CHAPTER ONE: INTRODUCTION

1.1 General Statement

Geophysical surveying is the application of methods of physics to the study of the solid Earth. It is the only branch of the earth sciences that can truly 'look' deep into the Earth's interior (Mussett & Khan, 2000). Geophysical surveying, although sometimes prone to major ambiguities or uncertainties of interpretation, provides a relative rapid and cost-effective means of deriving areally distributed information on subsurface geology (Kearey et al., 2002). The great advantage of geophysical surveying is that it can be used to make observations about the subsurface inaccessible to direct geological study, using measurements taken at the surface. Geophysics is especially important in the study of the thrust-fold-belts where the amount of fault offset is not obvious at the surface. The New England Fold Belt (NEFB) of eastern Australia is an area, which could benefit from the results of geophysical surveying. The subsurface structure of the NEFB is poorly known due to major faults, young cover rocks and a paucity of deep drill hole information.

During much of the Palaeozoic and Mesozoic, the New England Fold Belt was a convergent plate margin at the eastern edge of the Gondwana continent. Most authors infer the presence of a west-dipping subduction zone during much of this time (e.g. Leitch, 1974; Korsch, 1977; Cawood, 1982; Murray, 1997). The southern part of the New England Fold Belt (Fig 1.1) includes the Tamworth Belt representing a forearc basin, and the Tablelands complex corresponding to the accretionary wedge. The Belt is thrust westward over the Permian-Triassic Sydney-Gunnedah Basin, along the Mooki Fault that forms the western edge of the Tamworth Belt. To the east, the Peel Fault, which trends north-northwest, separates a less deformed Tamworth Belt in the west from the more deformed rocks of the Tablelands Complex. The ancient volcanic arc is now largely unexposed, either concealed beneath the Gunnedah Basin and/or the Tamworth Belt, or removed by erosion or strike-slip faulting or by a combination of these. Although numerous structural models of the southern part of the New England Fold Belt have been proposed (e.g. Leitch, 1975; Harrington and Korsch, 1985a, b; Murray et al., 1987; Aitchison and Flood, 1995), several key geological questions are still unanswered, including the subsurface geometry of the Mooki and Peel Faults, and the position of the volcanic arcs.

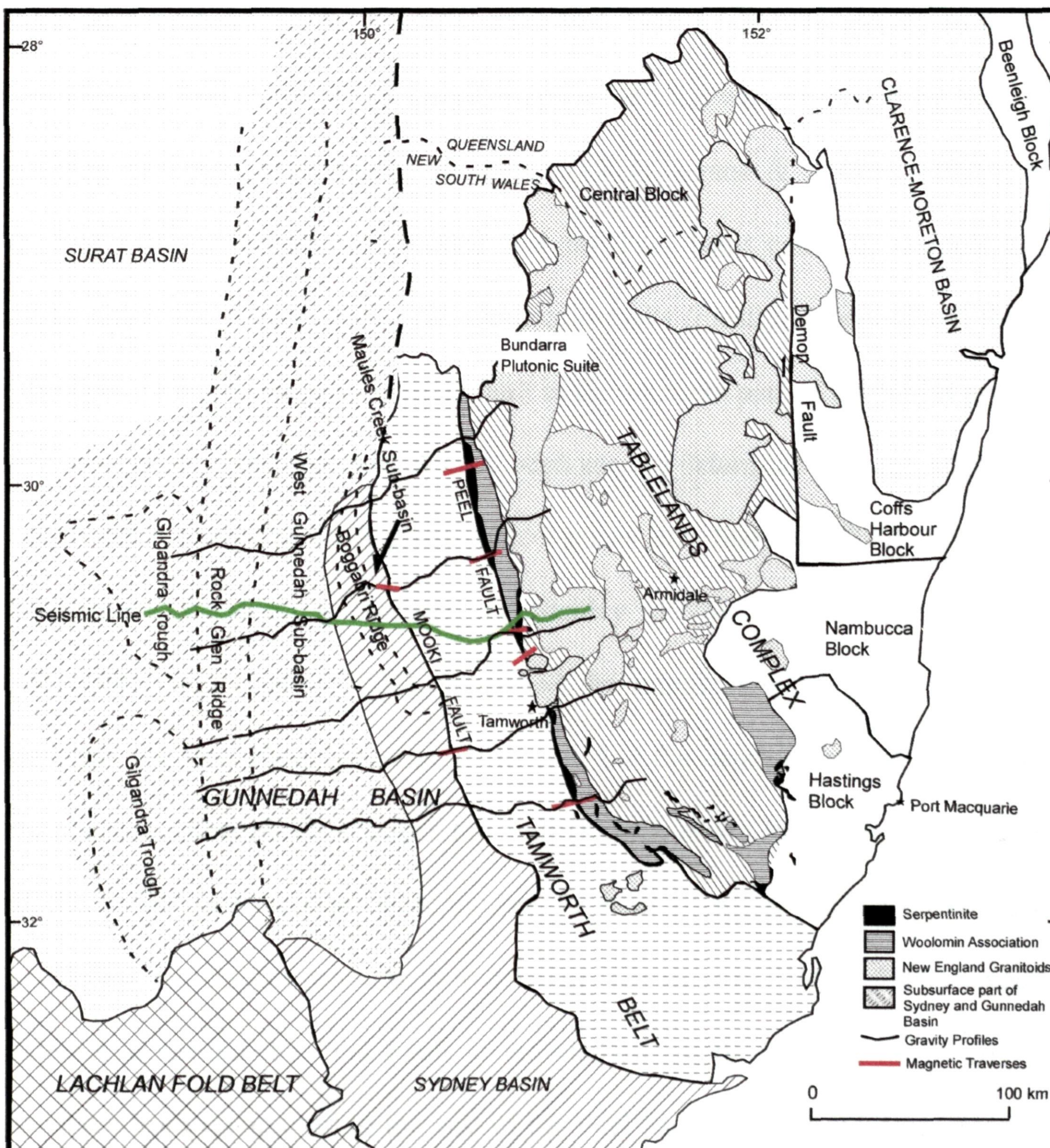

Figure 1.1 Generalised structural units of the Southern New England Fold Belt and the Gunnedah Basin. Also shown are locations of the gravity profiles and magnetic traverses conducted in this study and the seismic line BMR91-G01.

The structure of the Peel Fault (Fig 1.1) has been investigated by several authors over the past three decades (e.g. Scheibner & Glen, 1972; Ramsay & Stanley, 1976; Blake & Murchey, 1988a, b), but there is still little agreement about the detailed geometry of the subsurface. The classic pioneering geological studies of Benson (1913, 1914a, b, 1915a, b, 1917, 1918a, b, c, 1920), and the later studies of Voisey (1958), Crook (1963), Leitch (1974), and Blake & Murchey (1988a, b) showed the Peel fault as dipping steeply to the east. An eastward dipping Peel Fault with a concave-upward profile was suggested by Scheibner and Glen (1972), and was accepted by Runnegar (1974), Evans & Roberts (1980), and Cao (1994). Rod (1974), however, suggested that the Peel Fault could be vertical or steeply dipping to either the west or east. The ultramafic rocks (mostly serpentinized) that outcrop along the Peel Fault and associated splay fault in the southern part of the NEFB produce large magnetic anomalies. These anomalies have been used to determine both the angle of dip and down dip extent along the fault plane (e.g. Ramsay and Stanley, 1976). Interpretation of ground magnetic surveys by Ramsay and Stanley (1976), and later by Woods (1988), Edwards (1996) and Carter (2002), supported the initial view of Benson (1913), which indicated that the Peel Fault dips steeply to east, but the depth extension of the ultramafic rocks is not clear. Seismic data across the fault (Korsch et al., 1993a, b, 1997) revealed just beneath the surface outcrop position of the Peel Fault, at a depth of about 1 km (0.4 s TWT), a strong reflector, which dips moderately to the west. Korsch et al. (1993a, 1997) suggested that this reflector truncated the Peel Fault at the shallow depth. To date no outcrops of this reflector have been recognised to the east of the Peel Fault, and the published magnetic data (e.g. Ramsay and Stanley, 1976) do not generally support a truncation of the serpentinite bodies at such a shallow depth.

The Mooki Fault (Fig 1.1), which is the boundary between the Tamworth Belt and the Sydney-Gunnedah Basin, was initially inferred to have a dip of 40°-50° to the east (Carey, 1934a), but magnetic survey by Ramsay & Stanley (1976) indicated an easterly dip of 25°. The 25° dip has subsequently been supported by the new seismic data (Glen & Brown 1993; Korsch et al. 1993a, b, 1997). The source of the magnetic anomaly is still disputed. Ramsey and Stanley (1976) suggested the magnetic anomaly may be a composite of several anomalies caused in the south by a pluglike correlative of the Tertiary Warrigundi Intrusives and in the north by a 20 km long dyke of hawaiite presumed to be genetically related to the Tertiary Nandewar alkaline volcanic complex. Qureshi et al. (1990) argued that the anomaly might be due to the Early Permian Boggabri Volcanics in the Gunnedah Basin faulted against older rocks by the Mooki Thrust. Another suggestion for the source of these large positive anomalies may be the contrast between higher magnetic susceptibilities of the Currabubula Formation and Werrie Basalt to those of the Permian and Triassic sediments (Sydney-Gunnedah basin) to the west (Scheibner and Webster, 1982; Greentree and Flood, 1999). These opposing concepts could conceivably imply multiple sources for Mooki Fault magnetic anomalies.

The Palaeozoic volcanic arcs, produced over the west dipping subduction zone, are now largely missing except for the limited outcrops of Early-Middle Devonian, Early and Late Carboniferous volcanics (Cawood and Flood, 1989; Liang, 1989, 1991; R. H. Flood, Personal communication). The arc was inferred to lay to the west of the forearc accumulation (Tamworth Belt) during much of the Palaeozoic period (Leitch, 1974; Cawood, 1982), and is possibly at least partly concealed beneath the Sydney-Gunnedah Basin and/or Tamworth Belt, but much may have been removed by erosion or strike-slip faulting or by a combination of these (Leitch, 1974; McPhie, 1987; Buck, 1988; Korsch et al., 1997). The longitudinal gravity high within the Sydney-Gunnedah-Bowen Basin, referred to as the Meandarra Gravity Ridge, has been linked to the buried volcanic arc (e.g. Day et al., 1978), but has more recently been interpreted as volcanics of a rift origin developed after the arc had ceased or migrated to east.

The gravity anomaly over the Tamworth Belt (Namoi Gravity High) has been connected with the ancient volcanic arc (Murray et al., 1989) and the folded Devonian rocks of the Tamworth Belt (Bramall and Qureshi, 1984). The position(s) of the ancient volcanic arc is presently still unclear.

1.2 Aims and Scope of This Study

The main purpose of this study is to use geophysical methods (new gravity and magnetic surveys and previous seismic data to construct a 2D subsurface geometry across the Gunnedah Basin and Tamworth Belt, and thus better understand the structure of the upper crust and the tectonic development of the New England Fold Belt. Practical focus has been placed on the geometries of both the Mooki and Peel faults, and sources of the magnetic anomalies along them, the source and their subsurface geometries of the Meandarra Gravity Ridge and Namoi Gravity High.

New ground magnetic surveys were undertaken to better define the magnetic anomalies along the Peel and Mooki faults. Magnetic surveys comprised six lines across the Peel Fault and two across the Mooki Fault. Ground magnetic data is compared with extracted lines from the regional magnetic dataset from the Department of Mineral Resources, NSW, to examine the reliability of the extracted data. Both ground magnetic data and extracted lines are modelled to constrain the dip and depth of the Mooki and Peel faults. Serpentinite along the Peel Fault has been sampled to determine the contribution of the remanent magnetisation to the anomalies in order to improve the magnetic modelling over the Peel Fault. The possible sources of the magnetic anomaly over the Mooki fault will be reviewed.

The seismic data from Geosciences Australia was used to interpret and compare the results from the new ground magnetic data. The seismic data was particularly used in modelling of the gravity data in order to best constrain the subsurface geometry of the Peel

Fault, to answer the disputations of the previous geophysical surveys, i.e. it is or is not truncated at shallow depth (Korsch, 1993a, b, 1997), and to constrain the gravity modelling.

Five gravity survey profiles across the Gunnedah Basin and Tamworth Belt were conducted. Density measurements were made on fresh samples collected from both the field and fresh samples from drill core. On the basis of both the magnetic modelling and reinterpretation of the seismic data, together with collected density information of subsurface rocks, 2D gravity models of the subsurface shape of the Tamworth belt and Gunnedah Basin were constructed to test possible subsurface geometries. The gravity models were also used to infer the possible source of the Meandarra gravity ridge within the Gunnedah Basin and the Namoi Gravity High over the Tamworth Belt.

This study proposes a new tectonic model for the New England Fold Belt during the Late Devonian to Early Permian, on the basis of new geophysical evidence.

CHAPTER TWO: GEOLOGICAL SETTING OF THE NEW ENGLAND FOLD BELT

2.1 Introduction

The Tasman Orogenic Belt that constitutes most of Tasmania, Victoria, New South Wales and the eastern part of Queensland is divided into two major subdivisions: (i) the Lachlan-Thompson Fold Belt (LTFB) of Early to Late Palaeozoic age in the west, and (ii) the New England Fold Belt (NEFB) of Early Palaeozoic to Early Mesozoic age in the east (e.g. Leitch, 1974, 1975; Scheibner, 1987; Woodward, 1995). The two fold belts are separated by the marginal marine and terrestrial clastic and volcanic units of the earliest Permian to Triassic Sydney-Gunnedah-Bowen Basin (e.g. Scheibner, 1987) (Figure 2.1).

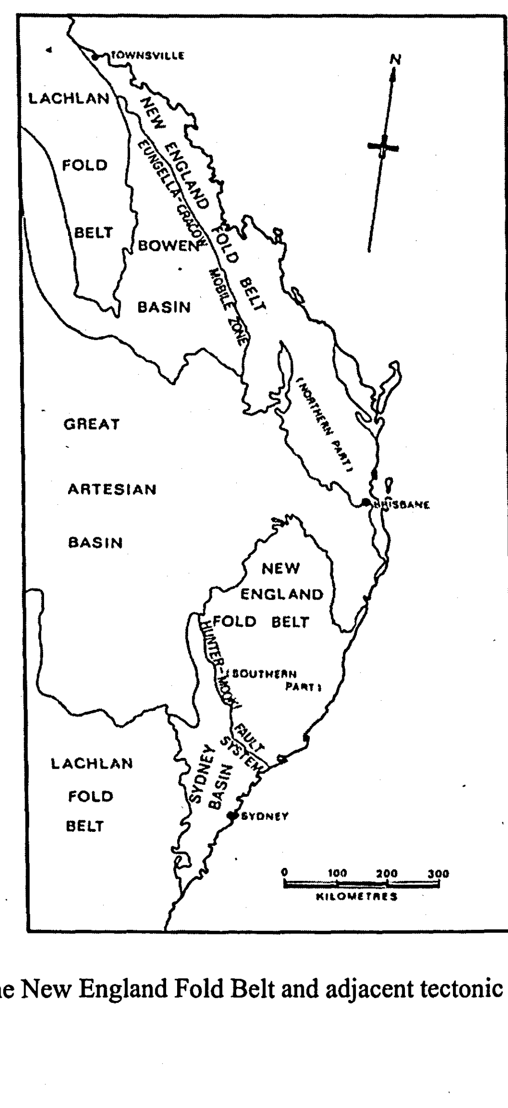

Figure 2.1 The New England Fold Belt and adjacent tectonic units. (After Leitch, 1974)

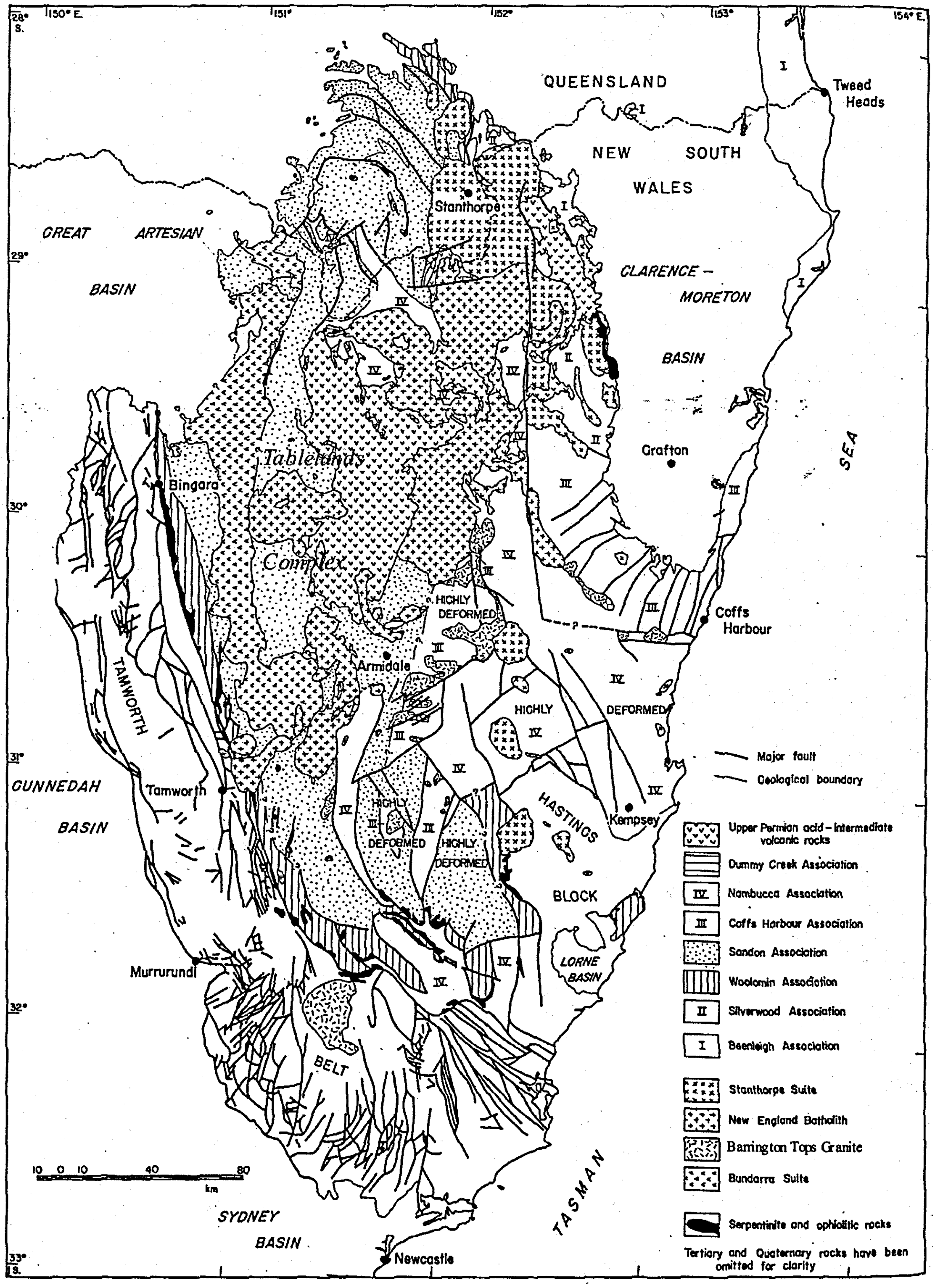

Figure 2.2 Major structural elements of the Southern New England Fold Belt. (After Korsch, 1977)

The structure of the Peel Fault (Fig 1.1) has been investigated by several authors over the past three decades (e.g. Scheibner & Glen, 1972; Ramsay & Stanley, 1976; Blake & Murchey, 1988a, b), but there is still little agreement about the detailed geometry of the subsurface. The classic pioneering geological studies of Benson (1913, 1914a, b, 1915a, b, 1917, 1918a, b, c, 1920), and the later studies of Voisey (1958), Crook (1963), Leitch (1974), and Blake & Murchey (1988a, b) showed the Peel fault as dipping steeply to the east. An eastward dipping Peel Fault with a concave-upward profile was suggested by Scheibner and Glen (1972), and was accepted by Runnegar (1974), Evans & Roberts (1980), and Cao (1994). Rod (1974), however, suggested that the Peel Fault could be vertical or steeply dipping to either the west or east. The ultramafic rocks (mostly serpentinized) that outcrop along the Peel Fault and associated splay fault in the southern part of the NEFB produce large magnetic anomalies. These anomalies have been used to determine both the angle of dip and down dip extent along the fault plane (e.g. Ramsay and Stanley, 1976). Interpretation of ground magnetic surveys by Ramsay and Stanley (1976), and later by Woods (1988), Edwards (1996) and Carter (2002), supported the initial view of Benson (1913), which indicated that the Peel Fault dips steeply to east, but the depth extension of the ultramafic rocks is not clear. Seismic data across the fault (Korsch et al., 1993a, b, 1997) revealed just beneath the surface outcrop position of the Peel Fault, at a depth of about 1 km (0.4 s TWT), a strong reflector, which dips moderately to the west. Korsch et al. (1993a, 1997) suggested that this reflector truncated the Peel Fault at the shallow depth. To date no outcrops of this reflector have been recognised to the east of the Peel Fault, and the published magnetic data (e.g. Ramsay and Stanley, 1976) do not generally support a truncation of the serpentinite bodies at such a shallow depth.

The Mooki Fault (Fig 1.1), which is the boundary between the Tamworth Belt and the Sydney-Gunnedah Basin, was initially inferred to have a dip of 40°-50° to the east (Carey, Murray, 1987). The Tablelands Complex consists largely of Devonian and Carboniferous deep marine sedimentary rocks, including a metabasalt-chert-argillite association interpreted as part of a Middle-Late Palaeozoic subduction-accretion complex (Cawood 1982; Fergusson, 1984), that formed between the forearc basin and the trench further east. A subduction related volcanic chain is inferred to lie west of the Tamworth Belt palaeogeographically during much of the Palaeozoic, on the basis of palaeocurrents from volcaniclastic sequences of the Tamworth Belt (e.g. Leitch, 1974; Day et al., 1978; McPhie 1987; Buck, 1988). The presence of Late Carboniferous silicic lavas in the Tulcumba Ridge (Liang, 1989; 1991), and Early to Middle Devonian andesite volcanic rocks west of the Peel Fault (Cawood and Flood, 1989), indicates a shift in both the position and geochemical character of the magmatic arc during the Palaeozoic (Cawood and Flood, 1989). The Late Devonian and Carboniferous arcs are generally buried under the present Sydney-Gunnedah-Bowen Basin (Leitch, 1975; Buck, 1988; Korsch et al. 1997) or they may have been overridden by the thrusts of the Tamworth Belt (Murray, 1989; Woodward, 1995).

The accretionary wedge deposits and possible accreted terranes of the southern part of the New England Fold Belt are intruded by the Late Carboniferous to Early Triassic New England Batholith (Shaw and Flood, 1981).

The Peel Fault that faults both Late Carboniferous and Early Permian rocks (Chappell, 1961; Crook, 1963) has been inferred to have both reverse and sinistral strike-slip displacements (Scheibner 1976; Corbett, 1976; Shaw & Flood 1981; Offler & Williams 1987 Aitchison and Flood, 1992). It is stitched by the Moonbi Adamellite of the New England Batholith dated by Chappell (1978) at 250 Ma.

2.2 Tamworth Belt

The Tamworth Belt consists of a succession of mainly forearc and volcanic arc rocks ranging in age from Middle - Late Cambrian to Early Permian (e.g. Leitch, 1974, Korsch, 1977; Cawood, 1980; Murray, 1987). Volcaniclastic sediments in the Woolomin area 30 km south of Tamworth, with Middle or early Late Cambrian limestone at the base and Ordovician limestone clasts towards the top (Cawood 1976, 1980, 1983), form the oldest forearc succession rocks of the Tamworth Belt. These early Palaeozoic rocks that consist of siltstone, sandstone, conglomerate, and tuff derived from a western source and were deposited in a submarine-fan complex (e.g. Cawood and Leitch, 1985), are bounded by the Peel Fault to the east and Devonian strata of the Tamworth Group to the west. Leitch (1974) has given a very complete summary of the lithostratigraphy of the Tamworth Belt, which is listed in Table 2.1.

The Tamworth Belt that is folded into a series of elongated, commonly doubly-plunging anticlines and synclines associated with numerous faults (Korsch, 1977), was referred to as the "western belt of folds and thrusts" by Voisey (1959). Woodward (1995, p110) wrote that, 'the Tamworth Belt shows all the obvious characteristics of a fold thrust belt, (i) long linear faults and folds in the western part repeating a limited stratigraphic interval; (ii) folded faults with windows in internal positions exposing strata typical of more external thrust sheets; and (iii) juxtaposition of rocks of similar ages but of dissimilar facies, or tectonic history, indicating major telescoping'.

The Tamworth Belt is intruded by the Inlet Monzonite north of Tamworth and by the Barrington Tops Granite in the Barrington Tops National Park area south of Nundle (Korsch, 1977). Emplacement of granites was dated at a K-Ar biotite age of 263 Ma (Cooper et.al, 1963), a Rb-Sr biotite age of 262 Ma (Hensel et al, 1985) and an average $^{207}Pb/^{206}Pb$ age of 281 Ma (Kimbrough et al, 1993) for the Barrington Tops Granite and a K-Ar age of 250 Ma for the Inlet Monzonite (Shaw and Flood, 1981).

Table 2.1 Major stratigraphic subdivisions of the Tamworth Belt.

(Compiled from Leitch, 1974; Woodward, 1995; Brown et al., 1992)

Age	Formation	Lithology	Thickness	Reference
Permian-Triassic		Sandstone, conglomerate, tuff, and mafic and silicic volcanics.	1000 m+	Woodward, 1995.
Late Carboniferous	Currabubula Formation	Paraconglomerate and orthoconglomerate with subordinate pebbly sandstone, mudstone, felsic ashflow and airfall tuff.	750 m+	Brown et al., 1992; Woodward, 1995.
Early Carboniferous	Merlewood Formation	Coarse, crossbedded feldspathic and lithic sandstone, minor conglomerate, mudstone, limestone	750 m	Brown et al., 1992; Woodward, 1995.
Early Carboniferous	Namoi Formation	Thinly bedded mudstone and siltstone with minor conglomerate.	1000 m	Brown et al., 1992; Woodward, 1995.
Early Carboniferous	Luton Formation	Lithic, felspathic and calcareous sandstones and siltstone interbedded with mudstones.	1200 m	Woodward, 1995 Leitch, 1974
Late Devonian	Mandowa Formation	Massive mudstone	400 m	Leitch, 1974
	Keepit Conglomerate	Conglomerate	130m	Leitch, 1974
Late Devonian	Baldwin Formation	Orthoconglomerate, turbidite sandstone, fine terrigenous sediments.	2800 m	Chappell, 1968 Crook, 1961b
Early-Middle Devonian	Tamworth Group	Radiolarian argillites, graywackes with coralline limestone lenses, graywackes, greywacke rudites with limestone.	3000 m	Crook, 1961a
Middle-Late Cambrian-Ordovician		Siltstone, sandstone, conglomerate, limestone and tuff.	1350m	Cawood, 1976, 1980

Glen and Brown (1993) presented a cross section interpretation of the Tamworth Belt near Manilla (Figure 2.3). The cross section was drawn along the BMR91-G01 seismic section, and is consistent with the surface geology of the belt and the upper crust interpretation of the seismic data collected by the BMR (Glen et al, 1993, Korsch et al, 1993a, b, 1997).

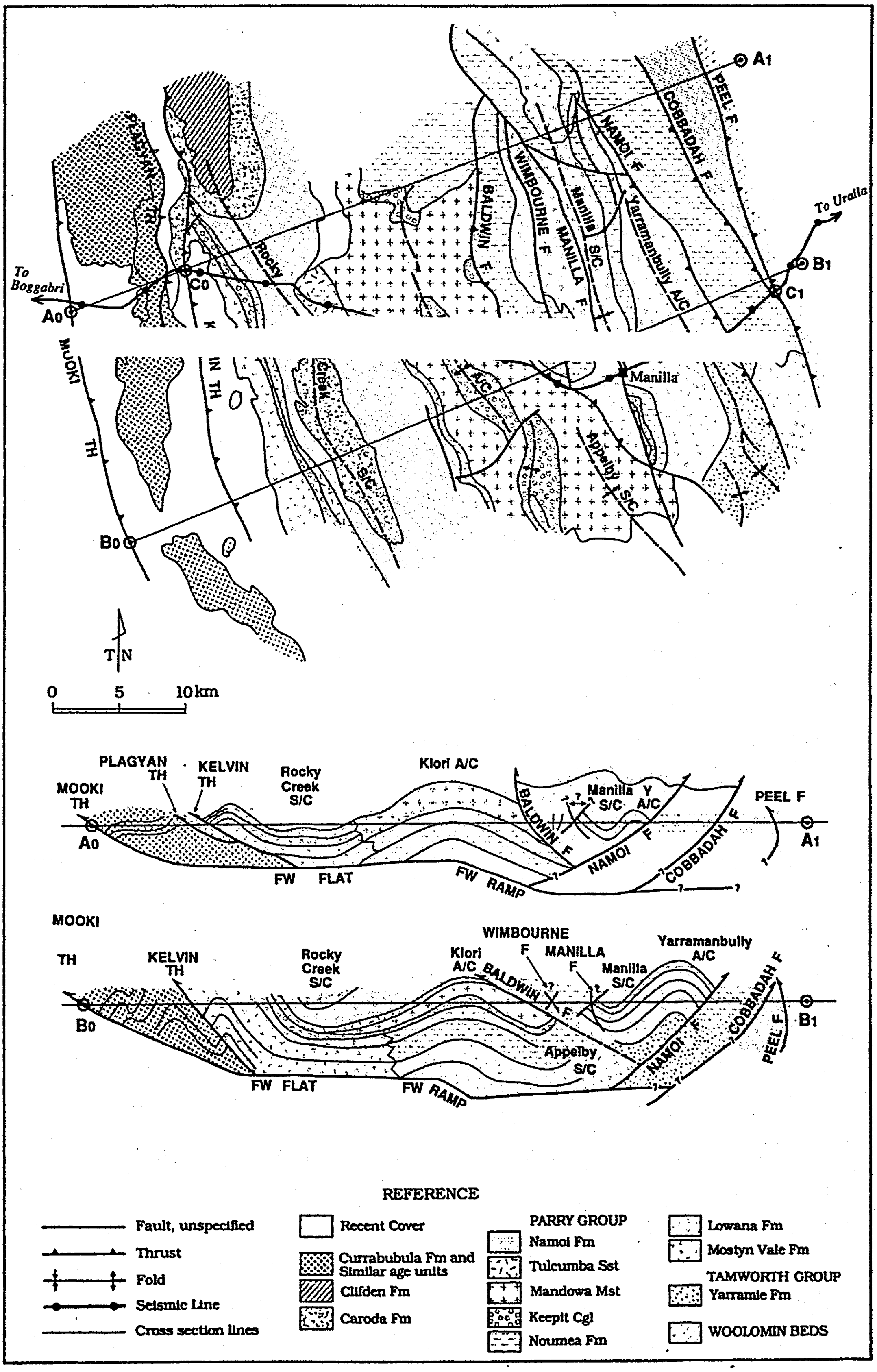

Figure 2.3 Cross Section of the Tamworth Belt near Manilla (After Glen and Brown, 1993).

2.3 The Hunter-Mooki Fault system

The Hunter-Mooki Fault System (HMFS) that forms the western boundary of the southern New England Fold Belt, was referred to as the "border thrust" by Voisey (1959). The HMFS thrusts Late Carboniferous and Early Permian rocks of the Tamworth Belt over Permian rocks of the Sydney-Gunnedah Basin (e.g. Carey, 1934a, b; Leitch, 1974; Murray 1987). The Mooki Fault was documented by Carey (1934a, b) from the west side of the Werrie Basin north of Quirindi. He noted the irregularity of the thrust surface and inferred that the faults dip east at 40°-50°. The continuation of the thrusts was mapped by Hanlon (1948a, b, 1950) in the Murrurundi-Temi and Narrabri areas. Liang (1991) concluded that overturning of the Currabubula Formation at Tulcumba Ridge to the north of the Mooki Fault may be the result of fault-propagation folding related to a thrust step-up angle of ~30° from a décollement (Figure 2.4). Magnetic anomaly interpretation, based on curve fitting using intrusive dyke models, have suggested an easterly dip of 25° for the Mooki Fault to the north of Tulcumba Ridge (Figure 2.5) (Ramsay & Stanley 1976). Interpretation of the BMR91-G01 seismic data supports this geometry (Figure 2.6) (Korsch et al 1993a, 1997; Glen & Brown 1993). It would seem that there is now good agreement that the Mooki Fault dips to east at an angle of 25-30°.

The continuation of the Mooki Fault in the Hunter Valley in the south is referred as the Hunter Fault, and was recognised as a major fracture by Osborne (1928a, b, 1929) and Raggatt (1929), who both suggested that the faults dip east at angles between 15° and 30°. Roberts and Engel (1987) suggested the fault dips to east at an angle of 13°-14° on the basis of NSW Department of Mineral Resources drilling (Muswellbrook East Drilling Program).

The Kelvin Fault which occurs to the east of the Mooki Fault on the Manilla map sheet has thrust the Late Devonian Mostyn Vale Formation westward over the late Carboniferous Currabubula formation. To the south, the two thrusts converge, suggesting that the Mooki fault is a diverging frontal splay fault in front of the main Kelvin fault (Korsch et al, 1993a, 1997). Korsch et al. (1993a, 1997), on the basis of the interpretation of seismic data, suggested that the Mooki-Kelvin Fault thrust westward over the Sydney-Gunnedah Basin at least 10 km. Glen and Brown (1993) suggested 25 km of displacement on the Mooki thrust. In a series of cross sections across the Tamworth belt, Woodward (1995) shows 45-58 km of displacement on the combined Mooki-Kelvin thrust system. The work of Woodward (1995) was questioned by Korsch et al. (1997) and Roberts et al. (2004) due to his assumption that the NEFB extended westward beneath the Lower Permian volcanic and sediments of the Gunnedah Basin.

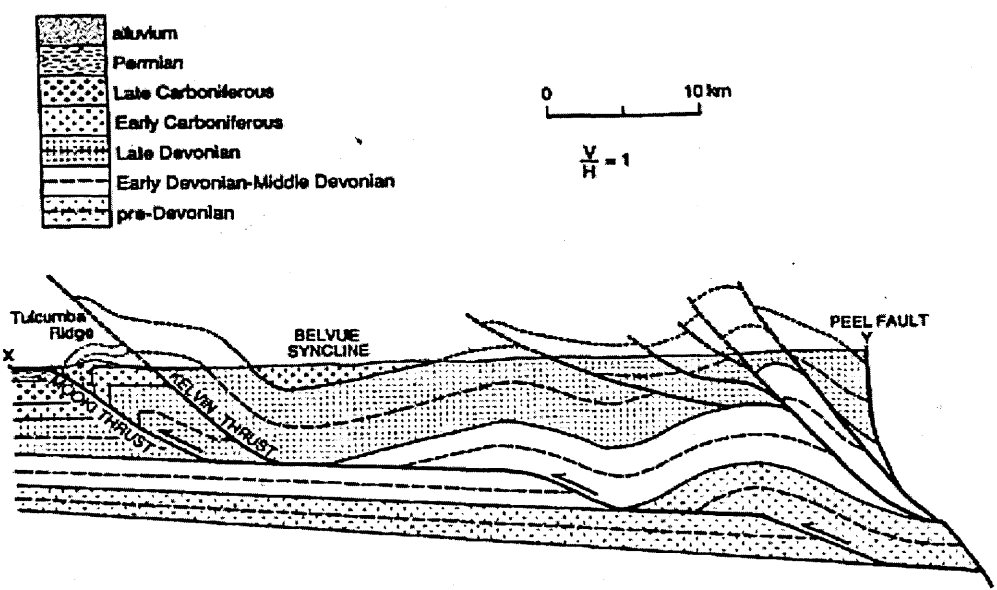

Figure 2.4 Cross-sections across the Tamworth Belt at Tulcumba Ridge, showing the geometry of folding spatially associated with the Mooki Fault. (After Liang, 1991).

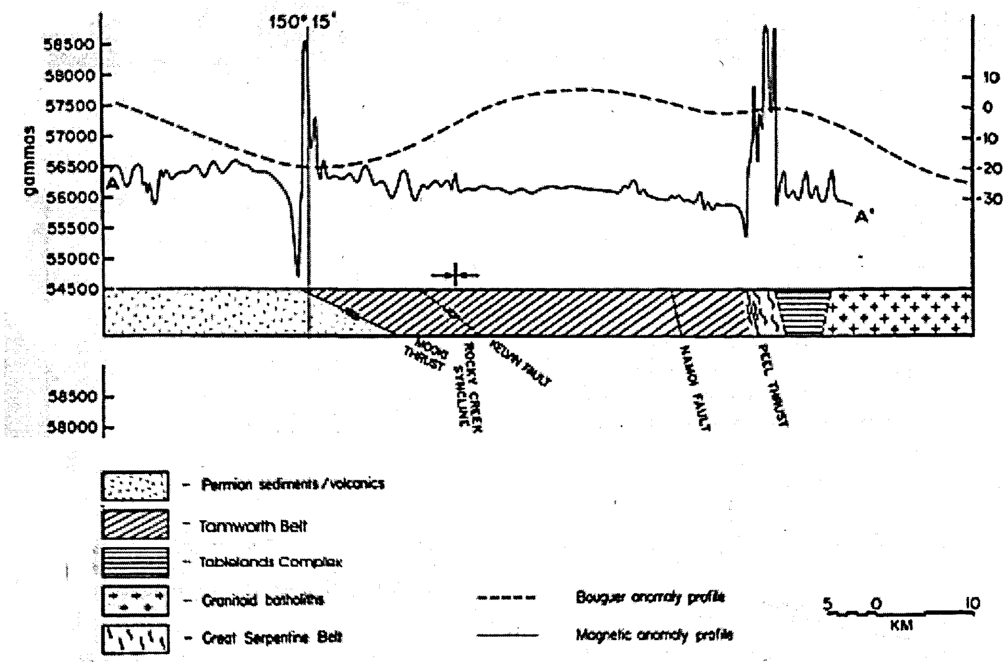

Fig 2.5 Subsurface geometry of the Mooki and Peel Faults inferred from ground magnetic data. (Adapted from Ramsay and Stanley, 1976)

2.4 The Peel Fault system

The Peel Fault, the present-day boundary between the Tamworth Belt (forearc basin) and the Tablelands Complex (accretionary wedge), has long been recognized as a fundamental geological discontinuity in eastern Australia. From near Warialda in the north, to the Liverpool Range in the south, the Peel Fault system outcrops as a north-northwest-trending structure with a straight trace (Benson, 1913; Voisey, 1939; Leitch, 1974, 1979). Further south, the fault swings towards the east and breaks up into a series of splay faults that constitute the Manning Fault system (Voisey, 1939). Serpentinite and lesser amounts of gabbroic and doleritic rocks occur along the fault system and immediately to the east (e.g. Benson, 1913; Leitch, 1979).

The Peel Fault is interpreted as dipping steeply to the east (Benson 1913, 1915; Crook 1963; Leitch 1974). Blake and Murchey (1988a, b) reported a steep dip to the east for the Peel Fault 35 km north of Manilla. In the Kootingal area Edwards (1996) used magnetic data to infer a dip of 70°E for the Peel Fault. Near Chrome Hill and Hanging Rock, the foliation in the schistose serpentinite along the fault dips at 70°E (Crook, 1963).

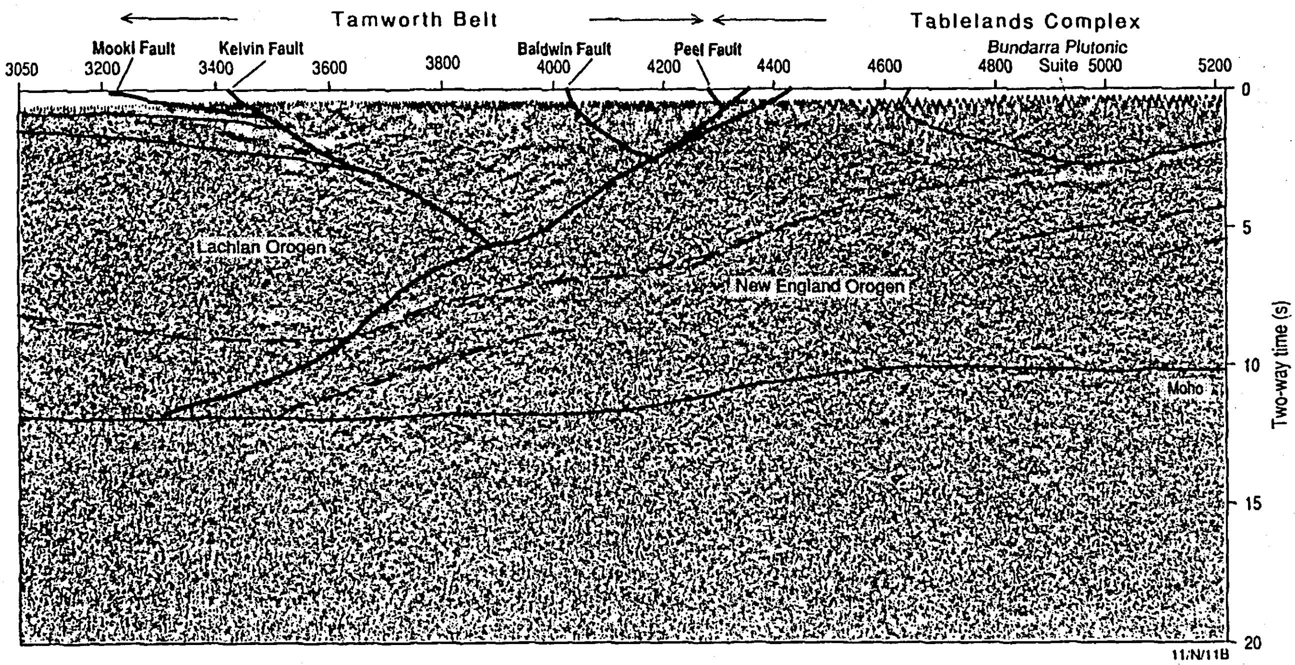

Figure 2.6 Subsurface geometry of the Tamworth Belt, including the Peel and Mooki faults as inferred from the BMR91-G01 seismic profile. (After Korsch, 1997)

The nature of the movement on the Peel Fault is not well constrained. Early interpretations by Benson (1913) suggested that it was a reverse fault with the more highly deformed Woolomin association thrust over the sediments of the Tamworth Belt to the west. Later researchers continued to consider it to be a steeply dipping reverse fault, based on the geological evidence (Voisey, 1959; Chappell, 1961). Crook (1963) suggested that strike-slip movement may also have been important, and later Corbett (1976) suggested a sinistral strike-slip movement that has also been generally accepted (e.g. Cawood, 1982; Korsch, 1982; Offler and Williams, 1987; Collins, 1991). The major strike slip movement is inferred to have occurred in the Permian time (e.g. Katz, 1986; Allan and Leitch, 1990; Collins, 1991; Aitchison and Flood, 1992). Scheibner and Glen (1972) revived the thrust model after they reviewed the nature and evolution of the Peel Fault. They believed that the Peel Fault formed as an "obduction zone" but had a dextral wrench movement of up to 150 km in Early Permian on the basis of tectonic analysis of blocks forming the NEFB.

The thrust model for the Peel Fault has been supported by Runnegar (1974), Scheibner (1973, 1976), Evans & Roberts (1980), Harrington & Korsch (1985a), and Cao (1994). Rod (1974), on the other hand, suggested that the fault could be vertical or steeply dipping to either the west or east. In general, the history of movement on the Peel Fault is still far from decided. The magnetic survey by Ramsay and Stanley (1976) indicated the Peel Fault has a dip of 65° to the east (Figure 2.5). These authors suggested the fault dipped easterly at this angle to a depth of at least 5 km, and perhaps 7.5km. This earlier view is contradicted by interpretations of deep seismic data that revealed a west-dipping structure seeming to truncate the east-dipping Peel Fault at a shallow depth (1 km) (Glen and Brown, 1993; Korsch et al, 1997; Figure 2.6).

The ultramafic rocks (mostly serpentinites) along the Peel Fault and adjacent parts of the Tablelands Complex formed the Great Serpentine Belt of NSW and were documented by the pioneering studies of Benson (1913, 1914a, b, 1915a, b, 1917, 1918a, b, c, 1920). The term "Great Serpentine Belt" has subsequently been restricted to the serpentinite masses lying adjacent to the Peel Fault (e.g. Leitch, 1979) and this usage has been kept for this study. The serpentinite masses have also been referred as the Weraerai terrane by Flood and Aitchison (1988). Spatially, the serpentinites concentrate along the Peel Fault and its associated splay faults, and extend some 300 km from near Warialda in north-eastern NSW to Nundle and then swing southeast to the coast (Voisey, 1959) with outcrop width varying from less than 10 m to approximately 10 km (Yang and Seccombe, 1996). Both massive and schistose serpentinites are recognised. Leitch (1979) points out that, some massive serpentinites retain pseudomorphs after pyroxene, and rarely relic pyroxene and olivine crystals. The schistose serpentinite is characterised by the presence of penetrative, anastomosing, commonly curved, shear surfaces from which relic textures and phrases are absent. The massive variety generally occurs as blocks within the schistose material giving rise to serpentinite breccias.

Benson (1926) concluded that the serpentinite was emplaced as intrusions along the Peel Fault. Thereafter, the extensive studies by Crook and Felton (1975), Cross (1983) and Rogers (1986) argued that the serpentinite is part of an ophiolite sequence representing ancient oceanic crust and upper mantle material, and must have been tectonically emplaced as the serpentinite rocks are little altered or metamorphosed. Scheibner and Glen (1972) proposed that the serpentinite represents slices of ocean floor on which the Woolomin flysch wedge was deposited, and that the ocean floor was formed during the Middle and Upper Silurian. The initial stage of the model, however, is still uncertain. Cawood (1982) suggested emplacement of the serpentinite involved cold-solid diapiric rising of mantle-derived serpentinite during the early Permian with the inclusion of rocks from different structural levels of the subduction complex with the serpentinite.

Aitchison et al (1992a) published a 530±6 Ma $^{206}Pb/^{238}U$ age for an included block of plagiogranite from a serpentinite body near Upper Bingara in the northern part of the serpentinite belt, and also ascribed this age to the serpentinite. Thereafter Aitchison and Ireland (1995, p20) suggested that, 'Weraerai terrane represents only portions of a Cambrian intra-oceanic arc rift sequence over which younger terranes..., may have been thrust westward as a series of thin-skinned nappes over older basement terranes during the process of their accretion to the margin of Gondwana'. Fukui et al. (1995) reported middle Ordovician (K-Ar ages of 465-480 Ma) high PT metamorphic rocks from serpentinite melange near Glenrock station and Pigna Barney River in the southern part of the serpentinite belt, and suggested that the serpentinite melange may be a remnants of the oceanic crust developed east of Gondwana and accreted to eastern Australia during the Devonian-Carboniferous.

Table 2.2. Stratigraphic units of the Tableland Complex (Compiled from Leitch, 1974; Korsch, 1977)

Age	Stratigraphic unit	Lithology	Deposition environment and reference.
Permian	Dummy Creek Association	Conglomerates, with minor sandstones and mudstones.	Shallow-water marine to terrestrial deposits. (Campbell, 1961, McKelvey and Gutsche, 1969; Korsch, 1977)
Early Permian	Nambucca Association	Diamictite mainly. Orthoconglomerates, greywackes, siltstones, mudstones and rarer acid to basic volcanic horizons and limestone members.	Deepwater marine mass movements (Leitch, 1974; Korsch, 1977)
Carboniferous	Coffs Harbour Association	Monotonous, thick sequences of greywacke, siltstone and argillite.	Turbidity current deposits (Korsch, 1971, 1977)
Carboniferous	Beenleigh Association	Greywackes, argillites, cherts, and rare basic lavas and conglomerates.	From continental shelf deposits to deepwater marine turbidites. (Fleming et al., 1974; Korsch, 1977)
Late Devonian-Early Carboniferous	Sandon Association	Greywackes, mudstones with minor cherts, jaspers, intermediate to basic volcanics and rare limestones and conglomerates.	Turbidity current deposits. Deepwater origin (Leitch, 1974; Korsch, 1977)
Early Devonian	Silverwood Association	Andesitic lavas, tuffs, agglomerates along with sandstones, mudstones, chert, and rare limestones.	Mass movement (Telford, 1972; Korsch, 1977)
Ordovician-Silurian	Woolomin Association	Chert, jasper, basic volcanic, rare sandstone, argillite and limestone lenses	Abyssal plain sediments (Chappell, 1961; Leitch, 1974; Korsch, 1977; Furey, 2000)

2.5 The Tablelands Complex and New England Batholith

The Tableland Complex of Korsch (1977), also known as zone B by Leitch (1974), is largely an accretionary wedge in the context of modern plate tectonic models (e.g. Leitch, 1974; Korsch, 1977; Murray et al., 1987; Roberts and Angel, 1987). Leitch (1974) recognised seven major lithological associations within the Palaeozoic sediments of the Tablelands Complex on the basis of lithological type. Korsch (1977) developed these lithological associations, defined a different areal distribution from that proposed by Leitch (1974) and proposed informal names (Table 2.2). The metasedimentary rocks of the Tableland Complex have suffered at least prehnite-pumpellyite grade of regional metamorphism (Korsch, 1977).

The accretionary wedge is intruded by granitoids, collectively termed the New England Batholith. The New England Batholith extends for approximately 340 km from south of Tamworth in the southern New England Fold Belt, to north of Stanthorpe in Queensland (Figure 2.2) (Chappell, 1978). It ranges between granite and gabbro in composition. Adamellite is the predominant rock type, granodiorite is common, and more mafic rocks are volumetrically insignificant (Leitch, 1974). Shaw & Flood (1981) grouped the New England Batholith into five suites and a group of leucoadamellites on basis of geochemical, mineralogical and isotopic characteristics. Two of five suites have S-type characteristics, whereas the remainder are I-type (Table 2.3).

The New England Batholith was emplaced in two major periods of plutonism, the first during the Upper Carboniferous and second during the Upper Permian and Triassic. The granitoids belonging to the first period of plutonism are inferred to have formed by partial melting of the deepest parts of a wedge of trench-complex sedimentary rocks (S-type), and divided into the Bundara Plutonic Suite dated by Shaw and Flood (1982) with an age of about 275 Ma, and the Hillgrove Plutonic Suite of whole rock Rb/Sr age of 295±25 Ma (Flood and Shaw 1977) to 310±12 Ma (Hensel et al, 1985). Landenberger et al (1993, 1995) suggested that the difference between ages of the Hillgrove Plutonic Suite and their error uncertainties may be due to variations in the initial isotopic composition of the suite and advised a Rb/Sr biotite age of 297 Ma for the emplacement of the Hillgrove Plutonic Suite and a faulting/uplift age of 256-266 Ma. The second and major period of plutonism of the New

England Batholith took place about 255-225 Ma ago, forming the Clarence River Plutonic Suite, the Moonbi Plutonic Suite and the Uralla Plutonic Suite and some other minor suites (Shaw & Flood, 1981).

Table 2.3. Subdivision of the New England Batholith. (After Shaw & Flood, 1981)

Association	Area		Lithology
	Km^2	%	
Bundarra Plutonic Suite	3,400	23	Adamellite, leucoadamellite, minor granite
Hillgrove Plutonic Suite	1,650	11	Subequal adamellite and granodiorite
Clarence River Plutonic Suite	910	6	Granodiorite and tonalite
Moonbi Plutonic Suite	1,800	12	Adamellite and minor monzonite
Uralla Plutonic Suite	1,700	11	Adamellite, granodiorite, minor tonalite and diorite
Leucoadamellites	4,440	30	Leucoadamellite
Triassic Granodiorites	300	2	Granodiorite
Unspecified Plutons	750	5	?
Total	14,950	100	

2.6 The Sydney- Gunnedah Basin

The Sydney-Gunnedah-Bowen Basin (Figure 2.7) is a foreland basin (e.g. Jones et al., 1984; Veevers et al., 1993; Tadros, 1993). To the west and the south, the deposits of the basin unconformably overly the deformed and metamorphosed Ordovician to Devonian Lachlan Fold Belt strata, to the north overly the Thomson Fold Belt, and to the east the basin tectonically underlies Devonian to Carboniferous New England Fold Belt strata, along the east dipping Hunter-Mooki Thrust (e.g. Korsch et al, 1997). The Sydney-Gunnedah Basin evolved from being a back-arc basin with respect to the Palaeozoic arc to a fore-deep basin associated with the westward prograding Tamworth Thrusts (Woodward, 1995). Geographically, the basin extends for 1700 km along the eastern margin of Australia, and is divided into three sections: the Bowen Basin to the north, Sydney Basin to the south, and the Gunnedah Basin in between. The Gunnedah Basin is some 350 km long and up to 200 km wide, and covers an area of 50 000 km^2 (Tadros, 1993).

2.6.1 Structural subdivision of the Gunnedah Basin

The Gunnedah Basin has three north-north-westerly oriented sub-basins lying between meridional basement ridges. These sub-basins have influenced the tectonic development of the basin throughout its history (Tadros, 1993). The meridional basement ridges include the Boggabri Ridge in the east, which separates the eastern Maules Creek Sub-basin from the central West Gunnedah Sub-Basin, and the Rocky Glen Ridge in the west, separating the West Gunnedah Sub-Basin from the western Gilgandra Sub-basin (Tadros, 1993) (Figure2.8).

The Maules Creek Sub-basin is a remnant Early Permian structural basin (e.g. Tadros, 1988). The Mooki Fault System forms the eastern boundary of the sub-basin at the surface, and the eastern flank of the Boggabri Ridge forms its western margin (Tadros, 1993). The interpretation of the BMR91-G01 seismic profile indicates that the Tamworth Belt has been thrust at least 10 km over the east margin of the Maules Creek Sub-basin (Korsch et al., 1993a, 1997). The basement in the Maules Creek basin is inferred to be the Boggabri Volcanics (e.g. Thomson, 1986; Korsch et al., 1993b, 1997).

The Boggabri Ridge (Russell, 1981), to the west of the Maules Creek Sub-basin, is a north-south structure, and is composed of the Late Carboniferous to Early Permian silicic and mafic Boggabri volcanic rocks (Martin, 1990; Tadros, 1993). It crops out to the west and south of Gunnedah, north of Boggabri and in the Deriah Forest area to the east of Narrabri (e.g. Tadros, 1993). Borehole data indicate that Boggabri volcanic rock is continuous in the subsurface between the outcrops, forming a prominent south to south-westerly trending basement ridge extending from south-east of Bellata south through the Deriah Forest area to

Baan Baa, Boggabri to south-east of Gunnedah where the ridge is truncated by the Mooki Fault System (e.g. Russell, 1981; Hill, 1986; Tadros, 1993).

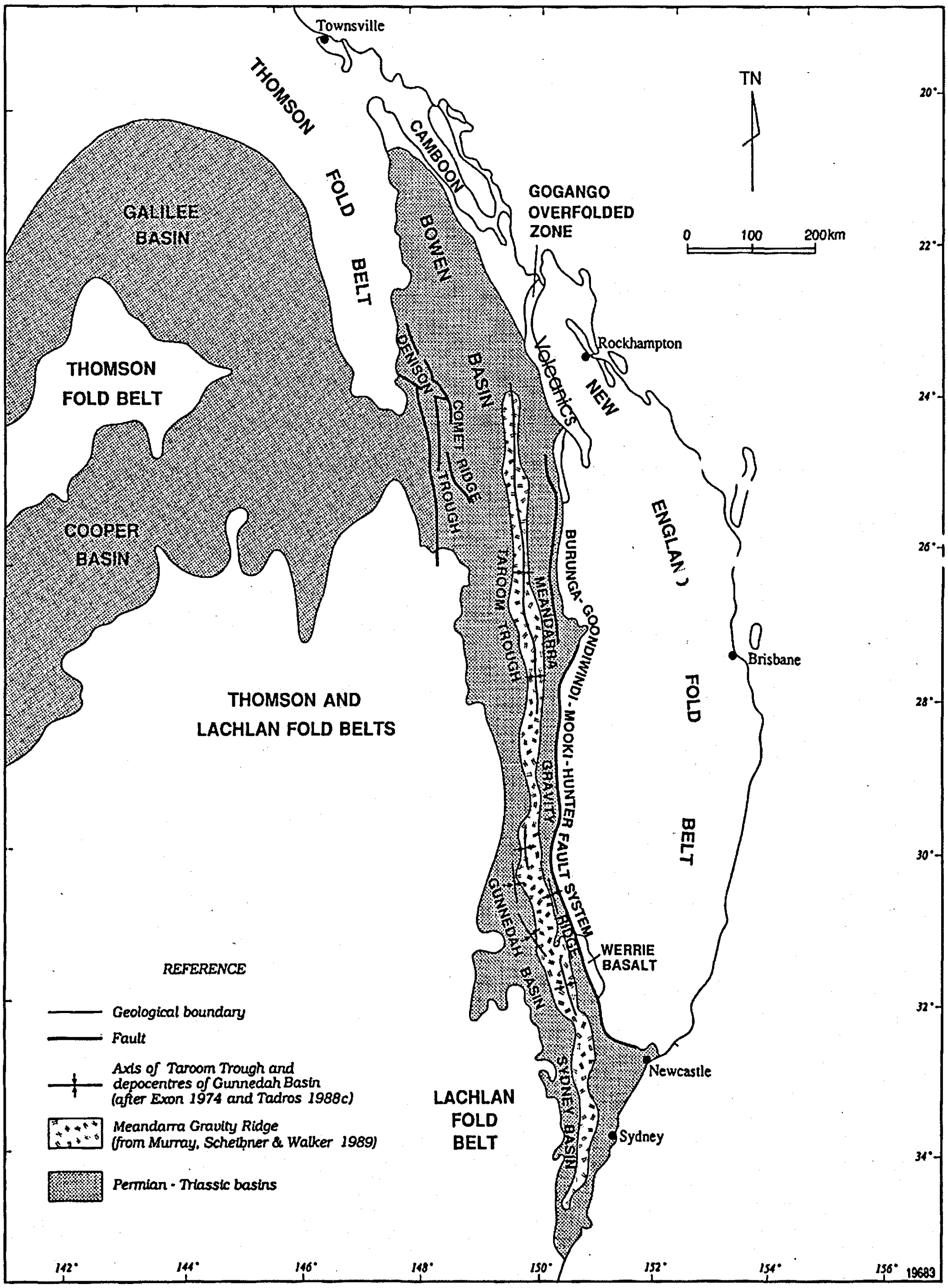

Figure 2.7 Structural setting of the Sydney-Gunnedah-Bowen Basin, showing its relationship to the New England and Lachlan Fold Belts (Taken from Figure 5.1, p48, Tadros, 1993).

The West Gunnedah Sub-basin (Tadros, 1988; Korsch et al., 1997) west of the Boggabri Ridge extends over the entire length of the Gunnedah Basin, from Moree in the north to the Mount Coricudgy Anticline in the south, and is the largest and most prominent of the sub-basins (Tadros, 1988, 1993). The basin is also divided by a number of west-south-west-trending transverse structural highs or ridges into a series of north-north-westerly oriented troughs (Tadros, 1988, 1993)

The Rocky Glen Ridge (Yoo 1988) forming the west limit of the West Gunnedah Sub-basin is present in the Coonabarabran area, and is composed of a group of Ordovician-Carboniferous metasediments, Carboniferous granitic and volcanic rocks (Tadros, 1993). Silicic ignimbrite and air fall tuff form the eastern flank of the Rocky Glen Ridge. Yoo (1988) suggested that the Ridge extends south to Dunedoo and north to Wee Waa to form a north-south-trending ridge, corresponding to a gravity high.

The Gilgandra Sub-basin is inferred, on the basis of the sparse drilling, to extend to the Mount Forster structural zone in the west, to the north-east-trending Cobar-Inglewood Kink Zone in the north and to the south of Dunedoo in the south (Yoo, 1988; Tadros, 1993). The basin is divided by a transverse basement high into northern and southern troughs, Pilliga Trough and Tooranweenah Trough (Tadros, 1993) (Figure 2.8). The interpretation of gravity data for the basin indicates that the troughs have gravity lows (Yoo, 1988).

2.6.2 Stratigraphy

Most strata within the Gunnedah Basin were deposited on basal Early Permian volcanic rocks. West of the Boggabri Ridge the strata dip gently towards the basin axis. East of the Boggabri Ridge the strata dip gently to the east but steeper dips occur near the Hunter-Mooki

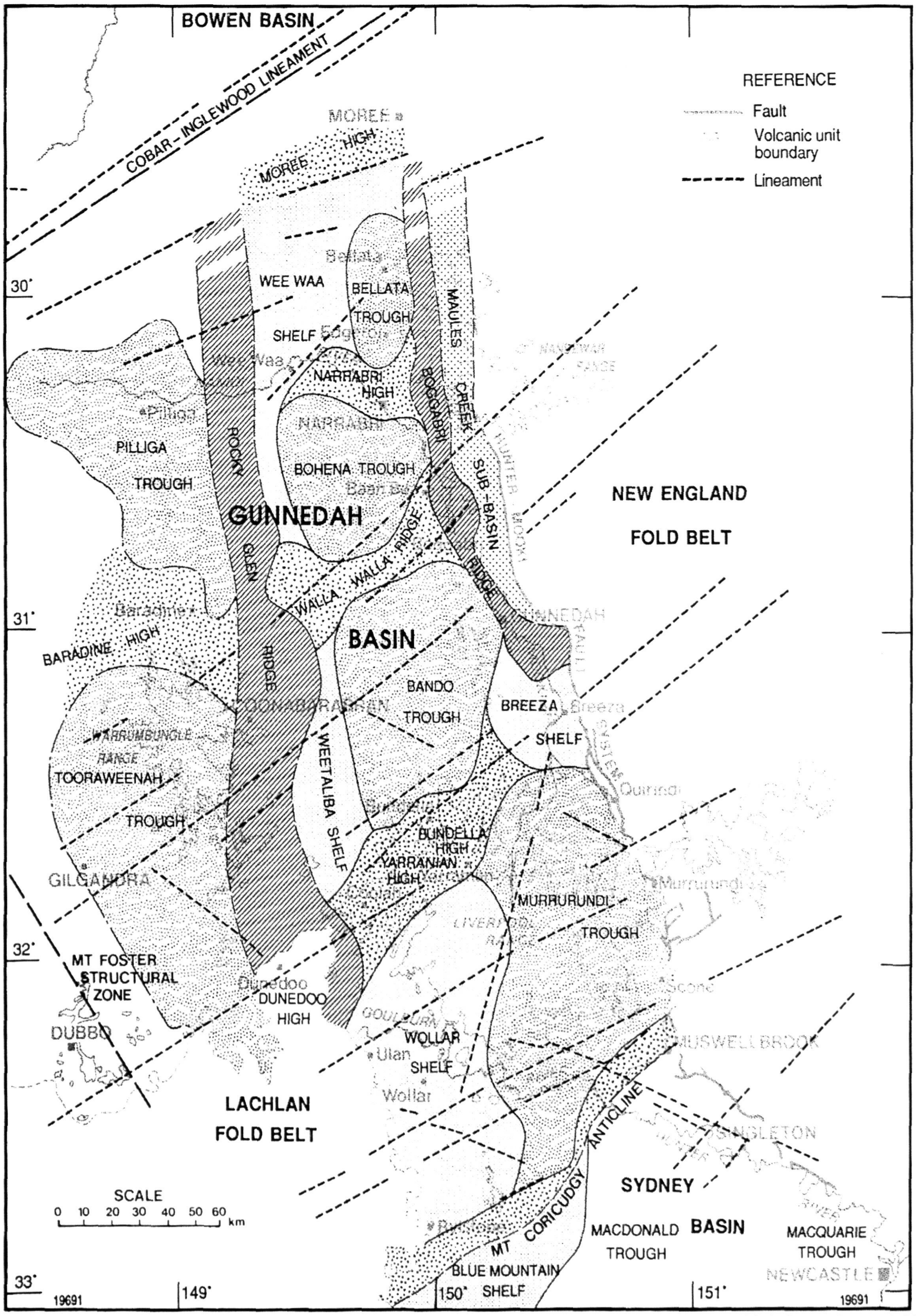

Figure 2.8 Structural subdivision of the Gunnedah Basin. (Taken from Figure 6.15, p74, Tadros, 1993)

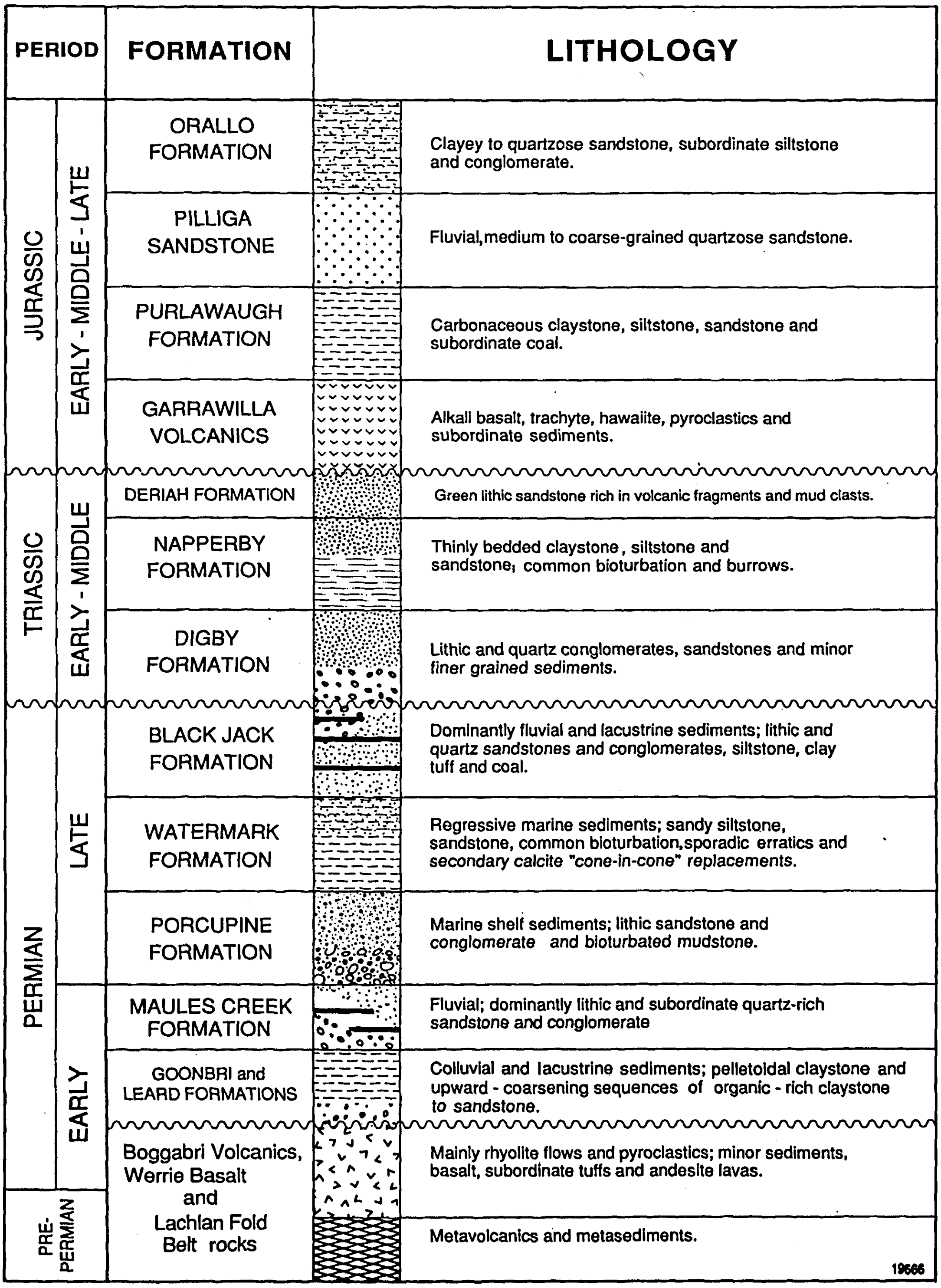

Figure 2.9 Stratigraphic sequence of the Permo-Triassic Gunnedah Basin. (After Figure 1.6, p11, Tadros, 1993)

Fault System (Tadros, 1993). The drill holes in the Gunnedah Basin indicate that localised palaeo-topographic variations within the basal volcanic units had an important bearing on the geometry of the overlying sedimentary sequence (Tadros, 1982, 1993; Thomson, 1986;

Martin, 1990, 1993). Tadros (1993) has given a very complete summary of the Permian-Jurassic lithostratigraphy of the Gunnedah Basin, which is listed in figure 2.9

2.6.3 The Tectonic Evolution of the Gunnedah Basin

In discussions of the tectonic evolution of the Gunnedah Basin, it has to be associated with the Sydney-Bowen Basin as a whole because, apart from local variations, the basins forming segments of a large single structure share similar histories, and their formation was controlled by the same tectonic processes and they evolved as one large elongate basin (Tadros, 1993). Veevers and Powell (1990) suggested that the Sydney-Gunnedah-Bowen Basin developed from a transitional tectonic or reactivation tectonic setting, the initiation of which was closely related to an active plate margin in the east, which in turn perhaps has been related to a new stage in Pangean cycle. Even though the evolution history of the basin has been difficult to unravel from surface investigations of the remnant basin, several theories have been proposed for the origin of the Sydney-Gunnedah-Bowen Basin. Harrington (1982) reviewed these earlier tectonic theories. Up-to-date reviews are those of Murray (1990), Scheibner (1993) and Tadros (1993). Murray (1990) grouped earlier proposed tectonic models and suggested five basin-origin mechanisms for the origin of the Sydney-Gunnedah-Bowen Basin (Table 2.4). The volcanic rift origin for the Sydney- Gunnedah Basin is generally accepted with gravity paradox, i.e. the Meandarra gravity Ridge coincides with the axis of the Taroom Trough, the site of maximum sedimentation. (eg. Tadros 1988, 1993; Scheibner 1989, 1993; Murray et al, 1989; Murray 1990).

The rifting of the Sydney-Gunnedah Basin was inferred to be Early Permian (e.g. Korsch et al., 1988; Finlayson and Fielding, 1990; Tadros, 1993; Scheibner and Veevers, 2000), and subsequently the basin became the foredeep or foreland basin of the New England Fold Belt when the tectonic setting of the region changed in the Middle Permian (Tadros 1988, 1993;

Jones et al, 1984; Scheibner, 1993). The Sydney-Gunnedah Basin was depressed under the advancing New England Orogenic Belt, strong deformation and uplift of the Devonian, Carboniferous and Early Permian rocks in the New England region resulted in tectonic stacking of the thrust sheets. The compressive deformation intensified towards the end of the Permian and caused major uplift and basin tilting (Tadros, 1986, 1993). Final basin inversion occurred during the Middle Triassic, with the New England Fold Belt thrust further over the Sydney-Gunnedah-Bowen Basin.

Table 2.4 Tectonic Models that have been proposed for the origin of the Sydney-Gunnedah-Bowen Basin (Modified from Scheibner, 1993, which is modified from Murray 1990)

Tectonic model	Process	References
Mantle diapirs	Rifting, volcanism and thermal subsidence following emplacement of mantle diapers into the crust	Brownlow 1981, 1982, 1988a, b.
Thermal collapse	cooling of crust/mantle after initial thinning	Murray 1990 and others
Rifting	(i) Formation of a tensional volcanic rift between the Lachlan and New England Fold Belts (Gunnedah and Sydney Basins), transitional to a volcanic arc in the north (Bowen Basin)	Scheibner, 1973, 1976; Murray, 1990.
	(ii) Precursor rift to Mesozoic break-up failed arm (?aulacogen) model	Harrington, 1982.
Back-arc extension	Unspecified back-arc extension adjacent to a calc-alkaline volcanic arc (Bowen Basin)	Fielding, 1990.
Back-arc spreading	Subsidence of craton behind eastward-migrating island arc system (Bowen Basin)	Battersby, 1981.