments are predominantly cryptocrystalline chert and lensoidal slate fragments. Quartz grains show varying degrees of undulose extinction. Where the rocks have undergone contact metamorphism, brown biotite commonly occurs. Granitic clasts up to 15 cm across occur in quartzose sandstone within the Bald Ridge Formation in Bald Ridge Creek (Figure 2.6e).

Near the top of the Bald Ridge Formation in the type locality, lenses of actinolite schist and conglomerate occur. The actinolite schist is massive and ranges from approximately 5 m to 50 m thick. Original sedimentary structures are not preserved, the schistosity being the only planar fabric observed. Schistose rocks are dominated by radiating actinolite needles with subordinate muscovite, plagioclase and well-rounded poly- and monocrystalline quartz grains.

The conglomerates are massive with bedding and other sedimentary structures generally not discernible. Where determined, beds are between 5 and 30 m thick. They are commonly poorly sorted, with subrounded to rounded clasts dominated by actinolite schist and lithic sandstone but also including quartz sandstone within a green to grey matrix of either quartz sandstone, actinolite schist or a mix of the two. The clasts range from <1.0 cm to 20 cm in size, are weakly imbricated and matrix supported. Lithic sandstone clasts are dominated by subangular plagioclase with subangular volcanic rock fragments and angular volcanic quartz grains. Bedding orientations near contacts between the quartz turbidites, actinolite schist and conglomerate are parallel to the contacts. In the eastern Thompson Creek area at G729174 6242697 and 729255 6242717 (Abercrombie 1:50 000 topographic sheet, 8730 II and III) quartz turbidites young into the actinolite schist. Therefore these three rock types occurring near the top of the Bald Ridge Formation are considered interbedded (Figure 2.1).
RELATIONSHIPS AND THICKNESS
The lower boundary of the Bald Ridge Formation is not exposed in the type locality. It is exposed in
the Golspie area (Enclosure 6A), where at least 10 m of thin-bedded quartz turbidites overlie the
Warbisco Shale, and are in turn overlain by sandstone beds 60-100 cm thick and turbidites with
10-20 cm basal sandstone layers and 60-100 cm mudstone layers.

In the Oberon area, Late Ordovician volcaniclastic sandstone of the Triangle Formation is overlain by
quartz turbidites of the Bald Ridge Formation at a sharp contact along Hampton Road (Enclosure 2).
Bedding orientations are similar across the contact and imply that the units are conformable.

The upper contact of the Bald Ridge Formation was only observed in the Thompson Creek area,
where it is an angular unconformity overlain by both the Fosters Creek Conglomerate and the
Kangaloolah Volcanics on the western margin of the Thompson Anticline (Enclosure 4). The contact
between the Bald Ridge Formation and the Kangaloolah Volcanics outcrops on Bald Ridges Road at
GR721000 6245700 (Abercrombie 1:50 000 topographic sheet, 8730 II and III; Figure 2.6d). Here, the
contact is discordant to bedding orientations in the underlying Bald Ridge Formation indicating that
the rocks of the Bald Ridge Formation have been folded prior to deposition of the overlying
Kangaloolah Volcanics, followed by subsequent folding of the contact itself. Directly below the
contact, 30-40 cm of the uppermost Bald Ridge Formation is rich in metamorphic biotite, thought to
have formed by hydrothermal alteration.

The greatest thickness of the unit is exposed in the Thompson Creek area, where a minimum thickness
of 1200 m is estimated (Enclosure 4) and includes 30-70 m of actinolite schist and conglomerate.
Elsewhere in the Oberon-Taralga region, a minimum thickness of 300-400 m is exposed in the
Crookwell River and Silent-Oakey Creeks areas and along Hampton Road in the Oberon area. In the
Golspie area, minimum thicknesses range from 45 m up to 550 m.
AGE
No fossils have been recovered from the Bald Ridge Formation in the Oberon-Taralga area. The unit overlies the Warbisco Shale in both the Crookwell River and Golspie areas (Enclosure 6A, 7), and overlies the uppermost Triangle Formation in the eastern Oberon area (Enclosure 2), indicating that it is younger than the Late Ordovician. In the western Thompson Creek area the Bald Ridge Formation unconformably underlies the Fosters Creek Conglomerate and Kangaloolah Volcanics of late Wenlock/early Ludlow age (Pogson & Watkins 1998; Enclosure 4). Therefore, the Bald Ridge Formation is between Late Ordovician and earliest Late Silurian in age.

ENVIRONMENT OF DEPOSITION
The Bald Ridge Formation was deposited as massive, coarse to fine-grained sandstone and thick-bedded turbidites in a deep marine setting. The thickness and massive texture common in these turbidite layers, along with the presence of conglomerate and coarse sandstone at the bases of some turbidite layers, implies extremely rapid deposition without traction effects (Mutti & Ricci Lucchi 1975). Comparison with the submarine fan facies model of Mutti and Ricci Lucchi (1975) indicates that these turbidites are similar to facies A, B and C, and based on this model, the Bald Ridge Formation would have been deposited primarily within channels on the mid- to outer parts of a submarine fan.

CORRELATIVES
There are many names in the stratigraphic literature for latest Ordovician to early Silurian quartz-rich sandstone and mudstone units in the Eastern Lachlan Fold Belt. In New South Wales, these include the Black Mountain Sandstone with the basal State Circle Shale, and the laterally equivalent Murrumbateman Creek Formation on the Canberra 1:100 000 Sheet (Abell 1991). The Yalmy Group is described from the Cooma region in New South Wales and from eastern Victoria where it has been divided into a number of different units based on the variation in the ratio of quartz sandstone and mudstone (Lewis et al. 1994; VandenBerg et al. 1992). These latest Ordovician to early Silurian units
have been described as "clean orthoquartzites" (VandenBerg 1992), and often exhibit a distinct bimodality of the sandstone component (Abell 1991, VandenBerg 1992).

The age of all these Bald Ridge Formation correlatives is poorly constrained, and is primarily based on fossiliferous units lying stratigraphically above or below the units. However, graptolites of Llandovery age have been recovered from mudstone horizons in the Canberra and Bendoc areas (State Circle Shale, Öpik 1954, 1958; Strusz & Henderson 1971; Strusz & Jenkins 1982; Yalmy Group, VandenBerg et al. 1992), and a shelly fauna, no older than Wenlock, was described from the top of the Mundoonen Sandstone in the Yass district (Crook et al. 1973).

2.3 ORDOVICIAN TO EARLY SILURIAN STRATIGRAPHY
MOLONG VOLCANIC PROVINCE
Within the Oberon-Taralga region, the rocks of the Molong Volcanic Province are all of deep-marine origin. The stratigraphic nomenclature previously used for the rocks of this province within the Oberon-Taralga region is summarised in Table 2.1, but herein they are divided into two units, the Triangle Formation and the Rockley Volcanics. Within the Triangle Formation, a number of members have been recognised. These members range in age from Early Ordovician to the Late Ordovician, and possibly extend into the Early Silurian. The Rockley Volcanics are restricted to the Late Ordovician-?Early Silurian and have been grouped with other Late Ordovician volcanic suites of the Molong Volcanic Province within the Cabonne Group (Pogson & Watkins 1998). These units of the Molong Volcanic Province have only been identified in the Oberon area and not in the other study areas described herein within the Oberon-Taralga region (Enclosures 1, 2, 3, Figure 1.4). In the Oberon area, the most plentiful outcrop occurs along creeks and in some road cuttings. Outcrop is generally sparse and locally totally absent, preventing a complete understanding of the rocks.

2.3.1 Triangle Formation
Stanton (1956, p. 134) formally introduced the Triangle Group as "consisting of low-grade
metamorphic products of shales, greywackes, isolated chert lenses and very minor tuffaceous bands." He recognised that this group was the oldest unit in the Rockley district (Figure 2.1) and described it as underlying all other units in the area. The type section was identified where plentiful outcrops of the unit occurred along Triangle Creek, approximately 6 km south-southwest of Rockley (see also Fowler & Iwata 1995).

Between 1958 and 1975, several unpublished honours theses covering relatively small areas of Palaeozoic rocks within, and surrounding the Rockley district, resulted in a proliferation of new stratigraphic names (Binns 1958; Nicholson 1960; Edenborough 1967; Cottle 1972; Brownlow 1973; Caracciolo de Vietri 1975; Tegart 1975; Fowler 1987). In two cases the complete inversion of Stanton's (1956) stratigraphic succession was suggested, making the Triangle Group the youngest Palaeozoic rocks in the area (Anderson 1960; Caracciolo de Vietri 1975). During mapping of the first edition of the Bathurst 1:250 000 and Taralga 1:100 000 geological map sheets, the Triangle Group was once again placed at the base of the Palaeozoic succession, and its areal extent was increased to include all of the ?Ordovician quartz-rich turbidites occurring further east and south on these sheets (Packham 1968, 1969; Scheibner 1970, 1973). Pickett (1982) discussed the problems posed by the confusion over the stratigraphic succession in the area south of the Bathurst Granite (Figure 1.4) and concluded that the Palaeozoic stratigraphy and structure were in need of thorough revision.

Fowler (1987) included only the feldspathic, quartz-poor pelites, greywackes and minor cherts around Rockley within the Triangle Formation (Triangle Group of Fowler 1987), and introduced the name "Carinya Formation" for crystal-rich volcanioclastics occurring to the north and east of Rockley. In contrast to the remainder of the Triangle Group, which is characterised by plagioclase, volcanic lithic fragments and very minor amounts of quartz, the Carinya Formation was composed of clinopyroxene, plagioclase and detrital hornblende.

Quartz-rich turbiditic rocks continued to be referred to as part of the Triangle Group following the use
of the name on the Taralga 1:100 000 sheet by Scheibner (1970, 1973; Fergusson & VandenBerg 1990; Scott 1991). During mapping of the second edition of the Bathurst 1:250 000 geological map sheet, the Triangle Group was redefined as the Triangle Formation and placed within the Early to Middle Ordovician Kenilworth Group (Wyborn & Henderson 1993; Wallace & Stuart-Smith 1994; Raymond et al. 1997; Pogson & Watkins 1998). The Kenilworth Group was introduced for intermediate to mafic volcanic and volcaniclastic sequences of probable Middle Ordovician age underlying the Cabonne Group and outcropping along the Molong High and southern margins of the Hill End Trough. This group was defined as conformably overlying, or in fault contact with the Adaminaby Group, and in the Rockley and Oberon areas, is overlain by the Rockley Volcanics (Pogson & Watkins 1998).

Herein, the status of the Triangle Formation is retained (as defined by Pogson & Watkins 1998). However, it is removed from the Kenilworth Group, which is restricted to the Early and Middle Ordovician, because mapping and palaeontological data collected from the Oberon area during this study indicate that the Triangle Formation ranges from the Early Ordovician to the Late Ordovician. Three members are recognised and mapped within the Triangle Formation in the Oberon area. They include the Budhang Chert Member, Gidyen Volcaniclastic Member and Mozart Chert Member. Additionally, the Late Ordovician part of the formation in the Oberon area has been left as undifferentiated Triangle Formation.

2.3.2 Budhang Chert Member (New Name)

**DERIVATION OF NAME, DISTRIBUTION AND TYPE SECTION**
This unit is introduced here for black chert and black cherty mudstone outcropping in the Oberon area along Brisbane Valley Creek for 1000 m upstream from its junction with Parlour Creek (Enclosure 1). The name 'Budhang', meaning black, is derived from the local Aboriginal (Wiradjuri) dialect. It is not known from anywhere else in the Oberon-Taralga region. Outcrops along Brisbane Valley Creek between GR756364 6258636 and 756380 6260090 (Edith 1:25 000 topographic sheet, 8830-II-N) are
Figure 2.7 Representative photographs of the Budhang and Mozart Chert Members, Triangle Formation. (a) - (c) Budhang Chert Member. (a) Outcrop of the Budhang Chert Member in the type section along the eastern bank of Brisbane Valley Creek west of Oberon at GR756393 6258779 (Edith 1:25 000 topographic sheet, 8830-II-N). (b) Tightly folded black and grey ribbon chert with thin mudstone interbeds, Brisbane Valley Creek (GR756379 6258716, Edith 1:25 000 topographic sheet, 8830-II-N). Hammer length is 28 cm. (c) Chert block in the centre of the photograph was collected from GR756379 6258716 (Edith 1:25 000 topographic sheet, 8830-II-N), serial sectioned (Sample R15975), and found to contain Bendigonian conodonts. Hammer is 28 cm long. (d) - (g) Mozart Chert Member. (d) Weathered and fractured chert and shaly mudstone in road outcrop on the Beaconsfield-O'Connell Road, north of Black Springs at GR754493 6258093 (Mt David topographic sheet, 8830-III-N). Hammer is 28 cm long. (e) Siliceous siltstone and mudstone of the Mozart Chert Member outcropping beside Mozart Road at GR758232 6255059 (Edith 1:25 000 topographic sheet, 8830-II-N). Hammer is 33 cm in length. (f) - (g) Mozart Chert Member in outcrop on the corner of Rockley-Oberon Road and Langs Road, west of Oberon at GR757761 6266141 (Oberon 1:25 000 topographic sheet 8830-1-S). (f) Photograph showing chert bands with thin mudstone interbeds. Darriwilian conodonts were identified in serial sections of chert from this outcrop (see Chapter 3, sample R15973). Arrows (in pink) are drawn onto the chert prior to removal of the sample from the outcrop so that the bedding orientation can later be determined for serial sectioning. The chisel is 20 cm long. (g) Same outcrop as in photograph 2.7f, but here beds of clay-rich mudstone can be seen interbedded with the ribbon chert.
nominated as the type section of the Budhang Chert Member (Enclosure 1, Figure 2.7a).

LITHOLOGY
The Budhang Chert Member consists of thin-bedded chert beds and siliceous mudstone beds approximately 1 to 10 cm thick. These beds commonly display internal plane lamination. Occasional siliceous shaly lamina occurs between the chert beds (Figure 2.7b, c). The chert, siliceous mudstone and shaly interbeds are all black to dark grey, weathering to pale grey. At a microscopic scale, carbonaceous organic material is abundant in the chert, and this accounts for the dark colour. Conodont elements and partly dissolved radiolarian tests along with small, disseminated pyrite cubes also occur in the chert. The upper part of the unit is dominated by siliceous mudstone.

RELATIONSHIPS AND THICKNESS
The lower boundary of the Budhang Chert Member is not exposed. The upper contact with the overlying Gidyen Volcaniclastic Member is poorly exposed, but is marked by the appearance of thin-bedded, fine- to medium-grained, feldspathic sandstone and mudstone outcropping along a forest road northeast of the junction of Brisbane Valley and Parlour Creeks (GR756136 6260090, Edith 1:25 000 topographic sheet, 8830-II-N, Enclosure 1). This contact is probably conformable. Further to the south it is very poorly exposed and is traced on the basis of a change from siliceous, pale grey to buff soil, float and weathered outcrop of the Budhang Chert Member, to clay-rich, orange/red soil, float and weathered outcrop of the Gidyen Volcaniclastic Member. Other contacts along the eastern and southern margins of the northerly trending anticline that contains the Budhang Chert Member are faulted against younger members of the Triangle Formation and Rockley Volcanics (Enclosure 1). The unit has a minimum true thickness of 120 m estimated from cross-section F-F’ (Enclosure 1).

AGE
Conodont elements of *Paracordylodus gracilis* have been identified in thick sections of the Budhang Chert Member (see Chapter 3). These microfossils indicate that the chert is Early Ordovician or
Lancefieldian to Bendigonian in age (La3-Be2). The age of mudstone in the upper part of the unit is not known.

ENVIRONMENT OF DEPOSITION
The Budhang Chert Member is a pelagic deposit predominately composed of siliceous and carbonaceous organic material with only a minor hemipelagic component. The fine grain size and planar lamination indicates that deposition occurred in a quiescent deep-marine environment affected by minor current activity and with minimal terrestrial sedimentary input. A deep-marine environment of deposition off the shelf margin is also supported by the presence of *P. gracilis* (Landing 1976).

CORRELATIVES
The Budhang Chert Member has been correlated with a basal unit of siltstone and chert of the Triangle Formation in Triangle Creek south of Rockley (Fowler & Iwata 1995; Figure 2.1). This correlation is based on similarities in lithology and stratigraphic position; however, Early Ordovician fossil control is lacking from the Triangle Creek area. It is also correlated with fine-grained, deep-marine Early Ordovician units from both the Junee-Narromine and Molong Volcanic Belts of the eastern Lachlan Fold Belt (Figure 1.3). Sherwin (1993) reported Early Ordovician (La3-Be1) graptolites from siltstone beds within the Yurrimbah Chert Member of the Nelungaloo Volcanics near Parkes (Junee-Narromine Belt) and also from the Hensleigh Siltstone 25 km south of Wellington (Molong Volcanic Belt). Bendigonian conodonts have also been recovered from shallow marine limestone redepósited as allochthonous blocks within the Hensleigh Siltstone during the Bendigonian (Percival *et al.* 1999).

2.3.3 Gidyen Volcaniclastic Member (New Name)

DERIVATION OF NAME, DISTRIBUTION AND TYPE SECTION
The Gidyen Volcaniclastic Member is named from the local Aboriginal (Wiradjuri) dialect, with the
Figure 2.8 Photographs of outcrops and cut hand specimens representative of the Gidyen Volcaniclastic Member, Triangle Formation. (a) Photograph shows interbedded siliceous black shale, mudstone and fine-grained volcaniclastic sandstone near the contact between the Budhang Chert Member and overlying Gidyen Volcaniclastic Member northeast of the junction between Brisbane Valley and Parlour Creeks. (GR756127 6259908; Edith 1:25 000 topographic sheet, 8830-II-N). Hammer is 33 cm in length. (b) Irregularly jointed and lichen-covered outcrop typical of the Gidyen Volcaniclastic Member, located in a paddock beside the Goulburn Road at GR755861 6253541 (Edith 1:25 000 topographic sheet, 8830-II-N). Geological hammer is 33 cm long. (c) Cut hand specimen R16498 collected from Yellow Waterhole Creek at GR759173 6256487 (Edith 1:25 000 topographic sheet, 8830-II-N). This photograph shows interbedded siltstone and mudstone overlain by fining-upward, crystal-rich volcaniclastic sandstone. (d) Cut and polished hand specimen R16496 showing chert, mudstone and lithic sandstone pebble conglomerate that occurs at the base of some turbiditic flows in the Gidyen Volcaniclastic Member. Collected in Brisbane Valley Creek at GR755604 6260140 (Edith 1:25 000 topographic sheet, 8830-II-N). (e) Alternating interbedded fine-grained crystal-rich volcaniclastic sandstone and black mudstone in sample R16499 from near the top of the Gidyen Volcaniclastic Member. Collected from a paddock beside Goulburn Road at GR757410 6254768 (Edith 1:25 000 topographic sheet, 8830-II-N).
word 'Gidyen' meaning 'green'. The name is introduced here for volcaniclastic conglomerate, sandstone and mudstone conformably overlying the Budhang Chert Member in Brisbane Valley Creek and elsewhere in the western Oberon area (Enclosure 1). A type section is nominated along Brisbane Valley Creek downstream of the junction with Parlour Creek between GR756000 6261849 and 755303 6260485 (Edith 1:25 000 topographic sheet, 8830-II-N).

LITHOLOGY

The Gidyen Volcaniclastic Member is composed of grey-green, thick- and thin-bedded, mafic to intermediate volcaniclastic turbiditic conglomerate and sandstone with interbedded dark grey to black mudstone. It is generally characterised by poor outcrop except along creek beds, with the exposures typically altered by greenschist metamorphism and displaying cleavage. Bed thicknesses and grain size of sandstones generally increase up the section.

At the base of the unit, overlying the Budhang Chert Member in Brisbane Valley Creek (GR756136 6260090, Edith 1:25 000 topographic sheet, 8830-II-N), fine volcaniclastic sandstone horizons approximately 5 to 10 cm thick occur and are interbedded with black mudstone and shale (Figure 2.8a). Several metres of black shale overlie these basal thin-bedded volcaniclastic units, and are in turn overlain by alternating beds of volcaniclastic sandstone and mudstone that range 0.4 - 2.0 m thick.

Above these thin-bedded basal units, conglomerate beds are rare, but occasionally occur as thin, 10-30 cm beds of polymictic pebble conglomerate at the base of thick Bouma A division sandstone beds (Figure 2.8d). Clasts within the conglomerate are generally matrix supported and comprise chert, lenticular mudstone and intraformational mafic volcanic rock fragments with plagioclase microphenocrysts in a fine-grained volcaniclastic matrix. Sandstone beds range from 1-2 cm (Figure 2.8e) up to several metres thick (Figure 2.8b). In beds less than 1 m thick, sandstones are well sorted, generally fine-grained and grading is common. Sandstone beds have sharp bases and diffuse upper
contacts with overlying mudstone beds as shown in sawn faces of several samples (Figure 2.8c, e). The fine sandstone is composed of small (<1 mm) subhedral albitised feldspar crystal fragments and abundant angular to subrounded volcanic lithic fragments, whereas massive coarse-grained sandstone contains 1-2 mm euhedral to subhedral pyroxene, plagioclase and minor primary hornblende crystal fragments, along with subrounded volcanic rock fragments containing tabular plagioclase microphenocrysts in a pilotaxitic groundmass. A thin section from one sample located just below the upper boundary with the Mozart Chert Member, contains rare subrounded quartz grains exhibiting undulose extinction (sample R16487). Common alteration products in the sandstones include actinolite, chlorite, metamorphic amphibole and albitised plagioclase.

**RELATIONSHIPS AND THICKNESS**

The lower contact of the Gidyen Volcaniclastic Member is a conformable boundary with the underlying Budhang Chert Member, and is placed at the first occurrence of volcaniclastic sandstone observed in the sequence. West of the Brisbane Valley Fault in Brisbane Valley Creek (Enclosure 1), the upper contact with the overlying Mozart Chert Member is marked by a distinct change from volcaniclastic sandstone and interbedded black mudstone to blue-grey, thin-bedded (2-5 cm), siliceous mudstone and siltstone of the Mozart Chert Member exposed at GR754820 6261328 (Edith 1:25 000 topographic sheet, 8830-II-N). South of here, exposure is very poor and the contact is mapped by a distinctive colour change from red/orange to buff/grey in the weathered outcrop, soil and float. Southwest of the Vulcan Fault, a gradational contact runs approximately parallel to the Goulburn Road in an area of limited exposure. Coarse-medium grained volcaniclastic sandstone outcrops west of Goulburn Road at GR757043 6255371 (Edith 1:25 000 topographic sheet, 8830-II-N) and fine eastwards into medium to fine-grained lithic sandstone and mudstone east of Goulburn Road at GR757982 6255602 (Edith 1:25 000 topographic sheet, 8830-II-N). Further east the unit fines into pale siliceous siltstone and mudstone of the Mozart Chert Member occurring on Mozart Road at GR758175 6254905 (Edith 1:25 000 topographic sheet, 8830-II-N, Enclosure 1).
Between Brisbane Valley and Parlour Creeks, the Mozart Chert Member is not present and the Gidyen Volcaniclastic Member is in contact with the overlying upper volcaniclastics of the Triangle Formation (Enclosure 1). These two units are very similar and the contact between them is difficult to establish.

A true thickness of 200 m is estimated for the Gidyen Volcanic Member from Brisbane Valley Creek in the type area (Enclosure 1). East of the Brisbane Valley Fault, a minimum thickness of 450 m is estimated (Enclosure 1).

**AGE**
No age-specific fossils have been found within the Gidyen Volcaniclastic Member. Its stratigraphic position between the Budhang Chert Member and the Mozart Chert Member indicates that it is of Early to Middle Ordovician age (Figure 2.1).

**ENVIRONMENT OF DEPOSITION**
The Gidyen Volcaniclastic Member was deposited in a deep-marine environment as indicated by the fine-grained, laminated and siliceous nature of mudstone that probably formed by hemipelagic settling and the interbedded graded volcaniclastic sandstones of turbidite origin. Bed thicknesses and grain size of sandstone increases up the section, consistent with an increase in volcanism and/or uplift in the source area supplying the volcanic detritus. The lack of exposures along bedding planes has prevented measurement of sedimentary structures, such as groove or flute castes, which would indicate the direction of sediment transport.

**CORRELATIVES**
In the region west and northwest of the Oberon area the Gidyen Volcaniclastic Member is equivalent to the Triangle Formation that includes both the feldspathic sandstones from the Triangle Formation type section in Triangle Creek (Fowler & Iwata 1995) and crystal-rich volcaniclastic sandstone with
clinopyroxene, plagioclase and primary hornblende, from north of Rockley (Figure 1.4; Fowler 1987). The Gidyen Volcaniclastic Member is also correlated with the Coombing Formation mapped and described from west of the Copperhannia Thrust in the Blayney region (Figure 1.3; Wyborn 1998a). The Coombing Formation is considered Middle Ordovician in age because it is overlain by the Weemalla Formation, which contains Late Darriwilian graptolites (Wyborn 1998a). However, in contrast to the Gidyen Volcaniclastic Member, the Coombing Formation conformably overlies quartz turbidites of the Adaminaby Group (Wyborn 1998a). Locally sandstones of the Coombing Formation contain small amounts of quartz as has been found for one sample in the Gidyen Volcaniclastic Member (see above).

2.3.4 Mozart Chert Member (New Name)

DERIVATION AND TYPE SECTION
The Mozart Chert Member is a new unit introduced for interbedded deep-marine chert, siliceous mudstone and siltstone found in the Oberon area. The name is derived from the Mozart Parish in which the unit occurs. This unit is equivalent to the uppermost chert and siltstone unit described from the Triangle Formation type section in Triangle Creek (Fowler & Iwata 1995). This type section is retained, but a second section, representative of this unit in the Oberon area has also been nominated along Mozart Road, where overturned and steeply dipping, interbedded siltstone and mudstone occur between GR758175 6254963 and 758325 6254813 (Edith 1:25 000 topographic sheet, 8830-II-N; Enclosure 1; Figure 2.7e). The siltstone and mudstone are overlain by bedded chert, which is exposed in the Oberon Council quarry at GR758267 6254713 (Edith 1:25 000 topographic sheet, 8830-II-N; Enclosure 1).

DISTRIBUTION
The Mozart Chert Member is widely distributed west of the Vulcan Fault in the western part of the Oberon area, where it occurs in several plunging folds (Enclosure 1; Figure 2.7d, f, g).
LITHOLOGY
The Mozart Chert Member consists of pink and white weathered siliceous mudstone, siltstone and thin-bedded (5-10 cm) grey to honey-coloured ribbon cherts. In thin-section, the chert comprises cryptocrystalline quartz, partially dissolved radiolarian tests, and opaque organic material with small blebs of iron oxide and pyrite cubes. In outcrops on the Oberon Rockley Road in the northwestern part of the Oberon area (Enclosure 1), chert bands and 2-5 cm thick mudstone interbeds are interbedded with 0.5 to 2.0 m, fine-grained, massive siltstone layers (Figure 2.7g). X-ray diffraction analysis of a siltstone sample from this area indicates the presence of kaolinite, quartz, and goethite. The kaolinite is probably derived from the weathering of feldspars, whereas goethite has probably developed from the weathering of pyrite (Deer et al. 1992). Other siltstone samples collected from the Mozart Chert Member in Brisbane Valley Creek (sample R16488, GR754864 6261497, Edith 1:25 000 topographic sheet, 8830-II-N) and from the base of the undifferentiated Triangle Formation volcanics in Native Dog Creek (sample R16489, GR751342 6262260, Mt David 1:25 000 topographic sheet, 8830-III-N, Figure 2.9a) also contain substantial quantities of albitised feldspar in addition to quartz and sericite (Enclosure 1). Siltstone of the Mozart Chert Member is therefore considered to contain an appreciable intermediate volcaniclastic component.

RELATIONSHIPS AND THICKNESS
The Mozart Chert Member has conformable contacts with the underlying Gidyen Volcaniclastic Member (see above) and with overlying undifferentiated volcaniclastic sandstone and mudstone of the Triangle Formation (Enclosure 1). West of the O'Connell Road, the Mozart Chert Member is also overlain by coarse-grained mafic volcanioclastics of the Rockley Volcanics (Enclosure 1). Where observed in Brisbane Valley Creek, Native Dog Creek and on Chain-of-Ponds Road, the Mozart Chert Member is sharply overlain by volcaniclastic sandstone and mudstone of the upper Triangle Formation (Enclosures 1, 3). A similar sharp but conformable relationship is inferred for the rest of the Oberon area.
The thickness of the Mozart Chert Member is variable across the Oberon area. A minimum thickness of 600 m is exposed north of the Oberon and Sloggetts Granites and to the west of the Brisbane Valley Fault. Between the Brisbane Valley Fault and Parlour Creek, the Mozart Chert Member has a true thickness of only 250 m and lenses out further south at GR756485 6256818 (Edith 1:25 000 topographic sheet, 8830-II-N; Enclosure 1). East of Parlour Creek the unit has a true thickness of 650 m estimated from a cross section south of Mozart Road (Enclosure 1).

AGE
A chert sample from the corner of Langs Road and the Oberon Rockley Road yielded Pygodus conodont elements, indicating a late Middle to early Late Ordovician age (Da3-Gi1) for this unit (sample R15973; Enclosure 1; Figures 2.7f, g). Other samples collected from this unit, including at the Mozart Road Quarry and from along the O'Connell Road (Figure 2.7d), were either barren or only yielded fragments of conodont elements that were not age diagnostic.

ENVIRONMENT OF DEPOSITION
The presence of recrystallised radiolaria is consistent with deposition of chert in the Mozart Chert Member by pelagic and hemipelagic settling in a deep-marine environment. Feldspar and clays derived from altered feldspar in siltstone indicates a probable volcanic source. Their mode of deposition probably involved fine-grained material settling from fine plumes of volcanic-derived mud. The volcanic source was probably in the Molong Volcanic Belt that was active during the latest Middle Ordovician (Pogson & Watkins 1998; Meakin & Morgan 1999).

CORRELATIVES
Within the Molong Volcanic Province, the Mozart Chert Member is correlated with the unnamed cherty siltstone unit from the top of the Triangle Formation in Triangle Creek. This unit has yielded a late Middle to early Late Ordovician conodont fauna (Fowler & Iwata 1995) indicating the two units are of a similar age. Graptolites recovered from the basal portion of the Weemalla Formation in the
Blayney district (Figure 1.3) indicate that it has a Darriwilian to Gisbornian age, similar to that of the Mozart Chert Member (Wyborn et al. 1998). The base of the Weemalla Formation is marked by the incoming of coarse volcaniclastic pebbly sandstone and debris flows, which conformably overlie the fine-grained Coombing Formation (Wyborn et al. 1998). This coarsening upwards contrasts to the Oberon area where the succession fines upwards from the Gidyen Volcaniclastic Member into the overlying Mozart Chert Member. The Mozart Chert Member is also correlated with the deep-marine Numeralla Chert of the Quartzose Sedimentary Province.

2.3.5 Late Ordovician-?Early Silurian mafic volcaniclastics of the Triangle Formation
In this section mafic volcaniclastic rocks from the upper Triangle Formation above the Mozart Chert Member in the Oberon area are described. In the northeastern part of the Oberon area they are conformable on rocks of the Quartzose Sedimentary Province (Enclosures 1, 2; Figure 2.1). Two different facies are recognised and referred to as Facies A and B.

**DISTRIBUTION, LITHOLOGY, RELATIONSHIPS, THICKNESS AND AGE**

**Facies A**
Facies A occurs in the northwestern part of the Oberon area between the Native Dog Fault and Foleys Creek (Enclosure 1). It consists of thin- to very thick-bedded layers (1 cm up to tens of metres thick) composed of alternating fine-grained feldspathic and lithic sandstone, siltstone, black mudstone and chert (Figure 2.9a, b). A representative section occurs along the northwest-southeast trending section of Foleys Creek before its junction with Native Dog Creek (GR751909 6262090 to 752485 6261667, Mt David 1:25 000 topographic sheet, 8830-III-N, Enclosure 1).

Where fresh, the fine-grained, massive volcaniclastic sandstone and siltstone is dark grey to green, but weathers to an orange or creamy-white colour depending on the amount of iron-staining. The thick beds of fine volcaniclastic sandstone grade into mudstone and cherty shale. The fine volcaniclastic sandstone is composed of poorly to moderately sorted, euhedral to subhedral plagioclase crystals and
angular to subangular intraformational lithic clasts containing plagioclase laths, orthoclase, tiny pyroxene grains and quartz. Mudstone beds are dark coloured, but are compositionally similar to the fine sandstone and siltstone with moderately to well-sorted subangular to subrounded grains of plagioclase, pyroxene and quartz. They contain varying proportions of very fine black material, which is probably carbonaceous, and finely disseminated brown chlorite alteration is commonly present throughout the rock.

Facies A is conformable on the underlying Mozart Chert Member and is distinguished from it by the incoming of volcaniclastic siltstone and fine sandstone. The contact is poorly exposed east of Native Dog Creek between GR751660 6261982 and 751928 6261946 (Mt David 1:25 000 topographic sheet, 8830-III-N; Enclosure 1). Overlying and in part contiguous to Facies A are coarse-grained crystal-rich volcaniclastics of the Rockley Volcanics; the upper boundary of Facies A is marked by a basal conglomerate to the Rockley Volcanics (see below). Facies A varies in thickness from approximately 200 m in Foleys Creek (cross-section B-B’, Enclosure 1) up to 400 m further south between Hope and Native Dog Creeks (cross-section C-C’-C”, Enclosure 1). Although no fossils were found in Facies A, the base must be of early Late Ordovician age, based on its position above the Mozart Chert Member.

Facies B
Facies B outcrops to the northeast, east and southeast of Facies A in the western half of the Oberon area and east of the Oberon South Fault, in the northeastern part of the Oberon area (Enclosure 1). South of the Sloggetts and Oberon Granites, it outcrops west of the Vulcan Fault, and north of the two granites it is exposed in road cuttings west of the Oberon South Fault (Enclosure 1; Figure 2.9c). Facies B also occurs as small fault slivers south of Hampton Road (east of the Duckmaloi River at GR769682 6259385, west of the Duckmaloi River at GR768200 6255287, both Edith 1:25 000 topographic sheet 8830-II-N; Enclosure 1).
Figure 2.9 Photographs representative of Upper Ordovician undifferentiated volcaniclastics of the Triangle Formation. (a) - (b) Cut and polished hand specimens showing graded crystal-rich lithic sandstone and black mudstone layers indicative of secondary deposition as thin-bedded turbidites. (a) Sample R16489 collected from the southern side of Native Dog Creek, GR751438 6262294 (Mt David 1:25 000 topographic map sheet, 8830-III-N). (b) Sample R16490 collected from the eastern side of Native Dog Creek at GR753652 6258099 (Mt David 1:25 000 topographic map sheet, 8830-III-N). (c) Chert and mudstone pebble conglomerate from the base of the Late Ordovician volcaniclastics in an outcrop beside the Fish River on Hampton Road at GR767950 6265400; (Oberon 1:25 000 topographic sheet, 8830-1-S). (d) Cut and polished hand specimen R17999 showing cross-bedded crystal-rich volcaniclastic sandstone from beside the Fish River, north of Hampton Road at GR770268 6266112 (Oberon 1:25 000 topographic sheet, 8830-1-S). (e) Outcrop photograph of thin-bedded turbidites from undifferentiated volcaniclastics of the Triangle Formation, Edith Road beside the Fish River at GR765650 6265725 (Oberon 1:25 000 topographic sheet, 8830-1-S). Hammer for scale in bottom centre of photograph is 33 cm in length. (f) Cut and polished face of sample R17998 a chert and mudstone pebble conglomerate from the base of the Late Ordovician volcaniclastics collected from outcrop shown in photograph 2.9c beside the Fish River at GR766976 6264510 (Oberon 1:25 000 topographic sheet, 8830-1-S). (g) Bedded volcaniclastic sandstone and mudstone outcropping beside Hampton Road at GR769611 6264647 (Oberon 1:25 000 topographic sheet 8830-1-S). Compass is 7 cm across.
Facies B consists of thick to very thick beds (0.5-30 m) of crystal-rich and lithic sandstone interbedded with thin-bedded lithic siltstone commonly exhibiting cross-lamination and shaly mudstone (Figure 2.9d). Sandstones are fine- to coarse-grained. Conglomerates occur at the base of sandstone beds. They are composed of pebbles of intraformational crystal-rich volcanioclastic sandstone, chert and mudstone and cobbles of mafic pyroxene and plagioclase porphyritic basalt in a crystal-rich volcanioclastic matrix (Enclosure 2, Figure 2.9e, f). Most sandstone beds are massive or weakly graded, consistent with deposition by mass flow processes (turbidites and megaturbidites).

The volcanioclastic sandstone is dominated by relict euhedral to subhedral clinopyroxene, plagioclase and minor primary hornblende crystals, together with volcanic rock fragments and rare chert fragments. Facies B is less feldspathic than Facies A and has a greater abundance of clinopyroxene crystals. In the area east the Oberon South Fault, sandstones include coarse-grained rounded polycrystalline quartz grains with granoblastic textures. In the area east of the Deep Creek Fault, samples R16491, R16492 and R16493 contain subhedral to anhedral zircon grains. Within granite aureoles, clinopyroxene has been altered to actinolite, chlorite and epidote. Plagioclase has been albitioned and altered to chlorite. Fine-grained biotite is also abundant.

No complete reference section can be nominated for this facies due to the incomplete exposure. Outcrops along the Oberon Rockley Road at GR755038 6266344 and 755713 6266546 (Oberon 1:25 000 topographic sheet, 8830-1-S, Enclosure 1) are representative. Another representative section is along Hampton Road between GR769552 6264697 and 769672 6264592 (Oberon 1:25 000 topographic sheet, 8830-1-S).

Facies B conformably overlies the Mozart Chert Member and this contact is poorly exposed west of the Vulcan Fault (north of Mozart Road at GR760236 6255204, Edith 1:25 000 topographic sheet, 8830-II-N; Enclosure 1). East of Brisbane Valley Creek, the Mozart Chert Member is absent, and mafic to intermediate volcanioclastic of Facies B conformably overlie the Gidyea Volcanioclastic.
Member. There is very little difference between these two members and the contact is only inferred (Enclosure 1). East of the Vulcan Fault, Facies B conformably and sharply overlies Warbisco Shale of the Quartzose Sedimentary Province, exposed in a road cutting at GR763000 6266848 (Oberon 1:25 000 topographic sheet, 8830-1-S; Enclosures 1, 2). The upper contact of Facies B, as for Facies A, is at a conglomerate that marks the base of the Rockley Volcanics in Brisbane Valley Creek (Enclosure 1). East of Oberon, the upper contact at GR769658 6264606 (Oberon 1:25 000 topographic sheet, 8830-1-S) is at the base of the overlying Bald Ridge Formation. The poor exposure and the lack of age-specific fossils prevent an accurate assessment of the nature of this contact, but bedding relations imply it is conformable (Enclosure 2). East of the Brisbane Valley Fault Facies B has a true thickness of 570 m (cross-section G-G', Enclosure 1) and thickens towards the southeast, with a minimum thickness of 1200 m on the western side of the Vulcan Fault (cross-section H-H', Enclosure 1). East of Oberon, along Hampton Road only a minimum thickness of 190 m can be estimated for Facies B (cross-section E-E', Enclosure 2).

No age data has been found within this facies. As for Facies A, the occurrences west of the Vulcan Fault (Enclosure 1) are conformable above the Mozart Chert Member and therefore the base of the unit is at least early Late Ordovician. Further east, the base of Facies B is possibly younger because the base of the Warbisco Shale is commonly late Gisbornian or slightly younger (see Section 2.3.2), although no fossils have been recovered from the unit in the Oberon area.

Facies B is overlain by the Rockley Volcanics, and where these are absent, it is unconformably overlain by coarse-sandstone and conglomerate of the Upper Silurian Fosters Creek Conglomerate or fine-grained siltstone and shale of the Upper Silurian Campbells Formation (Enclosure 1). Its upper age limit is relatively poorly constrained (see below). South of Hampton Road, volcanioclastics of Facies B occur as small fault slivers between quartz turbidites of the undifferentiated Adaminaby Group and Bald Ridge Formation (Enclosure 1 and see above).
ENVIRONMENT OF DEPOSITION
Facies A and B are interpreted as being deposited from mass flows, i.e. turbidites and megaturbidites, in a deep marine environment. They are thought to have formed as a volcaniclastic apron that extended off the eastern margin of the Molong High (see Chapter 5).

CORRELATIVES
Mafic volcanics of probable Late Ordovician to Early Silurian age are widely developed in the region south of Bathurst and are considered equivalents of Facies A and B of the Triangle Formation. North of Rockley, the Triangle Formation is abundant and dominated by mafic volcanics with abundant clinopyroxene crystals (formerly included within the “Carinya Formation” by Fowler 1987).

South of the Oberon area, inferred Late Ordovician to Early Silurian mafic volcanics contain interbedded conglomerate, sandstone and black mudstone. These rocks have been reported from the Jaunter, Isabella and Porters Retreat areas, and along the northern edge of the Taralga 1:100 000 map sheet (Figures 1.4; Shiels 1959; Dunnett 1961; Packham 1969; Scheibner 1970; Scott 1991). Sandstones in these units are crystal-rich and contain clinopyroxene, plagioclase and hornblende. They have previously been correlated with the Rockley Volcanics (Dunnett 1961; Scheibner 1970) and the Carinya Formation (Scott 1991). In both the Jaunter and Isabella districts, these mafic volcanics overlie or are interbedded with thick-bedded turbidites that contain quartz sandstone with bimodal textures (Dunnett 1961; Scott 1991). Possible Warbisco Shale has been mapped at the base of the quartz turbidites (Scott 1991). These quartz turbidites are more typical of the ?Lower Silurian Bald Ridge Formation, which in the Thompson Creek area also is interbedded with actinolite schist and conglomerate near the top of the unit.

Regionally, the uppermost Triangle Formation is correlated with other Upper Ordovician mafic volcaniclastic units occurring on the Molong High. The Weemalla Formation conformably overlies the Coombing Formation 20 km southwest of Blayney (Figure 1.3), and ranges from late Darriwilian
to late Eastonian in age. It comprises both volcaniclastic and primary volcanic rocks deposited in a shallow-marine environment and composed predominantly of feldspathic and lithic sandstone and conglomerate, along with tuffs and clinopyroxene porphyritic basalt flows and sills (Wyborn et al. 1998). Resedimentation of rocks initially deposited as part of the Weemalla Formation, may be the source of some of the upper Triangle Formation detritus.

Overlying the Coombing Formation in the Blayney area (Figure 1.3) are the upper Middle to lower Upper Ordovician Blayney Volcanics, which consist of porphyritic basalt flows, breccias and pillow lavas predominantly composed of augite and plagioclase (Wyborn 1998b). These volcanic rocks have a similar mineralogy to the upper Triangle volcaniclastics in the Oberon area and are another possible source of volcaniclastic detritus for volcaniclastics deposited in the Oberon area during the early Late Ordovician.

In the Mudgee area north of the Bathurst Granite, massive volcaniclastic sandstone, siltstone and chert have been described (the Coomber Formation of Fergusson & Colquhoun 1996). Although unfossiliferous, the Coomber Formation also conformably overlies thin-bedded quartz turbidites occurring at the top of the Adaminaby Group considered to be of late Middle to early late Ordovician age (Fergusson & Colquhoun 1996). However, the source of these volcaniclastics was probably different to the upper Triangle Formation, because these volcaniclastics are more potassium-rich, and have a higher (bright red) radiometric signature than the volcaniclastics in the Oberon area.

2.3.6 Rockley Volcanics

DERIVATION, NOMENCLATURE AND PREVIOUS WORK
The name Rockley Volcanics was derived from the village of Rockley, south of Bathurst, west of which these rocks are at their thickest extent (Stanton 1956). Although no type section was ever established, Stanton (1956) described the Rockley Volcanics as consisting predominantly of "andesitic pyroclastics" with occasional flows or sills. He considered them to lie conformably above
Figure 2.10 Photographs representative of the Rockley Volcanics in the Oberon Area. (a) Cut and polished face of poorly sorted crystal-rich sandstone. Crystals are predominantly clinopyroxene with some relict olivine and cross-cutting veins are composed of calcite. Sample R15963 collected from north Oberon at GR765824 6267080 (Oberon 1:25 000 topographic sheet 8830-1-S). (b) Conglomerate composed of subangular to rounded clasts of mafic volcanics or volcaniclastics which retain primary volcanic textures from Hope Creek, GR750974 6258867 (Mt David 1:25 000 topographic sheet, 8830-III-N). Geological hammer head is 18 cm in length. (c) Altered and deformed mafic volcaniclastic pebble conglomerate beside Chain-of-Ponds Road, GR752083 6262646 (Mt David 1:25 000 topographic sheet, 8830-III-N). Head on sledgehammer is 20 cm wide. (d) Sample R16494 collected from paddock beside Chain-of-Ponds Road (GR752130 6262475; Mt David 1:25 000 topographic sheet, 8830-III-N). The crystals in this sample are altered to epidote. (e) Cut and polished face of sample R16497 - conglomerate containing subangular to rounded clasts of mudstone and fine volcaniclastic siltstone occurring at the base of the Rockley Volcanics in Native Dog Creek, GR750223 6258730 (Mt David 1:25 000 topographic sheet, 8830-III-N). (f) Cut and polished face of sample R16497 - breccia/conglomerate containing angular clasts of chert, mudstone and fine volcaniclastic sandstone occurring at the base of the Rockley Volcanics in Brisbane Valley Creek, GR755323 6254596 (Edith 1:25 000 topographic sheet, 8830-II-N). (g) Bedded layers of altered volcaniclastics rich in serpentine minerals outcropping on the southern side of Native Dog Creek, GR751221 6262176 (Mt David 1:25 000 topographic sheet, 8830-III-N). Hammer is 28 cm long.
the Triangle Formation and below the Campbells Formation (Triangle Group and Campbells Group of Stanton 1956).

During compilation of the first edition of the Bathurst 1:250 000 sheet, the Rockley Volcanics were mapped east and southeast to include “andesites and tuffs” in the Oberon and Jaunter areas (Packham 1966, 1969), and on the northern margin of the Taralga 1:100 000 sheet (Scheibner 1970, 1973; Figure 1.3). Fowler (1987) remapped the Rockley Volcanics in the Rockley district and divided them into two units, the Rockley Volcanics and the now abandoned "Carinya Formation", based on the Carinya Formation containing primary hornblende whereas the Rockley Volcanics were more mafic, containing olivine and no primary hornblende.

During mapping of the second edition of the Bathurst 1:250 000 sheet, the Rockley Volcanics, including the "Carinya Formation" of Fowler (1987), were subdivided into three unnamed units referred to as Ocrc, Ocru and Ocrs along with ‘undifferentiated’ Rockley Volcanics termed Ocr (Raymond et al. 1997; Stuart-Smith et al. 1998). All of these subunits together with other Late Ordovician to earliest Silurian volcanic and volcaniclastic sequences of the Molong and Rockley-Gulgong Volcanic Belts were included within the Upper Ordovician Cabonne Group (Raymond et al. 1997; Stuart-Smith et al. 1998).

Of all the subunits recognised within the Rockley Volcanics, only Ocrc and the undifferentiated Ocr were recognised east of the Native Dog Fault in the Oberon area (Stuart-Smith et al. 1998). These units have been replaced by the Mozart Chert Member, the Gidyen Volcaniclastic Member, the upper undifferentiated Triangle Formation and the Rockley Volcanics in the Oberon area (Enclosure 1).

**DISTRIBUTION**

In the Oberon area, the Rockley Volcanics occurs in three places (Enclosure 1). The most extensive area of outcrop occurs between the Native Dog and Brisbane Valley Faults along Native Dog and Hope
Hope Creeks. They also occur on the eastern side of the Brisbane Valley Fault along Brisbane Valley Creek and they form a narrow north-south trending ridge of outcrop just east of Oberon adjacent to the Oberon South Fault.

**LITHOLOGY**

In the Oberon area, the Rockley Volcanics comprises volcaniclastic breccia and conglomerate with associated coarse- to fine-grained sandstone and black mudstone (Figure 2.10a, b, d). The conglomerates are matrix supported and contain angular to rounded porphyritic basalt clasts (Figure 2.10b). The sandstone ranges from coarse- to fine-grained and commonly contains large euhedral clinopyroxene crystals (up to 7 mm) some of which are pseudomorphs after olivine, and smaller plagioclase crystals (up to 5 mm in length). No primary volcanic rocks were identified, but primary volcanic textures preserved in boulders and cobbles indicate that the clastic rocks are all derived from primary mafic volcanic rocks. The Rockley Volcanics in Brisbane Valley Creek comprise coarse- to fine-grained sandstone, with clinopyroxene and plagioclase crystals and lack breccia and conglomerate.

A conglomerate commonly occurs at the base of the Rockley Volcanics in both Native Dog and Brisbane Valley Creeks. In Native Dog Creek, the conglomerate is composed of pebbles of lenticular mudstone in a matrix of black mudstone (Figure 2.10e). In Brisbane Valley Creek, the basal conglomerate comprises angular basaltic rock fragments in a volcaniclastic matrix (Figure 2.10f).

Serpentine-rich rocks with growths of secondary fibrous amphibole were observed on Chain-of-Ponds Road at GR752083 6262646 (Rockley 1:25 000 topographic sheet, 8830-4-S; Enclosure 1; Figure 2.10c, g). The presence of relict bedding indicates that these rocks are clastic rather than altered primary igneous rocks.

The Rockley Volcanics in the Oberon area have been affected by regional greenschist facies
metamorphism. Metamorphic mineral phases include albite, epidote, amphibole, chlorite, calcite and titanite, whereas the only relict phase preserved is clinopyroxene (see Chapter 5). Primary porphyritic and pilotaxitic groundmass textures are still preserved in volcanic lithic fragments.

RELATIONSHIPS AND THICKNESS
The lower contact between Facies A of the uppermost Triangle Formation and the Rockley Volcanics occurs in Native Dog Creek over approximately 5 m between GR751529 6259913 and 751516 6259858 (Mt David 1:25 000 topographic sheet, 8830-III-N; Enclosure 1). Rapid thickness changes in the upper Triangle Formation above the Mozart Chert Member and below the Rockley Volcanics apparent from the map pattern around Native Dog Creek (see Enclosure 1) are consistent with an irregular contact between both units. In the Oberon area, the Rockley Volcanics are unconformably overlain by Tertiary Basalt and Quaternary river sediments. Adjacent to the Native Dog Fault, the Rockley Volcanics dip unconformably beneath the Upper Silurian Bells Creek Volcanics.

In the vicinity of Native Dog Creek, a minimum thickness of 200-300 m is estimated for the Rockley Volcanics. Extreme thickness variations were noted in the Rockley district, with a maximum thickness estimated at 1500 m and lensing out elsewhere (Fowler 1987).

CORRELATION AND AGE
No age-specific fossils have been found within the Rockley Volcanics. Age constraints are therefore based on stratigraphic relationships. Within the Oberon area, the Rockley Volcanics conformably overlie the undifferentiated mafic to intermediate volcanioclastics of the Triangle Formation which also lack age-specific fossils, but the lower part of this succession must be no older than Gisbornian (see above). In the Rockley area (Figure 2.1), the Rockley Volcanics overlie the latest Darriwilian-Gisbornian chert of the Triangle Formation (Fowler & Iwata 1995), which is correlated with the Mozart Chert Member (see above). The Rockley Volcanics are unconformably overlain by the Upper Silurian Bells Creek Volcanics. Thus, the upper limit of the unit is poorly constrained. Therefore, a
general Late Ordovician to Early Silurian age is recognised for the Rockley Volcanics, although given the differences in underlying units in the Rockley and Oberon areas its base may well be diachronous.

Regionally, the Rockley Volcanics are correlated with other Late Ordovician mafic volcanic and volcanoclastic units in the Molong Volcanic Province, all of which are included within the Cabonne Group. Units occurring west of the Oberon area on the Molong High include the Blayney Volcanics, Byng Volcanics, Weemalla Formation and Forest Reefs Volcanics. In the Blayney Region (Figure 1.3), the Blayney Volcanics consist of porphyritic basalt flows, breccias and pillow lavas predominantly composed of augite with subordinate olivine and plagioclase (Wyborn 1998). The Blayney Volcanics are of late Middle Ordovician to early Late Ordovician age. The Byng Volcanics occur south and east of Orange (Figure 1.3) and comprise augite, plagioclase and olivine porphyritic basalt to basaltic-andesite flows, and associated volcanoclastic sandstone and siltstone (Scott 1998). They conformably overlie the Blayney Volcanics and based on their stratigraphic position are considered to have an early Late Ordovician age (Scott 1998). The Weemalla Formation occurs 20 km southwest of Blayney (Figure 1.3) and comprises feldspathic and lithic sandstone and conglomerate, limestone, calcareous mudstone along with tuffs and clinopyroxene porphyritic basalt flows and sills (Wyborn et al. 1998). Fossils indicate an age from late Darriwilian (Da3/4) to late Eastonian (Ea3/4). Conformably above the Weemalla Formation, the Forest Reefs Volcanics constitute a major Late Ordovician volcanic complex in the Blayney area, composed of basaltic lava, breccia, matrix supported conglomerate and associated volcanic sandstone and ash (Wyborn 1998).

North of the Bathurst Batholith, the Sofala Volcanics are composed of undifferentiated minor basaltic andesite with associated volcanoclastic products, chert and limestone. Graptolites, conodonts and corals have been recovered from chert and mudstone units at both the base and near top of the Sofala Volcanics (Watkins 1998). These faunas are indicative of Gisbornian to Eastonian ages for the base of the Sofala Volcanics and a Bolindian (Bo3) age near the top (Packham 1968; Pickett 1978; Rickards et al. 1998; Percival 1999a).
Volcaniclastic rocks containing pyroxene and plagioclase crystals occur near the top of the Sofala Volcanics in the Palmers Oakey district (Figure 1.4), where they are overlain by late Early Silurian limestone and shale (Bischoff & Fergusson 1982). Parts of the Rockley Volcanics in the Oberon area may also be equivalent to these Latest Ordovician to Early Silurian volcaniclastics, but are not distinguished from the Late Ordovician rocks.

ENVIRONMENT OF DEPOSITION
In the Oberon area, all exposures of the Rockley Volcanics lack any massive flows, pillow basalts and hyaloclastites that would provide evidence for primary volcanism. The abundance of massive conglomerate and breccia and the lack of evidence for shallow marine and subaerial environments are indicative of deposition as debris flows and turbidites (Cas & Wright 1987, p. 308-327). These were probably deposited down-slope from the volcanic vents, from which the clast types were originally erupted, in a deep-marine environment. The presence of well-rounded cobbles, in addition to many angular clasts, indicates that at least some abrasion of clasts has occurred in a subaerial to littoral environment prior to resedimentation in deeper water.

Although no evidence of primary volcanism is apparent from the Oberon area, further west in the Rockley area Fowler (1987) identified several possible eruptive centres. Pillow lavas, surrounded by mafic breccias occur in Badgers Creek (Figure 1.4). A cluster of peridotite and picrite pipe-like intrusions surrounded by abundant breccia is exposed at Dogs Rocks, 5 km southeast of Rockley (Figure 1.4). These are possible shallow intrusions beneath a volcanic vent in the Rockley Volcanics (Fowler 1987). Both of these sites are 6-10 km west of the Oberon area and represent a potential source region for the volcaniclastics of the study area. Primary volcanism on the Molong High during the Late Ordovician included volcanic suites within the Blayney Volcanics, Byng Volcanics and Forest Reefs Volcanics that are compositionally similar to the Rockley Volcanics. Limestone bodies interbedded within the Molong High volcanic units indicate that a shallow marine environment existed on parts of the Molong High during the Late Ordovician. These eruptive centres on the
Molong High also represent a potential source region for volcanioclastics of the Oberon area.

2.3.7 Fosters Creek Conglomerate

DERIVATION AND TYPE SECTION
The name "Fosters Creek Conglomerate" is derived from the Fosters Creek homestead, which occurs in the type area (Figure 1.3; Wallace & Stuart-Smith 1994). The type section is in Byrnes Creek, 20 km southwest of Rockley, where the Fosters Creek Conglomerate occurs as a steeply-dipping succession composed of thickly bedded oligomictic boulder conglomerate containing meta-feldspathic quartz sandstone clasts, sandstone and siltstone (Wyborn, Henderson et al. 1998). Further north in the Rockley area, clasts in the conglomerate consist of mafic and ultramafic volcanics derived from the underlying Rockley Volcanics (Wyborn, Henderson et al. 1998).

DISTRIBUTION
Within the Oberon-Taralga Region, the Fosters Creek Conglomerate has been recognised in the Thompson Creek and Oberon areas (Enclosures 1, 4). In the Oberon area, conglomerate and sandstone of the Fosters Creek Conglomerate occur beside Native Dog Creek and north of the Goulburn Road (Enclosure 1). In the Thompson Creek area, the Fosters Creek Conglomerate occurs on Bald Ridge Road and beside Thompson Creek on the western margin of the Thompson Anticline (Enclosure 4).

LITHOLOGY
In the Thompson Creek area, the Fosters Creek Conglomerate is composed of conglomerate with subangular to subrounded clasts of medium to fine-grained quartz sandstone. Clasts are between 2 and 20 cm in diameter and are matrix supported (Figure 2.11b). The matrix is also composed of fine grained quartz sandstone.

In the Oberon area, the conglomerate is composed of rounded chert clasts that are flattened within the S2 cleavage. The conglomerate is matrix supported and the clasts range in size from 1 cm up to 20 cm
Figure 2.11 (a) Conglomerate containing flattened chert and quartz sandstone clasts of Late Silurian age, Chain-of-Ponds Road (GR752255 6263526, Rockley 1:25 000 topographic sheet, 8830-4-S). Hammer head is 18 cm wide. (b) Fosters Creek Conglomerate containing cobbles and clasts of fine-grained quartz sandstone from the western side of the gateway to “Forest Hills” property at GR723655 6241376 (Abercrombie 1:50 000 topographic sheet, 8730-II and III). Hammer is 33 cm in length.
in diameter (Figure 2.11a). The matrix is composed of fine-grained quartz sandstone with larger 1-2 mm well-rounded and highly spherical monocrystalline and polycrystalline quartz grains; thus, the matrix has a bimodal texture similar to that of the Bald Ridge Formation. North of the Goulburn Road at GR761525 6268000 (Oberon 1:25 000 topographic sheet, 8830-1-S), clast sizes within the conglomerate are smaller than elsewhere, with chert clasts only up to 1-2 cm in size.

RELATIONSHIPS AND THICKNESS
In the western part of the Thompson Creek area, the Fosters Creek Conglomerate is a thin unit up to 5 m thick, occurring at the base of the Kangaloolah Volcanics and unconformably overlying quartz turbidites and actinolite schist of the Bald Ridge Formation (Enclosure 4).

In the Oberon area, this unit overlies volcanioclastics of both the upper Triangle Formation and the Rockley Volcanics with an angular unconformity. Both the Upper Silurian Bells Creek Volcanics and the Campbells Formation conformably overlie it. At Native Dog Creek, the thickness of the Fosters Creek Conglomerate is 20 m thick and further north lenses out completely before the Beaconfield-O‘Connell Road (Enclosure 1).

AGE
An undescribed and unpublished shelly fauna was collected from weathered outcrops of the Fosters Creek Conglomerate north of the Goulburn Road at GR761525 6268000 (Oberon 1:25 000 topographic sheet, 8830-1-S) by McMahon (1959). This fauna was re-examined by D. Strusz, who indicated an age tentatively estimated at Wenlock to early Ludlow (D. Strusz, personal communication 1997).

ENVIRONMENT OF DEPOSITION
The conglomerate is locally derived as indicated by the types of clasts. Clasts are typically well rounded with a high sphericity consistent with deposition in a fluvial environment rather than a beach
environment. These features are consistent with this unit representing a localised deposit resulting from the reworking of local basement rock types following uplift during the late Early Silurian.

2.4 DEPOSITIONAL HISTORY OF THE OBERON-TARALGA REGION DURING THE ORDOVICIAN TO MIDDLE SILURIAN

The Ordovician to Early Silurian depositional history of the Oberon-Taralga region is characterised by a deep-marine environment with volcanoclastic sedimentation in the Molong Volcanic Province and siliciclastic deposition in the Quartzose Sedimentary Province (Figures 2.12, 2.13). An overall convergent margin tectonic setting is recognised with the Molong Volcanic Province forming an island arc related to a west-dipping subduction zone (Glen et al. 1998). Deep-marine sedimentation has therefore taken place in a frontal arc to forearc basin setting associated with the island arc. The history of each province is considered in turn before the depositional relationships between them are examined.

2.4.1 Molong Volcanic Province

The rocks of the Molong Volcanic Province formed in two depositional settings (Webby 1976; Glen et al. 1998). These are (1) a deep-marine setting where volcanic and volcanoclastic deposits are interbedded with pelagic and hemi-pelagic deposits of chert and black mudstone/shale, and (2) a shallow marine to emergent setting where subaerial to submarine erupted volcanic and related volcanoclastic deposits are associated with shallow-marine limestone. The volcanic edifices shed volcanogenic detritus onto surrounding volcanoclastic aprons in the deep-marine basin (Fergusson & Colquhoun 1996; Murray & Stewart 2001, Chapter 3). The western part of the Molong Volcanic Province include volcanic centres in both the Molong and Junee-Narromine Volcanic Belts that mark the location of the island arc (Glen et al. 1998).

The oldest unit in the Molong Volcanic Province of the Oberon area is the Bendigonian Budhang Chert Member of the Triangle Formation. This unit formed in a phase of pelagic to hemi-pelagic
Figure 2.12 Stratigraphic columns recognised from various localities within the Oberon-Taralga Region. Oberon east and Oberon west columns are from the Oberon area referred to in this study. Thompson Creek area, Silent-Oakey Creeks area, Golspie area and Crookwell area are also related to this study. Stratigraphic column shown from the Rockley area is from Fowler (1987) and Fowler and Iwata (1995). Column for east of Jaunter is from Scott (1991). Column for the Isabella area is from Dunnett (1961). Column for the area east of Taralga is from Whalan (1986) and Roots et al. (1986). See Enclosures 1 to 7 for legend.
deposition. This matches the record from other parts of the Molong Volcanic Province south of Wellington and near Parkes where fine-grained deposition occurred at the same time (Sherwin 1993; Percival et al. 1999). In contrast to these other localities, no evidence of volcanism accompanying deposition of the Budhang Chert Member has been found. Deep-marine sedimentation continued for some time, as indicated by the organic-rich, black siliceous mudstone and shale overlying the chert.

In the Oberon-Taralga region two major pulses of volcanioclastic deposition occurred that are separated by a phase of pelagic to fine-grained deposition (Mozart Chert Member, Figure 2.13). The first major pulse of volcanioclastic sedimentation is shown by deposition of the Gidyen Volcaniclastic Member. The timing of the initiation of this pulse of volcanioclastic material is poorly constrained because of the lack of age control between the Bendigonian conodonts of the lower Budhang Chert Member and the late Darriwilian - earliest Gisbornian Mozart Chert Member. Either there has been a long hiatus between the Bendigonian and the Darriwilian rocks or the Budhang Chert Member and overlying Gidyen Volcaniclastic Member represent over 25 Ma of reasonably continuous deposition. The former alternative is favoured, because elsewhere in the Molong Volcanic Province only Darriwilian and younger ages have been established for the bulk of the volcanic succession (Packham et al. 1999).

The volcanic detritus comprising the Gidyen Volcaniclastic Member was probably derived from the Molong and Junee-Narromine Volcanic Belts. Possible sources for the Gidyen Volcaniclastic Member include volcanioclastic conglomerate and breccia of the Early Ordovician Mitchell Formation of the Molong Volcanic Belt, and/or the younger Darriwilian Walli, Fairbridge and/or basal part of the Goonumbla Volcanics from the Molong and Junee-Narromine Volcanic Belts (Figure 1.3). The abundant immature volcanic detritus in the Gidyen Volcaniclastic Member is consistent with synchronous volcanism in the source area. Given the late Darriwilian - earliest Gisbornian age of the overlying Mozart Chert, the late Darriwilian Walli, Fairbridge and/or basal part of the Goonumbla Volcanics are considered more likely as the source of the Gidyen Volcaniclastic Member.
Figure 2.13 Stratigraphic diagram for the Oberon-Taralga region, showing the time and space relationships between units of the Molong Volcanic Province (on left) and the Quartzose Sedimentary Province (on right). The Ordovician sea level curve is based on data from the Australian cratonic basins (from Nicoll & Webby 1996).
The first pulse of volcaniclastic deposition represented by the Gidyen Volcaniclastic Member was followed by pelagic and fine-grained sedimentation of the Mozart Chert Member. This phase of fine-grained deposition also occurred in the Rockley area (Fowler & Iwata 1995). Further west on the Molong High the Weemalla Formation of the same Darriwilian-Gisbornian age is marked by coarse volcaniclastic deposition (Wyborn et al. 1998). The base of the Weemalla Formation contains autochthonous shallow-marine limestone horizons in contrast to the deep-marine Mozart Chert Member. Only minor amounts of volcaniclastic material were deposited in the Mozart Chert Member.

Deposition of the Mozart Chert Member was followed by the second major pulse of volcaniclastic deposition in the Oberon area. This began in the early Late Ordovician with widespread deep-marine turbidite and megaturbidite deposition of immature volcanic detritus of the uppermost Triangle Formation and the Rockley Volcanics. These volcaniclastics occur over a much wider region than the earlier pulse and extended eastwards to overlie the Warbisco Shale of the Quartzose Sedimentary Province (Figures 2.12, 2.13). They are probably related to increased volcanic activity on the volcanic highs. The Triangle Formation reflects a mafic to intermediate volcanic source, whereas the Rockley Volcanics are more mafic in character. Similar areas with predominantly mafic and mafic to intermediate volcaniclastic deposition occur elsewhere in the Molong Volcanic Province (Fergusson & Colquhoun 1996; Glen et al. 1998; Watkins 1998; Meakin & Morgan 1999).

The timing of cessation of deposition associated with the upper pulse of volcaniclastic sedimentation is poorly constrained. Deposition may extend into the early Silurian. An upper constraint is provided by the unconformity with the overlying late Early to Late Silurian Fosters Creek Conglomerate and Bells Creek Volcanics.

2.4.2 Depositional History of the Quartzose Sedimentary Province

The stratigraphy of the Quartzose Sedimentary Province in the Oberon-Taralga region comprises a fining upward succession from Middle Ordovician quartz turbidites of the undifferentiated Adaminaby
Adaminaby Group, through the Numeralla Chert, and thin-bedded turbidites of the Bumballa Formation overlain by the Late Ordovician Warbisco Shale. This is similar to the regional stratigraphy recognised from east of Goulburn in the Shoalhaven River, southern New South Wales and eastern Victoria (VandenBerg & Stewart 1992; Glen 1994; Fergusson 1998b). The base of the Quartzose Sedimentary Province is not observed in the Oberon-Taralga region. No evidence was found that the undifferentiated Adaminaby Group is any older than late Middle Ordovician as has been documented on the south coast of New South Wales and in Victoria.

The Numeralla Chert is recognised as having a widespread distribution across the Eastern Belt of the Lachlan Fold Belt (VandenBerg & Stewart 1992; Glen 1994; Stewart & Fergusson 1995; Murray & Stewart 2001; see Chapter 3). Conformably above and in part contiguous to the Numeralla Chert, the Bumballa Formation occurs in all of the areas where the Quartzose Sedimentary Province units were mapped. The widespread distribution of the Numeralla Chert and Bumballa Formation during the late Darriwilian and early Gisbornian indicates a reduction in the sediment supply to the middle and outer parts of the submarine fan. Chert deposition varies between the middle Darriwilian and late Darriwilian to early Gisbornian.

The Warbisco Shale is the highest unit in the fining upward succession of the Quartzose Sedimentary Province. This was followed by a return to thick-bedded turbidite deposition of the ?Lower Silurian Bald Ridge Formation.

2.4.3 Relationship between the Provinces

Early-Middle Ordovician
No clear stratigraphic relationship exists between the two lithological provinces during the Early and early Middle Ordovician. The Bendigonian Budhang Chert Member is much older than the Middle Ordovician thick-bedded turbidites of the undifferentiated Adaminaby Group. In the Oberon area, Early Ordovician rocks are only observed in the Molong Volcanic Province, where they indicate a
Figure 2.14 Schematic block diagrams of the palaeogeography of the Oberon-Taralga region throughout the Ordovician. (a) The Oberon/Taralga region is situated in a deep-marine pelagic environment outboard of and not influenced by sedimentation from the volcanic chain. Terrigenous quartz turbidites deposition in a marginal sea separates an active volcanic islands from the Gondwanan continent. (b) Deposition of quartz turbidites indicates the continental margin submarine fan has extended into the Oberon-Taralga region following a Bendigonian to Darriwilian stratigraphic hiatus across both lithological provinces in the north of the Eastern Belt of the Lachlan Fold Belt. The quartz turbidites were deposited adjacent to and outboard of the eastern margin of the Molong Volcanic Province where volcaniclastic deposition derived from the Junee-Narromine and Molong Volcanic Belts occurred. (c) Volcanism shifted to seamounts to the east away from the Junee-Narromine and Molong Volcanic Highs, where minor primary volcanics were interbedded with limestone clasts and large volumes of volcaniclastic detritus derived from the topographic highs and deposited as turbidites in the basins on the flanks of the highs in what is now the eastern portion of the Molong Volcanic Belt and the Rockley-Gulgong Volcanic Belt. Anoxic conditions resulted in black shale deposition to the east and south away from the island volcanic chain.
(a) Bendigonian

(b) Darriwilian

(c) Eastonian

Ocean
Mafic Volcaniclastics
Black Shale

Quartz Turbidites Pelagic deposits Oceanic Crust
prolonged period of deep-marine pelagic and minor hemipelagic sedimentation outboard of the volcanic high to the west (Figure 2.14a).

**LATE MIDDLE ORDOVICIAN (EARLY TO MIDDLE DARRIWILIAN)**
The Gidyen Volcaniclastic Member is of probable late Middle Ordovician age, and it overlies the Early Ordovician pelagic deposits of the Budhang Chert Member in the Molong Volcanic Province. As discussed above, either there has been a long hiatus in deposition or undocumented continuous deposition has occurred, possibly in the upper part of the Budhang Chert Member. The undifferentiated Adaminaby Group quartz turbidites of the Quartzose Sedimentary Province are contemporaneous with the Gidyen Volcaniclastic Member. At present, these units are separated from each other by the Vulcan Fault in the Oberon area (Enclosure 1). They are lithologically distinctive; and the lack of any interfingering between these units implies that they were deposited relatively distant from each other. Rocks of the Quartzose Sedimentary Province were deposited eastward and outboard of the volcanic high from which the volcanic detritus was being shed (Figure 2.14b).

**LATEST MIDDLE TO LATE ORDOVICIAN (LATE DARRIWILIAN TO BOLINDIAN)**
The Gidyen Volcaniclastic Member and the Adaminaby Group are both overlain by late Darriwilian to early Gisbornian chert (Mozart Chert Member and Numeralla Chert respectively). Fossils recovered from these chert units are the same species and indicate that the cessation of deep-marine sedimentation was simultaneous across both lithological provinces. Sedimentation in the Quartzose Sedimentary Province recommenced in the early Late Ordovician with deposition of the Bumballa Formation. Overlying the Bumballa Formation, deposition is dominated by black shales (Warbisco Shale) deposited in an anoxic environment, which persisted from the late Gisbornian until the early Bolindian (Figure 2.14c).

The second pulse of volcaniclastic turbidite sedimentation derived from the volcanic highs was initiated in the early Gisbornian. In the Oberon and Rockley areas, the volcaniclastics of the
uppermost Triangle Formation overlie the Mozart Chert Member of the Molong Volcanic Province, and east of the Vulcan Fault, they overlie the Warbisco Shale of the Quartzose Sedimentary Province. The thickness of the uppermost Triangle Formation volcanioclastics in the Oberon area is quite variable, thickening from the northwest towards the southeast adjacent to the Vulcan Fault (Enclosure 1). This may reflect the location of the thickest part of a submarine wedge of volcanioclastics in the southern part of the Oberon area derived from the volcanic highs to the west. Farther east where the volcanioclastic apron overlies units of the Quartzose Sedimentary Province it is relatively thin (<300 m). Although the volcanioclastic apron overlies the Warbisco Shale, it was probably synchronous with deposition of the upper Warbisco Shale elsewhere. This is indicated by the thin beds of black shale interbedded with the volcanioclastic beds, which must have been deposited rapidly during slower background sedimentation of black shale. Late Ordovician units of the Molong Volcanic Province overlie and interdigitate with the black shale unit of the Quartzose Sedimentary Province east of the Vulcan Fault. Thus the two provinces were adjacent by the Late Ordovician.

**LATE MIDDLE TO LATE ORDOVICIAN REDUCTION IN SEDIMENT SUPPLY TO THE SUBMARINE FAN**

The reduction in sediment supply to the submarine fan of the Quartzose Sedimentary Province has been attributed by Jones *et al.* (1993) and Colquhoun *et al.* (1999) to increased sea levels during the late Middle to Late Ordovician. This is feasible for the mid Darriwillian part of the Numeralla Chert, as a significant rise in sea level occurred at this time according to the Australian Ordovician sea level curve (Figure 2.13; Nicoll & Webby 1996). Rising sea level at this time may have been a factor in the abrupt cessation of sand deposition to the outer margins of the turbidite fan.

In the late Darriwillian sea level was falling and this coincides with widespread fine-grained deposition in both lithological provinces of the Oberon-Taralga region (Figure 2.13). Following the drop in sea level volcanioclastic turbidite sedimentation derived from the volcanic highs increased significantly. Deposition of the Late Ordovician volcanioclastics overlies units of both lithological provinces in the Oberon area. Although sedimentation in the Quartzose Sedimentary Province also
recommenced on the submarine fan in the early Gisbornian following the drop in sea level, the sedimentation rate was greatly reduced compared to that of the Middle Ordovician. Sea level rose during the early Gisbornian contemporaneously with the increased volcaniclastic sedimentation (Figure 2.13), implying that factors other than sea level changes were more significant in controlling sediment supply. Following deposition of the Bumballa Formation in the early Gisbornian, another slight sea level rise occurred and the succession is dominated by pelagic black shale deposition. However, there is no obvious relationship between the black shale deposition and sea level throughout the rest of the Late Ordovician.

In the Western Belt of the Lachlan Fold Belt, deposition of quartz sandstone was continuous from the early Gisbornian until the latest Eastonian (VandenBerg & Stewart 1992). Detritus was still being supplied from the Delamerian mountain chain, but it no longer reached the Eastern Belt of the Lachlan Fold Belt during the Late Ordovician. Deformation in the central Lachlan Fold Belt during the Benambran Orogeny (Crook et al. 1973; Collins & Hobbs 2001) would have formed a barrier to the sediment supply from the west. Palaeocurrent directions indicate rocks deposited on the submarine fan during the Gisbornian were sourced from the south rather than the Delamerian mountain chain to the southwest (Powell 1983; Jones et al. 1993; Fergusson & Tye 1999). The initiation of the Benambran Orogeny in the mid Ordovician (Foster et al. 1999) is thought to be the major cause of sedimentation changes observed in the Quartzose Sedimentary Province of the Eastern Belt during the late Middle and early Late Ordovician.

**EARLY SILURIAN**

Overlying the late Ordovician units of both lithological provinces are quartz turbidites of the Bald Ridge Formation. This unit is probably latest Ordovician to Early Silurian in age, based on its stratigraphic relationships. The quartz-rich detritus is most likely derived from further west in the Lachlan Fold Belt, following uplift and erosion of the Early to Middle Ordovician Adaminaby Group and its correlatives. Further, the inclusion of small rounded granite clasts near the top of the unit are
indicative of the unroofing of granitic rocks in the provenance area during the Early Silurian. Horizons with a mixed provenance occur at the top of the Bald Ridge Formation, where they are dominated by subrounded crystal and lithic fragments along with subordinate rounded metamorphic quartz grains. This composition signifies a mixing of both volcanic and quartz-rich detritus, and may indicate derivation from more than one source area. These horizons are probably indicative of the reworking of mafic volcanics and volcanioclastics following continued uplift as part of the Benambran Orogeny.

In the Thompson Creek area, the Fosters Creek Conglomerate and the silicic Kangaloolah Volcanics overlie the Bald Ridge Formation with an angular unconformity. The quartz turbidites of the Bald Ridge Formation were folded prior to deposition of the overlying units during the early Late Silurian. This unconformity is also related to the Benambran Orogeny that had widespread effects in the Lachlan Fold Belt (Crook et al. 1973; Collins & Hobbs 2001).

2.5 CONCLUSIONS
Ordovician to Early Silurian stratigraphic units of both the Molong Volcanic Province and the Quartzose Sedimentary Province occur in the Oberon-Taralga Region, where they were deposited in a deep-marine environment. The stratigraphy of the Quartzose Sedimentary Province comprises a conformable, quartz-rich fining-upward succession composed of thick-bedded turbidites of the undifferentiated Adaminaby Group, Numeralla Chert, thin-bedded turbidites of the Bumballa Formation overlain by the Warbisco Shale. This succession was deposited on the mid to outer margins of a submarine turbidite fan. This stratigraphy is similar to that of other areas of the Eastern Belt of the Lachlan Fold Belt (Fergusson & VandenBerg 1990; VandenBerg & Stewart 1992; VandenBerg et al. 1992; Glen 1994; Fergusson & Colquhoun 1996; Fergusson 1998a, 1998b). No evidence exists to support the presence of Early to early Middle Ordovician Quartzose Sedimentary Province rocks in the Oberon-Taralga Region.
The reduction in the sediment supply to the submarine fan during the late Middle to Late Ordovician is related primarily to the uplift associated with the early part of the Benambran Orogeny in the central and western Lachlan Fold Belt. This formed a barrier diverting sediments derived from the eroding Delamerian mountain chain away from the Eastern Belt of the Lachlan Fold Belt. Deposition of the Numeralla Chert is partly related to a sea level rise during the middle Darriwilian, coincident with the onset of the reduction in sediment supply to the submarine fan.

The Molong Volcanic Province in the Oberon area is represented by the Triangle Formation and the Rockley Volcanics. These are composed of deep-marine pelagic deposits and volcaniclastic turbidite and megaturbidites. The Early Ordovician is characterised by pelagic sedimentation (Budhang Chert Member), which either continued until the Darriwilian or was followed by a hiatus in deposition between the Bendigonian and Darriwilian. The first major pulse of mafic volcaniclastic material (Gidyen Volcaniclastic Member) was deposited in the Darriwilian and was contemporaneous with volcanism on the Molong High. This was followed by a return to pelagic deposition as shown by late Darriwilian-early Gisbornian chert and siltstone of the Mozart Chert Member. Deposition of the Mozart Chert Member was synchronous with deposition of the Numeralla Chert in the Quartzose Sedimentary Province and may reflect the rise in sea level in the mid Darriwilian.

In the Gisbornian, a second cycle of mafic volcaniclastic turbidite and megaturbidite deposition began and was synchronous with volcanism on the volcanic high. The Late Ordovician volcaniclastics overlie the Mozart Chert Member of the Molong Volcanic Province and the Warbisco Shale of the Quartzose Sedimentary Province east of the Vulcan Fault in the Oberon area. The mafic Rockley Volcanics were also deposited in the Late Ordovician, overlying and possibly interfingering with the upper Triangle Formation.

The Quartzose Sedimentary Province formed east of the Molong Volcanic Belt during the late Middle Ordovician. Volcanism along the Molong High produced voluminous volcaniclastic deposition in the
Darriwilian and throughout the Late Ordovician as shown by widespread shallow marine to deep-marine deposition. During the Late Ordovician the volcaniclastic apron associated with emergent volcanism in the western part of the Molong Volcanic Province reached its maximum extent. Volcaniclastic deposits extended across the Warbisco Shale of the Quartzose Sedimentary Province in the Oberon area. Thus for at least for the Late Ordovician, these two provinces developed adjacent to each other and were not subsequently juxtaposed by faulting as implied by Glen and Wyborn (1997).

Deposition of the mixed detritus of the Bald Ridge Formation is related to the development of uplift in the central and neighbouring parts of the Lachlan Fold Belt (Benambran Orogeny). This occurred in the latest Ordovician and Early Silurian and its effects included deformation in the Thompson Creek area as shown by the angular unconformity between the Bald Ridge Formation and the overlying Fosters Creek Conglomerate.
CHAPTER 3
AGE AND PALAEOGEOGRAPHIC SIGNIFICANCE OF ORDOVICIAN CONODONTS FROM THE LACHLAN FOLD BELT

3.1 CONODONT BIOFACIES
Two major global Ordovician conodont biogeographic provinces have been recognised. These are: (1) the ‘North American Mid-Continent Province’, consisting of faunas recovered from shallow-marine rock associations from low to middle palaeolatitudes; and (2) the ‘Anglo-Scandinavian-Appalachian Province’, later known as the ‘North Atlantic Province,’ containing faunas from deeper and/or cooler marine rocks deposited either off the shelf margins in an open-ocean environment or from higher palaeolatitudes (Sweet et al. 1959). After consideration of conodont data from beyond Laurentia, it became clear that these province names were inappropriate and the two provinces were reinterpreted as two distinct faunal realms resulting from variations in water temperature, salinity and bathymetry, rather than as two geographically separate provinces (Tipnis, Chatterton & Ludvigsen 1978; Sweet & Bergström 1984; Miller 1984). Herein the terms Warm Faunal Realm (WFR) and Cold Faunal Realm (CFR) are used (after Miller 1984) instead of North American Mid-Continent Province and North Atlantic Province respectively.

Based on the understanding that provinciality within Ordovician conodonts results from variations in environmental conditions, a range of biofacies have been established within each of the faunal realms during the Middle and Late Ordovician (Sweet & Bergström 1984). Within the WFR, six major biofacies were recognised during the Late Ordovician Velicuspis Chronozone. The Velicuspis Chronozone was defined by Sweet (1982), and includes rocks between the level of the first appearance of *Oulodus veliscuspis* and that of the first appearance of *O. robustus*. In Australia, this chronozone falls within the Eastonian Ea3 graptolite zone. These biofacies represent a continuum from nearshore, shallow-marine biotopes to offshore, deeper marine biotopes; from shallowest to deepest they are: the *Aphelognathus-Oulodus* Biofacies, the *Pseudobelodina* Biofacies, the *Plectodina*
Biofacies, the *Phragmodus undatus* Biofacies, the *Amorphognathus superbus-ordovicicus* Biofacies and the *Dapsilodus mutatus-Periodon grandis* Biofacies (Sweet & Bergström 1984).

Within the Late Ordovician CFR, data from Britain, Baltoscandia and the Mediterranean were used to define three distinct biofacies, the *Hamarodus europaeus-Dapsilodus mutatus-Scabbardella altipes* (HDS) Biofacies, the *Phragmodus undatus-Icriodella-Plectodina* Biofacies and the *Amorphognathus-Plectodina* Biofacies (Sweet & Bergström 1984). These biofacies do not represent a series of depth-related biofacies comparable to those from the WFR, but are characterised by intergradational recurrent species associations whose distribution was probably controlled by local environmental conditions, of which the controlling factors are not currently understood (Sweet & Bergström 1984).

A further two biofacies, the *Periodon-Dapsilodus-Pygodus* and the *Prioniodus (Baltoniodus)-Dapsilodus-Pygodus* Biofacies were recognised from the Middle Ordovician of Baltoscandia where they were well developed. These biofacies intergrade laterally and both include *Protopanderodus* and *Eoplacognathus*. Sweet and Bergström (1984) also suggested these biofacies may have been replaced in Europe during the Late Ordovician by the HDS Biofacies.

### 3.2 PREVIOUS WORK ON CONODONTS WITHIN THE LACHLAN FOLD BELT

Within the Lachlan Fold Belt of southeastern Australia, Ordovician rocks have been assigned to two distinct lithological provinces, the Molong Volcanic Province and the Quartzose Sedimentary Province (Figure 1.3). Various macrofaunas from limestone of the Molong Volcanic Province have been studied over many years (e.g. Etheridge 1895, 1909; Hill 1957; Ross 1961; Pickett 1970; Webby 1971, 1972, 1973, 1975, 1977; Percival 1992), but only scant attention has been paid to the conodont faunas of the Molong Volcanic Province. Packham (1967) and Pickett (1978, 1985) described small conodont faunas from these limestone units in order to establish age control. Only recently has the conodont fauna of this volcanic and limestone association been studied in any biostratigraphic detail (Savage 1990; Trotter & Webby 1995; Zhen & Webby 1995; Percival 1999; Percival et al. 1999).
Conodont faunas from the Quartzose Sedimentary Province have also been studied primarily to provide stratigraphic age control in an otherwise poorly fossiliferous succession of quartz turbidites and chert (Stewart 1988; Stewart & Fergusson 1988, 1995; VandenBerg & Stewart 1992). However, Nicoll (1980) described a small Middle Ordovician (mid-late Llanvirn) conodont fauna from Canberra as having a North Atlantic faunal aspect (CFR), which differed from the two previously described from New South Wales (Packham 1967; Pickett 1978) which were both classified as having a mid-continent (WFR) aspect (Nicoll 1980). Using Oulodus, Plectodina, Phragmodus and Belodina as indicative of the WFR and Eoplacognathus, Pygodus, Prioniodus and Periodon as indicative of the CFR, Nicoll (1989) classified conodont faunas from the Australian cratonic basins, Tasmania, western New South Wales and Queensland as indicative of the Warm Faunal Realm and tentatively classified the few known faunas from Victoria and eastern New South Wales as associated with the Cold Faunal Realm.

Since 1988, conodont faunas have been more commonly used for dating chert of the Quartzose Sedimentary Province of the Lachlan Fold Belt in a technique developed by I.R. Stewart. Conodont faunas have also been described from shallow water limestone of the Molong Volcanic Province (Savage 1990; Trotter & Webby 1995; Zhen & Webby 1995). Over 140 conodont localities are known from both lithological provinces of the Lachlan Fold Belt, although only a small number of elements are reported from the bulk of these localities (see Appendix D).

### 3.3 AIMS, METHODS AND TIMESCALE TERMINOLOGY

The aim of this chapter is to describe eight conodont faunas from eight new localities, seven from the Oberon-Taralga Region where both lithological provinces occur (Figures 3.1; 3.2b) and one from the Mudgee area, just east of that described by Stewart and Fergusson (1995; Figure 3.2a). The faunas of the new occurrences are listed in Table 3.1, together with details of the stratigraphic assignment, age and catalogue details.
Figure 3.1 Generalised regional geology of the Oberon-Taralga-Crookwell district. Ordovician rocks are represented as the two lithological provinces. The map areas of Figures 3.2b, c, d and e are also outlined.
These new occurrences, along with most other conodont faunas known from the Lachlan Fold Belt, are placed into one of three age categories: Lancefieldian-Bendigonian, Darriwilian-Gisbornian or Gisbornian-Eastonian. All the Lachlan Fold Belt faunas are listed in Appendix D; where using the presence of *Oulodus*, *Plectodina*, *Phragmodus* and *Belodina* as indicative of the WFR and *Eoplacognathus*, *Pygodus*, *Prioniodus* and *Periodon* as characteristic of the CFR (Sweet & Bergström 1974; Nicoll 1989), the fauna of each locality has been classified as belonging to either the WFR, CFR, or unknown. The distribution, faunal realm and lithology of each locality is shown broken down into the three Ordovician time slices listed above, and when the distribution of conodonts together with their faunal realm affinities and the stratigraphy are considered, the development of volcanic islands and surrounding carbonate platforms offshore of the Gondwanan margin throughout the Ordovician is better understood (see sections 3.5 and 3.6).

### 3.3.1 Timescale
The Ordovician timescale used is that produced for Australia (Young & Laurie 1996) and this timescale is correlated with the Baltoscandia conodont zones, North American conodont zones and the Victorian conodont zones as on Chart 2 (Nicoll & Webby 1996) and summarised in Figure 3.3 herein. A tripartite system of Lower, Middle and Upper Ordovician is used (following Webby 1998) so that the Lower Ordovician includes the Australian stages Lancefieldian, Bendigonian and Chewtonian, the Middle Ordovician consists of the Castlemainian, Yapeenian and Darriwilian stages, and the Upper Ordovician, the Gisbornian, Eastonian and Bolindian stages. The Middle and Late Ordovician faunas of the Lachlan Fold Belt are also tentatively correlated with the Middle and Late Ordovician Biofacies previously outlined (Sweet & Bergström 1984).

### 3.3.2 Methods
Chert samples were selected from several localities in the general area between Oberon, Taralga and the Crookwell River, and one from Bara Creek, east of Mudgee (Figure 3.2). This area has been affected by regional deformation, which has produced abundant folds with a range of wavelengths.
Figure 3.2 Geologic maps and chert sample localities of each area studied within the Oberon/Taralga/Crookwell district. (a) Bara Creek, east of Mudgee. Sample Lu5 and map from Fergusson and Colquhoun (1996). (b) Samples collected from cherts southwest of Oberon. (c) Chert sample R16474 collected from a quarry northeast of Isabella. (d) Samples from chert bed at the junction of Silent and Oakey Creeks, Abercrombie National Park. (e) Locality of sample R16472 from roadside outcrop beside Crookwell River.
Table 3.1 Age and stratigraphic assignment of each chert sample. Stratigraphy indicates both lithological province (QSP or MVP) and formation.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Grid Ref.</th>
<th>Stratigraphic Assignment</th>
<th>Conodont Fauna</th>
<th>Age Assigned</th>
<th>Locality No. (Appendix D)</th>
</tr>
</thead>
<tbody>
<tr>
<td>R15973</td>
<td>GR757794</td>
<td>MVP</td>
<td>Pygodus sp.,</td>
<td>Da3-Gi1</td>
<td>Locality 138</td>
</tr>
<tr>
<td></td>
<td>6266095</td>
<td>Mozart Chert Member</td>
<td>Periodon sp.</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oberon 8830</td>
<td>1:100 000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>R15975</td>
<td>GR756379</td>
<td>MVP</td>
<td>Paracordylodus</td>
<td>Be1-Be2</td>
<td>Locality 135</td>
</tr>
<tr>
<td></td>
<td>6258716</td>
<td>Budhang Chert Member</td>
<td>gracilis, Indet.</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oberon 8830</td>
<td>1:100 000</td>
<td>conoform clusters, Indet</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>paraconodont clusters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R16472</td>
<td>GR718650</td>
<td>QSP</td>
<td>Pygodus sp.,</td>
<td>Da3-Gi1</td>
<td>Locality 136</td>
</tr>
<tr>
<td></td>
<td>6190123</td>
<td>Numeralla Chert</td>
<td>Periodon sp.</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Crookwell 8729</td>
<td>1:100 000</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>R16473</td>
<td>GR754370</td>
<td>MVP</td>
<td>?Pygodus sp.,</td>
<td>Ordovician</td>
<td>Locality 139</td>
</tr>
<tr>
<td></td>
<td>6259794</td>
<td>Undifferentiated Triangle Fmn.</td>
<td>indet. broken</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oberon 8830</td>
<td>1:100 000</td>
<td>denticulate element</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R16474</td>
<td>GR748350</td>
<td>MVP</td>
<td>?Pygodus sp.,</td>
<td>Ordovician</td>
<td>Locality 140</td>
</tr>
<tr>
<td></td>
<td>6240750</td>
<td>Undifferentiated Triangle Fmn.</td>
<td>indet.</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oberon 8830</td>
<td>1:100 000</td>
<td>denticulate element</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R16475</td>
<td>GR747330</td>
<td>QSP</td>
<td>Histiodella sp.,</td>
<td>Da2-Late</td>
<td>Locality 137</td>
</tr>
<tr>
<td></td>
<td>6211698</td>
<td>Numeralla Chert</td>
<td>Cordylyodus horridus,</td>
<td>Ordovician</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Taralga 8829</td>
<td>1:100 000</td>
<td>Periodon sp.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R16476</td>
<td>GR747302</td>
<td>QSP</td>
<td>Periodon sp.</td>
<td>Da2-Late</td>
<td>Locality 137</td>
</tr>
<tr>
<td></td>
<td>6216936</td>
<td>Numeralla Chert</td>
<td>Periodon sp.</td>
<td>Ordovician</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Taralga 8829</td>
<td>1:100 000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>R16477</td>
<td>GR761750</td>
<td>QSP</td>
<td>Pygodus sp.,</td>
<td>Da3-Gi1</td>
<td>Locality 141</td>
</tr>
<tr>
<td></td>
<td>6391175</td>
<td>Numeralla Chert</td>
<td>Periodon aculeatus</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mudgee</td>
<td>1:100 000</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
All the chert outcrops were strongly cleaved in at least one direction, with the cleavage direction quite variable across the region. The bedding orientation was marked on the sample prior to its collection from the outcrop. Chert samples were collected away from any mesoscopic fold hinges and were as unfractured and unweathered as possible.

Samples were cut parallel to bedding planes, then mounted on glass slides and sectioned for later microscopic examination (approximately 10 - 15 sections of each sample). Each section was ground until transmitted light 'just' penetrated through the chert. In translucent cherts this was approximately 0.8 - 1.0 mm thick, whereas darker more opaque cherts were ground thinner. Once light was able to penetrate through the section, it was examined under a microscope in both transmitted and reflected light. Conodont elements and radiolarians could often be seen at this stage but were usually still embedded too deeply within the chert to be identified, so further 'thinning' of the section was required in order to see the fossil more clearly. Sections were prepared in the normal way by polishing on a rotary lap or glass plate using silicon carbide powder.

For provinciality studies, use of this method resulted in only a partial fauna being described from a rock sample. Acid digestion (HF) of the chert was not used because many of the elements are preserved only as moulds, whereas other elements have been extended and fractured (e.g. AMF105961 - Figure 3.5e), so normal processing would have resulted in very few if any complete elements being recovered. Because each fossil is viewed in only two dimensions, compound elements could often be identified only to the generic level. Coniform conodonts cannot be reliably identified to the generic level and, therefore, only their presence is noted. Several taxa, such as *Pygodus*, have short enough ranges that samples can be dated with reasonable accuracy within the Ordovician stages.

The S, M and P notation used to classify the conodont elements is that of Sweet (1981), and all elements referred to herein have been catalogued (AMF Number), and are lodged at the Australian Museum in Sydney. ‘R’ numbers relate to the University of Wollongong rock sample catalogue.
Figure 3.3 Chart shows Australian and British Ordovician stages and graptolite zones correlated with conodont zones and ranges within Australia (after Nicoll & Webby, 1996).
3.4 NEW FOSSIL LOCALITIES
Conodont elements have been identified from cherts of the Triangle Formation (Molong Volcanic Province) and the Numeralla Chert, part of the Adaminaby Group (Quartzose Sedimentary Province) (Figure 3.4). Each sample of chert has been allocated a rock sample or ‘R’ number. The conodonts recovered from each chert sample are listed together with the stratigraphic assignment, grid reference and Locality Number (Appendix D) for each sample in Table 3.1.

3.4.1 Sample R15975
A dark grey to black chert (Budhang Chert Member of the Triangle Formation) collected from beside Brisbane Valley Creek at GR756379 6258716 (Oberon 1:100 000 topographic sheet, 8830; Enclosure 1; Figure 3.2b), contained a conodont fauna consisting of abundant Paracordylodus gracilis (Lindström) elements including S (AMF105956, AMF105955; Figure 3.5b, 3.5c), M (AMF105958, AMF105960; Figure 3.5d, g) and P (AMF105961, AMF105979; Figure 3.5e, f) elements, along with indeterminate paraconodont clusters and indeterminate coniform conodonts (AMF105957; Figure 3.5a).

AGE INDICATED
The range of P. gracilis (Figure 3.3) is uppermost Tremadoc to early Arenig, top Paltodus deltifer zone to low Oepikodus evae zone of the Baltoscandia Conodont Zones (Sweet 1988). Within the Victorian sequence the P. deltifer zone corresponds to the top of the Cordylodus spp. zone (La2) and the O. evae zone ranges from Bendigonian Be3 to Chewtonian. In Victoria P. gracilis occurs in association with Bendigonian Be1 graptolites at Dargo (Stewart & Fergusson 1988) and in association with a fragmentary tetraraptid of La3-Be1 age on the Mornington Peninsula (Stewart 1988). Therefore, based on the occurrence of P. gracilis in other parts of the Lachlan Fold Belt and the occurrence of paraconodonts, strata at this sample site are considered to be early Arenig (La3-Be2).
Figure 3.4 Correlation of generalised stratigraphic columns across the Lachlan Fold Belt. Bendigo-Ballarat Zone modified from VandenBerg and Stewart (1992), where graptolite (●) and conodont (*) age control indicates continuous deposition from the Lancefieldian to Darriwilian. Stratigraphy from Junee-Narromine and Molong Volcanic Belts modified from Pickett & Percival (2001), Lyons et al. (2000) and Raymond et al. (1997), and stratigraphy of (3) Bara Creek modified from Fergusson and Colquhoun (1996).
<table>
<thead>
<tr>
<th>British Stages</th>
<th>Victorian Stages &amp; Graptolite Zones</th>
<th>Bendigo-Ballarat Zone</th>
<th>Junee-Narromine Volc. Belt</th>
<th>Molong Volcanic Belt</th>
<th>Oberon West</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ashgill</td>
<td>Bolindian</td>
<td>Bo5() Bo4() Bo3() Bo2() Bo1()</td>
<td>(Angular unconformity with Kerrie Conglomerate, = Sil? Dev?)</td>
<td></td>
<td>Cabonne Group</td>
</tr>
<tr>
<td>Caradoc</td>
<td>Eastonian</td>
<td>Ea4() Ea3() Ea2() Ea1()</td>
<td>Riddell Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Gisbornian</td>
<td>Gi2() Gi1()</td>
<td>Goonumba Volcanics</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Llanvirn</td>
<td>Darriwilian</td>
<td>Da4() Da3() Da2() Da1()</td>
<td>Gunningblad Shale Fmn.</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Yapeenian</td>
<td>Ca4() Ya2() Ya1()</td>
<td>Billabong Creek Limestone Formation</td>
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<td></td>
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<tr>
<td>Arenig</td>
<td>Castlemainian</td>
<td>Ca3() Ca2() Ca1()</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Chewtonian</td>
<td>Ch2() Ch1()</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bendigonian</td>
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<td>Warendian</td>
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Oberon West:
- Rockley Volcanics
- Mozart Chert Member
- Gidyen Volcanic Member
- Budhang Chert Member
- Triangular Formation
- Cabonne Group
- Gunningblad Shale Fmn.
- Billabong Creek Limestone Formation
- Kenilworth Group
- Yarrimba Formation
- Hensleigh Siltstone
- Nelungaloo Volcanics
- Mitchell Formation
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<td>*(1) R15975</td>
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FAUNAL AFFINITY
No genera previously outlined for determining provinciality were identified from this sample. However, *P. gracilis* is a characteristic species of the CFR in Europe (Landing, 1976), but it has also been recovered from the lower Deep Kill Shale of the Taconic allochthon, eastern New York (Landing 1976), the Emanuel Formation in the Canning Basin, Western Australia (Nicoll *et al.* 1993) and from the Hensleigh Siltstone in New South Wales (Percival *et al.* 1999; Figure 3.4). All of these are associated with the WFR; however, they represent a deeper and/or colder water biofacies of that realm.

3.4.2 Sample R15973
Conodont elements including a single *Pygodus* sp. haddingodiform element (AMF105973; Figure 3.6f) and several *Periodon* sp. elements (AMF105976; Figure 3.6g) were identified in sections from a large chert outcrop occurring at the intersection of the Oberon Rockley Road and Langs Road, GR757794 6266095 (Oberon 1:100 000 topographic map, 8830; Figure 3.2b). Brown to honey coloured chert bands 4-10 cm thick are interbedded with pink-orange mudstone and belong to the Mozart Chert Member of the Triangle Formation (Figure 3.4). The bedding-cleavage intersection angle varies across the outcrop causing the chert to be quite fractured and friable in places.

3.4.3 Sample R16472
A single poorly preserved *Pygodus* sp. pygodiform element (AMF105974; Figure 3.6i) and more abundant *Periodon* sp. elements (AMF105975; Figure 3.6h) were recovered from a thin 2-3 cm chert band at the very top of the undifferentiated Adaminaby Group (Quartzose Sedimentary Province) along the Crookwell - Binda Road on the northern side of the Crookwell River bridge at GR718650 6190123 (Crookwell 1:100 000 topographic map, 8729; Figure 3.2f).

3.4.4 Sample R16477
Elements of both *Pygodus* and *Periodon aculeatus* (Graves & Ellison) have also been identified from
Figure 3.5 All specimens from Sample R15975. (a) Indet. coniform conodont AMF105957; (b)-(g) *Paracordylodus gracilis*. (b) S element, AMF105956; (c) S element, AMF105955; (d) EM element, AMF105958; (e) P element, AMF105961; (f) P element, AMF105979; (g) M element, AMF105960. All scale bars = 250 m except (d) and (g) where bar = 130 m.
Figure 3.6 All scale bars = 250μ. (a) - (d) From sample R16475; (a) - (c) *Cordylopus horridus*; (a) AMF105972; (b) AMF105970; (c) AMF105971. (d) *Histiodella* sp. AMF105969. (e) From sample R16476; *Periodon* sp., P element, AMF105977. (f) - (g) From sample R15973; (f) *Pygodus* sp., haddingodiform element, AMF105973; (g) *Periodon aculeatus*, P element, AMF105976. (h) - (i) From sample R16472; (h) *Periodon* sp., AMF105975; (i) ?*Pygodus* sp., pygodotentiform element, AMF105974. (j) From sample R16473; top ?*Pygodus* sp., S element, AMF105959. (k) From sample R16474; Indet. denticulated element, AMF105978.
a grey-honey-coloured, thin-bedded chert (Numeralla Chert) collected from beside Bara Creek east of Mudgee (GR761750 6391175 Mudgee 1:100 000 Topographic Sheet 8832, Figure 3.2a). *Pygodus* sp. elements identified from sections of this sample include three pygodontiform elements, two superimposed (AMF105962; Figure 3.7e) and the other by itself (AMF105966; Figure 3.7c), one haddingodiform element (AMF105965; Figure 3.7f) and a single S? element (AMF105967; Figure 3.7g). Elements belonging to the *Periodon aculeatus* apparatus included several S elements (AMF105963, AMF105964; Figures 3.7a, b) and at least one M element (AMF105968; Figure 3.7d).

**Age of samples R15973, R16472 and R16477**

*Pygodus serra* (Hadding) and *P. anserinus* (Lamont & Lindström) are characteristic of the Baltoscandia *P. serra* and *P. anserinus* conodont zones respectively. *P. serra* occurs in Victoria and New South Wales associated with Da3-Da4 graptolites (Stewart & Fergusson 1988) and was correlated with the Victorian Darriwilian latest Da3-Da4 zones by Webby and Nicoll (1989). *P. serra* was also recovered from the Wahringa Limestone, New South Wales, in association with representatives of Faunal Assemblages OT10 and OT11 (Percival *et al.* 1999). The base of the *P. anserinus* zone was originally defined as the level of first appearance of *P. anserinus* which overlaps with the occurrence of *P. serra* (see Bergström 1971). In Australia, *P. anserinus* occurs with early Gisbornian *Nemagraptus gracilis* zone graptolites in Victoria (Stewart 1988) and was correlated with the late Da4-Gi1 interval (Chart 2, Nicoll & Webby 1996; Figure 3.3). The occurrence of elements belonging to either *P. serra* or *P. anserinus* indicates that the age of these cherts is late Darriwilian to early Gisbornian (latest Da3-Gi1; equivalent to latest Llanvirn-early Caradoc; Figure 3.3).

**Faunal realm affinities of samples R15973, R16472 and R16477**

All these samples contain *Periodon* and/or *Pygodus*, both of which are indicative of the CFR and, therefore, all are classified as part of that realm. They can also be tentatively correlated with the Middle Ordovician *Periodon-Dapsilodus-Pygodus* biofacies of Baltoscandia (Sweet & Bergström 1984).
3.4.5 Samples R16475 and R16476

A conodont fauna was also identified in thick sections of the samples R16475, GR747330 6211698 and R16476, GR747302 6216936 (both Taralga 1:100 000 topographic map, 8829; Figure 3.2d) which were collected from 2.5 m of grey, thinly bedded and finely laminated Numeralla Chert (Figure 3.4). The fauna consisted of three *Cordylodus horridus* (Barnes & Poplawski) elements (AMF105972, AMF105970, AMF105971; Figures 3.6a, b, c) and a single *Histiodella* sp. Pa element (AMF105969; Figure 3.6d), along with abundant *Periodon* sp. elements (AMF105977; Figure 3.6e).

AGE OF SAMPLES R16475 AND R16476

The occurrence of *Cordylodus horridus* in Australia is rather limited. It was first recorded amongst the conodont fauna occurring with Da3 graptolites from Surprise Gully, Lancefield (Stewart 1988). The species was named and described by Barnes and Poplawski (1973) from the Mystic Formation in Quebec, where *C. horridus* occurred in two samples with a lower Llanvirn (mid-Darriwilian) fauna including *Belodella erecta* (Rhodes & Dineley), *Histiodella sinuosa* (Graves & Ellison) and *Periodon aculeatus*. Nicoll (1992) discussed the evolutionary development of *C. horridus* as part of the *C. angulatus* (Pander) lineage and noted that it occurs in the Llanvirn but is not very widely distributed. In contrast, Löfgren (1995, 1997) argued for *C. horridus* to be considered as part of the *Paroistodus* lineage after *C. horridus* elements were found in Arenig samples from Talubacken, central Sweden. A summary of *C. horridus* localities and associated faunas given by Löfgren (1997) indicates that the currently understood age range is from the *O. evae* zone of early-mid Arenig through to the early Llanvirn (Chewtonian to Darriwilian, Da2/Da3; Figure 3.3).

McHargue (1982) discussed the evolution of *Histiodella* after recovering over 9000 elements from the Joins Formation in Oklahoma. He described *Histiodella* as a biostratigraphically valuable genus that evolved rapidly through several species during the late Early Ordovician and, as such, species of *Histiodella* are able to be distinguished on the basis of characteristics of denticleations on the upper margin of the bryantodontiform (Pa) element. McHargue (1982) noted that the sequence of species
Figure 3.7 (a) - (g) From sample R16477 and all scale bars = 250 m. (a), (b) and (d) *Peridon aculeatus*; (a) S element, AMF105963; (b) S element, AMF105964; (d) M element, AMF105968. (c), (e)-(g) *Pygodus* sp.; (c) pygodontiform element, AMF105966; (e) 2 superimposed pygodontiform elements, AMF105962; (f) haddingodiform element, AMF105965; (g) S? element,
indicates a trend toward increased denticulation at progressively higher levels in the stratigraphic section. *H. altifrons* (Harris) occurs at the base and only exhibits smooth Pa element morphotypes, followed by *H. minutiserrata* (Mound), which exhibits both smooth and minutely serrated Pa elements, then *H. sinuosa* and *H. serrata* (Harris) which both have smooth, minutely serrate and denticulate Pa elements.

Only one *Histiodella* element (AMF105969; Figure 3.6d) was recovered from the chert in Oakey Creek, and this is determined only to generic level; this single Pa element is denticulated and therefore must belong to *H. sinuosa* or a younger species. *Histiodella sinuosa* and *H. holodentata* are both zone fossils in the North American Conodont Zones, correlating with the mid Da2 and the top of the Da2-Da3 Victorian graptolite zones respectively. Therefore a maximum mid-Darriwilian age (Da2) is indicated for sample R16475. Unfortunately the other elements recovered do not provide a minimum age.

**Faunal Affinity**
The presence of *Periodon* elements indicates this fauna belongs to the CFR. *Cordylocus horridus*, *Histiodella* and *Periodon* have been found together in the Deep Kill Shale, where they were considered to be associated with dominantly (North Atlantic Province) CFR taxa (Landing 1976), although *Histiodella* is cosmopolitan and commonly associated with WFR taxa. These two samples from Oakey Creek may represent a mixing of faunas from the WFR into a deep-water biofacies.

### 3.4.6 Samples R16473 and R16474
Samples of chert from two other outcrops thought to belong to the Mozart Chert Member of the Triangle Formation (Figures 3.2b, c, 3.4) were sampled and serial sectioned (Samples R16473, GR754370 6259794 and R16474, GR748350 6240750; both on Oberon 1:100 000 topographic map, 8830; Figures 3.2b, c). These samples had been re-silicified, and only those elements figured were preserved (AMF105959, AMF105978; Figures 3.6j, k). Figure 3.6j shows a broken denticulated
Figure 3.8 All known conodont localities within the Lachlan Fold Belt. Publications cited and faunas shown are listed by the corresponding number in Appendix D.
element (bottom) and a possible Pygodus sp. S element. Based on the stratigraphic position of these samples in the Triangle Formation, and without more convincing age diagnostic fossils, no more definitive age other than Ordovician could be determined for these sites. The faunal affinity of these sites also could not be determined.

3.5 FAUNAL AFFINITIES OF LACHLAN FOLD BELT CONODONT FAUNAS
The new conodont localities reported in this paper are listed along with all other known conodont localities from the Lachlan Fold Belt of NSW and Victoria in Appendix D, and are plotted on Figure 3.8, which shows the widespread distribution of conodont localities across the Lachlan Fold Belt. The faunas from these localities have then been divided into three Ordovician time slices (Lancefieldian - Chewtonian; Darriwilian - Early Gisbornian; Gisbornian - Eastonian) and these are plotted on Figures 3.9a, 3.9b and 3.9c, where the faunal affinity and the lithology of the rock samples are also indicated.

3.5.1 Lancefieldian - Chewtonian (Figure 3.9a)
Lachlan Fold Belt conodont faunas of Lancefieldian - Chewtonian age are known only from deep-marine chert and shale, except for those from the Digger Island Formation in southern Victoria and the Hensleigh Siltstone in Central New South Wales. Although Cordylodus rotundi (Pander), Onetodus sp. and Drepanodus spp. have been reported from the Digger Island Formation, descriptions and illustrations of these conodonts have never been formally published (Kennedy 1971). However, the trilobite fauna of this formation is considered to be Lancefieldian (La1) (Jell 1985) therefore, these conodonts together with those reported from Narooma (Bishoff & Prendergast 1987) are the oldest reported from the Lachlan Fold Belt. Of the 21 conodont localities included in this map, all except Locality 143 are from siliceous, deep-marine rocks. Six samples of the 21 localities contain Prioniodus indicative of the CFR (Localities 42, 43, 58, 64, 65 and 143 in Appendix D) and none of the samples contains any genera indicative of the WFR. A further seven samples (localities 35, 37, 58, 66, 67, 85 and 86) all contain Oepikodus evae, zone fossil of the Baltoscandia province (CFR) and primarily found associated with CFR taxa. However, faunas from localities 35, 74 and 143 also
Figure 3.9a Conodont localities identified for the Earliest Ordovician (Lancefieldian to Bendigonian).
contain *Bergstroemognathus*, a genus unknown from the Baltic and normally only recovered in association with other CFR taxa from deeper-water lithologies of what has previously been considered part of the Mid-Continent Province (e.g. the Deep Kill Shale from New York and the Marathon area of southwestern Texas: Bergström 1990). The paraconodont *Clavohamulus* is also unknown from the Baltic, but occurs in the Deep Kill Shale (Bergström 1990) and is also known from localities 38, 54 and 135 as well as from Melville Point (I. Stewart, unpublished data). *Paracordylodus gracilis*, also known predominantly only from the CFR, but also from deeper water areas of the WFR (Nicoll et al. 1993), occurs at several localities (64, 65, 73, 75, 135 and 143).

The overall affinities for the few known deep-marine localities of this age are to the CFR. Those taxa with some affinity to the WFR (*Bergstroemognathus, Clavohamulus, P. gracilis*) have previously been recovered from strata associated with deep-marine environments off low latitude shelf margins. These faunas may be representative of a similar palaeogeographic setting for eastern Australia during Early Ordovician time.

Two faunas from the Hensleigh Siltstone, one from allochthonous shallow-marine limestone within the Hensleigh Siltstone (locality 143) and the other from authochthonous laminated siltstone were described as being very similar; and most closely resembled the fauna described by McTavish (1973) from the Emanual Formation, Canning Basin (Percival et al. 1999). Although the fauna from locality 143 was recovered from limestone, and represents a shallower water facies, the faunas include the CFR genera *Prioniodus*, indicating a CFR affinity.

### 3.5.2 Darriwilian - Gisbornian (Figure 3.9b)

Of the conodont faunal localities shown on Figure 3.9b, most of the faunas (63 of 75 localities) are from siliceous, deep-marine rocks collected from both the Quartzose Sedimentary Province and the Molong Volcanic Province. Of these 63 localities, 59 include *Pygodus* and/or *Periodon*, indicative of the CFR. Two of these localities (63 and 137) along with locality 71 also include *Histiodella,*
Figure 3.9b Conodont localities identified for the Late Middle Ordovician to early Late
*Spinodus spinatus* (Dzik) and *Cordylodus horridus*. These genera are normally associated with the WFR and the faunal associations identified from the Lachlan Fold Belt sites closely resemble those recovered from the Deep Kill Shale of New York (Landing 1976).

The remaining 12 of the 75 faunas were recovered from limestone of the Molong Volcanic Province. Most of these faunas (10 of 12) contain the CFR indicator taxa *Pygodus* and/or *Periodon* along with associated CFR genera, *Belodella, Walliserodus, Dapsilodus* and *Protopanderodus* (localities 68-70, 123, 124, 126 & 144 - 146). However, localities 144 - 146 also contain genera associated with the WFR (*Plectodina, Phragmodus* and *Belodina*), and therefore, are classified as being a mix of the two faunal realms. *Belodina* and *Plectodina* were also recovered from samples 142 and 148, but these samples did not contain any taxa associated with the CFR. Therefore, of the 12 faunas recovered from limestones of Darriwillian to Gisbornian age, seven are classified as being of the CFR, two are of the WFR and three exhibit affinities with both faunal realms.

Samples of deep-marine siliceous rocks from the Quartzose Sedimentary Province around Cobar also contain faunas indicative of both the WFR and the CFR (*Pygodus, Periodon, Phragmodus* and *Pseudobelodina* from localities 92 & 94 – 101: Iwata et al. 1995).

Sweet and Bergström (1984) identified two biofacies, the *Periodon-Dapsilodus-Pygodus* biofacies, and the *Prioniodus (Baltoniodus)-Dapsilodus-Pygodus* biofacies, both of which are well developed in the Middle Ordovician of Baltoscandia and grade laterally into each other. The CFR localities identified in the Lachlan Fold Belt from both lithological provinces are most likely to be associated with either one or both of these two biofacies. This tentative association is based on the dominance of Middle to Late Ordovician species of *Pygodus* and *Periodon* occurring in the Lachlan Fold Belt samples. So far the associated genera *Dapsilodus* and *Eoplacognathus* have not been recognised from the Lachlan Fold Belt chert localities. However, *Eoplacognathus* is probably not able to be identified accurately in two-dimensions using the thick-section method.
Figure 3.9c Conodont localities identified for the Late Ordovician (Gisbornian to Eastonian).
The faunas with WFR affinities (localities 63, 71, 92, 94, 95, 98 - 101, 137, 142, 144 - 146, 148) are from both limestone and chert, and occur in three broad geographical areas; around Cobar, Narooma and the Molong Volcanic Belt (Figure 3.9b). Two localities (63, 71) are from Narooma on the south coast of New South Wales (Figure 1.3) where Middle-Late Cambrian limestone has been reported (Bishoff & Prendergast 1987). Therefore, these conodont faunas may indicate this area remained bathymetrically shallower during the Early to Middle Ordovician than areas of the surrounding basin which were dominated by CFR taxa throughout this time.

The other faunas with WFR affinities can be divided into two groups. Those from the Cobar area are all from deep-marine rocks of the Quartzose Sedimentary Province, and the faunas from the Molong Volcanic Belt are all from limestone of the Molong Volcanic Province. These results indicate that the Molong Volcanic Belt was a positive topographic feature by the Darriwilian/Gisbornian stages, but remained separated from the Gondwanan continent. Conodonts from deep-marine sediments in the Cobar area indicate the presence of a deep-marine basin between the Molong Volcanic Belt and the Gondwanan continent further to the west.

3.5.3 Gisbornian - Eastonian (Figure 3.9c)
Conodont localities of this age are concentrated in the limestone units of the Molong Volcanic Province (36 localities of 47) where the taxa predominantly indicate affinities with the WFR. Occurrences of *Belodina* and *Phragmodus* are common, but also associated with these taxa at several localities is the CFR indicator *Periodon grandis* (Etherington) (see localities 88, 89, 110 - 112, 146, 147, 150, 151) and *Scabbardella altipes* subsp. B. of Zhen and Webby (1995), an indicator in the HDS biofacies of the Late Ordovician CFR (Sweet & Bergström 1984).

The biofacies of the conodont assemblages recovered from the limestone were discussed by
both Trotter and Webby (1995) and Zhen and Webby (1995). Trotter and Webby (1995) noted that the conodont assemblage from the Eastonian Malongulli Formation (Cabonne Group; Figure 3.4) reflects derivation from both warm/shallow and cooler/deeper zones, and was probably formed initially as periplatform-zone deposits at the outer margins of an island platform.

Zhen and Webby (1995) recognised variations in the biofacies throughout the Cliefden Caves Limestone Group in central-western New South Wales (Cabonne Group; Figure 3.4). Three conodont biofacies were recognised within the lowermost Fossil Hill Limestone which they interpreted as reflecting environments ranging from near-shore restricted marine waters of the inner, fringing, island shelf, through to mid-shelf shoals, to more offshore, open marine waters of the outer island shelf. Biofacies within the overlying Belubula and Vandon Limestones were not as clearly recognised; however, a large number of forms with a (CFR) North Atlantic aspect were noted. The presence of at least two, separate, deeper biofacies associations off the island slope, the first occupying well-lit and oxygenated waters and the second, representing a deeper, probably cooler, water mass were suggested as the source of these CFR forms (Zhen & Webby 1995).

Gisbornian - Eastonian conodont faunas identified from the deeper marine rocks are restricted to the eastern side of the Molong Volcanic Belt (localities 2-4, 16, 26-31, 93). Black graptolitic shales are the most common deep-marine rocks of Late Ordovician age found in the Lachlan Fold Belt. Because graptolites are recovered with reasonable ease, conodonts are not usually looked for. Therefore, this lack of conodont localities within the deep-marine strata of this age is probably due more to a lack of study, rather than a lack of occurrence. Over half of the collection (6 of 11 localities) contain *Periodon grandis*, indicative of the CFR. Three of these localities (16, 30, 31) also contain the WFR form *Phragmodus*, and *Plectodina furcata* (Hinde) was also identified from locality 30. These may indicate some mixing of faunal realms with faunas from shallower biofacies being transported downslope after death or, faunas of the WFR living higher in the water column than the CFR forms, then conodonts from both faunal realms being deposited together on the ocean floor.
Overall, there is a greater WFR aspect to the faunas of this time period. However, this result may reflect a sampling bias and the few deep-marine faunas known indicate the CFR may still dominate in the deeper areas of the basin.

3.6 ORDOVICIAN PALAEOGEOGRAPHY OF THE LACHLAN FOLD BELT
The causes of provincialism probably include both ecological and zoological aspects, but the factors of major importance for conodont provincialism are ecological and include a twofold differentiation of faunas based on water temperature and depth. Palaeogeographic reconstructions by Scotese and McKerrow (1990) and Li and Powell (2001) both placed eastern Australia within 30° of the equator throughout the Ordovician. Therefore, variations in water temperature across the Lachlan Fold Belt would not have been greatly influenced by latitudinal variations, but rather by other factors including water depth, proximity to cold oceanic currents, upwelling zones and the Gondwanan continental margin. Changes in relative sea level would also have been an influencing factor in the shallow waters of the island platform during the Late Ordovician.

3.6.1 Early Ordovician
During the Early Ordovician, all Lachlan Fold Belt conodont faunas reflect a CFR aspect. This is consistent with stratigraphic and sedimentologic data which indicate a widespread quartz-turbidite fan prograding from the west-southwest to the east-northeast across the Lachlan Fold Belt throughout the Early Ordovician (VandenBerg & Stewart 1992).

In the Molong Volcanic Belt, the Early Ordovician Mitchell Formation (Figure 3.4) consists of volcanioclastic breccia, conglomerate and tuff and, although unfossiliferous, is conformably overlain by the predominantly fine-grained Hensleigh Siltstone (Figure 3.4) which contains allochthonous limestone clasts and Bendigonian graptolites and conodonts (Packham 1969; Percival et al. 1999). A similar stratigraphy was also reported from the Nelungaloo Volcanics, Junee-Narromine Volcanic Belt (Krynen et al. 1990; Percival et al. 1999), except that rather than allochthonous limestone lenses,
graptolites indicating a La3 to Be1/2 age were recovered from the Yarrimbah Chert Member (Figure 3.4). These stratigraphic relationships indicate that volcanism had commenced in both the Junee-Naromine and Molong Volcanic Belts by Bendigonian time. Although much of the northeastern Lachlan Fold Belt was dominated by deep-marine lithologies and CFR affinities during the Early Ordovician, large limestone lenses within the Hensleigh Siltstone (up to 100 m in length; Ian Percival, personnel communication 1999) indicate parts of the Molong Volcanic Belt had reached shallower water depths during this time.

3.6.2 Middle Ordovician
Only sparse fossil evidence exists for early to middle Middle Ordovician (Castlemainian and Yapeenian) deposition in the Lachlan Fold Belt of New South Wales. Within the Quartzose Sedimentary Province, Darriwilian (Da2-Da3) graptolites have been found southeast of Braidwood (Jenkins 1981) and both graptolites and conodonts have been identified from the Pittman Formation near Canberra (Öpik 1958, p. 15; Nicoll 1980). Conodont faunas indicating a Da2 to basal G1 age have also been identified from the Numeralla Chert (localities 63, 71, 137, 141). Both the Numeralla Chert and the Warbisco Shale overlie the unfossiliferous quartz turbidites of the undifferentiated Adaminaby Group, and the Pittman Formation is chronologically and lithologically equivalent to both the Adaminaby and Bendoc Groups (Figure 3.4).

Graptolites indicate that the entire Middle Ordovician is preserved within the Bendigo-Ballarat Zone of the Lachlan Fold Belt. Within this zone, the Middle Ordovician is represented by the Castlemaine Group (Figure 3.4), which is dominated by quartz-sandstone with minor lithic and feldspathic greywacke, interbedded with massive mudstone and black shale (VandenBerg & Stewart 1992). These rocks were probably deposited in a different environment on the turbidite fan than the open-ocean mid-fan environment inferred here for the eastern Lachlan Fold Belt. It is therefore likely that the Middle Ordovician (Castlemainian-Yapeenian stages) of the eastern Lachlan Fold Belt is represented by the unfossiliferous quartz turbidites underlying the black shales and cherts of the Warbisco Shale.
and Numeralla Chert; however, there is not yet any direct evidence of this.

Within the Molong Volcanic Province, a similar lack of Castlemainian-Yapeenian faunas has been attributed to a hiatus due to either:

- removal of shallow water deposits by erosion, or
- a lack of shallow water deposition (Percival et al. 1999).

Prior to the Darriwilian, the only evidence of shallow-marine deposition in Molong Volcanic Province rocks are the large allochthonous limestone lenses preserved in the Hensleigh Siltstone. As previously mentioned, these indicate that shallow-marine conditions existed around volcanic islands within the Molong Volcanic Belt by the Early Ordovician. However; there is no indication of shallow-marine to subaerial deposition of volcaniclastic detritus or deep-sea ash deposits in the Quartzose Sedimentary Province rocks. Therefore, even though the volcanoes in the Molong Volcanic Belt were at shallow water depths during the Early Ordovician, proximal deposition of volcaniclastic detritus erupted from this volcanic centre continued until the Darriwilian. It is further inferred that the volcanic edifices that erupted the other Early Ordovician volcanic deposits (Nelungaloo Volcanics and Gidyen Volcanic Member) formed well below sea level, and continued to erupt episodically and grow towards sea level during the Early and Middle Ordovician.

The reappearance of fossils in the record during the Darriwilian coincides with significant changes in the depositional environment. Within the Quartzose Sedimentary Province turbidite sedimentation decreased and was replaced by widespread chert and shale deposition.

On the volcanic highs of the Molong Volcanic Province, volcanism increased volumetrically, and carbonate platform deposits became more widespread. Subaerially erupted volcanic and volcaniclastic detritus along with the presence of oncolites in limestone lenses (Pickett 1984) provide evidence that some of the volcanoes had become emergent by late Darriwilian, although the majority of the volcanic edifices probably remained submergent until the Late Ordovician. Conodont faunas continued to
reflect a CFR aspect in the deep marine Triangle Formation (Fowler & Iwata 1995), and both CFR affinities and mixing of CFR and WFR affinities in the shallow marine limestones accumulated around the volcanic islands (Pickett 1984, 1985; Percival et al. 1999). The faunal affinities of this shallow, equatorial limestone may have been influenced by the upwelling of deep thermohaline currents. Contourites were reported from the Shoalhaven Gorge near Bungonia (Figure 1.3), and were attributed to contour currents moving northwards along the eastern margin of Gondwana from areas of subpolar downwelling (Jones et al. 1993). They related these contour currents to global climatic changes during the late Darriwilian to early Gisbornian.

Conodont faunas of deep marine rocks of the Quartzose Sedimentary Province also continued to exhibit a CFR affinity during the late Middle Ordovician except in the northwestern part of the Lachlan Fold Belt. Around Cobar (Figure 3.9b) conodont faunas identified in deep-marine chert and shale (Iwata et al. 1995) reflect a mix of both the WFR and the CFR. WFR conodonts in these rocks may be the result of one or several of the following: the Cobar region was probably closer to the shallow warm waters of the Gondwanan margin to the west; and/or the rising volcanic highs deflected the flow of cold ocean water away from the eastern margin of Gondwana, resulting in the marginal sea between the Gondwana margin and the chain of volcanic islands, which contained conodonts of various faunal affinities living at different depths and temperatures in the water column, but accumulating together in pelagic sediments after death.

3.6.3 Late Ordovician
By the early to middle Late Ordovician (Gisbornian to Eastonian) widespread accumulation of shallow marine limestone occurred in the shallow waters of the Molong and Rockley-Gulgong Volcanic Belts. Zhen and Webby (1995) have been able to identify most of the biofacies identified from the Late Ordovician WFR of North America within the Cliefden Caves Limestone Group of New South Wales (Cabonne Group; Figure 3.4). The presence of these shallow water biofacies along with widespread volcanic and volcanioclastic detritus indicate that many areas of the Molong Volcanic
Belt were either emergent or shallow marine during the Gisbornian - Eastonian. Minor variations in water depth are recognised, but probably resulted from eustatic changes in the sea level and/or local tectonic events (Zhen & Webby 1995).

Within the Quartzose Sedimentary Province widespread black shale and associated dysaerobic conditions dominated during the Late Ordovician; however, in the deep-marine areas surrounding the volcanic highs, volcanioclastic deposits interdigitate with black shale (see Chapter 2). Only a few conodont faunas are known from deep marine rocks of this age and, although these still predominantly reflect the CFR, some faunas do include species and genera from the WFR (Localities 16, 30, 31). These few specimens may either have been redeposited downslope from the volcanic islands (possibly as fecal deposits), or both WFR and CFR faunas lived at different levels in the water column surrounding the high and were later deposited together in the pelagic sediments.
CHAPTER 4
STRUCTURE

4.1 INTRODUCTION
The Oberon-Taralga region lies within the Capertee and Hill End Zones of the Eastern Belt of the Lachlan Fold Belt (Figure 1.2). The region is characterised by north-south trending anticlinoria containing Ordovician and Early Silurian rocks, synclinoria with Late Silurian and Devonian units and several granite batholiths (Figure 4.1). Most investigators have recognised at least two deformation events affecting the Ordovician-Early Silurian rocks in the region. The earliest event occurs locally within the Oberon, Rockley, Jaunter and Taralga districts and resulted in east-west to northwest-southeast trending folds devoid of cleavage (Figure 1.4; Binns 1958; Scheibner 1970; Scott 1985, 1991; Fowler 1989; Glen 1998). The second deformation event is the strongest to have affected the region and produced dominant north to northeast trending regional folds, faults and associated cleavage.

The objectives of this chapter are to describe and correlate the structural features from the five study areas within the Oberon-Taralga region. These areas were selected for detailed mapping as they potentially provided the most useful information on the stratigraphy and stratigraphic relationships of the Ordovician to Early Silurian rocks. Firstly, the structure of each study area is described separately, using the terminology DO, DT, DS, DG and DC to refer to the deformation events recognised within the Oberon, Thompson Creek, Silent-Oakey Creeks, Golspie and Crookwell River areas respectively (e.g. DO2 = Oberon area, second deformation event). These events are then correlated across the region and discussed in terms of regional deformation events, referred to as D1, D2 and D3.

Total Magnetic Intensity (TMI) and radiometric images cover the region and provide constraints on structures and the distribution of stratigraphic units. These images were compiled from confidential geophysical data acquired by Geoterrex Pty Ltd, for the New South Wales Department of Mineral Resources, the Australian Geological Survey Organisation (AGSO) and Geoterrex Pty Ltd. Both the
Figure 4.1 Map of the Oberon-Taralga region showing the distribution of the Molong Volcanic Province, Quartzose Sedimentary Province and the main structural features. Also shown are the location of the five areas discussed in the text (see enclosures 1-14): (1) Oberon area, (2) Thompson Creek area, (3) Silent-Oakey Creeks area, (4) Golspie area, (5) Crookwell River area. Also shown are the locations of Enclosures 2, 3, 9 and 10 in the Oberon Area.
Figure 4.2. Cross-section from along the Chain-of-Ponds Road oriented north-northeast to south-southwest between GR751062 6261625 and 751200 6261900 (Mt David 1:25,000 topographic sheet, 8830-III-N). Note variation in interlimb angles of F1 folds from near isoclinal to open. See Enclosure 3 for location. HS = VS.
data and images are commercially available to individuals and organisations but are not generally available and can only be referred to generally in this thesis.

4.2 STRUCTURAL ANALYSIS OF THE OBERON AREA
Within the Oberon area, Ordovician to Early Silurian rocks have been subjected to three episodes of deformation: DO1, DO2 and DO3 (Enclosures 8, 9, 10).

4.2.1 Deformation DO1
This deformation is recognised on the basis of pre-cleavage outcrop-scale folds occurring along Chain-of-Ponds Road in the western part of the Oberon area adjacent to Native Dog Fault and along the Hampton Road in the northeastern part of the area (Enclosures 1-3). Along Chain-of-Ponds Road most F1 folds are northeast trending with steep plunges (GR751062 6261625 and GR751200 6261900, Mt David 1:25 000 topographic sheet, 8830-III-N; Enclosure 3; Figures 4.2, 4.3). Whereas in the Hampton Road sites (GR771090 6263773, Oberon 1:25 000 topographic sheet 8830-1-S; Enclosure 2) they plunge moderately to the north and south (Figure 4.4a). F1 folds are upright, tight to open and typically angular. They lack axial planar cleavage but are cross cut by the S2 cleavage (Figure 4.4b, c).

On the map-scale it has not been possible to map out the development of F1 folds. Anomalous bedding orientations occur scattered throughout the Oberon area and include common steep dips to both the north and south amongst the dominant north-south structural grain (Figure 4.5). This is also supported by variation in F2 fold plunges as indicated by bedding/S2 cleavage intersection lineations which range from sub-horizontal to vertical in both northerly and southerly directions (Figure 4.5l). Deformation DO1 occurred after deposition of Early Silurian units, because F1 folds are known to affect Ordovician to Early Silurian rocks in the Oberon area.
**Eastern Limb:**
S0 = 137/61/E  
S2 = 144/80/E

**Western Limb:**
S0 = 107/90  
S2 = 160/50/E
Axial Surface = 121/70/E  
Hinge = 110/40  
Interlimb Angle = 30°

**Fold 1:**  
Hinge = 030/81  
Axial Surface = 054/70/SE  
Interlimb Angle = 77° (open)

**Fold 2**  
Hinge: 050/80  
Axial Surface: 060/80/SE  
Interlimb Angle: 65° (close)

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**Figure 4.3** Map view outcrop sketches of F1 folds occurring along Chain-of-Ponds Road in outcrops of cherty mudstone of the Mozart Chert Member, Triangle Formation, showing S2 cleavage cutting across bedding and F1 fold orientations. (a) Steeply inclined tight fold plunging steeply towards the southeast and crosscut by southeast trending S2 cleavage. (b) Two steeply inclined sub-vertical, northeast trending close to open folds. Both folds are crosscut by the younger east-southeast trending S2 cleavage, with no S1 cleavage observed. Location of (a) and (b) at GR751120 6261700, Mt David 1:25 000 topographic sheet 8830-III-N.
Figure 4.4 (a) Equal area lower hemisphere stereographic projection showing orientations of F1 folds from Chain-of-Ponds Road and Hampton Road. Blue squares = axial surfaces of F1 folds on Hampton Road, blue triangles = plunges of F1 fold hinges - Hampton Road, red squares = axial surfaces of F1 folds on Chain-of-Ponds Road and red triangles = plunges of F1 fold hinges - Chain-of-Ponds Road. (b) Photograph of mesoscale F1 folds in the Mozart Chert Member, Chain-of-Ponds Road. Geological hammer is 33 cm in length. (c) Accompanying sketch of mesoscale F1 folds in (b), solid thin lines represent bedding trends with some thick beds filled in black, thick solid lines are small faults and fractures, and dashed lines indicate direction of S2 cleavage.
4.2.2 Deformation D02
Deformation D02 has affected all rocks of Late Ordovician and Silurian age and is the strongest deformation event recognised in the Oberon area. It is characterised by widespread meridional trending map-scale folds and outcrop-scale folds.

OUTCROP-SCALE F2 FOLDS
F2 outcrop-scale folds are northerly trending, steeply inclined and have variable plunges from sub-horizontal to vertical (Figures 4.6, 4.7). They are tight to close with angular to rounded hinges. Common F2 outcrop-scale folds are exposed in the Lower Ordovician Budhang Chert Member along Brisbane Valley Creek between GR756450 6259400 and 756350 6258750 (Edith 1:25 000 topographic sheet 8830-II-N, Enclosures 1, 8), in the Mozart Chert Member in Native Dog Creek (Figure 4.7d, f; Mt David 1:25 000 topographic sheet 8830-III-N), along Hampton Road (GR770525 6264102, Oberon 1:25 000 topographic sheet 8830-I-S) and along the Duckmaloi River (GR768376 6253763, Edith 1:25 000 topographic sheet 8830-II-N, Enclosures 1, 2, 8, 9).

MAP-SCALE F2 FOLDS
Map-scale F2 folds are shown in Figures 4.8, 4.9 and on the Oberon and Hampton Road geological maps (Enclosures 1, 2, 8, 9). They are tight to close, angular or arrow-headed folds, which plunge gently to moderately to both the north and south. Wavelengths are up to 2000 m, and parasitic folds with ~500 m wavelengths also occur (Figure 4.9c, d). West of the Vulcan Fault in the Oberon Map area, the axial plane orientations of F2 folds are predominantly upright or steeply inclined towards the west (Enclosures 1, 8, Figure 4.9a-g). Whereas, F2 folds east of the Oberon South Fault, along Hampton Road and the Duckmaloi River, are either upright or their axial planes dip towards the east (Enclosures 1, 2, 8, 9, Figures 4.8a-h, 4.9h-j).

S2 CLEAVAGE
A regional S2 cleavage is associated with the F2 folds (Enclosures 8, 9, 10, Figures 4.8, 4.9) and
occurs as a generally steeply dipping, northerly trending slaty structure that is either parallel or subparallel to bedding in the less competent black shale and mudstone units. It commonly displays a silvery sheen due to strongly aligned micas. In the more competent quartz and volcaniclastic sandstone units, the intensity of the cleavage is heterogeneous, occurring as both a slaty cleavage and as a spaced cleavage. Within the turbiditic units, the cleavage is commonly refracted across sandstone to mudstone beds. Within the coarse-grained volcaniclastics of the Triangle Formation, S2 is weak and difficult to discern in hand-specimen. It is defined by the presence of weakly flattened crystals and amygdales.

4.2.3 Deformation D03
This deformation produced F3 folds which affect F2 folds, S2 cleavage and some faults across the whole area (Enclosures 1, 2, 8, 9). The F3 folds are open to gentle, devoid of any cleavage and trend east-northeast and east-southeast. S2 cleavage is locally warped into near east-west strikes and less commonly is folded from steep to gentle dips (Enclosures 8, 9, 10).

4.2.4 Faults
Faults occurring in the Oberon area are north and northeast trending map-scale faults shown on the Oberon and Hampton Road geological and structural maps (Enclosures 1, 2, 8, 9). The existence of many of these faults has been inferred from map relationships at contacts in the area. Their existence is supported by interpretation of RGB radiometric and magnetic (TMI) images. They are typically marked by areas of poor to non-existent exposure. Reliable information on the dips of faults and kinematic criteria has been difficult to find.

Native Dog Fault
This fault occurs near the western boundary of the Oberon Map area (Enclosures 1, 8), where it juxtaposes Middle Ordovician Triangle Formation, the overlying early Upper Silurian Fosters Creek Conglomerate and Bells Creek Volcanics, against Upper Silurian rocks of the Campbells Formation.
Figure 4.5 Equal area, lower hemisphere stereographic projections. (a) Poles to bedding (S0) orientations from measured section along Chain-of-Ponds Road, n = 52. (b) Poles to bedding (S0) orientations from measured section along Hampton Road, n = 110. (c) All poles to bedding (S0) orientations from the Oberon area including Chain-of-Ponds and Hampton Roads, n = 423. (d) Poles to S2 cleavage orientations from measured section along Chain-of-Ponds Road, n = 65. (e) Poles to S2 cleavage orientations from measured section along Hampton Road, n = 198. (f) Poles to all S2 cleavage orientations from the Oberon area including Chain-of-Ponds and Hampton Roads, n = 852. (g) Bedding/S2 cleavage intersection lineations for measurements recorded from measured section along Chain-of-Ponds Road, n = 39. (h) Bedding/S2 cleavage intersection lineations for measurements recorded from measured section along Hampton Road, n = 75. (i) All bedding/S2 cleavage intersection lineations from the Oberon area including Chain-of-Ponds and Hampton Roads, n = 248.
In outcrop the fault could be traced for approximately 500 m north of Native Dog Creek by a fault breccia exposed in a small tributary along Johns Gully (Enclosure 1). The fault breccia is composed of angular clasts (< 4 cm) of mafic volcanics, cherty mudstone, quartz porphyry and slate derived from the surrounding rock units and contained within a fine-grained grey/brown matrix with a scaly cleavage. On the eastern side of Johns Gully, the Fosters Creek Conglomerate is abruptly cut by the fault. South of Native Dog Creek the fault is difficult to trace due to lack of outcrop, but is inferred to continue south-southwest on the basis of the RGB radiometric image.

According to Glen (1998) the Native Dog Fault is a steep west-dipping thrust, marked by a 10 m wide shear zone. If the fault is west-dipping, then younger rocks occur in the hanging wall over older Ordovician rocks in the footwall and it would therefore have developed as a normal fault.

**BRISBANE VALLEY FAULT**
The Brisbane Valley Fault occurs to the south of the Sloggetts Granite and trends to the northeast (Enclosure 1). West of the fault is the Lower Ordovician Budhang Chert Member that forms a map-scale anticlinal structure and to the east is a map-scale F2 syncline cored by the Upper Ordovician Rockley Volcanics. Close to the fault, bedding orientations in both the Budhang Chert Member and the Rockley Volcanics are commonly oblique to the fault and dip steeply towards the east. S2 cleavage is also commonly oblique to the fault, but is either upright or dipping steeply towards the west. The fault is thought to be a reverse fault dipping steeply to the west, placing the older members of the Triangle Formation over Upper Ordovician Triangle Formation and Rockley Volcanics.

**VULCAN FAULT**
The Vulcan Fault is the most significant fault occurring in the Oberon area and occurs between the two Ordovician lithological provinces. It occurs to the north and south of the Oberon Granite, which is therefore, a stitching pluton of Carboniferous age. The fault juxtaposes Middle to Upper Ordovician Quartzose Sedimentary Province units on the east, against Middle to Upper Ordovician Triangle
Figure 4.6 (a) Equal area lower hemisphere stereographic projection showing orientations of F2 folds from Hampton Road and Brisbane Valley Creek. Blue diamonds = Axial surface of F2 folds on Hampton Road, blue stars = trend and plunge of F2 fold hinges - Hampton Road, red diamonds = axial surface of F2 folds in Brisbane Valley Creek, red stars = trend and plunge of F2 fold hinges - Brisbane Valley Creek. (b) Contoured equal area lower hemisphere stereographic projection of all F2 axial planes from the Oberon area.
Figure 4.7 Photographs and accompanying sketches of outcrop-scale F2 folds from Brisbane Valley and Native Dog Creeks. (a) North trending, vertically plunging F2 folds in black chert of the Budhang Chert Member, Triangle Formation, Brisbane Valley Creek (GR756433 6258885, Edith 1:25 000 topographic sheet, 8830-II-N). Hammer is 28 cm long. (b) Accompanying sketch of F2 folds in (a). Some chert beds shown in black. (c) South-southeast trending, steeply plunging F2 folds in the Budhang Chert Member, Triangle Formation, Brisbane Valley Creek (GR756379 6258716, Edith 1:25 000 topographic sheet, 8830-II-N). Hammer is 28 cm long. (d) Accompanying sketch of F2 folds in (c). Some chert beds are shown in black. (e) Northwest trending, moderately plunging F2 fold in the Mozart Chert Member, Triangle Formation, Native Dog Creek (GR751344 6261847, Mt David 1:25 000 topographic sheet, 8830-III-N). Note small fault planes. Hammer is 33 cm long. (f) Accompanying sketch of F2 folds in (e). Some thicker chert beds are shown in black, thinner beds are shown by thin lines. Note duplication of thicker chert bed by contraction fault at a low-angle to bedding and in the core of the fold.
Figure 4.8 Hampton Road cross-sections. For locations of sections and legend see the Hampton Road Geological Map (Enclosure 2). The horizontal scale = the vertical scale on all sections. (a) Hampton Road A-A’. (b) Hampton Road B-B’. (c) Hampton Road C-C’. (d) Hampton Road D-D’. (e) Hampton Road E-E’-E’’. (f) Hampton Road F-F’. (g) Hampton Road G-G’. (h) Hampton Road H-H’.
1100
1050
1000
950

(a) A

1100
1050
1000
950

(b) B

HS = VS

metres
0 50 100 150 200

Metres above sea level
HS = VS
Figure 4.8c, d.

Metres above sea level
Figure 4.8e. Metres above sea level
Figure 4.9 All cross-sections for the Oberon Area. For locations and legend see the Oberon Geological Map (Enclosure 1). (a) Oberon A-A’. (b) Oberon B-B’. (c) Oberon C-C’-C”. Note cross section (b) is drawn at a different scale to the other cross-sections. (d) Oberon D-D’. (e) Oberon E-E’ and F-F’. (f) Oberon G-G’. (g) Oberon H-H’. (h) Oberon J-J’. (i) Oberon K-K’. (j) Oberon L-L’
Metres above sea level

Figure 4.9a.
Figure 4.9b, c.
Figure 4.9d, e, f, g.
Metres above sea level

Metres

0 1000 2000

Metres

HS = VS

Figure 4.9h, i, j.
Figure 4.10  Scaly cleavage occurring in the Numeralla Chert and the Bumballa Formation outcropping on the northeastern side of the Vulcan Fault (GR759993 6257995, Edith 1:25 000 topographic sheet 8830-II-N). Photograph is looking east, S2 cleavage dipping towards the northeast. Hammer for scale is 28 cm long.
Formation to the west. The fault plane itself is not exposed in this area, but ubiquitous quartz veins occur in the general area. The fault crosses the Goulburn Road at GR760000 6258000 (Edith 1:25 000 topographic sheet, 8830-II-N), where a break in outcrop of approximately 10 m occurs. To the northeast of the fault, the Numeralla Chert and Bumballa Formation outcrop. Almost all traces of bedding have been destroyed and the outcrop displays a scaly cleavage (Figure 4.10).

Bedding orientations across the fault on Goulburn Road are oblique to the fault. South of the Oberon Granite, the fault trace is curved through several gentle folds, the significance of these features is unknown but is possibly related to folding by F3 folds. East of the fault, there is little outcrop of Quartzose Sedimentary Province rocks. In TMI images, the Vulcan Fault is a magnetic lineament parallel to and coincident with the mapped lithological boundaries except in the area immediately north of the Black Springs Granite where the magnetic lineament is located approximately 1 km west of the mapped fault trace.

Glen (1998) considered that the Vulcan Fault dipped east and resulted in Lower to Middle Ordovician quartz turbidites being placed over younger Upper Ordovician "Rockley Volcanics" (herein Triangle Formation). Reverse slip was inferred from two hanging wall anticlines and a footwall syncline all occurring in the "Rockley Volcanics" to the north of Oberon.

**OBERON SOUTH FAULT**

This fault occurs to the east of Oberon and is followed to the south approximately parallel to the Shooters Hill Road then south-southwest along a tributary of the Fish River (Enclosures 1, 8). In the Oberon area, it juxtaposes Middle to Upper Ordovician Adaminaby Group, Warbisco Shale and Triangle Formation volcanoclastics against Upper Ordovician and ?Lower Silurian Rockley Volcanics, Triangle Formation and Bald Ridge Formation. Further south its location is speculative and based on distinction between the Adaminaby Group and Bald Ridge Formation (see Chapter 2). West of the fault, both bedding and S2 cleavage orientations dip steeply towards the east; whereas, east of the
Figure 4.11 (a) Map-view sketch showing outcrop-scale F1 fold at GR727422 6241986, Abercrombie 1:50 000 topographic sheet, 8830-II and III, (this locality is within area shown in Figure 7.11; see Enclosure 11). (b) Equal-area, lower-hemisphere stereographic projection of the F1 fold data shown in (a). Red squares = axial planes; blue diamonds = fold hinges; black dots = poles to S2 cleavage. Great circle is the average S2 orientation.
Figure 4.12 See Enclosure 11 for location. (a) Enlarged map of part of the Thompson Creek area showing axial traces of F1 folds on the limb of an F2 syncline, and the location of cross sections H-H' and J-J'. (b) Cross section H-H' showing southwesterly vergent F1 folds. (c) Cross section J-J' showing southwesterly vergent F1 folds.
fault bedding predominantly dips and youngs towards the west and S2 cleavage is moderate to steep and dips to both the east and west. Based on this data, the fault is considered to be dipping steeply to the east and therefore is possibly a normal fault.

**Deep Creek Fault and other unnamed faults mapped along Hampton Road**
The faults mapped along Hampton Road trend northerly to northeasterly. These faults are locally exposed along Hampton Road but away from the road are mapped on the basis of the distribution of the rock units. In road cuttings, these faults are marked by zones 10-50 m wide with angular cleaved rock fragments, derived from adjacent rock units, within a fine-grained matrix. Cleavage commonly intensifies towards these zones, obscuring bedding and sedimentary features. Where coarse-grained outcrops of quartz sandstone or volcaniclastic sandstone occur adjacent to the fault zones, they are either intensely cleaved or jointed. The faults dip at moderate to steep angles towards both the east and west subparallel to S2 cleavage. The Hampton Road faults are considered to have been active during deformation D02 based on their relationship to S2 cleavage.

**4.3 Structural Analysis of the Thompson Creek Area**
Early Silurian rocks in Thompson Creek form the core of the Thompson Anticline (Stanton 1956) and have been affected by three deformation events. The second of these deformation events (DT2) was the most intense, whereas the first and third deformation events (DT1 and DT3) were relatively mild.

**4.3.1 Deformation DT1**
Deformation DT1 produced rare outcrop folds and more common map-scale folds. They are recognised by the cross-cutting S2 cleavage and local refolding by F2 folds (Figures 4.11, 4.12). F1 folds are typically tight with southeast-trending axial planes that dip at moderate to steep angles towards the northeast. Outcrop F1 folds occur at GR727422 6241986 (Abercrombie 1:50 000 topographic sheet, 8730-II and III, Figure 4.11). Map-scale F1 folds occur in Thompson Creek near the centre of the Thompson Anticline, where the creek runs north-south (GR727000 6242000 and
Figure 4.13 Equal-area lower-hemisphere stereographic plots for data from the Thompson Creek area. (a) Poles to bedding (S0) data, n = 188. (b) Poles to S2 cleavage, n = 250. (c) F2 folds, n = 28. Red squares = poles to axial planes; blue diamonds = fold axes; black dots = poles to S2 cleavage. (d) Bedding (S0)/S2 cleavage intersection lineations, n = 169. (e) Poles to fault planes, n = 14. (f) Pink triangles = F3 axial planes, n = 9; blue diamonds = poles to late joints, n = 53.
Later F2 folding has resulted in F1 trends varying from southeast to south (Figure 4.12). A wider distribution of F1 folds is indicated in the Bald Ridge Formation by complicated bedding orientations; stereographic plots show a wide scatter in the strike of steeply dipping bedding planes compared to S2 (Figure 4.13a, b). Deformation DT1 post-dates deposition of the ?Lower Silurian Bald Ridge Formation, but it is not known whether it affects the Kangaloolah Volcanics occurring on the margins of the Thompson Anticline (Enclosures 4, 11).

**4.3.2 Deformation DT2**

Deformation DT2 has affected all the rocks of Silurian age in Thompson Creek, and is the strongest deformation event recognised in this area. It is characterised by widespread meridional trending folds and map-scale faults.

**Outcrop-scale F2 folds**

Outcrop-scale F2 folds are widespread, with hinges of small (< 5 m) folds common (Figure 4.14a-e). F2 folds generally have rounded hinges with planar but short limbs, and plunge towards the north and south (Figure 4.13c). The axial planes vary in strike from northwest to northeast, are upright or dip steeply towards the east, indicating a slight westward vergence. Variation in strike of the axial planes is due to weak F3 folding. Interlimb angles are tight to open. The F2 folds have an associated axial planar S2 cleavage. Small contraction faults affecting a single sandstone bed occur in the hinges of F2 folds (Figure 4.15a-d) and resemble those associated with tangential longitudinal strain on the inner bend of a folded competent layer (Ramsay & Huber 1987, p.459-461).

**Map-scale F2 folds**

Map-scale folds are shown in Figure 4.16 and on the Thompson Creek geological and structural maps (Enclosures 4, 11). They occur as tight to close, upright folds with rounded to angular hinges and planar limbs. The axial planes are either upright or steeply dipping, and trend north-south with minor
Figure 4.14 Photographs and accompanying sketches of outcrop-scale F2 folds from the Thompson Creek area. (a) - (c) Northeast trending, moderately to steeply plunging F2 folds in quartz sandstone of the Bald Ridge Formation, Thompson Creek (GR728297 6242147, Abercrombie 1:50 000 topographic sheet, 8830-II and III). Hammer is 33 cm long. (b) - (d) Sketches of F2 folds in (a) and (c). (e) North to northeast trending moderately to steeply plunging F2 folds. These F2 folds affect the contact between the Bald Ridge Formation and the overlying Kangaloolah Volcanics along Bald Ridge Road (GR720943 6244685, Abercrombie 1:50 000 topographic sheet 8830-II and III). The measuring tape case at the base of the outcrop is 35 cm from the base to the top of the handle.
Figure 4.15 Photograph of a contraction fault affecting a single sandstone bed in the Bald Ridge Formation at Collinore (GR729276 6242137, Abercrombie 1:50 000 topographic sheet, 8830-II and III). Upper bed terminates just left of photograph. Hammer is 33 cm in length. (b) Sketch of photograph in (a). Solid lines show the outlines of sandstone beds. (c) The photograph shows a bedding-parallel contraction fault affecting a single sandstone bed in the hinge of an F2 fold outcropping in Thompson Creek (GR729155 6242229, Abercrombie 1:50 000 topographic sheet 8830-II and III). Swiss Army knife is 9 cm long. (d) Sketch of photograph shown in (c). Solid thick lines are outlines of the sandstone bed, solid thin lines indicate fractures and quartz veins, and the broken thin lines indicate the direction of S2. (e) Broken, quartz sandstone clasts in the shear zone of the Forest Hills Fault at GR723897 6241366, Abercrombie 1:50 000 topographic sheet, 8730-II and III). The clasts are extended in the S2 direction and show jigsaw fits. (f) Thick north trending quartz vein occurring along the fault on the eastern side of the Thompson Creek area. The photograph is taken looking towards the north. The Bald Ridge Formation outcrops west of the quartz vein (left of photograph) and phyllites of the Campbells Formation outcrop to the east (right of photograph). Person included as a scale.
Figure 4.16 Cross-sections from the Thompson Creek area. See Enclosure 5 for legend and location of sections. HS = VS (a) Cross Section A-A'. (b) Cross-section B-B' and B''-B''''. (c) Cross-section C-C'. (d) Cross-section D-D' and D''-D'''. (e) Cross-section E-E'. (f) Cross-section F-F'''. (g) Cross-section G-G'.
Heights are in metres above sea level.

Figure 4.16a, b, c.
variation caused by weak F3 folds. Bedding-S2 cleavage intersection lineations ($L^0_2$) indicate plunges are variable, trending towards both the north and south at moderate to steep angles (Enclosure 11, Figure 4.13d). Wavelengths of F2 folds are up to 2000 m, but are commonly less (Figure 4.16a, c, d, e, f).

**S2 Cleavage**
A regional S2 cleavage is axial planar to the F2 folds (Enclosures 4, 11, Figure 4.16). In mudstone, it is a northerly trending, steeply dipping slaty cleavage, generally either close to or parallel to bedding (Figure 4.13a, b). Within mudstone, the S2 surface commonly exhibits a silvery sheen due to strongly aligned micas; whereas, within sandstone, the cleavage is disjunctive with a closer spacing in fine-grained rocks. Cleavage is refracted from sandstone to mudstone layers and the intensity increases adjacent to faults. Where the rocks have undergone more intense contact metamorphism, typically on the flanks of the anticline, the cleavage is not as well developed as normal and in some places, is difficult to discern at all.

**4.3.3 Deformation DT3**
This deformation produced F3 kinks and folds. The presence of F3 map-scale folds is indicated by minor smooth changes in the orientation of both the F2 folds and S2 cleavage across the whole area (Enclosures 4, 11). The axial traces of the F3 folds trend approximately east-west and dip gently. Outcrop-scale F3 kinks occur in the more ductile mudstone and actinolite schist, and the axial planes of these dip at angles of less than 30°. Outcrop-scale F3 kinks also occur in quartz phyllites of the Campbells Formation on the eastern side of the Thompson Anticline. They have wavelengths ranging from 1 to 3 cm, interlimb angles of approximately 90° and their axial planes dip gently towards the west and crenulate the S2 cleavage (Figures 4.13f, 4.17e, f). These F3 kinks are notably abundant east of the Collinore Fault on the eastern flank of the Thompson Anticline and indicate that this structure was reactivated during the DT3 deformation.
4.3.4 Late Joint Surface
Prominent, gently dipping joints occur in all competent sandstone and volcanic units in the Thompson Creek area (Figure 4.14e). These joints cross-cut and post-date S2 and have similar gentle dips to the axial planes of the F3 kink folds (Figure 4.13f).

4.3.5 Faults

Outcrop-scale faults were observed across the Thompson Creek Area and commonly occur close to map-scale faults (Figure 4.13e). They trend from northeasterly to northwesterly and offset bedding and quartz veins with displacements ranging from 20 cm up to 3 m. Dips on the fault planes are predominantly steep and moderate towards the east. Apparent offsets along northeast trending faults are sinistral, whereas apparent offsets along northwest-trending faults are dextral. If these faults developed in a strike-slip regime these apparent offsets would indicate a northerly compression direction.

Map-scale faults

Forest Hills Fault
The Forest Hills Fault outcrops in the western part of the Thompson Creek area and is north-northeasterly trending (Enclosures 4, 11). It is characterised by shear zones containing small (<20 cm) angular clasts composed of quartz sandstone. These are commonly extended in the direction of S2 but retain jigsaw fits (Figure 4.15e). The dark fine-grained matrix is strongly cleaved parallel to the S2 direction and S2 intensifies close to the fault. This fault is visible on the TMI magnetic image, where it has apparent dextral offsets of the Kangaloolah Volcanics to the south (Raymond et al. 1997).

Collinore Fault
The Collinore Fault occurs on the eastern margin of the Thompson Anticline, where it is marked by changes in lithology and is oblique to bedding (Enclosures 4, 11). An increase in strain is
Figure 4.17 (a) Boudinaged sandstone beds in shear zone shown in Thompson Creek at GR727245 6240669 (Abercrombie 1:50 000 topographic sheet, 8830-II and III; Enclosure 12). Hammer is 33 cm long. (b) Large quartz vein in Bald Ridge Formation parallel to the S2 cleavage occurring in Thompson Creek at GR725893 6240565 (Abercrombie 1:50 000 topographic sheet, 8830-II and III). This quartz vein occurs to the southwest, along strike from the shear zones shown on Enclosure 12. (c) Photograph of boudinaged, poorly cleaved sandstone blocks within mudstone containing a slaty cleavage along Bald Ridge Creek (GR723110 6243696, Abercrombie 1:50 000 topographic sheet, 8830-II and III). These rocks have undergone dextral shear associated with the fault crossing Bald Ridge Creek (Enclosures 4, 11). Hammer is 33 cm long. (d) Accompanying sketch of boudinaged sandstone blocks in (c). Solid lines indicate S2 cleavage in mudstone and speckled pattern indicates sandstone blocks. (e) Weak F3 kinks dip gently towards the west (right of photograph) and is weakly crenulating the steeply dipping S2 cleavage approximately 100 m east of the quartz vein and inferred fault shown in (4.14f). Top of the hammer is 4 cm across. (f) Gently dipping F3 kink folds, crenulating S2 in a hand specimen collected from beside the inferred fault shown in (4.14f). The kinking is stronger close to the fault.
demonstrated by stronger S2 development as the fault is approached. The fault is marked by the presence of abundant vein quartz. A large, north-south oriented quartz vein occurs at the contact between the Bald Ridge Formation and the Campbells Formation (Figure 4.15f). The Collinore Fault is a strong magnetic lineament on the TMI image.

**Thompson Creek Fault**
This north-northeast trending fault cuts through the centre of the Thompson Anticline (Enclosures 4, 11) and coincides with a change in the trend of the S2 cleavage from predominantly northerly to the west of the fault, to north-northeast on the eastern side. Rocks in this vicinity show significant deformation with a 3 to 4 m wide melange zone observed at GR725983 6240721 and 725908 6240651 (Abercrombie 1:50 000 topographic sheet, 8830-II and III). Boudinaged angular, weakly cleaved sandstone blocks occur within a mudstone matrix exhibiting a slaty cleavage at GR726436 6240376 and 726580 6240203 (Abercrombie 1:50 000 topographic sheet, 8830-II and III), where fishtails indicate that the sandstone blocks have undergone dextral rotation. Movement on this fault has also resulted in the axial traces of F1 folds being dextrally offset (Figure 4.12).

**Unnamed map-scale faults**
Several unnamed north to northeast oriented map-scale faults occur in the Thompson Creek area. These are characterised by shear zones <5 m wide containing blocks of quartz sandstone in a fine-grained matrix, well developed S2 cleavage and the occurrence of strong jointing and large quartz veins (Figure 4.17a, b). The shear zones are aligned parallel to S2 cleavage. Boudinaged sandstone beds within cleaved mudstone were also observed along Bald Ridge Creek (Figure 4.17c, d). This indicates movement occurred on these faults during the DT2 deformation.

### 4.4 STRUCTURAL ANALYSIS OF THE SILENT-OAKEY CREEKS AREA
The Silent-Oakey Creeks area is dominated by a complicated set of north-trending folds and related faults (Enclosures 6A, 6B). A number of stratigraphic units have been mapped across the area,
including two subunits of the Bumballa Formation, the overlying Bald Ridge Formation and the underlying undifferentiated Adaminaby Group, that have enabled recognition of faulted contacts. Structural relationships differ from elsewhere in the Oberon-Taralga region as the main-phase cleavage and tight folds are typically gently dipping. There are also areas with steeply dipping structures occurring amongst the more common gently to moderately dipping structures. Evidence for multiple deformation includes local refolded folds and downward facing folds.

Three deformation events referred to as DS1, DS2 and DS3 have been recognised in the Silent-Oakey Creeks area (DS indicating Deformation - Silent) and are described below. The area has been subdivided into eight domains, numbered from west to east and bounded by north-trending faults on the eastern and western margins, except for the boundary between Domains III and V which is a southeast trending fault (Figure 4.18). The map-scale structure of the Silent-Oakey Creeks area is described in terms of these eight domains.

### 4.4.1 Deformation DS1
This deformation is recognised on the basis of a few early easterly trending folds and more widely on the basis of overturning of strata that predates the main-phase deformation (DS2). Outcrop-scale F1 folds occur in Domain I and are east-west trending, tight to isoclinal and recumbent (GR746134 6216508, Fullerton 1:25 000 topographic sheet, 8829-IV-S; Figure 4.19a, b). The vergence direction of these folds indicates a north over south transport direction (Figure 4.19a); although the facing direction of these folds is unknown. No S1 cleavage is associated with these folds and they are transected by the S2 cleavage. Early recumbent folds are also indicated in Domain VII by the presence of a series of downward facing F2 folds.

### 4.4.2 Deformation DS2
This is the most significant deformation event in the Silent-Oakey Creeks area, and has resulted in north trending, moderately to gently plunging map- and outcrop-scale F2 folds with an axial planar S2
Figure 4.18 Structural Domains of the Silent-Oakey Creeks Area
Figure 4.19 (a) Sketch showing cross-section view of outcrop-scale east-west trending, recumbent F1 folds in Domain I at Gr746134 6216508 (Fullerton 1:25 000 topographic sheet, 8829-IV-S). Dashed line represents the axial surface. (b) Equal-area, lower-hemisphere stereographic projection of the F1 fold data shown in (a). Blue triangles = poles to axial planes; red diamonds = fold hinges; n=4.
cleavage (Figure 4.20). In Domain VII some of these folds are downward facing presumably due to interference with pre-existing F1 folds.

Hinges of F2 folds are exposed in Domains III, V, and VII, where they occur in the undifferentiated Adaminaby Group. F2 folds are tight to isoclinal and rounded to angular with long planar limbs. They are typically near recumbent to moderately inclined (Figures 4.21a, b, c, d, 4.22a, b). Examples occur at GR747172 6216991 and 747352 6218173 (Fullerton 1:25 000 topographic sheet, 8829-IV-S). In Domain VII the axial planes are upright or inclined at moderate to steep angles towards the west (Figure 4.21c, 4.23). Based on both outcrop measurements and bedding/S2 cleavage intersection lineations ($L^0_2$), the hinges of F2 folds in the Silent-Oakey Creeks area generally plunge towards the north-northeast and south-southwest at moderate to gentle angles (Enclosure 12, Figures 4.20, 4.21c).

The S2 cleavage is the main cleavage formed in the Silent-Oakey Creeks area. It is associated with F2 folds and affects rocks in all eight domains of the Silent-Oakey Creeks area. In Domains I, II, III, IV, V and VIII, the S2 cleavage varies from a strong slaty cleavage in mudstone, to a closely spaced cleavage of a few mm up to 1 cm in sandstones that is either parallel to or slightly oblique to bedding on fold limbs (Figures 4.20b, e, h, k, n, w, 4.21). S2 becomes stronger as faults are approached, becoming slaty in fine sandstone beds. In Domains VI and VII, the S2 cleavage is also slaty and commonly either parallel to or slightly oblique to bedding (Figure 4.20q, t). In these domains, S2 predominantly dips towards the west at steeper angles than in other domains.

In thin section, the cleavage is defined by 'P' domains (Stephens et al. 1979; Waldran & Sandiford 1988), composed of preferentially aligned white mica and chlorite, alternating with 'Q' domains of quartz, rare feldspar and white mica aggregates. Flattening of quartz grains in the 'Q' domain has occurred in rocks (R17991 and R17990) collected from the hanging wall within 200 m of the Silent Creek Fault in Domain III, and adjacent to the fault zone at GR748767 6218435 (Fullerton 1:25 000 topographic sheet, 8829-IV-S; R17992) in Domain VII. Flattened quartz grains were not observed
Figure 4.20 Stereonet plots of bedding (S0), S2 cleavage and S0/S2 intersection lineation (L°2) data from the Silent and Oakey Creeks area for each domain and for the total area.
Figure 4.21 (a) Photograph and (b) sketch of northwest trending recumbent fold in Domain V at GR747352 6218173. Knife is 2 cm wide. (c) Stereonet showing F2 folds from the Silent-Oakey Creeks area. Blue triangles = axial planes of folds in Domains III and V, blue diamonds = hinges of F2 folds in Domains III and V, red triangles = axial planes of folds in Domain VII and red diamond = hinges of fold in Domain VII. (d) Gently dipping beds in cliff beside Silent Creek. Reversals in younging direction occur but fold hinges were not found.
from anywhere else in the Silent-Oakey Creeks area.

4.4.3 Deformation DS3
Deformation DS3 formed gentle to open, outcrop- and map-scale F3 folds and a rare but associated S3 cleavage. F3 outcrop-scale folds were observed in Domains III, IV and VI (Figure 4.24). They are upright folds with rounded hinges that gently fold bedding and S2 cleavage. Axial planes dip at moderate angles towards the west and are generally devoid of any S3 cleavage (Figure 4.25). In Domains IV and VI however, they are associated with a rare, north-south oriented, steeply to moderately dipping S3 crenulation cleavage (Enclosures 5, 12).

4.4.4 Map-scale Structure
On the map-scale the area is dominated by easterly verging recumbent to locally steeply dipping F2 folds and associated contraction faults (Figure 4.22).

Domain I
Stratigraphic repetition of the different facies of the Bumballa Formation indicates that this domain consists of two moderately west-dipping imbricate thrust slices (Figures 4.18, 4.22a). Strata within both thrust slices are overturned towards the east with consistent easterly younging directions. These slices form part of the overturned limb of a major westerly inclined anticline that is cored to the west by the Adaminaby Group. The thrust faults trend north-south and dip towards the west parallel to bedding in the hanging walls (Figure 4.20b, c; Enclosure 5). The thrust fault on the boundary between Domains I and II is marked by an abrupt change from the Bumballa Formation to the Adaminaby Group further east. The Adaminaby Group is overturned and steeply westward dipping; it becomes more strongly cleaved as the fault is approached from the east (Figure 4.20b, c).

Domain II
This domain is dominated by a map-scale F2 syncline that is folded by steeply inclined, open F3 folds
Figure 4.22 Cross-sections from the Silent-Oakey Creeks area. See Enclosure 5 for cross-section localities and legend.
(a) Cross-section A-A'-A''
(b) Cross-section B-B'
(c) Cross-section C-C'
in the eastern part of the domain (Figures 4.22b, 4.24). The F2 fold is a tight near recumbent structure in the east and further west has a moderately dipping axial plane to the west. In cross section B-B’ (Figure 4.22b) two faults are shown along axial planes of F3 folds. These faults consist of sheared angular blocks of quartz sandstone within a fine-grained matrix and locally strongly cleaved angular sandstone rock fragments. The steep dips on the fault zones in Domain II contrasts with the moderate westerly dips of the imbricate thrust faults in Domain I. The map pattern indicates that the steeper faults have only had minor offsets along them.

**Domain III**

Domain III contains several map-scale F2 folds that are close to tight with angular hinges, long planar limbs, and plunge gently to moderately towards the north and south. They are overturned with axial planes dipping towards the west at moderate angles, giving them an eastern vergence direction (Figure 4.22b). From west to east there are five map-scale faults in Domain III including the boundary faults, and a number of outcrop-scale faults. The western boundary fault occurs as a zone of strongly cleaved sandstone and mudstone with the east facing recumbent F2 fold in Domain II faulted against the Adaminaby Group. Further east, several faults are recognised on the basis of stratigraphic omissions with Bumballa Formation faulted against the younger Bald Ridge Formation where the Warbisco Shale has been cut out. Also younger but overturned Bald Ridge Formation is thrust over older but also overturned Adaminaby Group. Slickenlines found in this fault zone indicate late dextral normal oblique-slip motion that presumably reflects late motion along the earlier formed west-dipping thrust fault. Small thrust faults have offset sandstone beds in the Adaminaby Group on the footwall side of the fault (Figure 4.19b).

At GR747143 6216917 (Fullerton 1:25 000 topographic sheet, 8829-IV-S) in the Adaminaby Group of Domain III a series of recumbent folds and associated bedding-parallel contraction faults occur (Figure 4.26a, b). Several small thrust planes were also observed at GR747203 6216972 (Fullerton 1:25 000 topographic sheet, 8829-IV-S, Figure 4.26c, d). These faults are parallel to bedding in the
Figure 4.23 (a) Photograph and (b) sketch showing east west section through Adaminaby Group along Silent Creek in Domain VII. Visible are downward facing isoclinal F2 folds commonly with faulted hinges. On the western end (right hand side) the fold is very open, and this may be an F3 fold.
Axial plane = 170/80/W
SO = 150/78/SW

SO = 010/85/E
S2 = 105/80/S

> = younging direction
hanging walls and all juxtapose downward facing beds of the Adaminaby Group. Further north in Silent Creek other small thrust faults of similar orientation were also observed (e.g. GR749410 6218499, Fullerton 1:25 000 topographic sheet, 8829-IV-S).

The fault at the eastern boundary of Domain III occurs in Oakey Creek at GR747694 6216516 (Fullerton 1:25 000 topographic sheet, 8829-IV-S) and is an abrupt change from strongly cleaved siltstone of the Bumballa Formation to strongly cleaved and jointed thick-bedded turbidites of the Adaminaby Group. The fault is steeply dipping and S2 cleavage either side of the fault has been rotated clockwise into the fault, indicating probable dextral strike-slip motion. This is presumably a late structure that has developed along an earlier thrust fault with Bumballa Formation thrust over the Adaminaby Group (Figure 4.22b).

DOMAIN IV
In Domain IV the Adaminaby Group dips and youngs to the east (Figure 4.22b).

DOMAIN V
Domain V is dominated by one large F2 anticline overturned to the east with faulted limbs (Figure 4.22a). West of this structure is a smaller F2 anticline cored by Bumballa Formation. The eastern boundary of the domain is a fault zone containing jointed and folded sandstone beds and pods of strongly cleaved Numeralla Chert that are surrounded by strongly cleaved and iron-stained shaly mudstone with polished surfaces (Figure 4.23c). Numerous quartz veins are also present. This and other north-trending faults in Domain V are major thrust faults that developed during the DS2 deformation.

The southern boundary of Domain V is a southeast-trending fault that juxtaposes Bumballa Formation against Adaminaby Group (e.g. GR747694 6217957, Fullerton 1:25 000 topographic sheet, 8829-IV-S). S2 cleavage is locally more intense near this structure. This fault is thought to have postdated the
Figure 4.24 The map shows the location of F3 folds in the Silent-Oakey Creeks area.
Figure 4.25 (a) Photograph and (b) sketch of Bald Ridge Formation beds offset on a small thrust fault beside the fault on the boundary of Domains III and VI in Oakey Creek. The face of the compass is 8 cm wide. (c) Photograph and (d) faulted F3 fold in Bald Ridge Formation, Oakey Creek. Hammer used for scale is 30 cm long. (e) Photograph and (f) sketch showing open, north-south oriented F3 fold. Note offsets of quartz veins indicative of flexural slip occurring along bedding planes. Hammer is 30 cm long.
Figure 4.26 (a) Photograph and (b) sketch showing recumbent F2 folds thrust on top of one another in Domain III. Transport direction is southwest over northeast. (c) Photograph and (d) sketch showing imbricate thrust faults in Adaminaby Group at GR747203 6216972 (Fullerton 1:25 000 topographic sheet, 8829-IV-S).
north-trending DS2 faults and is associated with DS3.

**Domain VI**
Domain VI is dominated by a west-dipping and east-verging section of Bald Ridge Formation that is faulted to the east against a west-dipping and overturned section of Adaminaby Group and Bumballa Formation (Figure 4.22a). The eastern boundary fault is parallel to bedding and S2. The fault is approximately 1-2 m wide, dips to the west at angles varying between 70° and 35°, and is composed of angular sandstone clasts up to 5 cm across in a matrix of cleaved mudstone.

**Domain VII**
In Domain VII, map-scale F2 folds are northerly-trending, steeply inclined, close to open and downward facing. These F2 folds plunge gently towards the north and south (Figure 4.20u, Enclosure 12). The fault at the boundary of Domains VII and VIII is marked by cleaved carbonaceous black shale. Local steeply dipping faults in this zone have offsets of a few metres and gently pitching slickenlines that indicate sinistral movement with a reverse component.

The occurrence of an early fault is known from the presence of a fault breccia in Domain VII at GR748767 6218435 (Fullerton 1:25 000 topographic sheet, 8829-IV-S). The breccia is composed of angular and broken quartz-sandstone clasts with jigsaw fits ranging from 0.5 to 5 cm in size, all contained within a fine-grained matrix. The breccia is approximately 1 m wide and trends southeast, parallel to bedding, but is transected by the S2 cleavage, which trends south-southwest and post-dates breccia formation.

**Domain VIII**
This domain also consists of several downward facing steeply inclined F2 folds (Figure 4.22a). Some outcrop-scale southeast-trending faults occur in Domain VIII; they are steeply dipping towards the southwest and slickenlines indicate reverse movement. Associated small-scale folds affect S2
indicating that these faults post-date F2 folds.

4.5 STRUCTURAL ANALYSIS OF THE GOLSPIE AREA
Middle to Late Ordovician and ?Early Silurian rocks outcrop in the Golspie area and are overlain by hill capping Tertiary basalts. Three deformation events are recognised; the second (DG2) was the most intense. The Golspie area has been divided into three domains based on variations in structural trends (Figure 4.27). Domain I is dominated by the Middle to Upper Ordovician Adaminaby and Bendoc Groups, which form northwest to north-trending structures (Enclosures 6A and 13A). Domain II is bounded to the east and west by faults and is composed of the Upper Ordovician Bendoc Group and ?Lower Silurian Bald Ridge Formation, which form a northeast trending antiformal syncline (Enclosures 6A and 13A). Domain III is mainly composed of ?Lower Silurian Bald Ridge Formation, with minor exposure of Ordovician units, with north to northeast trends (Enclosures 6B and 13B).

4.5.1 Deformation DG1
Deformation DG1 formed map-scale F1 folds in Domain III (Enclosure 6B, cross-sections O-O' and P-P', Figure 4.28). F1 map-scale folds are tight with angular to rounded hinges and planar limbs. They lack any S1 cleavage and have upright to steeply north-dipping axial planes that trend approximately east-west, but have been folded around younger north trending F2 folds. S2 cleavage transects the axial traces of the F1 map-scale folds (Enclosure 6B, 13B). No-outcrop-scale F1 folds were found in the Golspie area.

Although F1 folds are not recognised in Domains I and II their presence is supported by complicated bedding orientations. Stereographic plots show a wider scatter in steep bedding compared to the S2 cleavage orientations, which are more closely clustered (Figure 4.29a, b, d, e). Widespread F2 fold plunge variation and downward facing F2 folds also supports the widespread existence of pre-existing folds (Enclosures 13A, B, Figure 4.29c, f, i, l).

4.5.2 Deformation DG2
Deformation DG2 is characterised by widespread, approximately north trending map-scale and
outcrop-scale F2 folds and associated faults. These structures affect all of the Ordovician and Early Silurian rocks in all three domains.

**OUTCROP-SCALE F2 FOLDS**

F2 outcrop-scale folds were observed in Domains I and II. F2 outcrop-scale folds are mainly close to open with rounded to sub-angular hinges with planar limbs (Figure 4.30). They have weak axial planar S2 cleavage (Figure 4.31). F2 fold orientations are variable due to F3 folding (Figure 4.30b, c). Axial planes range from steeply inclined to recumbent (Figure 4.30c) with moderate to steep plunges (Figure 4.30b). Upright folds tend to have smaller interlimb angles, whereas recumbent folds are generally more open.

**MAP-SCALE F2 FOLDS**

Map-scale F2 folds are shown on Enclosures 6A and 6B, and in the accompanying cross sections (Figures 4.32a-m). They occur as close to open folds with rounded to angular hinges, planar limbs and are associated with the S2 axial planar cleavage. In Domain I axial planes of these folds generally trend southeast, are upright or dip steeply towards the east (Enclosure 6A, Figures 4.32a-e). Fold plunges are at moderate to gentle angles towards both the northwest and south-southeast (Figure 4.29c). On cross-section E-E’ the main structure is a recumbent synformal fold bounded by faults (Figure 4.32e).

Domain II is dominated by a northeast to southeast trending, steeply inclined syncline (Enclosures 6A, 13A, Figure 4.32f-i). Variation in fold tightness occurs along this structure due to later F3 folding. Overall this structure plunges gently towards the north, although locally plunges change from steep to gentle to the north-northeast and south-southwest (Enclosure 13A, Figure 4.29f). The syncline is flanked by anticlines, which have faulted limbs, and towards the south, the syncline is also cut by contraction faults, which have resulted in the removal of the Warbisco Shale (Enclosures 6A, 13A, Figure 4.32).

Map-scale folds in Domain III trend north to northeast (Enclosures 6B, 13B, Figure 4.32j-m). Plunge
Figure 4.27 Locality diagram showing the three structural domains discussed in the Golspie Area.
Figure 4.28 Geological cross-sections. Refer to Enclosure 6B - Golspie Geological Map B for legend and location of sections. (a) Golspie O-O’. (b) Golspie P-P’.
is predominantly towards the north at moderate to steep angles (Figure 4.29l). Wavelengths of the map-scale folds in Domain III are approximately 300 to 400 m. Although smaller parasitic folds commonly occur and individual fold wavelengths vary due to interference patterns between the different fold generations.

S2 CLEAVAGE
In Domains I and II, S2 cleavage occurs as a weak disjunctive cleavage in fine sandstone and mudstone, especially close to the hinge zones of F2 folds (Figure 4.31e). In some of the tighter folds cored by mudstone, a pencil cleavage was observed in the hinge zone (e.g. at GR747479 6209201, Golspie 1:25 000 topographic sheet 8829-III-N). Away from the hinges of the folds and in the hinges of the more open folds cleavage is lacking (Figure 4.30a).

In Domain III, S2 cleavage is stronger than in the other domains. It occurs both in the hinge zones and on the limbs of F2 folds. In sandstone beds it forms a disjunctive cleavage and forms a slaty cleavage in the mudstone interbeds.

S2 cleavage is variable in orientation (Enclosures 6A, 13A, Figure 4.29b). In Domain II, local variations from the dominant north-northeast trend occur. Dips are commonly greater than 50° although some are as low as 24° (Enclosures 6A, 13A, Figure 4.29e). S2 in Domain III is the least affected by deformation DG3. The orientation of S2 is predominantly north to northeast and dips are commonly greater than 65° towards the west (Enclosures 6B, 13B, Figure 4.29h).

4.5.3 Deformation DG3
Deformation event DG3 produced F3 folds, which gently fold or kink F2 folds, S2 cleavage and D2 faults across the whole area (Enclosures 6A, 6B, 13A, 13B). The folds are open to gentle and lack any S3 cleavage (Figure 4.31f, g). In Domain I they trend variably, with axial planes dipping at gentle angles towards the east and steep angles to the west (Figure 4.31i). Hinges plunge up to 20° towards the north and northeast (Figure 4.31h). In Domain II, F3 folds trend towards the east. Axial planes are
Figure 4.29 Equal area, lower hemisphere stereonet plots of poles to bedding (S0), poles to S2 cleavage and S0/S2 intersection lineation ($L^0_2$) data from the Golspie area for each domain and for the total area.
Figure 4.30 (a) Annotated photograph showing downward facing, southeast plunging F2 fold, Monkey Creek (GR748012 6205194; Golspie 1:25 000 topographic sheet 8829-III-N). Arrow indicates hinge. Red circles outline load castes on base of bed and the hammer for scale (blue circle) is 33 cm in length. (b) Stereonet plot of F2 folds in Domains I and II. Domain I shown in blue, Domain II shown in red. Separate girdles are calculated for data from each domain. (c) Stereonet plot of axial plane orientations of F2 folds in Domains I and II. Domain I shown in blue, Domain II shown in red. Separate girdles are calculated for data from each domain. (d) Photograph and (e) sketch of an open recumbent F2 fold in undifferentiated Adaminaby Group turbidites, Monkey Creek, GR748548 6206180 (Golspie 1:25 000 topographic sheet 8829-III-N). Width of the base of the backpack is 30 cm.
Figure 4.31 (a) Photograph and (b) sketch of recumbent F2 folds in thin-bedded turbidites of the Bumballa Formation, Monkey Creek, GR748512 6206313 (Golspie 1:25 000 topographic sheet 8829-III-N). 33 cm long geological hammer is in the centre of the photograph and circled in blue on the sketch. (c) Photograph and (d) sketch of moderately plunging upright F2 folds outcropping in Domain III above Burra Burra Creek (GR751047 6205152, Golspie 1:25 000 topographic sheet 8829-III-N). Width of view is approximately 10 m. (e) Disjunctive S2 cleavage in Bumballa Formation near the hinge of an F2 fold, Monkey Creek (GR748014 6205232, Golspie 1:25 000 topographic sheet 8829-III-N). Note the cleavage is perpendicular to bedding in sandstone and refracted closer to bedding orientation in mudstone (left hand side of photograph). Width of view is approximately 8 m. (f) Photograph and (g) sketch showing southeast dipping, gentle F3 folds in thin-bedded turbidites of the Bumballa Formation above Burra Burra Creek (GR747992 6206595, Golspie 1:25 000 topographic sheet 8829-III-N) (h) Stereonet plot of hinge orientations of F3 folds in Domains I and II. Data from Domain I is shown in blue, and Domain II data is shown in red. Separate girdles are calculated for data from each domain. (i) Stereonet plot of axial plane orientations of F3 folds in Domains I and II. Data from Domain I is shown in blue, and Domain II data is shown in red. Separate girdles are calculated for data from each domain.
Heights are in Metres above sea level

Figure 4.32c, d, e.
Heights are in Metres above sea level
Heights are in Metres above sea level
generally steep (>75°) and towards the south (Figure 4.31i). The hinges of F3 folds in Domain II also plunge at angles of between 40° and 60° (Figure 4.31h).

4.5.4 Faults
Most of the faults mapped in the area are based on stratigraphic evidence and are in general poorly exposed. In Domain I both map- and outcrop-scale faults trend north to northwest (Enclosures 6A, 6B, Figures 4.32a, d, e, h, i). Faults are locally marked by fault breccia (e.g. GR748887 6205798, Golspie 1:25 000 topographic sheet 8829-III-N). The fault between Domains I and II strikes northeast (Enclosures 6A, 13A). The dip on this fault is unknown. One possibility is that this fault dips steeply to the southeast with Domain I rocks in the hanging wall (Figure 4.32g, h). The eastern boundary of Domain II is also bounded by a northeast trending fault (Enclosures 6B, 13B). This fault also strikes approximately parallel to a syncline in Domain III, but cuts obliquely across the structures in Domain II. It is interpreted as a steeply dipping reverse fault placing older Adaminaby Group rocks occurring in Domain III over younger units in Domain II.

4.6 STRUCTURAL ANALYSIS OF THE CROOKWELL RIVER AREA

4.6.1 Introduction
The Crookwell River area encompasses an 8 km east-west stretch of the Crookwell River, east of foliated plutons of the Wyangala Batholith. The whole area has been affected by three deformations, herein termed DC1 DC2 and DC3 (Deformation – Crookwell 1, 2 and 3). Structurally, the area is been divided into two domains (Domain I and II) that have distinctive patterns in bedding (S0), S2 cleavage and F2 folds (Figures 4.33, 4.34a, b, d, e). These domains are separated by an inferred north-northeast trending fault that juxtaposes Middle Ordovician Adaminaby Group to the west, against Upper Ordovician Bumballa Formation in the east (Figure 4.33).

4.6.2 Deformation DC1
Deformation DC1 is indicated by early S1 foliation (S1) and a map-scale F1 fold determined from bedding and S1 relationships in Domain I (Enclosure 7; Figure 4.33). The S1 foliation is sparsely
Figure 4.33 Location of the structural features and two structural domains in the Crookwell River area, separated by an inferred north-northeast trending fault.
preserved along the Crookwell River. It occurs in contact metamorphosed quartz-sandstone of the Adaminaby Group and Bumballa Formation, located up to 1000 m east of the Wyangala Batholith and also near foliated intrusive quartz/feldspar porphyries in the eastern part of the area (Enclosures 7, 14). F2 folds have folded S1 (Enclosures 7, 14, Figure 4.34c). In Domain I, S1 is a disjunctive cleavage, but in some places it occurs as a closely spaced jointed surface (Figure 4.35a, b, c). This surface has subsequently been recrystallised during contact metamorphism.

4.6.3 Deformation DC2
This deformation resulted in east-west shortening by north-south trending folds and faults.

Outcrop-scale F2 folds
Outcrop-scale F2 fold hinges are rarely exposed in the Crookwell River area. In Domain I, only one fold occurs (Figure 4.35d, e, f). It is a downward facing close fold, that plunges steeply towards the south. Outcrop-scale folds were also recorded from Domain II, where they formed tight to close folds with upright axial planes and angular hinges. The hinges plunge moderately towards the north and south (Figure 4.36a, b, c, d). In Domain I, bedding/S2 cleavage intersection lineations (L\(^0\)) indicate moderate to steep plunges ranging from 25° to 77° towards both the northeast and southwest (Enclosure 14, Figure 4.34e). Whereas in Domain II, bedding/S2 cleavage intersection lineations indicate fold plunges are generally at gentle to moderate angles towards the north-northeast or south-southwest (Enclosure 14, Figure 4.34e).

Map-scale F2 folds
Map-scale F2 folds occur in both domains (Enclosures 7, 14, Figure 4.37a, b). F2 folds are tight to close, angular folds with planar limbs and steeply inclined axial planes, dipping mainly to the west (Enclosures 7, 14, Figure 4.37a). Several orders of folds are developed with the longest having a half wavelength of ~500 m. S2 cleavage is axial planar to the F2 folds. In Domain I, F2 folds trend northeast, whereas in Domain II, they trend north to north-northeast with some hinges cored by the more ductile Upper Ordovician Bumballa Formation or Warbisco Shale, (Enclosure 7, Figure 4.37b).
Figure 4.34 Equal area, lower hemisphere stereographic plots of data from the Crookwell area. (a) Poles to bedding (S0), those from Domain I are shown in blue and those from Domain II in red. The dashed blue girdle is the mean S0 value for Domain I, and the dashed red girdle is the main S0 value in Domain II. (b) Contour plot of poles to bedding for all the Crookwell River area. (c) Poles to S1 are shown as blue diamonds for Domain I and red diamonds for Domain II. Pink crosses are bedding/S1 intersection lineations (L_0^1), and green dots are S1/S2 intersection lineations (L_1^2). The girdle of best fit is calculated (B - axis = 345/6) from only the poles to S1. (d) Poles to S2; blue dots are values from Domain I and red dots values from Domain II. Blue girdle is mean S2 for Domain I and red girdle is mean S2 for Domain II. (e) F2 fold orientations and S0/S2 intersection lineations (L_0^3). Blue symbols are data from Domain I and red symbols signify data from Domain II. Triangles are F2 fold hinges, squares are axial plane orientations and crosses are S0/S2 intersection lineations. (f) Black diamonds are S3 data, and the girdle is the best fit based on S3 orientations only. Pink crosses are S2/S3 intersection lineations (L_3^2).
S0

DI N = 16
DII N = 32

S1, $L_1^0$, and $L_2^1$

S2

DI N = 20
DII N = 75

F2 and $L_2^0$

S3 and $L_3^2$
Figure 4.35 Diagrams showing various views on a downward facing F2 fold in quartz sandstone of the Adaminaby Group, Crookwell River, GR718518 6189996 (Crookwell 1:50 000 topographic map sheet, 8729-II and III). (a) Photograph showing poorly preserved S1 crossed by S2. Neck of hammer is 11 cm long. (b) Photograph and (c) drawing of photograph show the disruption of S1 by S2. The head of the geological hammer is 20 cm long and 2 cm wide. (d) Diagram shows the orientation of bedding (S0), S1 and S2 with respect to the orientation of the fold. Legend applicable to diagram is below (e) photograph, and (f) drawing of fold profile. Position of photograph Figure 4.38b is also shown.
S2 CLEAVAGE
S2 cleavage is well developed in all the units of the Crookwell River area including quartz-feldspar porphyries that are probably related to the Wyangala Batholith (Figure 4.35a, b, c). It predominantly trends north-northeast and dips steeply towards the west (Figure 4.34d). In the Adaminaby Group quartz sandstone the S2 cleavage is a disjunctive cleavage with spacing of between 0.2 and 1.0 cm parallel to the limbs of outcrop-scale folds. In the Bald Ridge Formation, S2 cleavage is rare. S2 has developed as a slaty cleavage in the fine-grained Bumballa Formation and Warbisco Shale, where it is predominantly bedding parallel. S2 is also bedding parallel in the ?Upper Silurian limestone. In granite of the Wyangala Batholith, S2 forms a gneissic anastomosing foliation around elongated quartz phenocrysts.

The orientation of Deformation DC2 structures vary between Domains I and II. The S2 cleavage and F2 map-scale folds trend and plunge predominantly towards the northeast in Domain I and towards the north or south in Domain II. Deformation DC2 has affected all the Ordovician and Silurian metasedimentary units in the Crookwell River area, as well as the Siluro-Devonian Wyangala Batholith.

4.6.4 Deformation DC3
This deformation is weakly developed. S3 crenulation cleavage occurs along the axial planes of gentle folds in S2. S3 is widely variable in orientation (Figure 4.34f).

4.6.5 Faults
Faults are inferred on the basis of stratigraphic omissions (e.g. the Warbisco Shale in Domain II) and anomalous younging relationships with older units younging away from younger units. For example, the fault at the boundary of the two domains has the Middle Ordovician Adaminaby Group to the west that youngs away from the younger Bumballa Formation east of the fault (Enclosures 7, 14). Faults are associated with quartz vein arrays and enhanced S2 cleavage development (Figure 4.37a;
Figure 4.36 (a) Photograph and (b) drawing of F2 chevron folds in Upper Ordovician Warbisco Shale. Photograph taken looking north, with the small folds plunging moderately towards the north. Knife is 11 cm long. (c) Photograph and (d) drawing of chevron folds in Late Ordovician Bumballa Formation. The outcrop is not in situ, therefore no orientations were measured for these folds. Hammer is 33 cm in length.
Enclosure 7). A poorly exposed shear zone has been mapped along the contact between the Wyangala Batholith and the Adaminaby Group on the western edge of the Crookwell River area (Enclosures 7, 14). The granite on the western side becomes more intensely foliated as the contact is approached.

4.7 CONCLUDING DISCUSSION
Three deformation events have been recognised from each of the study areas within the Oberon-Taralga region. These deformations however, cannot be correlated on a one-to-one basis across the Oberon-Taralga region. This is because regional constraints indicate that the major deformation in the north is of probable Early Carboniferous age (Fowler 1989), whereas in the south it is more likely of pre-Late Devonian age (Abell 1991; Powell & Fergusson 1979).

4.7.1 Early Deformation D1
An early deformation is present in all the study areas and is indicated by recognition of F1 folds, variable bedding orientations, variable plunges of F2 folds and the occurrence of downward facing F2 folds. Most of the early folds are devoid of S1 cleavage and have variable plunges predominantly towards the northeast and northwest. These structures post-date deposition of the ?Lower Silurian Bald Ridge Formation.

An early deformation event (D1) was reported from the Rockley area, immediately west of the Oberon area and northeast of the Thompson Creek area (Fowler 1989). In the Rockley area D1 produced mapscale F1 folds with half wavelengths of 1-2 km that affected all pre-Middle Devonian units. F1 folds include the Brownlea Anticline, Native Dog and Rockley Synclines. As for most of the study areas F1 folds in the Rockley area contain no cleavage and are overprinted by F2 north-trending folds with associated axial planar S2 cleavage. In contrast to the Oberon and Thompson Creek areas, no outcrop-scale F1 folds were observed in the Rockley area (Fowler 1989).
Figure 4.37 Cross Sections from the Crookwell River area. Refer to Enclosure 7 for legend and location of sections (a) Crookwell A-A’ (b) Crookwell B-B’.

HS = VS

Heights are in Metres above sea level.
Fowler (1989) explained the erratic fold trend pattern in the Rockley area by complex buckling caused by rapid thickness variations in the Rockley Volcanics. Due to the thin-bedded nature of the rocks in the Oberon, Thompson Creek and Golspie areas, complex buckling may explain the presence of outcrop-scale F1 folds in these areas. In contrast, the massive nature of the succession in the Rockley area only allowed map-scale F1 folds to develop.

Pre-cleavage F1 folds were also described from south of Burrara (Figure 4.1), where they affected all rock units of Ordovician to Late Silurian age (Dunnet 1961), and in the Jaunter area, where only the ?Lower Silurian Bald Ridge Formation and underlying units were affected (Scott 1991). In both of these areas, no unconformity was recognised between the Early and Late Silurian units, and the origin of the early folding was tentatively related to near surface gravity slides.

Early isoclinal, recumbent F1 folds, devoid of cleavage and downward facing F2 folds have also been reported from east of Queanbeyan to the south (Figure 1.3; Stauffer & Rickard 1966). These folds affect Ordovician but not Late Silurian rocks, trend northerly, at a slight angle to the north-northeast trending F2 folds and associated S2 axial planar slaty cleavage, and are characterised by very large wavelengths (kilometres). Further investigation has resulted in the recognition of an S1 spaced, stripy cleavage in rocks of higher metamorphic grades, which was related to development of the metamorphic complexes in the area (Carson & Rickard 1998). Therefore, the F1 folding east of Queanbeyan has been related to metamorphism at depth during the latest Ordovician - Early Silurian.

Features characteristic of gravity sliding or slumping of units include: (1) intraformational minor folds; (2) tight to isoclinal folds with no related cleavage; (3) wide dispersion of fold axes unrelated to polyphase deformation; (4) close association of contractional and extensional structures; and (5) association with olistostromes and debris flows (Webb & Cooper 1988). Early folds reported from the study areas herein have the first two of these features, but the lack of olistostromes, debris flows or extensional features suggests that these structures probably did not form in response to gravitational
The lack of cleavage in these folds is consistent with them having formed at shallow crustal levels and like the early structures observed in the low grade metamorphic rocks of the Queanbeyan area, they are thought to be related to deformation and/or metamorphism at depth.

The Do1 structures in the Bathurst 1:250 000 sheet area were interpreted as reflecting Middle Devonian thrusting (Glen 1998). Glen regarded the "Oberon Fault" along Hampton Road in the Oberon area as an early east-west trending structure, separating the Upper Ordovician Triangle Formation from the Adaminaby Group and Warbisco Shale package that had been folded by D2 north-south oriented structures (Glen & Wyborn 1997; Glen 1998). This interpretation was based on younging directions in presumed Adaminaby Group rocks on the Hampton Road suggesting that they young away from, rather than into, the presumed younger black shale and volcanics (Raymond et al. 1997; Glen 1998). Thus the "Oberon Fault" was considered to have enabled thrusting of the presumed Adaminaby Group northwards over the Triangle Formation during the D1 deformation (Do1 of Glen 1998).

Detailed mapping reported in this thesis has established that the so-called Adaminaby Group rocks of Glen (1998) are actually the ?Lower Silurian Bald Ridge Formation that are conformable on the underlying Upper Ordovician volcanioclastics of the Triangle Formation (Figure 2.1 – Oberon East). The stratigraphy has been folded into a series of north-south oriented folds, plunging variably towards the north and south and cut by several north-south contraction faults. Early F1 folds in this section are oriented north-northeast to northeast, and no evidence for an early easterly trending "Oberon Fault" has been found. Strain has been accommodated by movement along the black shale contact, but this is not of the orientation proposed for the "Oberon Fault".

The earliest constraint on the timing of deformation in the Oberon-Taralga region is indicated by early Late Silurian angular unconformities in the Oberon and Thompson Creek areas. In the western part of the Oberon area the Upper Silurian Fosters Creek Conglomerate and Bells Creek Volcanics
unconformably overlie the Upper Ordovician Triangle Formation and Rockley Volcanics. A similar unconformity exists in the Thompson Creek area where the Upper Silurian Fosters Creek Conglomerate and Kangaloolah Volcanics unconformably overlie the Lower Silurian Bald Ridge Formation. This unconformity was not recognised by Fowler (1989) in the intervening Rockley area. According to Raymond et al. (1997) the Fosters Creek Conglomerate and underlying unconformity does occur in the Rockley area and provides evidence of uplift and erosion in the region. No structures in the Oberon-Taralga region have been related to this unconformity (Benambran in the sense of VandenBerg et al. 2000). Although it is conceivable that some of the early deformation in the Oberon-Taralga region may be related to this unconformity no direct evidence is available to support this postulate.

4.7.2 Regional Deformation D2
The second deformation is the strongest deformation event recognised in the study areas, where it has affected all rocks of Ordovician to Early Silurian age and Late Silurian rocks in both the Oberon and Thompson Creek areas. This deformation event is characterised by north to northeast trending folds, associated axial planar cleavage (S2) and related faults. It has formed from strong east-west shortening.

In the northern part of the Oberon-Taralga region the timing of the major deformation is constrained by the age of the rocks affected and by the timing of the post-kinematic suite of Carboniferous plutons (Shaw & Flood 1993; Wallace & Stuart-Smith 1998) exposed in the Oberon area. These Carboniferous granites are associated with the Bathurst Batholith and they clearly cut across and postdate the D2 structures (Enclosure 1). The Burraga, Rockley and Native Dog Synclines (Figure 4.1) record continuous sedimentation from the Late Silurian to the Early Devonian (Raymond et al. 1997; Pogson & Watkins 1998). All of these units have been cleaved and folded by deformation D2 which resulted in plunging macroscopic F2 folds (\(\lambda/2 = 0.4-2 \text{ km}\)) with associated parasitic mesoscopic folds and development of the regional slaty cleavage (S2) (Fowler 1989; Pogson &
This fold style and orientation is typical of the D2 structures observed in all five of the study areas herein. For the northern Oberon-Taralga region the major D2 deformation is therefore constrained between the Early Devonian and Early Carboniferous.

The Davies Creek Granite occurs south of the Bathurst Batholith on the Oberon 1:100 000 geological sheet, where it forms two weakly foliated plutons along with early foliated and late massive, cross-cutting rhyolitic dykes (Fowler 1987, Fowler & Lennox 1992). Based on field and petrographic evidence, the Davies Creek Granite is thought to be a late syn-tectonic granite emplaced after the peak of syn-tectonic acid-dyke intrusion and towards the end of the regional cleavage forming event (Fowler 1987; Fowler & Lennox 1992). A syn-kinematic emplacement history is supported by a weak gneissic foliation occurring around the margins of the plutons which is parallel to the regional S2 cleavage, while outcrops occurring greater than 200 m from the pluton margins are massive (Fowler 1987).

The age of the Davies Creek Granite has not been determined; however, based on the relationships outlined above, it has to be between Middle Devonian and Early Carboniferous in age. It has been interpreted on the basis of regional D2 deformation relationships as Early Carboniferous in age (Fowler & Lennox 1992; Fowler 1996). Conversely, on the basis of similarities in foliation and deformation characteristics, Pogson and Watkins (1998) considered it to be a northerly extension of the Early Devonian Wolgorong Batholith (see below). However, the Davies Creek Granite is less deformed than the Wolgorong Batholith, which has a mylonitic foliation (Vernon et al. 1983), while the Davies Creek Granite is only weakly foliated to massive (Fowler & Lennox 1992).

North of the Bathurst Batholith, the Upper Devonian Lambie Group occurs in the cores of regional folds that formed in the major regional deformation of the Hill End, Capertee and Molong Zones (Powell et al. 1977). This regional deformation has been attributed to the Early Carboniferous Kanimblan Orogeny. South of the Bathurst Granite in the northern Oberon-Taralga region Fowler
Brownlow (1989, 1996) has attributed the regional deformation (including the D2 structures of the Oberon and Thompson Creek areas) to the Early Carboniferous deformation. New radiometric dates from north of the Bathurst Batholith indicate that the Middle Devonian metamorphic (thermal) maximum outlasted penetrative deformation, and has been estimated at 370 Ma (latest Givetian – timescale of Young & Laurie 1996; mid Fammennian – timescale of Tucker et al. 1998; Packham 1999). Thus the timing of the regional deformation of the Hill End Zone has been considered to be mid to Late Devonian and older than that developed in the adjacent Molong and Capertee Zones where the deformed Late Devonian rocks are preserved. Late Devonian rocks are not preserved within the northern Oberon-Taralga region. Until a more reliable age can be established for the Davies Creek Granite the D2 event in the Oberon and Thompson Creek areas is considered to have occurred between the Middle Devonian and Early Carboniferous. It may even reflect a combination of structures formed during the Tabberabberan and Kanimblan Orogenies that, without the further control gained by mapping the structures in the younger units are indistinguishable.

In the southern part of the Oberon-Taralga region the timing of the major deformation is even less well constrained. The increase in the angle of unconformity between rocks of the Upper Devonian Lambie Group and underlying Early Devonian units from the northeastern Lachlan Fold Belt east of Mudgee, to east of Taralga was detailed by Powell and Edgecombe (1978) and Powell and Fergusson (1979). The implication of this work is that the intensity of the pre-Late Devonian deformation increased southwards.

To south around Canberra and Braidwood, significant regional deformation is considered to be Middle Devonian, coincident with the main M2 metamorphic phase and with K-Ar biotite ages of 370-380 Ma east of Canberra (Abell 1991; Foster et al. 1999). West of the southern Oberon-Taralga region deformation along shear zones in the Early Silurian Wyangala Batholith has Ar-Ar biotite ages of $384 \pm 4$ Ma and $379 \pm 2$ Ma indicating a minimum Middle Devonian age for shearing along the eastern margin of this batholith (Lennox et al. 1998; Foster et al. 1999).
Situated to the east of the Crookwell River area and west of the Golspie and Silent-Oakey Creeks areas, the Wolongorong Batholith occurs as a north-south trending S-type granite with a strong north-trending foliation intruding Ordovician to Early Silurian rocks (Vernon et al. 1983). A Rb-Sr isochron defined what was regarded as an emplacement age for the Wologorong Batholith at 405 ± 11 Ma (Shaw et al. 1982). Five Rb-Sr biotite/whole rock ages ranged from 381 to 339 Ma, with the older ages (381, 379 Ma – Early Devonian) interpreted as minimum emplacement ages, and the younger ages (348, 339, 344 Ma - Carboniferous) considered to date a major period of regional metamorphism (Shaw et al. 1982). Based on these dates and the revised Devonian timescale of Tucker et al. (1998), the Wologorong Batholith was emplaced during the Early to Middle Devonian and underwent deformation and regional metamorphism in the Early Carboniferous.

In conclusion, constraints from stratigraphic and igneous rocks in the southern Oberon-Taralga region are too poor to provide tight constraints on the timing of regional deformation (D2 structures). The second deformation in the Crookwell River area is probably Middle Devonian based on Ar-Ar ages of deformation in the Wyangala Batholith, but potentially could be as old as early to mid Silurian (Collins & Hobbs 2001). In the Silent-Oakey Creeks and Golspie areas the timing of the D2 structures are poorly constrained. Biotite ages from the Wologorong Batholith imply a possible Early Carboniferous timing of regional deformation whereas radiometric ages around Canberra and relationships across the Lambian unconformity imply that significant Middle Devonian or older deformation may have affected these areas.

**4.7.3 Late Deformation D3**

The D3 deformation event was caused by north-south shortening and resulted in gentle folds in D1 and D2 structures in the Oberon, Thompson Creek, Golspie and Crookwell River areas. F3 folding is most intense in the Oberon area along Chain-of-Ponds Road where S2 cleavage has been rotated to east-west and southeast-northwest orientations compared to a predominantly north-south orientation.
elsewhere (Enclosures 8, 9, 10; Figure 4.5d, e, f; Figure 4.4d, e, f, and enclosures 12, 13 & 14). In the
Oberon area the Carboniferous Oberon, Sloggetts, Black Springs, Rossdhu and Duckmaloi Granites
(Shaw & Flood 1993; Wallace & Stuart-Smith 1998) appear to post-date this late deformation
(Enclosure 1) and indicate an earliest Carboniferous age for this deformation. Carboniferous plutons
do not occur in the other study areas and the timing of the F3 folds is poorly constrained.

Along the south coast of New South Wales late stage angular fold patterns about steep axes that
formed from north-south shortening have been described as “megakinks” (Powell et al. 1985). The
timing of “megakinking” was constrained to intra-Carboniferous based on megakink deformation in
the Early Carboniferous Budawang Synclinorium, with the upper age limit shown by the absence of
north-south shortening in the basal rocks of the Sydney Basin. The presence of the larger domains of
“megakinks” were extrapolated from the south coast to the entire Lachlan Fold Belt, where they were
related to major northwest-southeast and northeast-southwest trending lineaments. Some of these
lineaments had a demonstrated pre-Carboniferous history (Scheibner & Stevens 1974), therefore the
megakinks were thought to have nucleated on inhomogeneities provided by the lineaments (Powell
1984). One of the megakinks was noted as coinciding with the southernmost portion of the Lachlan
River Lineament also termed the Lachlan Transverse Zone occurring north of the Oberon area
(Powell 1984, Scheibner & Stevens 1974; Glen & Wyborn 1997). In the Oberon-Taralga region the
late deformation has produced both kinks and folds but at the map-scale angular to rounded folds are
recognised with wavelengths typically less than 2-3 km rather than megakinks in domains of
20 - 30 km.
CHAPTER 5
PETROGRAPHY, GEOCHEMISTRY AND TECTONIC AFFINITY OF MAFIC VOLCANIC AND VOLCANICLASTIC ROCKS OF THE OBERON-ROCKLEY DISTRICT

5.1 INTRODUCTION
The tectonic affinities based on the geochemistry of Ordovician mafic-intermediate volcanic and volcaniclastic rocks of the Molong Volcanic Province provide constraints on the tectonic development of the Lachlan Fold Belt. The Molong Volcanic Province has been interpreted as a north-south trending mafic volcanic island arc with an oceanic basement and separated from the Gondwanan margin (Cas 1983; Powell 1983; Packham 1987; Glen et al. 1998). Alternatively, Wyborn (1992a) emphasised the shoshonitic composition of these rocks and related the source of the mafic magmatism to melting of a metasomatized lithosphere, triggered by overturning within an intraplate tectonic setting and underlain by continental crust.

This chapter presents petrographic and whole-rock geochemical data obtained from mafic volcanic rocks of the Oberon-Rockley district located at the southern end of the Rockley-Gulgong Volcanic Belt (Figure 5.1). The aims of this chapter are to petrographically describe and geochemically characterise the Ordovician Molong Volcanic Province rocks of the Oberon-Rockley district (Rockley Volcanics, Triangle Formation) in order to resolve the tectonic affinity of this succession. Nd isotopic data have also been obtained to provide some constraints on the nature of the basement to the volcanic succession of the study area. These data are also compared with published geochemical data from the northern Rockley-Gulgong Volcanic Belt and the adjacent Molong Volcanic Belt.

5.2 SAMPLE COLLECTION, PREPARATION AND ANALYTICAL METHODS
In the Rockley area only lavas/shallow intrusions were sampled for geochemical analysis. Within the Oberon area, both the Rockley Volcanics and Triangle Formation comprise volcaniclastic deposits
Figure 5.1 Location of the four Ordovician mafic volcanic belts in the northeast Lachlan Fold Belt and the transverse zones of Glen et al. (1998) which cut obliquely across the volcanic belts. Dark grey stiple = outcropping Ordovician mafic volcanics; light grey stiple = Ordovician mafic volcanics delineated from aeromagnetic data; red outlines = Ordovician mafic volcanics delineated from gravity data. NTZ = Nyngan transverse zone; HRTZ = Hunter River transverse zone; LTZ = Lachlan transverse zone. The location of the Rockley/Oberon area covered in Figure 5.2 is also shown. Diagram modified from Wyborn (1992) and Glen et al. (1998).
Samples of the Rockley Volcanics from the Oberon area were collected from individual clasts within breccias and are therefore representative of primary volcanic rocks. Samples from the Triangle Formation are all from the basal Bouma A or B layers of turbidites (see Figure 5.2 for localities).

Thin sections were cut from samples in order to determine the extent of alteration. The least altered samples were broken up in a hydraulic press and obvious alteration minerals such as calcite and quartz from veins and amygdales were removed. The freshest material was then separated into two parts, the first powdered in a tungsten carbide mill for X-Ray Fluorescence and radiogenic isotope analysis, and the second sample was powdered in a chrome-steel mill for Instrumental Neutron Activation Analysis. Only analyses with less than 5 wt% loss on ignition (LOI) were considered because, analyses carried out by Müller et al. (1992) showed this limit marked an hiatus between apparently fresh and more weathered or altered samples.

XRF and INAA analyses were carried out at the Department of Geology, The Australian National University (analyst, Bruce Chappell). Isotope analyses were measured at the Centre for Isotope Studies, CSIRO North Ryde (Sydney).

5.3 PETROGRAPHY

5.3.1 Rockley Volcanics

Within the Oberon area, the Rockley Volcanics comprise basalt breccia, with associated volcaniclastic sandstone and black mudstone. Basaltic clasts contain relict phenocrysts set in a fine-grained pilotaxitic groundmass. Pyroxene and plagioclase are the dominant phenocrysts. Euhedral to subhedral clinopyroxene phenocrysts typically compose between 15% and 20% of the rock. Fresh clinopyroxene is twinned, has occasional oscillatory zoning and is colourless to weakly pleochroic. They range up to 7 mm across and show rare glomeroporphyritic textures (Figure 5.3a, b, c). Alteration to actinolite, chlorite, epidote, serpentine and calcite is common along fractures and at
Figure 5.2 Inset (a) shows the locality of the Oberon and Rockley map areas. The main map shows the geology and main structural features of the Oberon area, with the localities of geochemical samples referred to from this area. Inset (b) shows the geology, main structural features and sample localities from the Rockley area (after Fowler 1987).
crystal rims (Figure 5.3b, d). In places the entire phenocryst is completely altered to actinolite, calcite and chlorite. Euhedral plagioclase laths are less abundant, comprising up to 10% of the rock, and are altered to albite (Figure 5.3a, b). They range up to 5 mm in length and commonly show chlorite alteration along rims. Titaniferous magnetite and pseudomorphs after olivine are minor phenocryst phases. Amygdalae are present in the samples and consist of radiating chlorite (Figure 5.3a) and less common albite and calcite. The groundmass is predominantly composed of feldspar and pyroxene microlites together with interstitial chlorite and epidote (Figure 5.3a, d). Where the groundmass is altered, chlorite and calcite are ubiquitous. Whole-rock XRD analyses of the Rockley Volcanics samples indicate that serpentine, predominantly lizardite with minor antigorite, is present.

The Rockley Volcanics of the Rockley area consist of ultramafic and mafic rocks, together with felsic volcanics, all of which have been altered to greenschist grade (Fowler 1987). The ultramafic rocks occur as breccias and tuffs that are characterised by an assemblage of tremolite + chlorite + serpentine + talc. They have densely packed well-preserved pseudomorphs after pyroxene and olivine phenocrysts and a notable absence of feldspar (Fowler 1987). The mafic rocks include mainly intrusive rocks with a mineral assemblage of actinolite ± augite + clinozoisite/epidote + albite ± chlorite ± biotite ± serpentine ± talc. The felsic volcanics have rare feldspar phenocrysts and lack actinolite. They are characterised by the mineral assemblage albite + chlorite + talc + quartz + calcite. The samples analysed herein are all mafic rocks.

5.3.2 Triangle Formation
Volcaniclastic rocks from the Gidyen Volcaniclastic Member and undifferentiated volcaniclastic rocks within the Triangle Formation are lithologically similar. Volcanogenic derived rock and crystal fragments dominate the sandstone and fine conglomerate of the Triangle Formation, with rare quartz and mudstone clasts (Figure 5.3e, f). Rock fragments range from <1.0 mm up to 4 cm in length. The larger rock fragments are angular and elongate in one direction (up to 1 cm x 4 cm), while the smaller ones (<2-3 mm) are more rounded. Rock fragments are commonly sub-angular to rounded, dark grey,
porphyritic basalt-andesite clasts with a pilotaxitic groundmass texture (Figure 5.3e). Phenocrysts in these rock fragments include small tabular plagioclase laths. Some rock fragments are sedimentary but are composed entirely of fine-grained mafic volcanic clasts.

Crystal fragments commonly include pyroxene and plagioclase with rare hornblende (Figure 5.3e, f). Pyroxene crystals range up to 8 mm long, but are generally less than 4 mm. Plagioclase commonly occurs as albitised, tabular laths ranging up to 5 mm in length but is more commonly less than 3 mm in length. Primary yellow/green to brown pleochroic hornblende (<4 mm long) is rare, but occurs in many samples. The crystal fragments are predominantly euhedral to subhedral. Rare sub-rounded to rounded polycrystalline quartz clasts are also present in rocks from the eastern part of the Oberon area along Hampton Road.

5.3.3 Alteration
Both the Rockley Volcanics and the Triangle Formation have been altered by regional metamorphism to lower greenschist facies. Common alteration minerals include amphibole (actinolite/tremolite), chlorite and epidote (Figure 5.3e, f, g). The secondary amphibole exhibits a distinctive tremolitic pale green to bright teal-green pleochroic scheme, but retains relict ~90° cleavage angles of pyroxene (Figure 5.3h), and along with quartz and biotite, is especially prevalent in the aureoles of the Carboniferous granites within the Oberon area. Other grains retain relict clinopyroxene cores with amphibole and/or epidote alteration around the rims and along cleavage planes and fractures. Chlorite is abundant and occurs both in the matrix and as a replacement of clinopyroxene and plagioclase phenocrysts. Calcite is present in amygdales and veins as well as being ubiquitous in the matrix.

5.4 GEOCHEMISTRY
The results of whole rock geochemical analyses from the Rockley Volcanics and Triangle Formation are listed in Table 5.1. Interpretation of the geochemical data has been made with caution, because alteration of the rocks during deformation and low-grade metamorphism is known to cause mobilisation of elements. In particular, the low field strength elements (LFSE), including Si, Na, K,
Figure 5.3 (a)–(d) Cross-polarised photomicrographs of the Rockley Volcanics. Field of view in all photomicrographs is 3.58 mm x 2.66 mm. (a) Thin section of the Rockley Volcanics in Brisbane Valley Creek, showing twinned clinopyroxene (C) and plagioclase (P) phenocrysts in a pilotaxitic groundmass (G) composed predominantly of feldspar with minor pyroxene and oxides. Small amygdales (A) filled with secondary chlorite also occur. (b) Thin section of the Rockley Volcanics in Brisbane Valley Creek, showing clinopyroxene with oscillatory zoning (C) and euhedral plagioclase lathes with tapered twins (P). Note calcite and chlorite alteration along rims, fractures and outlining zoning in the clinopyroxene crystals. (c) Thin section of the Rockley Volcanics in Hope Creek showing glomeroporphyritic texture. Note calcite and chlorite alteration along rims and fractures of the crystals. (d) Thin section of the Rockley Volcanics in Hope Creek, showing slightly altered euhedral to subhedral clinopyroxene and small plagioclase phenocrysts in a felted groundmass. The clinopyroxene phenocryst at bottom left (C) exhibits both twinning and oscillatory zoning. Chlorite alteration of the groundmass can also be observed. (e) – (h) Photomicrographs of the Triangle Formation. The field of view is 3.58 mm x 2.66 mm for all of them. (e) and (f) Twinned and altered clinopyroxene (C) and plagioclase (P) crystal fragments and volcanic derived rock fragments (R) comprising the Triangle Formation. Both thin sections are from the Upper Ordovician Triangle Formation occurring along Brisbane Valley Creek. (g) Photomicrograph shows actinolite and chlorite alteration of crystal fragments and rock fragments in the Upper Ordovician Triangle Formation occurring along Hampton Road, east of Oberon. C = pyroxene crystal fragments, Cl = chlorite alteration and Ac = actinolite. (h) Photomicrograph taken in plane polarised light shows bright green pleochroism of tremolite (T) replacing pyroxene crystal fragments, and chlorite alteration of plagioclase crystal fragments (Cl-P) and the matrix (Cl-M).
Table 5.1 Representative analyses of the Rockley Volcanics and Triangle Formation from the Oberon-Rockley districts.

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| Nd 144/143 | 0.5128 | 0.51284 | 0.51278 | 0.51285 | 0.51288 | 0.51275 | 0.512703 |
| Sm 147/143 | 0.1413 | 0.1546 | 0.1693 | 0.1733 | 0.1845 | 0.169 | 0.1482 |
| Sr 450 | 0.70449 | 0.70419 | 0.70437 | 0.70459 | 0.70415 | 0.70439 | 0.70481 |
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<td>K₂O</td>
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<td>Total</td>
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<td>100.68</td>
<td>100.49</td>
<td>100.63</td>
<td>101.61</td>
<td>100.78</td>
<td>100.77</td>
<td>100.63</td>
</tr>
</tbody>
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| Ba        | 1210   | 500    | 450    | 90     | 320    | 360    | 575    | 905    |
| Rb        | 29     | 23     | 46     | 3      | 17     | 18     | 34     | 44     |
| Sr        | 615    | 411    | 595    | 720    | 306    | 760    | 555    | 271    |
| Pb        | 4      | 4      | 4      | 2      | <2     | <2     | 4      | 2      |
| Th        | 1.2    | 0.6    | 0.4    | 0.5    | 0.5    | 1.1    | 2.3    | 0.3    |
| U         | 0.5    | 0.3    | 0.3    | 0.3    | 0.3    | 0.7    | 0.7    | <1.00  |
| Zr        | 42     | 44     | 40     | 34     | 28     | 56     | 124    | 36     |
| Nb        | 4      | 6      | 2      | 6      | 4      | 12     | 20     | <2     |
| Y         | 14     | 20     | 16     | 14     | 13     | 16     | 22     | 13     |
| La        | 7.8    | 11.4   | 7.1    | 11     | 7.3    | 13.6   | 20     | 5.4    |
| Ce        | 16.4   | 20     | 13.4   | 18.6   | 14.4   | 27     | 40.5   | 11.4   |
| Nd        | 7.6    | 10.4   | 7.3    | 11.2   | 7.3    | 13.8   | 18.6   | 5.5    |
| Sm        | 2.25   | 2.85   | 2.1    | 2.7    | 2.15   | 3      | 4.2    | 1.94   |
| Eu        | 0.71   | 0.99   | 0.78   | 1.03   | 0.76   | 1.22   | 1.19   | 0.68   |
| Tb        | 0.34   | 0.44   | 0.36   | 0.39   | 0.34   | 0.42   | 0.57   | 0.26   |
| Ho        | 0.5    | 0.7    | 0.5    | 0.5    | 0.4    | 0.55   | 0.75   | 0.4    |
| Yb        | 1.68   | 2      | 1.73   | 1.53   | 1.51   | 1.7    | 2.2    | 1.41   |
| Lu        | 0.27   | 0.32   | 0.27   | 0.24   | 0.24   | 0.25   | 0.35   | 0.26   |
| Sc        | 42     | 46     | 38     | 37     | 46     | 28     | 42     | 34     |
| V         | 248    | 344    | 246    | 248    | 296    | 226    | 268    | 250    |
| Cr        | 378    | 76     | 154    | 166    | 86     | 22     | 124    | 484    |
| Ni        | 96     | 26     | 34     | 40     | 38     | 10     | 24     | 202    |
| Cu        | 74     | 108    | 52     | 4      | 4      | 14     | 108    | 84     |
| Zn        | 94     | 110    | 82     | 40     | 64     | 32     | 88     | 84     |
| Ga        | 13     | 14     | 14     | 15     | 15     | 16     | 13     | 12     |
| Hf        | 1.3    | 1.4    | 1.2    | 1.1    | 1      | 1.6    | 3.1    | 1.1    |
| Ta        | 0.8    | 1.5    | 0.5    | 1.7    | 0.9    | 3.4    | 5.2    | 0.7    |

| Sr87/86   | 0.70638 | 0.70572 | 0.70744 | 0.706015 | 0.705847 | 0.705176 | 0.70538 | 0.70885 |
| Rb 87/86  | 0.136   | 0.162   | 0.224   | 0.012   | 0.161   | 0.069   | 0.177   | 0.47   |
| Nd 144/143| 0.51275 | 0.51277 | 0.51281 | 0.512692 | 0.512783 | 0.512675 | 0.51267 | 0.51284 |
| Sm 147/143| 0.179   | 0.1657  | 0.1739  | 0.1458  | 0.1781  | 0.1314  | 0.1365  | 0.2133 |

| Sr 450    | 0.70551 | 0.70468 | 0.706   | 0.70594 | 0.70482 | 0.70474 | 0.70422 | 0.70584 |
| Nd 450    | 3.18    | 4.38    | 4.59    | 3.98    | 3.9     | 4.48    | 3.39    | 3.01    |
Figure 5.4 Harker variation diagrams. (a) Ba versus SiO$_2$. (b) K$_2$O versus SiO$_2$ with compositional fields of Peccerillo and Taylor (1976). (c) MgO versus SiO$_2$. 
Ca, Cs, Rb, Ba and Sr are considered particularly mobile (Pearce 1987). In contrast, according to observational and theoretical criteria, the high field strength elements (HFSE – Ti, Y, Zr, Nb, Hf, Ta), transition elements (Sc, V, Cr, Co Ni) and the rare earth elements (REE) are essentially immobile (Pearce 1996). The mobility of the LILE is demonstrated by the scatter of points on the Ba-SiO₂ plot (Figure 5.4a).

5.4.1 Rockley Volcanics
In all, ten samples from the Rockley Volcanics have been analysed; five of these are from the Rockley area and five from the Oberon area (Figure 5.1). The Rockley Volcanics are basalts to basaltic andesites ranging from 45 to 53 wt% SiO₂ (Figure 5.4a, b, c). They show considerable variation in potassium values but are generally high, with K₂O values ranging from 2 to 7 wt%. Samples from the Rockley area are generally more potassium-rich than those from the Oberon area (Figure 5.4b). Al₂O₃ values are variable, ranging from 9.36 to 13.95 wt% in the Rockley samples and 8.56 to 15.97 wt% in those from the Oberon area. MgO content varies from 5.45 to 14.90 wt% (Figure 5.4c), with Cr and Ni contents showing a positive correlation with MgO, with rocks from the Rockley district slightly higher in Cr, Ni and MgO content than those from the Oberon area (Figures 5.5a, b). This linear trend may be explained by olivine and pyroxene fractionation from the melt (Green 1980; Wilson 1993, Table 2.1, pg 17), which is consistent with the phenocrysts observed in these rocks. The high concentrations of MgO (>6.0 wt%) in some basalts is accompanied by high Ni, Cr and Sc contents, up to 340, 850 and 47 ppm respectively, and these are taken to indicate significant accumulation of olivine and pyroxene (Pearce 1996). The plot of Zr against MgO also shows a simple linear trend indicating Zr has been relatively immobile under low-grade regional metamorphic conditions (Figure 5.5c). TiO₂ values occur in a narrow range from 0.4 to 1.1 wt%, consistent with a high-K calc-alkaline to shoshonitic magmatic affinity but not an alkaline affinity which are > 1.3 wt% TiO₂ (Morrison 1980; Figure 5.5d).

Light rare earth elements (LREE) show a slight to moderate fractionation with a slight to moderate depletion evident in the Heavy Rare Earth Elements (HREE) with \((La/Yb)_N = 2.77-4.23\) in samples
Figure 5.5 Variation diagrams. (a) MgO versus Cr. (b) MgO versus Ni. (c) MgO versus Zr. (d) MgO versus TiO₂.
from the Rockley area and \((\text{La/Yb})_N = 4.53-6.13\) in rocks from the Oberon area (Figure 5.6a, b). Most of the samples from both areas are depleted in high field strength elements (HFSE), particularly Nb, Zr, Hf, and Y, relative to N-MORB, with several of the samples recording Nb values below the detection limit and these are plotted at half the detection limit (Figure 5.7a, b). The LILE (Ba, Sr, Rb, K) on both plots show considerable scatter, which is attributed to alteration; however, the Rockley Volcanics are generally enriched in Ba, Sr and Rb in addition to K, but they are not characterised by relative enrichment of Nb and Ti typical of rift-related, continental K-rich magmas (Rogers et al. 1995; Table 5.1). However, they do exhibit positive Ta anomalies. They have low V/Ti ratios, moderate Ti/Zr ratios, low Zr/Y ratios, and high Cr contents consistent with a subduction related assemblage (Shervais 1982; Pearce 1983, 1996; Figures 5.8a, b, 5.9a), and plot in oceanic arc-related fields on the tectonic discrimination diagrams for potassic rocks of Müller et al. (1992) (Figure 5.10a, b).

\(\varepsilon_{\text{Nd}} (450)\) values were calculated for three of the samples from the Rockley area and five from the Oberon area. The \(\varepsilon_{\text{Nd}} (450)\) values are moderately positive relative to CHUR, ranging between +6.43 and +4.29 for the Rockley area and +5.39 and +3.39 for those in the Oberon area, reflecting relatively unevolved mantle-derived magmas (Faure 2001, p.12-17; Rollinson 1993, p.255-256). Initial \(^{87}\text{Sr}/^{86}\text{Sr}\) values at 450 Ma are relatively low and quite variable (0.70415-0.70481), probably reflecting redistribution during interaction with seawater and low-grade metamorphism (Menzies & Seyfried 1979).

5.4.2 Triangle Formation - Gidyen Volcaniclastic Member and Upper Ordovician Volcaniclastics

Six samples have been analysed from the Triangle Formation. Of these, one is from the Middle Ordovician Gidyen Volcaniclastic Member (R15987) and five are from the Upper Ordovician undifferentiated Triangle Formation in the Oberon area (Figure 5.1). Of these five, one sample (R15986) is from Hampton Road. An additional sample was also collected from the Jaunter Volcanics
Figure 5.6 Chondrite normalised Rare Earth Element patterns for: (a) Rockley Volcanics from the Rockley area. (b) Rockley Volcanics from the Oberon area. (c) Triangle Formation from the Oberon area and one sample from the Jaunter Volcanics in the Jaunter area. Normalising values from Boynton (1984). Also shown for comparison are data representative of an average ocean island basalt (average OIB) data from Sun and McDonough (1989), average calc-alkaline island arc basalt (average calc-alkaline IAB), data from McCulloch and Gamble (1991) and average shoshonite (Sample TS37 – Rogers & Setterfield 1994).
Figure 5.7 Rock/MORB normalising patterns for: (a) Rockley Volcanics from the Rockley area, (b) Rockley Volcanics from the Oberon area, (c) Triangle Formation from the Oberon area and one sample from the Jaunter Volcanics in the Jaunter area. N-MORB values from Sun and McDonough (1989). Also shown for comparison are data representative of an average ocean island basalt (average OIB) data from Sun and McDonough (1989), average calc-alkaline island arc basalt (average calc-alkaline IAB), data from McCulloch and Gamble (1991) and average shoshonite (Sample TS37 – Rogers & Setterfield 1994).
(R16010, Jaunter area, Figure 2.12). Although rocks from the Triangle Formation were deposited as turbidites, they are dominated by volcaniclastic debris and their geochemistry is considered to be indicative of their primary magmatic source.

Triangle Formation volcaniclastics are basaltic with SiO₂ contents ranging from 48 to 54 wt% (Table 5.1, Figure 5.4a, b, c). K₂O values are generally moderate but variable ranging from 0.20 to 2.15 wt%.

The Jaunter Volcanics sample has the highest K₂O content with 2.42 wt% (Figure 5.4b). MgO content varies from 4.00 to 7.76 wt% in the Triangle Formation with 9.69 wt% in R16010 from the Jaunter Volcanics (Table 5.1, Figure 5.4c). TiO₂ values are low and occur in a narrow range from 0.6 to 0.96 wt%, and the Cr content is also relatively low and clustered except for R15965 and R16010 (Figure 5.5c, d).

The plot of REE has a similar pattern to those of the Rockley Volcanics with slight to moderate fractionation and slight to moderate depletion in the HREE (Figure 5.6c). Normalised La/Yb ratios for the Triangle Formation range from 2.77 to 5.40 with (La/Yb)ₙ = 2.58 in sample R16010 (Jaunter Volcanics). On the multi-element plot, these samples exhibit a similar spiked pattern to the Rockley Volcanics (Figure 5.7c). Scatter indicates mobility in the LILE probably due to alteration, although all the Upper Ordovician samples still show enrichment in Ba, Sr, Rb and K, whereas R15987 (Gidyen Volcaniclastic Member) has consistently lower, relatively unenriched values of all these elements (Table 5.1; Figure 5.7c). The HFSE, especially Nb, Zr, Hf and Y, are depleted relative to N-MORB (Figure 5.7c). Additionally, these rocks are also characterised by low V/Ti ratios, high Ba/La ratios, moderate Cr contents, moderate Ti/Zr ratios, high Ba/La ratios and low Zr/Y ratios (Figures 5.8a, b, 5.9a, b; Table 5.1). ɛ⁰Nd (450) values were calculated for each of the Triangle Formation samples and R16010; they are moderately positive relative to CHUR, ranging from +3.01 up to +4.59 (Figure 5.11a).
Figure 5.8 (a) V versus Ti/1000 diagram showing the arc characteristics of both the Rockley Volcanics and the Triangle Formation. Discrimination diagram and Fiji shoshonitic data from Shervais (1982), Tavua, Fiji field from Rogers and Setterfield (1994), other compiled fields from Rollinson (1993). (b) Cr versus Y plot shows volcanic arc basalt characteristics of both the Rockley Volcanics and the Triangle Formation. Discrimination diagram from Rollinson (1993) after Pearce (1982) showing the fields for MORB, volcanic-arc basalts (VAB) and within-plate basalts (WPB). Tavua, Fiji data from Rogers and Setterfield (1994).
Figure 5.9 (a) Ti/Zr versus Zr plot for the Rockley Volcanics and Triangle Formation. Comparative fields from the Tonga Arc and Lau Basin basalts (Gamble et al. 1993), Aleutian arc basalts (McCulloch & Perfit 1981; Romick et al. 1990), Marianas arc basalts (Hole et al. 1984; Woodhead 1989) and Vanuatu arc basalts (Gorton 1977) have been added. (b) La versus La/Yb plot for the Rockley Volcanics, Triangle Formation, along with comparative shoshonitic suites and calc-alkaline arc basalts. Shoshonitic fields for Tavua, Fiji from Rogers and Setterfield (1994); Poncho, Andes from Kay and Gordillo (1994); Marianas from Lin et al. (1989); Absaroka Mountains, Montana from Meen and Eggler (1987) and Vulsini, Italy from Rogers et al. (1985). Calc-alkaline fields for Aleutian arc from McCulloch and Perfit (1981) and Romick et al. (1990), and Vanuatu arc from Gorton (1977).
Figure 5.10 (a)-(b) Tectonic discrimination diagrams of Müllar et al. (1992) for arc-related potassic rocks showing the Rockley Volcanics and Triangle Formation predominantly plotting in the (a) oceanic arc and (b) late oceanic arc fields. Data from the Lower Silurian Gundaiy Formation (data from Jones et al. 1995) plots in the continental or post-collisional arc field. (c)-(d) Zr/Y versus Zr discrimination diagrams of Pearce (1983) for basalts showing the oceanic volcanic-arc characteristics of the Rockley Volcanics and Triangle Formation samples, and the within-plate or continental volcanic-arc characteristics of the Lower Silurian Gundaiy Formation (data from Jones et al. 1995). In (b) the fields are A, volcanic-arc basalts; B, MORB; C, within-plate basalts; D, MORB and volcanic-arc basalts; E, MORB and within-plate basalts. Fields from Pearce and Norry (1979) and Rollinson (1993). In (c)-(d) the fields are for continental and oceanic-arc basalts. Lines either side of line Zr/Y = 3, outline the field of overlap between the two basalt types.
5.5 DISCUSSION

5.5.1 Calc-alkaline to Shoshonitic Affinity of the Rockley Volcanics and Triangle Formation

The Rockley Volcanics and Triangle Formation have been classified previously as shoshonitic based on their high potassium content along with a 'red', high potassium signature relative to other rocks on radiometric images (Wyborn 1992a; Watkins 1998). Most samples from the Rockley Volcanics plot in the high-K to shoshonitic fields on the K\textsubscript{2}O versus SiO\textsubscript{2} plot of Peccerillo and Taylor (1976), although the data shows considerable scatter (Figure 5.4b). The characteristic features of the shoshonite association include Na\textsubscript{2}O + K\textsubscript{2}O > 5%; K\textsubscript{2}O/Na\textsubscript{2}O > 0.6 at 50% SiO\textsubscript{2}; enrichment in P, Rb, Sr, Ba, Pb, LREE; and low TiO\textsubscript{2} (<1.3%) (Morrison 1980). High total alkalis, low TiO\textsubscript{2} values and enrichment of LREE are characteristic of the Rockley Volcanics. Taken together, these characteristics support the high-K calc-alkaline to shoshonitic magmatic affinity of the Rockley Volcanics that has previously been identified. The characteristics used here however are more reliable than previous classifications that were based primarily on the high concentrations of K\textsubscript{2}O, which is more susceptible to mobility during alteration and metamorphism.

Samples of the Triangle Formation are also scattered on the K\textsubscript{2}O-SiO\textsubscript{2} diagram of Peccerillo and Taylor (1976) (Figure 5.4b), but predominantly plot in the calc-alkaline to high-K calc-alkaline fields. TiO\textsubscript{2} values of these rocks are also low (between 0.60 – 0.96 wt%), and they also exhibit enrichment in the LREE; however, the total alkali values are generally lower than for the Rockley Volcanics and range between 3.85 and 6.09 wt%.

Based on the above data, the Rockley Volcanics are classified as high-K calc-alkaline to shoshonitic basalts and basaltic andesites, whereas the Triangle Formation are considered to have a calc-alkaline to high-K calc-alkaline, but not shoshonitic magmatic affinity. When compared with other shoshonites, including the type shoshonites from the Absaroka Mountains, the Rockley Volcanics and Triangle Formation samples exhibit lower La/Yb ratios, indicating they are less fractionated, and
compare well with Miocene shoshonitic rocks erupted from Tavua Volcano in Fiji (Figure 5.9b).

5.5.2 Source of Triangle Formation detritus
The slight differences in petrographic and geochemical characteristics indicate that the volcaniclastic detritus of the Upper Ordovician Triangle Formation was not derived from the Rockley Volcanics but must have been supplied from mafic volcanics erupted elsewhere within the Molong Volcanic Province. Possible hornblende-bearing sources with appropriate medium to high-K calcalkaline affinities include the Oakdale Formation and the Byng Volcanics (Watkins 1998).

The sample from the late Middle Ordovician Gidyen Volcaniclastic Member has lower Rb, Sr, Ba and K values than other samples from the Triangle Formation, which indicates calc-alkaline magmatic affinities of its provenance. This is consistent with the geochemistry of Early-Middle Ordovician volcanics of the Molong Volcanic Province, which have low- to high-K calc-alkaline affinities (Glen et al. 1998; Watkins 1999); shoshonitic affinities have only been demonstrated for Late Ordovician rocks.

5.5.3 Tectonic Setting from basalt geochemistry
The Rockley Volcanics and Triangle Formation show enriched LREE and depleted HREE chondrite normalised patterns (Figures 5.6a, b), depletion in the HFSE (Nb, Zr, Ti, Y) relative to LFSE (Rb, Ba, K, Sr) and N-MORB, and depletion in Th relative to Rb and K (Figures 5.7a, b). These compositional signatures are characteristic of subduction-related basaltic magmas erupted in an oceanic island arc volcanic setting (Pearce 1983, 1996). The subduction-related affinities are further demonstrated on tectonic discrimination diagrams for potassic rocks, where they predominantly plot in the late oceanic arc field (Figure 5.10a, b). This tectonic setting is consistent with subduction-related geochemical signatures reported for other Middle to Late Ordovician mafic volcanics within the Dubbo and Bathurst 1:250 000 geological sheet areas in the Molong Volcanic Belt (Watkins 1998, 1999; Glen et al. 1998), from the Parkes area of the Junee-Narromine Volcanic Belt (Clarke 1990) and from the
Figure 5.11 (a) Epsilon Nd versus $^{87}\text{Sr}/^{86}\text{Sr}$ correlation diagram showing moderate to high $\varepsilon$Nd values calculated at 450 Ma for the Rockley Volcanics and Triangle Formation. For comparison values from other Ordovician mafic volcanics of the Lachlan Fold Belt calculated at 450 Ma. Goonumbla Volcanics from Whitford et al. (1993). Other Ordovician volcanics shown without $^{87}\text{Sr}/^{86}\text{Sr}$ values (on right abscissa) are from Wyborn and Sun (1993) and Raymond and Sun (1998), and the Silurian Gundary Formation also calculated at 450 Ma (Jones et al. 1995). Field of the mantle array as defined by modern oceanic basalts and the enriched mantle end member 1 (EM1) from Hart et al. (1986). (b) For comparison with 5.10a, Nd and Sr isotopic values of the Aleutian, Marianas-Izu, New Britain, Scotia, Lesser Antilles/Andes (Grenada-St Kitts-Dominica) and Indonesian arcs compared to MORB and oceanic island basalt values from Iceland and Hawaii. Modified from McCulloch and Perfit (1981).
Kiandra Volcanic Belt to the south (Owen & Wyborn 1979).

R15988 from the Rockley Volcanics in the Oberon area has higher Zr, Nb, Ce and REE values, a slightly lower Ti content and a higher Zr/Y ratio and plots within the continental arc field on the discrimination diagrams of Müller et al. (1992) (Figures 5.10a, c, d). The $\varepsilon_{Nd}$ value calculated for this sample (+3.39) is low compared with the other Rockley Volcanics samples, which may be indicative of greater crustal contamination (Figure 5.11a; Table 5.1). However, this sample occurs at the same stratigraphic level as R15982 and R15961 and there is no apparent reason for these differences.

5.5.4 Nature of the basement in the Oberon-Rockley district

$\varepsilon_{Nd}$ values of Ordovician mafic volcanic rocks provide constraints on contamination of mantle-derived magmas by material derived from melting of continental lithosphere (DePaolo & Wasserburg 1976; Faure 2001, p12; Figure 5.11a, b). $\varepsilon_{Nd}$ values calculated at 450 Ma for the Rockley Volcanics and Triangle Formation are compared with $\varepsilon_{Nd(450)}$ from the other Ordovician mafic volcanics from the Lachlan Fold Belt (Wyborn 1988; Whitford et al. 1993; Wyborn & Sun 1993; Raymond & Sun 1998), together with $\varepsilon_{Nd(450)}$ for Silurian mafic volcanics of the Gundary Formation from the Goulburn area (Jones et al. 1995; Figure 5.11a) $\varepsilon_{Nd(450)}$ for both the Rockley Volcanics and Triangle Formation are moderate to high, between +3 and +7, indicative of relatively unevolved, mantle derived magmas with little or no crustal contamination. These data are consistent with Pb isotopic signatures, which indicate derivation by mixing of Pb from enriched lithospheric and primitive asthenospheric mantle sources (Carr et al. 1995). Initial $^{87}\text{Sr}/^{86}\text{Sr}$ values are relatively low and quite variable in the Triangle Formation, with samples from the Gidyen Volcaniclastic Member and Hampton Road along with the Jaunter Volcanics sample all having higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Figure 5.11a). Some of the variability may reflect redistribution during interaction with seawater and low-grade metamorphism.

When compared with the $\varepsilon_{Nd(450)}$ values from the other Ordovician mafic volcanic, the Rockley Volcanics and Triangle Formation are slightly more evolved, with R15988 having the lowest values.
The values calculated from the Silurian rocks have lower values and overall the data plot as a continuum, with a gradual decrease in $\varepsilon_{Nd}(450)$ values from the Ordovician through to the Early Silurian (Figure 5.11a). This indicates that the magmas became more evolved over time with crustal input increasing throughout the Late Ordovician to Early Silurian interval. This is consistent with the change from an oceanic arc-related volcanic setting in the Late Ordovician to a continental arc-related volcanic setting during the Late Silurian (Figure 5.9a, c, d).

5.5.5 Shoshonitic Volcanism in Island Arcs
Shoshonitic rocks occur in most tectonic settings, and in modern island arc settings they are uncommon but widespread, and generally associated with perturbation to the normal tectonic setting of arc magmatism. Resolving the cause of K-rich magmatism however, is difficult due to the diversity of petrogenetic processes and sources that can result in shoshonitic rocks being produced (Box & Flower 1989). Notwithstanding this, they are most commonly associated with four tectonic settings: (1) above deep Wadati-Benioff zones, (2) arc-continent collision, (3) arc rifting, and (4) near arc termini (Box & Flower 1989).

Where K-rich magmatism occurs above deep Wadati-Benioff zones, the shoshonitic rocks form linear chains located farthest from the trench and are considered associated with melt derived from the deepest parts of the Benioff zone such as in the Sunda Arc (Whitford et al. 1979). K-rich arc magmatism sometimes occurs during, or following, continental underthrusting associated with arc-continent collision, such as in the Aeolian Arc of southern Italy, where the arc is built on continental crust (Varekamp & Kalaramides 1989). Shoshonitic volcanism related to arc rifting has been documented from the northern Mariana (Volcano) Arc, where it is thought to be related to propagation of the Mariana Trough spreading centre into the arc (Bloomer et al. 1989). Arc rifting along transform structures oblique to the arc is considered responsible for the production of shoshonites and the onset of rifting in the once continuous Vanuatu-Fiji-Tonga arc during the Late Miocene (Gill & Whelan 1989; Rogers & Setterfield 1994). In settings 2, 3 and 4 above, the
shoshonitic rocks form with no spatial trend, and are often co-magmatic with calc-alkaline and/or tholeiitic rocks. These settings are considered associated with either the initial (e.g. Marianas - Volcano Arc), or last phase of arc magmatism often following or associated with transcurrent faulting, rotation, splitting and break-up of the arc (e.g. Vanuatu-Fiji-Tonga Arc; Aeolian Arc, Italy), ultimately resulting in subduction zone failure or flipping, commonly with associated block faulting and uplift within the arc (Morrison 1980).

5.5.6 Tectonic Setting of the Molong Volcanic Province

High initial $^{87}\text{Sr}/^{86}\text{Sr}$ values, the dominance of shoshonites over calc-alkaline rocks and the broad geographical distribution rather than curvilinear belts of volcanic units, has lead previous investigators to argue that the modification of the lithospheric mantle responsible for the shoshonitic magmatism occurred during an earlier subduction event associated with the assembly of continental basement blocks (Owen & Wyborn 1979; Wyborn 1992a). As such, they related Ordovician volcanism to lithospheric heating or overturning prior to upwelling of unmodified asthenosphere. The implication being that subduction was not synchronous with volcanism during the Ordovician. However, unevolved mantle $\varepsilon_{\text{Nd}}$ values for basalts from the Oberon-Rockley district (+3.4 to +6.4) or elsewhere within the Lachlan Fold Belt (+4.2 to +8.0; Wyborn & Sun 1993), together with Pb isotopic data (Carr et al. 1995), do not support the presence of continental basement blocks beneath the Molong Volcanic Province.

Several lines of evidence support the simultaneous occurrence of subduction and arc volcanism contemporaneous with variation and enrichment of the magmatic source material during the Middle to Late Ordovician. This evidence includes the change from early formed plagioclase phenocrysts to early formed clinopyroxene phenocrysts as the magmatism becomes younger (Wyborn 1996); the decrease in $\varepsilon_{\text{Nd}}$ values from $> +6.5$ for late Middle Ordovician rocks to $< +6.5$ for Late Ordovician
rocks (Wyborn & Sun 1993); and the trend of Pb isotopic ratios along a very precise oceanic mantle mixing line (Carr et al. 1995). The Pb isotopic values however, lack an arc-like Pb isotopic signature involving subduction of crustal derived sediments or continental crust. Therefore constraining any Ordovician magmatic model involving contemporaneous subduction to one of recycling of metasomatized oceanic crust with minimal crustal derived sediments and the lack of continental crust (Carr et al. 1995). A tectonic model with a basement of oceanic crust is also supported by the $\varepsilon_{Nd}$ isotopic data.

Glen et al. (1998) suggested that eruption of the shoshonites was restricted to intersections of volcanic belts with transverse zones, which tapped the LREE enriched source material. They argued that the shoshonitic volcanism occurred following subduction of a seamount that resulted in uplift of the arc, a hiatus in volcanism and subsequent deposition of shallow marine limestone. As outlined in Chapter 3 however, deposition of shallow-marine limestone began around isolated volcanic edifices during the late Darriwilian synchronous with volcanism, and continued into the Late Ordovician (Percival et al. 1999). Further, the ages of many Late Ordovician volcanic and volcaniclastic units are inferred from sparse age data, and their age ranges are not constrained well enough to reliably delineate a volcanic hiatus during this period.

The change from calc-alkaline to shoshonitic volcanism was accompanied by a volumetric increase in volcanism, and does appear to have a spatial relationship with the transfer zones (Lachlan Transfer Zone and Hunter River Transfer Zone) of Glen et al. (1998). The Ordovician shoshonitic basalts are geochemically similar to shoshonites erupted in the Fiji islands at ~5.0 Ma during the late Miocene, following rifting of the Vanuatu-Fiji-Tonga arc along transform faults oblique to the arc between 7.0 and 5.0 Ma (Gill & Whelan 1989; Rogers & Setterfield 1994). In Fiji, shoshonitic rocks are most abundant close to the transform faults, but are documented up to 225 km away where they occur with
calc-alkaline and tholeiitic rocks of the same age. The tectonic setting of the Fiji Islands is further compared with the northeastern Lachlan Fold Belt in Chapter 6.

Wyborn (1992a) argued that the broad geographical distribution of the Ordovician volcanic units did not support a subduction related setting, which generally had curvilinear belts of volcanics. The Ordovician volcanic belts are presently spread over a width of approximately 200 km (Figure 5.1), although this is probably an underestimate of the original width following shortening during the Late Devonian and Early Carboniferous. Rifting and breakup of the Ordovician arc either synchronously with or immediately following shoshonitic magmatism in the northeastern Lachlan Fold Belt would account for a broad width. Miocene shoshonitic arc volcanics in the Fiji region also form a broad zone in excess of 400 km in width following splitting and rotation of the original Miocene arc (Gill & Whelan 1989).

5.6 CONCLUSIONS
(1) The Rockley Volcanics in the Oberon-Rockley district have high contents of LREE, LILE (Ba, K, R, Sr) and $K_2O + Na_2O$, and low TiO$_2$ concentrations, indicating high-K to shoshonitic magmatic affinities with oceanic island-arc characteristics and compare well with oceanic island arc shoshonitic basalts from Fiji, whereas the Triangle Formation are calcalkaline to high-K based on lower values of enrichment in LREE and LILE, low TiO$_2$, but lower (<5 wt%) $K_2O + Na_2O$ values than the Rockley Volcanics.

(2) Based on differences in geochemistry and mineralogy, the Triangle Formation was derived from a different source to that of the Rockley Volcanics.

(3) $\varepsilon_{Nd}$ values (+6.43 to +3.01) and Pb isotopic data (Carr et al. 1995) indicate that the magma source of the Ordovician volcanics was derived from an unevolved mantle source mixed with an enriched mantle source, but lacking contamination from melts derived from continental crust or subducted sediments. This implies that they are underlain exclusively by oceanic crust. The enrichment of the mantle source can be explained by addition of fluids or melt derived from subducted
metasomatised oceanic crust beneath the arc.

(4) The Rockley Volcanics and the source volcanoes of the Triangle Formation were formed above an active subduction zone in an oceanic island-arc tectonic setting that was active throughout the Middle to Late Ordovician.
CHAPTER 6

CONCLUSIONS

6.1 INTRODUCTION
The following discussion expands upon the conclusions of the preceding chapters (2, 3, 4, 5), and
considers the implications of them with regard to the palaeogeographic and tectonic relationships
between the two Ordovician lithological provinces. The tectonic history of the Oberon-Taralga region
throughout the Palaeozoic, with specific emphasis on the Ordovician to Early Silurian regional
palaeogeography and tectonic evolution of the Ordovician volcanic arc is discussed with reference to
current Lachlan Fold Belt tectonic models and a recent tectonic analogue. The chapter concludes with
a summary of the major thesis findings.

6.2 QUARTZOSE SEDIMENTARY PROVINCE
In the Oberon-Taralga region, the stratigraphy of the Quartzose Sedimentary Province comprises a
fining upward succession from Middle Ordovician quartz turbidites of the undifferentiated
Adaminaby Group, through the late Middle Ordovician Numeralla Chert and thin-bedded turbidites of
the Bumballa Formation, overlain by the Upper Ordovician Warbisco Shale. This stratigraphy is
comparable with that described from elsewhere within the Eastern Belt of the Lachlan Fold Belt
(Figures 1.2, 6.1). The rocks of the Adaminaby and Bendoc Groups in the Oberon-Taralga region are
considered an extension of those Groups mapped farther south, particularly in the Shoalhaven River
Gorge, Mallacoota Zone and in eastern Victorian in the southern part of the Cowra-Canberra-Buchan
Zone where they are well developed (VandenBerg & Stewart 1992; Lewis et al. 1994; Fergusson
1998b; VandenBerg et al. 2000; Figures 1.2, 6.1). A similar stratigraphy consisting of undifferentiated
Adaminaby Group quartz turbidites (Unit A), late Darriwilian chert (Unit B) and thin-bedded quartz
turbidites (Unit C) has also been described from the Mudgee area, where volcaniclastic rocks of the
Molong Volcanic Province conformably overlie it (Fergusson & Colquhoun 1996). Based on
sedimentological features, the continuity and lateral regularity of turbidite beds and conodont
biofacies within units of the Adaminaby Group, the succession is interpreted as having been deposited
in a deep-marine setting, on the middle to outer margins of a large terrestrial submarine fan distant from the continental detrital source.

In contrast to the areas along the south coast of New South Wales (Mallacoota and Narooma Zones), in eastern Victoria and the Western Belt of the Lachlan Fold Belt, no evidence has been found in the Oberon-Taralga region indicating the presence of undifferentiated Adaminaby Group quartz turbidites older than late Middle Ordovician (Figures 2.1, 6.1). Fossil evidence indicating the presence of Early to early Middle Ordovician quartz turbidites of the undifferentiated Adaminaby Group is also lacking in the Shoalhaven River area to the south and the Mudgee area north of the Oberon-Taralga area (Fergusson & VandenBerg 1990; Fergusson & Colquhoun 1996; Fergusson 1998b). The lack of the older parts of the Adaminaby Group in this area may be related to the Lower to late Middle stratigraphic hiatus documented from the shallow marine units of the Molong and Junee Narromine Volcanic Belts. The significance of this hiatus in the rock record is discussed later.

### 6.3 MOLONG VOLCANIC PROVINCE

Molong Volcanic Province rocks only occur in the Oberon area of the Oberon-Taralga region. They comprise deep-marine chert and mafic volcaniclastic deposits and all are included within the Triangle Formation. The basal Budhang Chert Member contains Early Ordovician (Bendigonian) conodonts, and is overlain by two major pulses of mafic volcaniclastic deposits; the Gidyen Volcaniclastic Member and undifferentiated Upper Ordovician Triangle Formation, separated by the late Darriwilian Mozart Chert Member. The timing of the onset of deposition of the Gidyen Volcaniclastic Member is poorly constrained and indicates that either there has been a long hiatus between deposition of the Bendigonian and Darriwilian, or together the Budhang Chert Member and Gidyen Volcaniclastic Member represent over 25 Ma of continuous deposition. The existence of a long hiatus is the favoured alternative because Darriwilian or younger ages only have been established for the rest of volcanic succession in the Molong Volcanic Province (Meakin & Morgan 1999, Packham et al. 1999; Pickett & Percival 2001). Deposition of the second pulse of volcaniclastic deposition began in the early Late
Ordovician, but its cessation is only constrained by the unconformably overlying Upper Silurian Fosters Creek Conglomerate. Therefore, deposition of volcanioclastic detritus into the deep-marine environment may have continued into the Early Silurian.

The stratigraphy recognised within the Molong Volcanic Province rocks occurring in the Oberon area correlates with that described from the Rockley area to the west (Fowler & Iwata 1995; Figure 2.1). These units in the Oberon area are considered an extension of these deep-marine facies within the southern part of the Rockley-Gulgong Volcanic Belt (Figure 1.2). Farther to the west in the Molong and Junee-Narromine Volcanic Belts, Early Ordovician pelagic and hemi-pelagic deposits (Hensleigh Siltstone and Yarrimbah Formation respectively) are comparable to the Budhang Chert Member occurring in the Oberon area. In these volcanic belts however, mafic volcanic rocks with geochemical signatures indicating island arc tectonic affinities underlie the Early Ordovician deep-marine units (Figure 6.1). No comparable Early Ordovician volcanics have been recognised in the Rockley-Gulgong Volcanic Belt.

As in the Rockley-Gulgong Volcanic Belt, a significant Early to Middle Ordovician stratigraphic hiatus occurs in both the Junee-Narromine and Molong Volcanic Belts (Meakin & Morgan 1999; Percival et al. 1999; Lyons et al. 2000). Late Middle Ordovician (Darriwilian) rocks are also the oldest rocks recognised from the Kiandra Volcanic Belt (Owen & Wyborn 1979).

By the late Middle to early Late Ordovician (Darriwilian to Gisbornian), the Molong and Junee-Narromine Volcanic Belts are characterised by shallow marine to subaerial conditions, with Darriwilian to Gisbornian shallow marine limestone and limestone clasts interbedded with primary volcanics and volcanioclastics at the base of the Fairbridge and Goonumbla Volcanics (Percival et al. 1999; Pickett & Percival 2001). Contemporaneously with late Darriwilian to Late Ordovician volcanism on the topographic highs, volcanioclastic detritus probably derived from the Molong Volcanic Belt, was deposited as a volcanioclastic apron into the surrounding deep marine basins, where
the volcaniclastic detritus is interbedded with deep-marine chert, black shale and middle Late Ordovician (?Eastonian) mafic volcanics and comprises the Rockley-Gulgong and Kiandra Volcanic Belts.

A hiatus in volcanism is recognised in the Junee-Narromine Volcanic Belt during the late Gisbornian to Eastonian, (Pickett & Percival 2001) and a similar hiatus in volcanism has been postulated for the western portion of the Molong Volcanic Belt, when reefal limestone deposition took place during the Eastonian (Packham et al. 1999). On the eastern side of the Molong Volcanic Belt and in the Rockley-Gulgong Volcanic Belt, the Eastonian is characterised by significant volumes of volcaniclastic deposition, together with continued volcanism resulting in eruption of minor lavas in the Byng, Rockley and Sofala Volcanics and the Oakdale, Burrangh and Tucklan Formations, in a deeper water environment off the topographic high. Limestone and volcaniclastic clasts derived from the adjacent volcanic high occur in breccia/conglomerate within these units. Volcanic activity recommenced in the western Molong and Junee-Narromine Volcanic Belts in the latest Eastonian to Bolindian, with primary volcanics recognised in the Goonumbla and Forest Reefs Volcanics and the Angalong and Cheesemans Creeks Formations (Pogson & Watkins 1998; Meakin & Morgan 1999; Packham et al. 1999; Lyons et al. 2000). The cessation of mafic volcanism within the Molong and Junee-Narromine Volcanic Belts during the latest Ordovician to earliest Silurian is marked by a phase of widespread plutonic activity.

6.4 PALAEOGEOGRAPHIC RELATIONSHIPS BETWEEN THE TWO ORDOVICIAN LITHOLOGICAL PROVINCES IN THE OBERON TARALGA REGION

6.4.1 Early Ordovician
Early Ordovician sedimentation within the Oberon-Taralga region is only known from the Molong Volcanic Province. Therefore, there is no clear stratigraphic relationship between the two provinces prior to the late Middle Ordovician.
6.4.2 Late Middle Ordovician
In the Oberon area, late Middle Ordovician rocks of both the Molong Volcanic Province and the Quartzose Sedimentary Province were deposited synchronously and are now juxtaposed along the Vulcan Fault. Rocks of both provinces are characterised by deep-marine depositional environments. However, those of the Molong Volcanic Province (Gidyen Volcaniclastic Member) are composed entirely of volcaniclastic detritus, whereas those of the Quartzose Sedimentary Province (undifferentiated Adaminaby Group) comprise craton-derived quartz turbidites. Based on the lithological distinctiveness of these units and the lack of any interdigitation between them, they are interpreted as having been deposited distant from each other.

Both the Gidyen Volcaniclastic Member and the undifferentiated Adaminaby Group are overlain by pelagic chert and siltstone units (Mozart Chert Member and Numeralla Chert respectively), indicating both provinces were affected by a reduction in sediment supply during the late Darriwilian to early Gisbomian, and both units contain the same species of conodonts, have similar deep-marine depositional environments and slow sedimentation patterns. The presence of fine-grained volcaniclastic detritus interbedded within the Mozart Chert Member and its complete absence in the Numeralla Chert indicates these units were also deposited at some distance from each other, with the Mozart Chert Member closer to the source of volcanic detritus.

6.4.3 Late Ordovician
Overlying the Numeralla Chert in the Quartzose Sedimentary Province are thin-bedded, deep-marine quartz turbidites (Bumballa Formation), succeeded by siliceous black shale (Warbisco Shale). Within the Molong Volcanic Province, the Mozart Chert Member is overlain by Upper Ordovician to Early Silurian mafic volcaniclastic turbidites thought to be derived from volcanic centres to the north and west. These Upper Ordovician to Early Silurian mafic volcaniclastic turbidites also overlie the Upper Ordovician Warbisco Shale in the eastern part of the Oberon area. The widespread deposition of the volcaniclastic detritus during the Upper Ordovician represents the distal expansion of a volcaniclastic
Stratigraphic relationships between Late Ordovician units of both lithological provinces east of the Vulcan Fault indicate that the two provinces were adjacent to each other within the forearc by the middle Late Ordovician. Movement of fault blocks within the forearc region of the Ordovician arc is considered to have occurred throughout the late Middle and Late Ordovician synchronous with an increase in magmatism and the west to east expansion of the volcanioclastic apron. Movement of fault blocks within the forearc is supported by the occurrence of proximally derived Rockley Volcanics occurring stratigraphically above units of the Quartzose Sedimentary Province east of Oberon, indicating this structural block was closer to the arc than it had been in the Middle to early Late Ordovician. In addition, small syndepositional thrust faults were observed in the Rockley Volcanics in Native Dog Creek implying that the area was under compression during the Late Ordovician. A similar stratigraphic relationship is also recognised in the Mudgee district, north of the Bathurst Granite, where volcanioclastics of the Upper Ordovician Coomber Formation conformably overlie the Bumballa Formation of early Upper Ordovician age (Fergusson & Tye 1999).

6.5 EARLY TO MIDDLE ORDOVICIAN STRATIGRAPHIC HIATUS
The presence of the stratigraphic hiatus in the shallow-marine platform areas of the volcanic highs has been attributed to either removal of shallow water deposits by erosion, or a lack of shallow water deposition throughout this time (Percival 1999a). The change from deep marine Early Ordovician deposits (Yarrimbah Formation and Hensleigh Siltstone) in the two western Volcanic Belts to late Middle Ordovician shallow marine limestone deposition indicates these areas underwent significant uplift during the period represented by the hiatus. Recent work in the Gunningbland area of the Junee-Narromine Volcanic Belt has demonstrated the presence of reworked chert and volcanic detritus at the base of the Gisbornian Billabong Creek Formation, indicating erosion of the underlying Ordovician section (Pickett & Percival 2001). Recognition of the stratigraphic hiatus in the deep marine deposits
of the Oberon area is significant. It indicates that the missing Early to late Middle Ordovician shallow-marine deposits were not simply eroded off the platform areas and redeposited into the surrounding deep-marine basins, but that the effects of this event were widespread, affecting sedimentation in both the shallow and deep marine environments across the entire arc. Quartzose Sedimentary Province rocks occurring in the forearc region east of the arc and immediately south of the volcanic highs are not considered to be older than Darriwilian in age, based predominantly on the relatively thin stratigraphic succession interleaved between Upper Ordovician black shales. However, the maximum age of the Ordovician turbidites in these areas is difficult to ascertain due to their lack of fossil content, but the lower parts of the Adaminaby Group exposed further south in the Eastern Belt are not known to occur. Therefore, the lack of Quartzose Sedimentary Province rocks older than ?Darriwilian may be the result of later deposition of the turbidite fan in this area or may also indicate rocks of this province were also affected by the hiatus forming event.

6.6 REDUCTION IN SEDIMENT SUPPLY TO THE TURBIDITE FAN
A reduction in the sediment supply to the submarine fan resulted in the widespread change from predominantly thick-bedded quartz turbidite deposition to hemi-pelagic chert and thin-bedded quartz turbidite deposition during the late Middle Ordovician. The cause of the reduction in the sediment supply was discussed in Chapter 2 (Section 2.4), where it was considered that a rise in sea level during the early Darriwilian may have been a factor in the onset of chert deposition at this time. However, it was concluded that factors other than eustatic sea level changes were more significant in controlling the sediment supply throughout the late Darriwilian and Late Ordovician. Continued turbidite fan sedimentation in the Western Belt of the Lachlan Fold Belt precludes the reduction in sediment supply being due to the peneplanation of the Delamerian Mountain Chain. A change in palaeocurrent directions in the Eastern Belt flowing to the northeast to flowing towards the north was recognised in the Shoalhaven River Gorge and the Burnt Hole Creek area south of Oberon (Fergusson & Tye 1999). Based on the above, the rapid reduction in sediment supply to the submarine fan in the Eastern Belt of the Lachlan Fold Belt during the Darriwilian is related to the onset of the Benambran
Orogeny, resulting in uplift in the Wagga-Omeo Zone blocking supply of sediment to the Eastern Belt whilst sedimentation continued uninterrupted within the western belt (Fergusson & Fanning in prep).

6.7 RECONSTRUCTING THE ORDOVICIAN ISLAND ARC AND THE TECTONIC SETTING OF THE OBERON-TARALGA REGION THROUGHOUT THE ORDOVICIAN-EARLY SILURIAN

6.7.1 Early Ordovician
Constraints on the tectonic setting of the Oberon-Taralga area are based on Early Ordovician units which are of limited extent. However, Lancefieldian to Bendigonian deep-marine pelagic sedimentation characterises deposition in the three northern volcanic belts, including the Rockley-Gulgong Volcanic Belt, with tholeiitic to calc-alkaline mafic volcanic rocks (Nelungaloo and Mitchell Volcanics) underlying the pelagic sediments in the Junee-Narromine and Molong Volcanic Belts respectively (Sherwin 1996; Meakin & Morgan 1999; Sherwin 2000). Geochemical analyses of the mafic volcanics indicate that they have oceanic-island arc tectonic affinities; thus an oceanic island existed offshore from Gondwana during the Early Ordovician (Meakin & Morgan 1999; Sherwin 2000). The presence of limestone clasts and well-rounded pebbles in conglomerate/breccia interbedded with the pelagic deposits indicate deposition via turbidite or debris flows from a sub-aerial to nearshore environment close to the volcanic source (Meakin & Morgan 1999).

Due to the stratigraphic hiatus between the Early and late Middle Ordovician units, it is not known whether the Early Ordovician volcanics represent the onset of continuous Ordovician volcanism, or whether there was a break in both magmatism and sedimentation during the Early Ordovician, with the re-establishment of arc volcanism and marginal basin sedimentation in the Middle Ordovician. Turbidite fan deposition continued uninterrupted in the southern part of the Eastern Belt during the Early to Middle Ordovician, with subduction occurring farther outboard and preserved in the Narooma accretionary prism (VandenBerg et al. 2000).
6.7.2 Late Darriwilian-Gisbornian
Following the Early to Middle Ordovician stratigraphic hiatus, arc volcanism, associated with westward subduction, had commenced by the late Middle Ordovician. Late Middle to Late Ordovician rocks of both lithological provinces occurring within the Oberon-Taralga region were deposited within a deep-marine environment in the forearc, east to southeast and outboard of an oceanic island arc. Quartzose turbidite and chert sedimentation (Wagga and Girilambone Groups) was also occurring in a marginal basin (Wagga Marginal Basin) west of and behind the arc during the late Middle Ordovician (Chapter 3). The Narragudgil Volcanics were erupted in this marginal basin during the late Darriwilian to Gisbomian, based on the age of the Mugincoble Chert, and have been related to minor sea-floor spreading. (Percival 1999b; Lyons et al. 2000). Formation of a new basin occurred between the present-day Junee-Narromine and Molong Volcanic Belts during the late Middle Ordovician allowing deposition of quartz turbidites of the Kirribilli Formation (Lyons et al. 2000). Evidence of spreading or rifting on this new basin floor is preserved as the basalts and dolerites of the Brangan Volcanics and ultramafic Wambidgee Serpentinite, that have been associated with late Darriwilian to Gisbornian chert (Percival 1999c; Lyons et al. 2000). Small discontinuous limestone reefs had developed around emergent volcanic edifices by the latest Darriwilian to Gisbornian in the Junee-Narromine Volcanic Belt and the western part of the Molong Volcanic Belt (Pickett & Percival 2001; Percival et al. 1999). This time also marks the onset of shoshonitic volcanism within the Junee-Narromine Volcanic Belt. By the middle Gisbornian, large volumes of volcanioclastic material were also deposited into basins adjacent to volcanic centres and the onset of shoshonitic volcanism had begun in the eastern part of the Molong Volcanic Belt.

6.7.3 Eastonian
A shift in the volcanic front away from the linear chains of volcanoes on the volcanic highs occurred during the Eastonian. During this time, the Junee-Narromine and western part of the Molong Volcanic Belts, are characterised by shallow marine limestone deposition followed by deeper marine shale deposition as the volcanic highs started to subside (Packham et al. 1999; Pickett & Percival 2001).
However, minor lavas may have been erupted during this time as there are some poorly dated mafic volcanic units present (Parkes Volcanics, Nash Hill Volcanics and the Goobang Volcanics). Within the eastern Molong and Rockley-Gulgong Volcanic Belts however, large volumes of volcaniclastic detritus derived from the adjacent western Molong Volcanic Belt were deposited along with pelagic black shale, and are interbedded with conglomerates containing allochthonous limestone clasts also derived from the volcanic highs. At the time same, in the eastern Molong and Rockley-Gulgong Volcanic Belts, minor lavas were erupted from intrabasinal seamounts away from the topographic highs (Rockley Volcanics, Sofala Volcanics, Mt Pleasant Basalt Member).

### 6.7.4 Bolindian to Early Silurian

During the latest Eastonian to early Bolindian volcanic activity was renewed in both the Junee-Narromine and western Molong Volcanic Belts and continued within the eastern Molong Volcanic Belt, whilst widespread volcaniclastic deposition continued in the surrounding deep-marine basins. Farther south away from the volcanic highs, pelagic black shale deposition continued in the early Bolindian. By the end of the Bolindian however, volcanic activity had ceased and widespread magmatic intrusive activity took place both within the arc and farther west in the Girilambone Group (Lyons *et al.* 2000). During the earliest Silurian, volcaniclastic deposition continued into the surrounding basins derived from still emergent areas of the volcanic highs, but within the Junee-Narromine Volcanic Belt and in the basins west of it, widespread deposition of deep marine shale (Cotton Formation) had begun. Within the forearc basin and south of the volcanic arc, a significant shift to quartzose turbiditic sedimentation occurred. The renewed quartz turbidite deposition is related to uplift of the Benambran highlands and recycling of Ordovician quartz turbidites from west of the volcanic arc, and has been related to the onset of west facing subduction between the Eastern and Western Belts of the Lachlan Fold Belt (Foster *et al.* 1999).
Figure 6.2 Present-day tectonic map of the south western Pacific modified after Meffre and Crawford (2001). Solid lines with solid triangles show active subduction zones. Dashed lines with open triangles show fossil subduction zones. All latitudes in degrees south. The shaded area is the 2500 m contour after Kroenke (1984). Areas above sea level in black.
6.8 MIocene/PlIocene Calc-Alkaline To Shoshonitec Island Arc Volcanism In Fiji - An Analogue Of The Ordovician Island Arc.

6.8.1 Evolution of the Fijian Arc
Calc-alkaline to shoshonitic volcanism occurred in the Fiji Islands during the late Miocene to Pliocene. Similarities in the geochemistry, tectonic setting and evolution of the New Hebrides-Fiji-Tonga arc make it a suitable analogue for comparison with the Lachlan Fold Belt Ordovician island arc and its volcanic products. The Fiji archipelago is situated on the northern half of a remnant arc fragment which today forms a north-south bathymetric high (Lau-Colville Ridge). The archipelago forms part of a triangular plate segment occurring between two transform faults that connects the opposite facing Tonga-Kermadec and New Hebrides oceanic island arc-subductions systems (Figure 6.2). The largest islands of the archipelago occur at the northern termination of the Lau-Colville Ridge, where it intersects the Hunter Fracture Zone and Fiji transform (Figure 6.2).

Subduction has been proceeding in the Fiji-Tonga region since at least the Eocene, and following rifting of the arc to form the South Fiji Basin as a back-arc basin in the Oligocene, steady-state subduction and predominantly medium-K andesitic arc magmatism continued along the Vitiaz Arc until the mid Miocene (~10 Ma; Gill et al. 1984; Figure 6.3a). Collision of the Ontong-Java Plateau with the Solomon Islands farther north in the arc at about 25 Ma is thought to have caused significant changes in the arc system, ultimately resulting in the cessation and breakup of the arc (Gill & Whelan 1989a; Clift & Dixon 1994). Although the exact timing of the beginning and end of the early rifting stage are not well defined, Vanuatu and Fiji are known to have separated due to propagation of a transverse rift (Hunter Fracture Zone) across the former Vitiaz Arc sometime between 8 and 5.5 Ma (Gill & Whelan 1989a; Figure 6.3b). This resulted in the reversal of subduction polarity along the New Hebrides Arc, opening of the North Fiji Basin, and the rifting of Vanuatu away from Fiji (Clift & Dixon 1994). Counterclockwise rotation of the northwestern end of the Lau Ridge accompanied separation. Westerly dipping subduction in the Tonga portion of the arc continued in the same direction.
Figure 6.3 Neogene tectonic evolution of the Vitiaz, New Hebrides and Tonga-Kermadec island arcs after Gill and Whelan 1989a, b, Greene et al. 1994 and Meffre and Crawford 2001. (a) Mature arc stage when the New Hebrides and Vitiaz arcs formed one arc (outer Melanesian Arc of Packham & Falvey 1971). (b) Initial splitting of the outer Melanesian arc is estimated with a hinge point north and est of Erromango Island (Musgrave & Firth 1993). (c) Vanuatu and Fiji began to separate when a transverse rift formed between them, caused subduction polarity to reverse in Vauatu. The volcanic front remained in the same place in the east but shifted to easter Vanuatu in the west. (d) Late rifting stage. Back-arc basins basins opened east of Fiji and the Lau-Colville Ridge, isolating them from Tonga as a remnant arc. (e) Present-day tectonic setting of the Tonga-Kermadec and New Hebrides Island arcs. Black triangles represent active subduction zones, white triangles represent inactive subduction zones. DER, d’Entrecasteauz Ridge; F, Fiji. LB, Lau Basin; L-CR, Lau-Colville Ridge; NC, New Caledonia; NFB, North Fiji Basin; NH New Hebrides; NLB, North Loyalty Basin; OJP, Ontong Java Plateau; SFB, South Fiji Basin; TK, Tonga-Kermadec arc; VA, Vitiaz arc; WTM, West Torres Massif.
Following the initial rifting of the arc, volcanism is known to have occurred from at least 25 different centres across the Fijian islands; ten of these centres were shoshonitic (including Tavua discussed in Chapter 5), 13 were calc-alkaline and two large areas were tholeiitic. The shoshonitic rocks were erupted between 5.5 and 3.0 Ma, beginning within 50 km of the transverse rift zone that broke the arc across strike and erupted from the ten volcanoes along three lineaments up to 225 km away from the Hunter Fracture Zone and Fiji Transform (Gill & Whelan 1989a). The enriched characteristics of the shoshonitic rocks were interpreted to be acquired at mantle sources rather than by crustal assimilation during ponding beneath thick crust. All of the Fijian volcanism generated during the early rifting phase shared traits common to most oceanic island arcs including low concentrations of HFSE and HREE, with high εNd values and relatively high concentrations of LFSE (Gill & Whelan 1989a).

ODP drilling within the Lau Basin indicates that back-arc basin rifting was fully established in the Lau Basin by approximately 3.0 Ma (Figure 6.3c). However, during the initial stages of basin rifting, the principal mode of volcanism was due to intrabasinal submarine edifices prior to the establishment of back-arc spreading and a fixed chain of arc volcanoes (Clift & Dixon 1994). Large volumes of continuous volcaniclastic sedimentation during this stage is thought to have occurred by proximal mass-flows and turbidity current deposition from the submarine intrabasinal seamounts and via airfall deposits from the still active Lau Ridge. The renewal of arc volcanism from the Tafua Arc on the Tonga Platform coincided with the propagation of the active Lau spreading centres into that area between 2.0 and 3.0 Ma (Clift & Dixon 1994; Figure 6.3d, e).

6.8.2 Similarities between the Lachlan Fold Belt Ordovician arc and the Fijian arc
Similarities between the Ordovician arc in the Lachlan Fold Belt and the Miocene-Pliocene arc in the Fiji region include: (1) the major change in chemistry from normal medium-K arc volcanism to shoshonitic volcanism, accompanied by a widening of activity away from the earlier linear chain; (2) the occurrence of a major across strike transform structure adjacent to the area of major shoshonitic
volcanism (Lachlan Transfer Zone); and (3) the complete cessation of arc volcanism along the main volcanic chain after the shoshonitic volcanic event.

6.8.3 Characteristics of tectonic collisions with modern arc systems
Across strike movement on the Hunter Fracture Zone and Fiji Transform (Figure 6.2) and subsequent breakup of the Vitiaz Arc is considered to have been initiated by collision of the Ontong-Java Plateau with the Vitiaz Arc. The impact of major collisions of arc systems with oceanic plateaus, submarine ridges and seamounts was investigated by Meffre and Crawford (2001). The results of this study showed that most collisions are transient events that leave little impact on the plate configuration and geometry. Within the New Hebrides Arc, collisions were found to have resulted in the trench, forearc and arc-backarc undergoing a cycle of rapid uplift followed by subsidence of between 2000 - 6000 m. Meanwhile, the intra-arc area underwent an opposing cycle with initial subsidence followed by uplift of up to 3000 m, with the complete cycle taking between 3.0 and 5.0 Ma. These cycles result in certain features being preserved in the geological record, including the development of an unconformity surface close to the subduction trench, throughout the arc, and in the backarc (Meffre & Crawford 2001). An angular unconformity was developed in the intra-arc basin due to tilting, and can result in a minor hiatus in sedimentation of up to 0.4 Ma, but commonly resulted in a major change in the style and major increase in the rate of sedimentation (e.g. from approximately 25 to 450 m/Ma; Staerker 1994). The development of thick limestone capping on the uplifted areas of the margins of the intra-arc basin produced discontinuous limestone reefs. Following collision in the modern arcs studied, it was found the intra-arc area was uplifted while the forearc and backarc subsided (Meffre & Crawford 2001). It was determined that the occurrence of multiple unconformity surfaces occurring within a short timespan (<3.0 Ma), limestone deposition and rapid changes in palaeo-depth were characteristic features of collisions that could be preserved within ancient arc-related successions (Meffre & Crawford 2001).

6.9 BENAMBRAN OROGENY
Traditionally, the onset of the Benambran Orogeny in the Eastern Belt of the Lachlan Fold Belt has
been placed near to the Ordovician/Silurian Boundary (VandenBerg 1999). However, Collins and Hobbs (2001) outlined several lines of evidence to suggest that the Benambran Orogeny could be separated into “pre-” and “type-Benambran” events that were possibly unrelated. The onset of the earlier “pre-Benambran” event is though to have occurred within the Eastern Belt of the Lachlan Fold Belt sometime during the Late Ordovician around 450 Ma (Eastonian). In the Western Belt or Whitelaw Terrane of Victoria (VandenBerg et al. 2000) however, the onset of the Benambran Orogeny is earlier than previously thought based on $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric mica ages ranging from 460 to 420 Ma (Foster et al. 1999). Sm-Nd and $^{206}\text{Pb}/^{238}\text{U}$ ages from the Arunta Block in central Australia also indicate a change in tectonic style and crustal evolution occurred during the Early to Middle Ordovician. Here, high grade lower crustal metamorphism occurred in an extensional setting throughout the Early Ordovician and is considered to have changed to compressional deformation as early as 450 Ma and resulted in the uplift of the intracratonic basins during the Late Ordovician (Shaw et al. 1991; Mawby et al. 1999; Hand et al. 1999; Buick et al. 2001). Onset of the Benambran Orogeny during the late Darriwilian has also been related to early deformation, with an eastward tectonic transport direction affecting Cambrian and pre-middle Ordovician Mathinna Group rocks occurring on the north coast of Tasmania. Whereas, Ordovician conglomerate overlying these units and younger Mathinna Group rocks farther east are unaffected by this deformation (Reed 2000).

The cause of the Benambran Orogeny is not known, however, Powell (1983, 1984) suggested that the onset was related to a change from decoupled Marianas-type subduction to coupled Chilean-type subduction occurring at the end of the Ordovician, resulting in a change in the convergence vector that in turn led to the formation of both the Omeo-Wagga Metamorphic Belt following the cessation of back-arc spreading and early east-west folds. VandenBerg et al. (2000) related the initiation of the Benambran Orogeny to oblique convergence on the plate margin east of and outboard of the Lachlan Fold Belt, initiating the Baragwanath Transform and southward movement of the Thomson Fold Belt over the northern extent of the Lachlan Fold Belt along the Olepoloko Fault. Glen and Walshe (1999) considered the Lachlan Transverse Zone to be an eastward extension of the Olepoloko Fault and
related Early Silurian east-west structures within the Bathurst and Dubbo 1:250 000 geological sheets to south over north thrusting along the Lachlan Transverse Zone.

Herein, the onset of the Benambran Orogeny is considered to have occurred during the Darriwilian. It is considered responsible for uplift between the Eastern and Western Belts of the Lachlan Fold Belt (Figure 1.2), forming a barrier to the sediment supplied to the turbidite fan in the Eastern Belt. The onset of Benambran deformation is also considered to have resulted in uplift and the subsequent widespread stratigraphic hiatus recognised within the volcanic arc and forearc regions of the Eastern Belt. These events rapidly followed rifting of the original island chain (Junee-Narromine and western Molong Volcanic Belts), contemporaneous formation of an intra-arc basin and changes in the chemistry of arc magmatism as well as an eastward shift in the volcanic front away from the volcanic high occurring within approximately 5-7 Ma.

The features recognised during the evolution of the Ordovician arc are analogous to those occurring in the Tonga-Fiji region following collision of the Ontong-Java plateau with the subduction trench. The collision resulted in a shift in the geometry of the overriding Australian plate and subsequent rifting of the Vitiaz arc by an across strike transform structure. Based on similarities in the evolution of the two arc systems, and recognition of those features characteristic of tectonic collisions within ancient arc-related successions, the onset of the Benambran Orogeny during the Darriwilian is interpreted as related to the collision of an oceanic plateau or ridge with the Ordovician subduction zone. This in turn is thought to have resulted in the reactivation by strike-slip movement of across-strike, northwest to southeast lineaments, principally the Lachlan Transfer Zone, recognised within the Lachlan Fold Belt, subsequently resulting in the rifting, breakup and ultimate cessation of Ordovician magmatism.

Glen et al. (1998) postulated that a hiatus in volcanism occurred in the Middle to Late Ordovician following subduction of a buoyant seamount or inactive ridge. The buoyancy and topography resulted
(a) Bendigonian

(b) Early Darriwilian

(c) Late Darriwilian/Gisbornian

(d) Eastonian
Figure 6.4 Proposed model of the tectonic evolution of the Ordovician oceanic island arc and forearc. (a) Terrigenous quartz turbidites deposition in a marginal sea separates an active oceanic island arc from the Gondwanan continent. Pelagic sediments are deposited outboard of the arc. (b) Subduction of an oceanic ridge or plateau occurs resulting in the onset of the Early Benambran event causes uplift of the arc and the formation of the large Bendigonian to Darriwilian stratigraphic hiatus across both lithological provinces in the north of the Eastern Belt of the Lachlan Fold Belt. Island arc volcanism occurs with the Junee-Narromine and western portion of the Molong Volcanic Belts comprising one island chain. By the Darriwilian, the continental margin submarine fan has extended to the northeast, with quartz turbidites deposited in the forearc basin. (c) Rifting of the arc forms an intra-arc basin in the Cowra trough with eruption of MORB interbedded with chert. The Molong and Junee-Narromine Volcanic Belts flank the intra-arc basin to the east and west respectively. Volcanism in these belts is characterised by the onset of high-K to shoshonitic lavas. Uplift occurs west of the marginal sea cutting off the sediment supply to the turbidite fan in the Eastern Belt, resulting in the widespread deposition of chert and thin-bedded clastic units. During the late Gisbornian, the volcanic front moves eastwards to the western part of the Molong Volcanic Belt, with the Junee-Narromine Volcanic Belt affected by a hiatus in volcanism, limestone deposition and subsidence of the arc platform. (d) The volcanic front shifted farther east resulting in seamount volcanism extending into the western part of forearc basin, contemporaneous with eastward extension of a volcaniclastic apron over units of the Quartzose Sedimentary Province. Rifting in the intra-arc basin has ceased and is followed by deposition of terrigenous sediments along the rifted basin. Erosion, subsidence and carbonate deposition dominate the topographic highs of the Junee-Narromine and western Molong Volcanic Belts. (e) Shoshonitic volcanism recommenced on the topographic highs of both the Junee-Narromine and western Molong Volcanic Belts and continued in the eastern Molong Volcanic Belt. Volcanism and associated plutonism ceased in the Junee-Narromine Volcanic Belt in the early Bolindian, and was succeeded by subsidence and deposition of marine siltstone and shale. Volcaniclastic deposition continued on the flanks of the Molong Volcanic Belt and in the Rockley-Gulgong Volcanic Belt throughout the Bolindian.
in a shallowing of subduction which caused a temporary halt to volcanism, as well as causing uplift and erosion of the older volcanic rocks as it passed underneath the arc, followed by subsidence, deposition of limestone and local turbidites then by the resumption of normal arc volcanism. Glen et al. (1998) also postulated that the arc has been rifted, however they suggested that this occurred as part of middle Palaeozoic extension, while the presence of quartz turbidites and chert between the rifted portions of the volcanic arc were the result of structural translocation from a backarc to apparent forearc setting. Although the collision aspects of model proposed by Glen et al (1998) are similar to those proposed here, the tectonic collision is thought to have occurred during the Darriwilian, and had more widespread effects within the Lachlan Fold Belt than previously recognised. The model of Glen et al. (1998) does not account for the across arc variation in the timing of volcanism and resultant stratigraphy nor the cessation of arc volcanism at the end of the Ordovician.

6.9.1 Where did the Collision occur?
The location of the collision with the subduction zone cannot be precisely known. However, the stratigraphic effects of the tectonic collision are greater in the northern part of the Lachlan Fold Belt than in Victoria where deposition is continuous and the Early to Middle Ordovician stratigraphic hiatus is not recognised in either the Benambra or Whitelaw Terranes. Therefore, the collision with the subduction trench is thought to have been east or northeast of the volcanic arc.

6.10 SUMMARY AND CONCLUSIONS
Ordovician rocks of the northeastern Lachlan Fold Belt can be separated into two distinct lithological provinces. The Quartzose Sedimentary Province comprises a fining upward sequence of craton-derived turbidites, together with pelagic chert and black shale, deposited on a terrigenous deep-marine outboard from Gondwana. The Molong Volcanic Province is characterised by mafic volcanic and volcaniclastic rocks with associated shallow marine limestone and deep-marine chert.
This thesis has focussed on the following. (a) Determining the stratigraphy within the Quartzose Sedimentary Province and comparing it with the regional stratigraphy delineated from farther south in the Eastern Belt where rocks of this province are well represented. (b) Delineating a stratigraphy within the deep-marine rocks of the Molong Volcanic Province, and determining the nature of the transition between the quartz turbidite succession (Quartzose Sedimentary Province) and the mafic volcanic belt (Molong Volcanic Province) in the Oberon-Taralga region. (c) Geochemically characterising the Ordovician mafic volcanic rocks of the Molong Volcanic Province in order to better understand the tectonic affinities of these rocks and the tectonic evolution of the northeastern Lachlan Fold Belt during the Ordovician. (d) Synthesising all the collected data for reconstruction and analysis of original basin of deposition of the northeastern Lachlan Fold Belt.

The major conclusions of this thesis are that the Oberon-Taralga region developed in the forearc region, outboard of an oceanic island arc. The regional stratigraphy recognised from the northern part of the Eastern Belt is similar to the regional stratigraphy of the Quartzose Sedimentary Province previously described from farther south. The stratigraphy of the Molong Volcanic Province rocks formed in a deep-marine basin adjacent to the volcanic arc and received pulses of volcaniclastic detritus from the arc interspersed with pelagic sedimentation. During the Middle Ordovician, rocks of the two lithological provinces were deposited distant from each other with those of the Quartzose Sedimentary Province considered to have been located further outboard of the arc than the deep-marine Molong Volcanic Province succession. However, the two provinces were juxtaposed by the middle Late Ordovician, when a rapid increase in volcaniclastic sedimentation due to collision of the subduction zone with an oceanic plateau resulted in the expansion of the volcaniclastic apron over units of the Quartzose Sedimentary Province.

Petrographic, geochemical and isotopic analyses of the volcanic arc succession show that these rocks have oceanic island arc characteristics and were derived from mantle sources with little or no crustal contamination. Volcanism is interpreted as having occurred synchronously with subduction, with the
arc region floored by oceanic crust; however, based on the shoshonitic characteristics of the magmatism, the arc is considered to have undergone a major perturbation during the late Darriwilian. Based on the stratigraphic sequence across the arc, a large Bendigonian to Darriwilian stratigraphic hiatus, followed by significant uplift with discontinuous limestone deposition and the shoshonitic character of the volcanic rocks, the arc is interpreted to have rifted and broken up during the Late Ordovician. This has been related to the onset of the Benambran deformation during the Darriwilian possibly following the collision of an oceanic plateau or ridge with the Ordovician subduction trench. Onset of the Benambran Orogeny also resulted in uplift between the Eastern and Western Belts, and is interpreted as the factor responsible for cutting off the sediment supply to the turbidite fan during the late Middle Ordovician resulting in widespread deposition of chert and black shale, and providing the source of renewed quartz turbidite sedimentation during the latest Ordovician and Early Silurian.

The Ordovician rocks of both provinces have been affected by three deformation events. D1 was a weak deformation resulting in early pre-cleavage folds. The timing of this event is poorly constrained; however it may be related to Late Ordovician “type-Benambran” deformation, corresponding to the formation of an angular unconformity in the Oberon and Thompson Creek areas. The second deformation event (D2) is the strongest event recognised in the study area,. It is characterised by north to northeast trending folds, an associated axial planar cleavage (S2) and related faults. Timing constraints on the age of this deformation are poor, and evidence from the different study areas may indicate it is diachronous across the Oberon-Taralga region. D3 was also a weak deformation event, caused by north-south shortening and resulted in gentle folding of D1 and D2 structures throughout the region. It is post-dated by intrusion of the Carboniferous granites in the Oberon area, indicating an Early Carboniferous age for this deformation in that area. The intrusion of granites associated with the Bathurst Batholith during the Middle Carboniferous marks the termination of tectonic development within the northeastern Lachlan Fold Belt.
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isotope geochemistry of Ordovician Igneous rocks from Goonumbla: A reconnaissance.


**Appendix A** Catalogue of rock samples referred to in this thesis.

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<th>UOW R. No.</th>
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Appendix C  Conodonts elements catalogued and held at the Australian Museum, Sydney. Figure numbers refer to figures in Chapter 3.

Sample R15975

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<td><em>Periodon aculeatus</em></td>
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### Sample R16473

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## Sample R16474

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# Appendix D Table of known Conodont faunas from the Lachlan Fold Belt of southeastern Australia

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<th>Sample No.</th>
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<th>Age</th>
<th>WFR/CFR</th>
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<td>Packham, 1967</td>
<td>Grey limestone and</td>
<td><em>Belodina compressa, Drepanodus sp.</em>, <em>Eobelodina fornicala, Falodus sp.</em>, <em>Microcoelodus sp.</em>, <em>Oistodus sp.</em>, <em>Ozarkodina sp.</em>, <em>Panderodus sp.</em>, <em>Phragmodus sp.</em></td>
<td>Gi-Ea</td>
<td>WFR</td>
<td>Map a, b or c</td>
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<td>G/II/2 Siliceous black shale</td>
<td><em>Panderodus sp.</em></td>
<td>Gi</td>
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<td>6</td>
<td>Nicoll, 1980</td>
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<td>Moulds of: <em>Pygodus serra, Periodon aculeatus aculeatus</em></td>
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<td>150 m below top Nelungaloo Volcanics</td>
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<td><em>Cordylocus proavus</em>, abundant paraconodont clusters and simple euconodonts</td>
<td>Cambrian-Early Ordovician</td>
<td>?</td>
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<td>Black chert</td>
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<td>poorly preserved paracondonts including Furnishina, Prooneotodus, ?Clavohamulus conform euconodonts</td>
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<td>75</td>
<td>VandenBerg and Stewart, 1992</td>
<td>168A</td>
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<td><em>?Drepanodus sp., Periodon cf. aculeatus, and a variety of simple cones</em></td>
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<td>184F Siliceous mudstone</td>
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<td>Trotter and Webby, 1994</td>
<td>Coppermine Creek Samples CM1-CM6 and CM8</td>
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<td>Geol. Survey of NSW, 1996</td>
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<td>Autochthonous limestone in the Oakdale Formation</td>
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<td>Limestone member of Sourges Shale</td>
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Palaeogeographic Significance of Ordovician Conodonts from the Lachlan Fold Belt, Southeastern Australia

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Conodont faunas from eight new localities within Ordovician rocks of the Lachlan Fold Belt are identified and discussed. Ordovician rocks of the Lachlan Fold Belt have been assigned to two distinct lithological provinces: the Quartzose Sedimentary Province and the Molong Volcanic Province. Three of the eight new faunas are from the deep-marine Numeralla Chert of the Quartzose Sedimentary Province and five are from chert members of the deep-marine Triangle Formation of the Molong Volcanic Province. One of the Triangle Formation samples contains the Bendigonian, Bel-2 (Early Ordovician) conodont Paracordylodus gracilis, the oldest fossil identified from deep-marine strata east of the Molong Volcanic Belt in the north-eastern Lachlan Fold Belt. All other new conodont faunas indicate either Darriwilian-Gisbornian (Middle-Late Ordovician) or generalised Ordovician ages.

Conodont faunas of both major Ordovician faunal realms (Warm Faunal Realm [WFR] and Cold Faunal Realm [CFR]) are found in the Lachlan Fold Belt. The distribution of these two faunal realms across the fold belt during the Early, Middle and Late Ordovician provides a framework for Ordovician palaeogeographic reconstructions of the Lachlan Fold Belt. Volcanism began in the Early Ordovician and continued episodically in at least three of the four volcanic belts in central-western New South Wales throughout the Ordovician; however, the volcanic edifices did not become emergent until the Darriwilian (late Middle Ordovician). Throughout the Early Ordovician and continuing to Darriwillian time, conodont faunas from most of the Lachlan Fold Belt exhibited CFR affinities. However, by Gisbornian-Eastonian time (middle Late Ordovician) many areas of the Molong Volcanic Province were emergent and partially surrounded by shallow marine limestone in which conodont faunas associated with the WFR became dominant. Offshore from the Volcanic Highs, volcaniclastic detritus formed widespread aprons which interfingered with siliceous black shale deposits in adjacent deep-marine basins. Conodont faunas from these Late Ordovician siliceous rocks continued to reflect CFR affinities.

Keywords: Ordovician, Lachlan Fold Belt, Conodonts, Palaeogeography, Stratigraphy

CONODONT BIOFACIES

Two major global Ordovician conodont biogeographic provinces have been recognised. These are: (1) the “North American Mid-Continent Province”, consisting of faunas recovered from shallow-marine rock associations from low to middle palaeolatitudes; and (2) the “Anglo-Scandinavian–Appalachian Province”, later known as the “North Atlantic Province,” containing faunas from deeper and/or cooler marine rocks that were deposited either off the shelf margins in an open-ocean environment or at higher palaeolatitudes (Sweet et al., 1959). After consideration of conodont data from beyond Laurentia, it became clear that these province names were inappropriate and the two provinces have been reinterpreted as two distinct faunal realms resulting from variations in water temperature, salinity

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and bathymetry, rather than as two geographically separate provinces (Tipnis, Chatterton and Ludvigsen, 1978; Sweet and Bergström, 1984; Miller, 1984). Herein the terms Warm Faunal Realm (WFR) and Cold Faunal Realm (CFR) are used (after Miller, 1984) instead of North American Mid-Continent Province and North Atlantic Province respectively.

Based on the understanding that provinciality within Ordovician conodonts results from variations in environmental conditions, a range of biofacies has been established within each of the faunal realms during the Middle and Late Ordovician (Sweet and Bergström, 1984). Within the WFR, six major biofacies were recognised during the Late Ordovician Velicuspis Chronozone. The Velicuspis Chronozone was defined by Sweet (1982), and includes rocks between the level of the first appearance of *Oulodus velicuspis* and that of the first appearance of *O. robustus*. In Australia, this chronozone falls within the Eastonian Ea3 graptolite zone. These biofacies represent a continuum from nearshore, shallow-marine biotopes to offshore, deeper marine biotopes; from shallowest to deepest they are: the *Aphelognathus–Oulodus* Biofacies, the *Pseudobelodina* Biofacies, the *Plectodina* Biofacies, the *Phragmodus undatus* Biofacies, the *Amorphognathus superbus–ordovicius* Biofacies and the *Dapsilodus mutatus–Periodon grandis* Biofacies (Sweet and Bergström, 1984).

Within the Late Ordovician CFR, data from Britain, Baltoscandia and the Mediterranean were used to define three distinct biofacies, the *Periodon–Dapsilodus–Pygodus and the Prioniodus (Baltoniodus)–Dapsilodus–Pygodus* Biofacies were recognised from the Middle Ordovician of Baltoscandia where they were well developed. These biofacies intergrade laterally and both include *Protopanderodus and Epilacognathus*. Sweet and Bergström (1984) also suggested these biofacies may have been replaced in Europe during the Late Ordovician by the HDS Biofacies.

**REGIONAL STRATIGRAPHY OF THE LACHLAN FOLD BELT**

Within the Lachlan Fold Belt of southeastern Australia, Ordovician rocks are assigned to two distinct lithological provinces (Figure 1; after Packham, 1969; VandenBerg and Stewart, 1992). The first, herein referred to as the Quartzose Sedimentary Province (quartz-rich greywacke and slate association of Packham, 1969; Quartzose Flysch Province of VandenBerg and Stewart, 1992), is an extensive deep-marine sedimentary province, characterised by Early and Middle Ordovician quartz-rich turbidites with interbedded chert horizons, that is thought to have formed as a Bengal-sized submarine turbidite fan (Fergusson and Coney, 1992). This is overlain by a widespread Late Ordovician graptolitic black shale facies known in many areas of the Lachlan Fold Belt as the Warbisco Shale (VandenBerg and Stewart, 1992). The second lithological province, referred to as the Molong Volcanic Province (after Fergusson and VandenBerg, 1990; Molong Volcanic Rise of Webby, 1976; Molong High of Packham 1987), is largely restricted to the eastern Lachlan Fold Belt, and comprises a northerly trending suite of ultramafic to intermediate volcanic and volcanioclastic rocks with fringing carbonate platforms, interpreted as part of an island arc separated from the Gondwanan margin by oceanic crust (Cas, 1983; Powell, 1983; Glen et al.,
FIGURE 1 Map showing the distribution of the two lithological provinces of the Lachlan Fold Belt. The map areas of Figures 2 and 3A are outlined. Modified from VandenBerg and Stewart (1992)
However, Wyborn (1992) emphasized the shoshonitic composition of these rocks, relating them to intraplate magmatism underlain by continental crust. Although the tectonic setting of these rocks remains in doubt, most researchers agree that the volcanic belt was separated from the Gondwanan continental margin by a marginal sea (Cas, 1983; Powell, 1983, Webby, 1992; Wyborn, 1992). Glen et al., (1998) subdivided the Molong Volcanic Province into four volcanic belts, the Junee–Narromine, Molong, Rockley–Gulgong and Kiandra Volcanic Belts. These names are referred to herein.

Various macrofaunas from limestone of the Molong Volcanic Province have been studied over many years (Etheridge, 1895, 1909; Hill, 1957; Ross, 1961; Pickett, 1970; Webby, 1969, 1971, 1972, 1973, 1975, 1977; Percival, 1992), but only scant attention has been paid to the conodont faunas of the Molong Volcanic Province. Packham (1967) and Pickett (1978, 1985) described small conodont faunas from these limestone units in order to establish age control. Only recently has the conodont fauna of this volcanic and limestone association been studied in any biostratigraphic detail (Savage, 1990; Trotter and Webby, 1995; Zhen and Webby, 1995). Over 140 conodont localities are now known from both lithological provinces of the Lachlan Fold Belt, although only a small number of elements are reported from the bulk of these localities (see Appendix 2).

**AIMS, METHODS AND TIMESCALE**

Conodont faunas described in this paper are from eight new localities, seven in the Oberon–Taralga district where both lithological provinces occur (Figure 2 and Figure 3B) and one in the Mudgee area, just east of that described by Stewart and Fergusson (1995; Figure 3A). The faunas are listed in Table I, together with details of their stratigraphic assignments, and ages. Catalogue details are provided in Appendix 1.

These new occurrences, along with most other conodont faunas known from the Lachlan Fold Belt, are placed in one of three age categories: Lancefieldian–Bendigonian, Darriwilian–Gisbornian or Gisbornian–Eastonian. All the Lachlan Fold Belt faunas are listed in Appendix 2, where the fauna of each locality has been assigned to the WFR or the CFR, or unknown,
FIGURE 2 Generalised regional geology of Oberon/Taralga/Crookwell district. Ordovician rocks of the two lithological provinces are differentiated. The map areas of Figures 3B, C, D and E are outlined.
FIGURE 3 Geologic maps and chert sample localities of each area studied within the Oberon/Taralga/Crookwell district. A. Bara Creek, east of Mudgee. Sample Lu5 and map from Fergusson and Colquhoun (1996). B. Samples collected from cherts southwest of Oberon. C. Chert sample R16474 collected from a quarry northeast of Isabella. D. Samples from chert bed at the junction of Silent and Oakey Creeks, Abercrombie National Park. E. Locality of sample R16472 from roadside outcrop beside Crookwell River.
using *Oulodus*, *Plectodina*, *Phragmodus* and *Belo­dina* as being indicative of the WFR and *Eoplacognathus*, *Pygodus*, *Prioniodus* and *Periodon* as being characteristic of the CFR (Sweet and Bergström, 1974; Nicoll, 1989). All the conodont sample localities are plotted on Figure 9A, showing the widespread distribution of conodont localities across the Lachlan Fold Belt. In Figure 9B-D, the location, faunal realm and lithology of each locality is shown, broken down into the three Ordovician time slices listed above. When the distributions of conodonts are considered together with their faunal realms and stratigraphy, the development of Ordovician volcanic islands and surrounding carbonate platforms offshore from the Gondwanan margin is better understood.

**Timescale**

The Ordovician timescale used is that produced for Australia and correlated with the Baltoscandia conodont zones, North American conodont zones and the Victorian conodont zones by Nicoll and Webby (1996), summarised here in Figure 5. A tripartite system of Lower, Middle and Upper Ordovician is used (following Webby 1998) so that the Lower Ordovician includes the Australian stages Lancefieldian, Bendigonian and Chewtonian, the Middle Ordovician consists of the Castlemainian, Yapeenian and Darriwilian stages, and the Upper Ordovician, the Gisbornian, Eastonian and Bolindian stages. The Middle and Late Ordovician faunas of the Lachlan Fold Belt are also tentatively correlated with
FIGURE 5 Correlation of Australian and British Ordovician stages and graptolite zones with conodont zones and ranges within Australia (after Nicoll and Webby, 1996)
the Middle and Late Ordovician Biofacies previously outlined (Sweet and Bergström, 1984).

**Methods**

Chert samples were selected from several localities in the general area between Oberon, Taralga and Crookwell, and one from Bara Creek, east of Mudgee (Figure 3). The study area has been affected by regional deformation which has produced abundant folds with a range of wavelengths. All the chert outcrops were strongly cleaved in at least one direction, with the cleavage direction quite variable across the study area. The bedding orientation was marked on the sample prior to its collection from the outcrop. Chert samples were collected away from any mesoscopic fold hinges and were as unfractured and unweathered as possible.

Samples were cut parallel to bedding planes, then mounted on glass slides and sectioned for later microscopic examination (approximately 10–15 sections of each sample). Each section was ground until transmitted light “just” penetrated through the chert. In translucent cherts this was approximately 0.8 – 1.0 mm thick, whereas darker more opaque cherts were ground thinner. Once light was able to penetrate through the section, it was examined under a microscope in both transmitted and reflected light. Conodont elements and radiolarians could often be seen at this stage but they were usually still embedded too deeply within the chert to be identified. Further “thinning” of the section was then required in order to see the fossil more clearly. Sections were prepared in the normal way by polishing on a rotary lap or glass plate using silicon carbide powder.

Use of this method produces only a partial fauna from a rock sample. Acid digestion (HF) of the chert was not used because many of the elements are preserved only as moulds, whereas other elements have been extended and fractured (e.g. AMF105961 – Figure 6E), so normal processing would have resulted in the recovery of very few if any complete elements. Because each fossil is viewed only in two dimensions, compound element genera could often be identified only to the generic level. Coniform conodonts cannot be reliably identified to the generic level, so only their presence is noted. Several taxa, such as *Pygodus*, have short enough ranges that samples can be dated with reasonable accuracy within the Ordovician stages.

The S, M and P notation used to classify the conodont elements is that of Sweet (1981), and all elements referred to herein have been catalogued (AMF Number), and are lodged at the Australian Museum in Sydney. “R” numbers relate to the University of Wollongong rock sample catalogue.

**NEW FOSSIL LOCALITIES**

Conodont elements have been identified from cherts of the Triangle Formation (Molong Volcanic Province) and the Numeralla Chert, the lowermost member of the Bendoc Group (Quartzose Sedimentary Province) (Figure 4). Each sample of chert has been allocated a rock sample or “R” number. The conodonts recovered from each chert sample are listed together with the stratigraphic assignment, grid reference and Locality Number (Appendix 2) for each sample in Table I.

**Sample R15975**

A dark grey to black chert (Budhang Chert Member of the Triangle Formation) collected from beside Brisbane Valley Creek at GR563587 (Figure 3B), contained a conodont fauna consisting of abundant *Paracordylodus gracilis* (Lindström) elements including S (AMF105956, AMF105955; Figure 6B, 6C), M (AMF105958, AMF105960; Figure 6D, 6G) and P (AMF105961, AMF105979; Figure 6E, 6F) elements, along with indeterminate paraconodont clusters and indeterminate coniform conodonts (AMF105957; Figure 6A).
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Grid Ref.</th>
<th>Stratigraphic Assignment</th>
<th>Conodont Fauna</th>
<th>Age Assigned</th>
<th>Locality No. (Appendix 2)</th>
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<td>MVP</td>
<td>Pygodus sp.,</td>
<td>Da3–Gi1</td>
<td>Locality 138</td>
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<td></td>
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<tr>
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<td>Triangle Fmn.</td>
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<tr>
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<td>Be1–Be2</td>
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<td>GR186901</td>
<td>QSP</td>
<td>?Pygodus sp.,</td>
<td>Da3–Gi1</td>
<td>Locality 136</td>
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<td>GR544592</td>
<td>MVP</td>
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<td></td>
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<td>Histiodella sp.,</td>
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<td></td>
<td></td>
<td></td>
<td>Periodon sp.</td>
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<td>Local 141</td>
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<td>Periodon aculeatus</td>
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**Age indicated**

The range of *P. gracilis* (Figure 5) is uppermost Tremadoc to early Arenig, top *Paltodus deltifer* zone to low *Oepikodus evae* zone of the Baltoscandia Conodont Zones (Sweet, 1988). Within the Victorian sequence the *P. deltifer* zone corresponds to the top of the *Cordylodus* spp. zone (La2) and the *O. evae* zone ranges from Bendigonian Be3 to Chewtonian. In Victoria *P. gracilis* occurs in association with Bendigonian Be1 graptolites at Dargo (Stewart and Fergusson, 1988) and in association with a fragmentary tetraraptid of La3–Be1 age on the Mornington Peninsula (Stewart, 1988). Therefore, based on the occurrence of *P. gracilis* in other parts of the Lachlan Fold Belt and the occurrence of paraconodonts, strata at this sample site are considered to be early Arenig (La3–Be2).
Faunal affinity

No genera previously outlined for determining provinciality were identified from this sample. *P. gracilis* is a characteristic species of the CFR in Europe (Landing, 1976), but it has also been recovered from the lower Deep Kill Shale of the Taconic allochthon, eastern New York (Landing, 1976), the Emanuel Formation in the Canning Basin, Western Australia (Nicoll et al., 1993) and from the Hensleigh Siltstone in New South Wales (Percival et al., 1999; Figure 4). All of these are associated with the WFR; however, they represent a deeper and/or colder water biofacies of that realm.

Sample R15973

Conodont elements including a single *Pygodus* sp. haddingodiform (P) element (AMF105973; Figure 7F) and several *Periolen* sp. elements (AMF105976; Figure 7G) were identified in sections from a large chert outcrop occurring at the intersection of Sewells Creek Road and Langs.
Road, GR577661 (Oberon 8830 1:100 000 topographic map; Figure 3B). Brown to honey coloured chert bands 4–10 cm thick are interbedded with pink-orange mudstone and belong to the Mozart Chert Member of the Triangle Formation (Figure 4). The bedding/cleavage intersection angle varies across the outcrop causing the chert to be quite fractured and friable in places.

Sample R16472

A single poorly preserved *Pygodus* sp. pygodontiform (P) element (AMF105974; Figure 7I) and more abundant *Periodon* sp. elements (AMF105975; Figure 7H) were recovered from a thin 2–3 cm chert band at the base of the Bendoc Group (Quartzose Sedimentary Province) along the Crookwell-Binda Road on the northern side of the Crookwell River bridge at GR186901 (Crookwell 8729 1:100 000 topographic map; Figure 3E).

Sample R16477

Elements of both *Pygodus* and *Periodon aculeatus* (Graves and Ellison) have also been identified from a grey-honey-coloured, thin-bedded chert (Numeralla Chert) collected from beside Bara Creek east of Mudgee (GR617912 Mudgee 1:50 000 Topographic Sheet; Figure 3A). *Pygodus* sp. elements identified from sections of this sample include three pygodontiform (P) elements, two superimposed (AMF105962; Figure 8E) and the other by itself (AMF105966; Figure 8C), one haddingodiform element (AMF105965; Figure 8F) and a single S? element (AMF105967; Figure 8G). Elements belonging to the *Periodon aculeatus* apparatus included several S elements (AMF105963, AMF105964; Figures 8A and 8B) and at least one M element (AMF105968; Figure 8D).

Age of samples R15973, R16472 and R16477

*Pygodus serra* (Hadding) and *P. anserinus* (Lamont and Lindström) are characteristic of the Baltoscandia *P. serra* and *P. anserinus* conodont zones respectively. *P. serra* occurs in Victoria and New South Wales associated with Da3–Da4 graptolites (Stewart and Fergusson, 1988) and was correlated with the Victorian Darriwilian latest Da3–Da4 zones by Webby and Nicoll (1989). *P. serra* was also recovered from the Wahringa Limestone, New South Wales, in association with representatives of Faunal Assemblages OT10 and OT11 (Percival et al., 1999). The base of the *P. anserinus* zone was originally defined as the level of first appearance of *P. anserinus* which overlaps with the occurrence of *P. serra* (see Bergström, 1971). In Australia, *P. anserinus* occurs with early Gisbornian *Nemagraptus gracilis* zone graptolites in Victoria (Stewart, 1988) and was correlated with the late Da4–G1 interval (Chart 2, Nicoll and Webby, 1996 and Figure 5). The occurrence of elements belonging to either *P. serra* or *P. anserinus* indicates that the age of these cherts is late Darriwilian to early Gisbornian (latest Da3–G1; equivalent to latest Llanvirn–early Caradoc; Figure 5) in age.

Faunal realm affinities of samples R15973, R16472 and R16477

All these samples contain *Periodon* and/or *Pygodus*, both of which are indicative of the CFR and, therefore, all are classified as part of that realm. They can also be tentatively correlated with the Middle Ordovician *Periodon–Dapsilodus–Pygodus* biofacies of Baltoscandia (Sweet and Bergström, 1984).

Samples R16475 and R16476

A conodont fauna was also identified in thick sections of the above samples (R16475, GR473170 and R16476, GR472169, both Taralga 8829 1:100 000 topographic map; Figure 3D) which were collected from 2.5 m of grey, thinly bedded and finely laminated Numeralla Chert (Figure 4). The fauna consisted of three *Cordyllum horridus* (Barnes and Poplawski) elements (AMF105970, AMF105971, and AMF105972; Fig-
FIGURE 8 Conodonts from sample R16477. A, B and D Peridont aculeatus: A. S element, AMF105963; B. S element, AMF105964; D. M element, AMF105968. C, E-G Pygodus sp.: C. pygodontiform (P) element, AMF105966; E. two superimposed pygodontiform (P) elements, AMF105962; F. haddingodiform (P) element, AMF105965; G. S? element, AMF105967. All scale bars = 250μm

ures 7B, 7C and 7A) and a single Histiodella sp. Pa element (AMF105969; Figure 7D), along with abundant Peridont sp. elements (AMF105977; Figure 7E).

Age of samples R16475 and R16476
The occurrence of Cordylodus horridus in Australia is rather limited. It was first recorded in the conodont fauna occurring with Da3 graptolites from Surprise Gully, Lancefield (Stewart, 1988). The species was named and described by Barnes and Poplawski (1973) from the Mystic Formation in Quebec, where C. horridus occurred in two samples with a lower Llanvirn (mid-Darriwilian) fauna including Belodella erecta (Rhodes and Dineley), Histiodella sinuosa (Graves and Ellison) and Peridont aculeatus. Nicoll (1992) discussed the evolutionary development of C. horridus as
part of the *C. angulatus* (Pander) lineage and noted that it occurs in the Llanvirn but is not very widely distributed. In contrast, Löfgren (1995, 1997) argued for *C. horridus* to be considered as part of the *Paroistodus* lineage after *C. horridus* elements were found in Arenig samples from Talubacken, central Sweden. A summary of *C. horridus* localities and associated faunas given by Löfgren (1995) indicates that the age range currently understood is from the *O. evae* zone of early-mid Arenig through to the early Llanvirn (Chewtonian to Darrriwilian, Da2/Da3; Figure 5).

McHargue (1982) discussed the evolution of *Histiodella* after recovering over 9000 elements from the Joins Formation in Oklahoma. He described *Histiodella* as a biostratigraphically valuable genus that evolved rapidly through several species during the late Early Ordovician. Species of *Histiodella* are distinguished on the basis of characteristics of denticulations on the upper margin of the bryantodontiform (Pa) element. McHargue (1982) noted that the sequence of species indicates a trend toward increased denticulation at progressively higher levels in the stratigraphic section. *H. altifrons* (Harris) occurs at the base and only exhibits smooth Pa element morphotypes, followed by *H. minutiserrata* (Mound), which exhibits both smooth and minutely serrated Pa elements, followed by *H. sinuosa* and *H. serrata* (Harris) which both have smooth, minutely serrate and denticulate Pa elements.

Only one *Histiodella* element (AMF105969; Figure 7D) was recovered from the chert in Oakey Creek. It is determined only to generic level; this single Pa element is denticulated and therefore must belong to *H. sinuosa* or a younger species. *Histiodella sinuosa* and *H. holodentata* are both zone fossils in the North American Conodont Zones, correlating with the mid-Da2 and the top of the Da2-Da3 Victorian graptolite zones respectively. Therefore a maximum mid-Darriwilian age (Da2) is indicated for sample R16475. Unfortunately the other elements recovered do not provide a minimum age.

**Faunal affinity**

The presence of *Periodon* elements indicates this fauna belongs to the CFR. *Cordylodus horridus*, *Histiodella* and *Periodon* have been found together in the Deep Kill Shale, where they were considered to be associated with dominantly (North Atlantic Province) CFR taxa (Landing, 1976), although *Histiodella* is cosmopolitan and commonly associated with WFR taxa. These two samples from Oakey Creek may represent a mixing of faunas from the WFR into a deep-water biofacies.

**Samples R16473 and R16474**

Chert from two other outcrops thought to belong to the Mozart Chert Member of the Triangle Formation (Figures 3B and 3C; see also Figure 4) was sampled and serially sectioned (Samples R16473, GR544592 and R16474, GR483407; both on Oberon 8830 1:100 000 topographic map; Figures 3B, 3C). These samples had been re-silicified, and only those elements figured were preserved (AMF105959, AMF105978; Figures 7J, 7K). Figure 7J shows a broken denticulated element (bottom) and a possible ?*Pygodus* sp. S element. Based on the stratigraphic position of these samples in the Triangle Formation, and without more convincing age diagnostic fossils, no age more definitive than Ordovician could be determined for these sites. The faunal affinity of these sites also could not be determined.

**FAUNAL AFFINITIES OF LACHLAN FOLD BELT CONODONT FAUNAS**

The new conodont localities reported in this paper are listed along with all other known conodont localities from the Lachlan Fold Belt of NSW and Victoria in Appendix 2, and plotted on Figure 9A. The faunas from these localities have then been divided into three age categories (Lancefieldian – Chewtonian; Darrriwilian –
Early Gisbornian; Gisbornian – Eastonian. These are plotted on Figures 9B–D, where the faunal affinities and the lithologies of the rock samples are also indicated.

Lancefieldian – Chewtonian (Figure 9B)

Lachlan Fold Belt conodont faunas of Lancefieldian–Chewtonian age are known only from deep-marine chert and shale, except for those from the Digger Island Formation in southern Victoria and the Hensleigh Siltstone in Central New South Wales. Although *Cordylodus rotundus* (Pander), *Onetodus* sp. and *Drepanodus* spp. have been reported from the Digger Island Formation, descriptions and illustrations of these conodonts have never been formally published (Kennedy, 1971). However, the trilobite fauna of this formation is considered to be Lancefieldian (Lai) (Jell, 1985) and, therefore, these conodonts together with those reported from Narooma (Bishoff and Prendergast, 1987) are the oldest reported from the Lachlan Fold Belt. Of the 21 conodont localities included in this map, all except Locality 143 are from siliceous, deep-marine rocks. Six samples of the 21 localities contain *Prioniodus* indicative of the CFR (Localities 42, 43, 58, 64, 65 and 143 in Appendix 2) and none of the samples contains any genera indicative of the WFR. A further seven samples (localities 35, 37, 58, 66, 67, 85 and 86) all contain *Oepikodus evae*, zone fossil of the Baltoscandia province (CFR) and primarily found associated with CFR taxa. However, faunas from localities 35, 74 and 143 also contain *Bergstroemognathus*, *Clavohamulus*, and *P. gracilis*. These genera are normally associated with the WFR and the faunal associations identified from the Lachlan Fold Belt sites closely resemble those recovered from the Deep Kill Shale of New York (Landing, 1976).

Darriwilian – Gisbornian (Figure 9C)

Of the conodont faunal localities shown on Figure 9C, most of the faunas (63 of 75 localities) are from siliceous, deep-marine rocks collected from both the Quartzose Sedimentary Province and the Molong Volcanic Province. Of these 63 localities, 59 include *Pygodus* and/or *Periodon*, indicative of the CFR. Two of these localities (63 and 137) along with locality 71 also include *Histiodella*, *Spinodus spinatus* (Dzik) and *Cordylyodus horridus*. These genera are normally associated with the WFR and the faunal associations identified from the Lachlan Fold Belt sites closely resemble those recovered from the Deep Kill Shale of New York (Landing, 1976).
FIGURE 9 A. All conodont localities within the Lachlan Fold Belt. The three plots, B-D represent different times during the Ordovician. B. Earliest Ordovician (Lancefieldian to Bendigonian). C. Late Middle Ordovician to early Late Ordovician (Darriwillian to Gisbornian). D. Late Ordovician (Gisbornian to Eastonian) time. Publications cited and faunas shown on the three figures are listed in Appendix 2.
The remaining 12 of the 75 faunas were recovered from limestone of the Molong Volcanic Province. Most of these faunas (10 of 12) contain the CFR indicator taxa Pygodus and/or Periodon along with associated CFR genera, Belodella, Wallasserodus, Dapsilodus and Protopanderodus (localities 68–70, 123, 124, 126 and 144 – 146). However, localities 144 – 146 also contain genera associated with the WFR (Plectodina, Phragmodus and Belodina), and therefore, are classified as being a mix of the two faunal realms. Belodina and Plectodina were also recovered from samples 142 and 148, but these samples did not contain any taxa associated with the CFR. Therefore, of the 12 faunas recovered from limestones of Darriwilian to Gisbornian age, seven are classified as being of the CFR, two are of the WFR and three exhibit affinities with both faunal realms.

Samples of deep-marine siliceous rocks from the Quartzose Sedimentary Province around Cobar also contain faunas indicative of both the WFR and the CFR (Pygodus, Periodon, Phragmodus and Pseudobelodina from localities 92 and 94–101: Iwata et al., 1995). Sweet and Bergström (1984) identified two biofacies, the Periodon–Dapsilodus–Pygodus biofacies, and the Prioniodus (Baltoniodus)–Dapsilodus–Pygodus biofacies, both of which are well developed in the Middle Ordovician of Baltoscandia and grade laterally into each other. The CFR localities identified in the Lachlan Fold Belt from both lithological provinces are most likely to be associated with either one or both of these two biofacies. This tentative association is based on the dominance of Middle to Late Ordovician species of Pygodus and Periodon occurring in the Lachlan Fold Belt samples. So far the associated genera Dapsilodus and Eoplacognathus have not been recognised from the Lachlan chert localities. However, Eoplacognathus is probably not able to be identified accurately in two-dimensions using the thick-section method.

The faunas with WFR affinities (localities 63, 71, 92, 94, 95, 98–101, 137, 142, 144–146 and 148) are from both limestone and chert. They occur in three broad geographical areas: around Cobar, Narooma and the Molong Volcanic Belt (Figure 9C). Two localities (63 and 71) are from Narooma on the south coast of NSW (Bishoff and Prendergast, 1987). Therefore, these conodont faunas may indicate this area remained bathymetrically shallower during the Early to Middle Ordovician than areas of the surrounding basin which were dominated by CFR taxa throughout this time.

The other faunas with WFR affinities can be divided into two groups. Those from the Cobar area are all from deep-marine rocks of the Quartzose Sedimentary Province, and the faunas from the Molong Volcanic Belt are all from limestone of the Molong Volcanic Province. These results indicate that the Molong Volcanic Belt was a positive topographic feature by the Darriwilian/Gisbornian stages, but remained separated from the Gondwanan continent. Conodonts from deep-marine sediments in the Cobar area indicate the presence of a deep-marine basin between the Molong Volcanic Belt and the Gondwanan continent further to the west.

Gisbornian – Eastonian (Figure 9D)

Conodont localities of this age are concentrated in the limestone units of the Molong Volcanic Province (36 localities of 47) where the taxa predominantly indicate affinities with the WFR. Occurrences of Belodina and Phragmodus are common, but also associated with these taxa at several localities is the CFR indicator Periodon grandis (Etherington) (see localities 88, 89, 110–112, 146, 147, 150 and 151) and Scabardella altipes subsp. B. of Zhen and Webby (1995), an indicator in the HDS biofacies of the Late Ordovician CFR (Sweet and Bergström, 1984).

The biofacies of the conodont assemblages recovered from the limestone were discussed by both Trotter and Webby (1995) and Zhen and Webby (1995). Trotter and Webby (1995)
noted that the conodont assemblage from the Eastonian Malongulli Formation (Cabonne Group; Figure 4) reflects derivation from both warm/shallow and cooler/deeper zones, occurring in rocks that probably accumulated as periplatform facies deposits at the outer margins of an island platform.

Zhen and Webby (1995) recognised variations in the biofacies throughout the Cliefden Caves Limestone Group in central-western New South Wales (Cabonne Group; Figure 4). Three conodont biofacies were recognised within the lowermost Fossil Hill Limestone, which they interpreted as reflecting environments ranging from near-shore restricted marine waters of the inner, island-fringing shelf, through mid-shelf shoals, to more offshore, open marine waters of the outer island shelf. Biofacies within the overlying Belubula and Vandon Limestones were not as clearly defined. However, a large number of forms with a CFR, North Atlantic aspect were noted. The presence of at least two, separate, deeper biofacies associations off the island slope, the first occupying well lit and oxygenated waters and the second, representing a deeper, probably cooler, water mass were suggested as the source of these CFR forms (Zhen and Webby, 1995).

Gisbornian-Eastonian conodont faunas identified from the deeper marine rocks are restricted to the eastern side of the Molong Volcanic Belt (localities 2–4, 16, 26–31 and 93). Black graptolitic shales are the most common deep-marine rocks of Late Ordovician age found in the Lachlan Fold Belt. Because graptolites are recovered with reasonable ease, conodonts are not usually sought. The lack of conodont localities within the deep-marine strata of this age is probably due more to a lack of study than a lack of occurrence. Over half of the collection (6 of 11 localities) contain Periodon grandis, indicative of the CFR. Three of these localities (16, 30 and 31) also contain the WFR form Phragmodus, and Plectodina furcata (Hinde) was identified from locality 30. These may indicate some mixing of faunal realms, with faunas from shallower biofacies being transported downslope after death, or faunas in which WFR forms lived higher in the water column than CFR forms, causing conodonts of both faunal realms to be deposited together on the ocean floor.

Overall, there is a greater WFR aspect to the faunas of this time period. However, this may reflect a sampling bias, as the few deep-marine faunas known indicate that the CFR was still dominating deeper areas of the basin.

### ORDOVICIAN PALEOGEOGRAPHY OF THE LACHLAN FOLD BELT

The causes of provincialism probably have both ecological and zoological aspects, but the factors of major importance for conodont provincialism are ecological, including a twofold differentiation of faunas based on water temperature and depth. Palaeogeographic reconstructions by Scotese and McKerrow (1990) and Powell and Li (1997) both place eastern Australia within 30° of the equator throughout the Ordovician. Therefore, variations in water temperature across the Lachlan Fold Belt would not have been greatly influenced by latitudinal variations, but rather by other factors including water depth, proximity to cold oceanic currents, upwelling zones and the Gondwanan continental margin. Changes in relative sea level would also have influenced the shallow waters of the island platform during the Late Ordovician.

### Early Ordovician

During the Early Ordovician, all Lachlan Fold Belt conodont faunas reflect a CFR aspect. This is consistent with stratigraphic and sedimentologic data which indicate a widespread quartz-turbidite fan prograding from WSW to ENE across the Lachlan Fold Belt throughout the Early Ordovician (VandenBerg and Stewart, 1992).

In the Molong Volcanic Belt, the Early Ordovician Mitchell Formation (Figure 4) consists of volcanioclastic breccia, conglomerate and tuff. It is
unfossiliferous, but it is conformably overlain by the predominantly fine-grained Hensleigh Siltstone (Figure 4) which contains allochthonous limestone clasts and Bendigonian graptolites and conodonts (Packham, 1969; Percival et al., 1999). A similar stratigraphy was reported from the Nelungaloo Volcanics, Junee-Narromine Volcanic Belt (Krynen et al., 1990; Percival et al., 1999), except that rather than allochthonous limestone lenses, graptolites indicating an La3 to Be1/2 age were recovered from the Yarrimbah Chert Member (Figure 4). These stratigraphic relationships indicate that volcanism had commenced in both the Junee-Narromine and Molong Volcanic Belts by Bendigonian time. Although much of the north-eastern Lachlan Fold Belt was dominated by deep-marine lithologies and conodonts of CFR affinities during the Early Ordovician, large limestone lenses within the Hensleigh Siltstone (up to 100 m in width; Ian Percival, pers comm.) indicate that parts of the Molong Volcanic Belt had reached shallower water depths during this time.

**Middle Ordovician**

Only sparse fossil evidence exists for early to middle Middle Ordovician (Castlemainian and Yapeenian) deposition in the Lachlan Fold Belt of New South Wales. Within the Quartzose Sedimentary Province, Darriwilian (Da2-Da3) graptolites have been found southeast of Braidwood (Jenkins, 1981) and both graptolites and conodonts have been identified from the Pittman Formation near Canberra (Öpik, 1958, p. 15; Nicoll, 1980). Conodont faunas indicating a Da2 to basal G1 age have also been identified from the Numeralla Chert (localities 63, 71, 137 and 141). Both the Numeralla Chert and the Warbisco Shale overlie the unfossiliferous quartz turbidites of the Adaminaby Group, and the Pittman Formation is chronologically and lithologically equivalent to both the Adaminaby and Bendoc Groups (Figure 4).

Graptolites indicate that the entire Middle Ordovician is preserved within the Bendigo–Balgarat Zone of the Lachlan Fold Belt. In this zone, it is represented by the Castlemaine Supergrupo (Figure 4), which is dominated by quartz-sandstone with minor lithic and feldspathic greywacke, interbedded with massive mudstone and black shale (VandenBerg and Stewart, 1992). These rocks were probably deposited in an environment on the turbidite fan different from the open-ocean mid-fan environment that is here inferred for the eastern Lachlan Fold Belt. It is therefore likely that the Middle Ordovician (Castlemainian–Yapeenian stages) of the eastern Lachlan Fold Belt is represented by the unfossiliferous quartz turbidites underlying the black shales and cherts of the Bendoc Group; however, there is not yet any direct evidence of this.

Within the Molong Volcanic Province, a similar lack of Castlemainian-Yapeenian faunas has been attributed to a hiatus, due to removal of shallow water deposits by erosion or lack of deposition (Percival et al., 1999). Prior to the Darriwilian, the only evidence of shallow-marine deposition in Molong Volcanic Province rocks is the occurrence of large allochthonous limestone lenses preserved in the Hensleigh Siltstone. As previously mentioned, these indicate that shallow-marine conditions existed around volcanic islands within the Molong Volcanic Belt by the Early Ordovician. However; there is no indication of shallow-marine to subaerial deposition of volcaniclastic detritus or deep-sea ash deposits in the Quartzose Sedimentary Province rocks. Therefore, even though volcanoes in the Molong Volcanic Belt were at shallow water depths during the Early Ordovician, proximal deposition of volcaniclastic detritus erupted from this volcanic centre continued until the Darriwilian. It is further inferred that the volcanic edifices that erupted the other Early Ordovician volcanic deposits (Nelungaloo Volcanics and Gidyen Volcanic Member) formed well below sea level, and continued to erupt episodically and grow towards sea level during the Early and Middle Ordovician.

The reappearance of fossils in the record during the Darriwilian coincides with significant
changes in the depositional environment. Within the Quartzose Sedimentary Province turbidite sedimentation decreased and was replaced by widespread chert and shale deposition.

On the highs of the Molong Volcanic Province, volcanism increased in volume and carbonate platform deposits became more widespread. Subaerially erupted volcanic and volcaniclastic detritus, along with the presence of oncolites in limestone lenses (Pickett, 1984), provide evidence that some of the volcanoes had become emergent by late Darriwilian, although the majority of the volcanic edifices probably remained submergent until the Late Ordovician. Conodont faunas continued to reflect a CFR aspect in the deep marine Triangle Formation (Fowler and Iwata, 1995), and conodonts of both CFR affinities and mixed CFR and WFR affinities occur in the shallow marine limestones accumulated around the volcanic islands (Pickett, 1984, 1985; Percival et al., 1999). The faunal affinities of this shallow, equatorial limestone may have been influenced by the upwelling of deep thermohaline currents. Jones et al. (1993) reported contourites from the Shoalhaven Gorge near Bungonia (Figure 1), which were attributed to contour currents moving northwards along the eastern margin of Gondwana from areas of subpolar downwelling. They related these contour currents to global climatic changes from the late Darriwilian to early Gisbornian.

Conodont faunas of deep marine rocks of the Quartzose Sedimentary Province also continued to exhibit a CFR affinity during the late Middle Ordovician except in the northwestern part of the Lachlan Fold Belt. Around Cobar (Figure 9C) conodonts identified in deep-marine chert and shale (Iwata et al., 1995) reflect a mix of both WFR and CFR affinities. WFR conodonts in these rocks may be the result of one or several of the following:

a. the Cobar region was probably closer to the shallow warm waters of the Gondwanan margin to the west; and/or

b. rising volcanic highs deflected the flow of cold ocean water away from the eastern margin of Gondwana, resulting in the existence of a marginal sea between the Gondwana margin and the chain of volcanic islands, where conodonts of various faunal affinities lived at different depths and temperatures in the water column, accumulating together in pelagic sediments after death.

Late Ordovician

By the early to middle Late Ordovician (Gisbornian to Eastonian), widespread accumulation of shallow marine limestone occurred in the waters of the Molong and Rockley–Gulgong Volcanic Belts. Zhen and Webby (1995) have been able to recognize most of the biofacies identified from the Late Ordovician WFR of North America within the Cliefden Caves Limestone Group of New South Wales (Cabonne Group; Figure 4). The presence of these shallow water biofacies together with widespread volcanic and volcaniclastic detritus indicates that many areas of the Molong Volcanic Belt were either emergent or shallow marine during the Gisbornian-Eastonian. Minor variations in water depth probably resulted from eustatic changes in the sea level and/or local tectonic events (Zhen and Webby, 1995).

Within the Quartzose Sedimentary Province widespread black shale and associated dysaerobic conditions dominated during the Late Ordovician. However, in deep-marine areas surrounding the volcanic highs, volcaniclastic deposits interdigitate with black shale (S. Murray, unpublished data). Only a few conodont faunas are known from deep marine rocks of this age and, although these still predominantly reflect the CFR, some faunas do include species and genera from the WFR (Localities 16, 30 and 31). These few specimens may either have been redeposited downslope from the volcanic islands (possibly as fecal deposits), or both WFR and CFR faunas lived at different levels in the
water column surrounding volcanic highs and were later deposited together in the pelagic sediments.

Acknowledgements

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References


Kennedy, D. J. (1971) Ordovician conodont faunas in southern Australia. ANZAAS 43rd Congress, Section C Abstracts, Brisbane, 43.


APPENDIX 1

Conodonts catalogued and held at the Australian Museum, Sydney.

Sample R15975

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Sample R16474

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**APPENDIX 2**

This appendix provides complete lists of conodont species so far recognized for each locality in the Lachlan Fold Belt. It includes locality numbers, literature references, sample numbers, a characterization of the lithology and the stratigraphic age of each locality. Each conodont fauna is assigned to the Warm Faunal Realm or the Cold Faunal Realm, or in some cases these are designated as “mixed” or “uncertain”. This Appendix is available from the authors and electronically at http://www.uow.edu.au/science/geosciences/research/researchdata.htm