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BULLETIN **232**

THE EROMANGA - BRISBANE GEOSCIENCE TRANSECT:

A GUIDE TO BASIN DEVELOPMENT ACROSS PHANEROZOIC AUSTRALIA IN SOUTHERN QUEENSLAND

COMPILED & EDITED BY D. M. FINLAYSON

F MINERAL RESOURCES, GEOLOGY AND GEOPHYSICS

DEPARTMENT OF PRIMARY INDUSTRIES AND ENERGY
BUREAU OF MINERAL RESOURCES, GEOLOGY AND GEOPHYSICS

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THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT:
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D. M. Finlayson

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ABSTRACT

Crustal dynamics throughout geological history have played an important role in the development of sedimentary basins. A basic knowledge of major crustal structures is, therefore, crucial to any interpretations aimed at modelling particular basin systems. This Bulletin contains papers by authors from a number of geoscience institutions and companies on various aspects of crustal and basin development along an 1100 km east-west transect in southern Queensland, the Eromanga-Brisbane Geoscience Transect. In particular, deep seismic profiling along this transect has enabled, for the first time, a 3-dimensional interpretation of deep structures and processes which have controlled the development of major basin systems in eastern Australia. Complete answers to all questions on basin development are still evolving, but the papers presented in this Bulletin, together with the 1:1 000 000 scale map folio, provide a much improved basis for further, detailed investigations.

The Eromanga-Brisbane Geoscience Transect crosses three major basement provinces in eastern Australia: 1) the Thomson Fold Belt under the central Eromanga Basin and its infra-basins, 2) the northernmost Lachlan Fold Belt under the Taroom Trough of the Bowen Basin and Surat Basin, and 3) the New England Fold Belt under the Clarence-Moreton Basin. Basement geology in this region has, until now, been only poorly understood because it is largely obscured by the Mesozoic cover rocks of the Eromanga, Surat and Clarence-Moreton Basins. However, the application of geophysical techniques (seismic methods in particular) in recent years has enabled a much better understanding of the crustal architecture and processes likely to have been involved in the development of the major basins. Such an understanding provides the framework for more detailed investigations directed primarily at economic resources of oil, gas, coal, groundwater and many minerals.

The precis paper at the end of this Bulletin should be consulted for a summary of geoscience results. It is evident from these results that the transect interpretation has now firmly established concepts of crustal-scale ramp structures, multiple intra-crustal detachment surfaces, strike-slip fault architecture, lower crustal magmatism/underplating, Moho remobilisation, and intra-crustal terranes into the geological reconstructions of southern Queensland. In so doing, it has played a major role in developing a better understanding of the sedimentary basins of eastern Australia.

CO-OPERATING INSTITUTIONS

The compilation of data, information and interpretations along the Eromanga-Brisbane Geoscience Transect involved earth scientists from a number of institutions in co-operation with those at the Australian Bureau of Mineral Resources, Geology and Geophysics, Canberra. These institutions were as follows:

Queensland Department of Resource Industries, Brisbane.
Department of Geology & Mineralogy, University of Queensland, Brisbane.
Department of Applied Geology, University of New South Wales, Sydney.
School of Earth Sciences, Macquarie University, Sydney.
Division of Exploration Geoscience, Commonwealth Scientific & Industrial Research Organisation, Sydney.
The Australian Museum, Sydney.
Department of Geology & Geophysics, University of Sydney.
Department of Geology, University of Wollongong.
Department of Geology & Geophysics, University of New England, Armidale.
Petroconsultants Australasia Pty. Ltd., Sydney.
LASMO Oil Co. Australia Pty. Ltd.
Department of Geology, La Trobe University, Melbourne.
School of Science & Technology, University of Western Sydney.

THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT - AN INTRODUCTION.

D. M. Finlayson

Bureau of Mineral Resources, Geology and Geophysics, Canberra.

The processes by which hydrocarbons and minerals are concentrated near the Earth's surface are linked closely to dynamic events occurring in the lithosphere, that outer region of the Earth which has evolved by processes described well by plate tectonic models (Maxwell, 1984). At the depths to which resources can be drilled or mined (usually less than 5 km) the local geology in sedimentary basins, orogenic belts and cratons is a product of large-scale tectonism throughout geological time. These events have affected regions over lateral distances of hundreds/thousands of kilometres and at depths of tens/hundreds of kilometres. Any attempt to understand the history of hydrocarbon and mineral provinces must, therefore, embrace data on a regional scale from a large volume of the outer part of the Earth.

In the last 20-30 years, the basins in southern Queensland have attracted major exploration activity in the search for oil, gas, coal and groundwater. There have been large industry exploration programs in the region, including considerable seismic reflection profiling and drilling. The Bureau of Mineral Resources, Geology and Geophysics (BMR) has undertaken regional geological and geophysical investigations which included seismic traverses, and gravity and magnetic surveys across the area. The Queensland Department of Resource Industries and the Commonwealth Scientific and Industrial Research Organisation (CSIRO) have made substantial

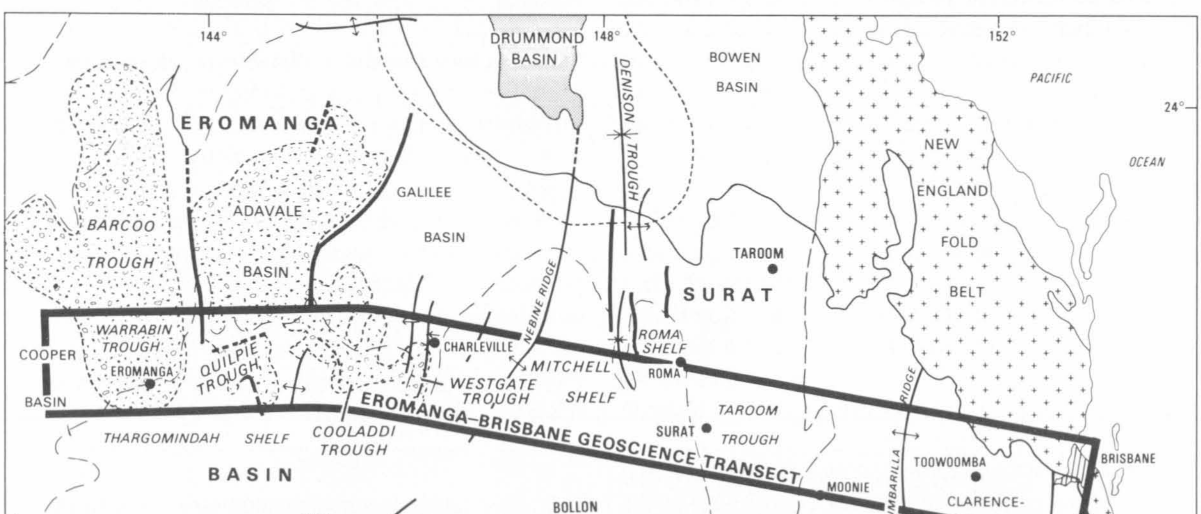

Fig. 1 The Eromanga-Brisbane Geoscience Transect area and simplified geology.

contributions to databases and interpretations, and numerous research projects completed by the University of Queensland, the University of New South Wales, Macquarie University, the University of New England, Sydney University, and others, have added to our knowledge of the region. Consequently, there is now a considerable amount of geoscience information from southern Queensland available for interpretation.

The main purpose of this Bulletin is to bring together data sets, recent interpretations and syntheses from the region 26° to 29° South, 142° to 154° East, along the Eromanga-Brisbane Geoscience Transect (Fig. 1). This area includes parts of basins and mineral provinces which are of major economic importance to Australia. Some of the data are presented in either map or diagramatic form at 1:1 000 000 scale. Interpretations of other data give new insights into the tectonic processes affecting the region. The information presented here provides a regional basis for detailed interpretations in areas of specific interest to the exploration industry.

Largely as a result of the work in the last 10 years, it is now possible to develop conceptual models for the development of basins in southern Queensland and present interpretations of the likely lithospheric processes that have affected the region. The regional database includes both geological and geophysical information and interpretations. One element of this database is a deep seismic reflection profile extending for 1100 km across southern Queensland which now enables the interpretation of structures and features to depths of about 60 km. This information on the depth-extent of geological structures places major constraints on three-dimensional evolutionary models for the region.

The concepts used to develop models for crustal evolution across Phanerozoic Australia are not generated in isolation. Rather they are often the product of ideas and models conceived from world-wide information. Consequently, comparisons of the eastern Australian data with analogues in other parts of the world is an important element in the development of tectonic models. The Global Geoscience Transects (GGT) project of the Inter-Union Commission on the Lithosphere (ICL) provides an international avenue for cooperation in the understanding of lithospheric evolution. Dr. J. W. H. Monger (Geological Survey of Canada), in the next chapter of this Bulletin, provides the background to the international aspects of the project. The Eromanga-Brisbane Transect in southern Queensland is included in the GGT catalogue of transects from all continents where information is being compiled with common guidelines to enable direct international comparisons.

Ideas on the development of Phanerozoic Australia are contained in many research papers, some of which are listed below in the References and Selected Bibliography. The authors of papers in this Bulletin have included numerous other references. Not all ideas and models presented in this Bulletin are complete or final. There are certainly differing views on the interpretations of some data, and there are some questions which can only be resolved by further field and laboratory work. One obstacle facing all research into the early Palaeozoic history of southern Queensland is the Mesozoic and Cainozoic cover which precludes direct examination of basement in many areas. The Great Artesian Basin covers about 20% of continental Australia, and along the Eromanga-Brisbane Transect there are few basement outcrops. However, basement cores and along-strike outcrops can be used to interpret much of the upper crustal basement in the transect area.

This Bulletin presents separate papers by various authors on different aspects of geoscience along the Eromanga-Brisbane Transect. The authors have been asked to concentrate on those topics which they think are particularly relevant for a guide to crustal developments affecting basin formation across Phanerozoic Australia in southern Queensland. Ideas on basin development are still evolving, hence there are likely to be differences in emphasis and interpretation. However, this does not preclude the presentation of interpretative cross-sections that extend down to at least the base of the Earth's crust, together with associated sketches of the probable way in which the crust evolved (two aims of the GGT project). Areas where differences in interpretation occur can then be highlighted for future research.

Some topics relevant to Phanerozoic developments and crustal structure are not included in this Bulletin. They include palaeomagnetism, electrical conductivity, heat flow and palaeogeothermometry, and volcanic geochemistry. In some cases there is a need for further work before recent review papers can be improved on. In other cases there are comprehensive recent publications. Selected papers on each of these topics are contained in the following references:

<u>Palaeomagnetism</u> - Goleby (1980), Klootwijk (1987), Klootwijk & Giddings (1988), Li (1987), Schmidt & others (1986), Schmidt & others (1987), Schmidt (1988).

<u>Heat flow and palaeogeothermometry</u> - Beeston (1986), Cull (1982), Cull & Conley (1983), Cook (1986), Kantsler & others (1983), Middleton (1979), Marshallsea & others (1985), Marshallsea (1988).

<u>Electrical conductivity</u> - Lilley & others (1980), Middleton (1979), Spence & Finlayson (1983), Woods & Lilley (1980).

<u>Volcanic geochemistry</u> - Ewart (1985, 1987), Ewart & others (1980,1988), Johnson (1989), Wellman & McDougall (1974).

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THE GLOBAL GEOSCIENCE TRANSECTS (GGT) PROJECT

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ABSTRACT

The Global Geoscience Transects (GGT) project is part of the International Lithosphere Program (ILP). It is modelled generally after the program for continent-ocean transects which were compiled during the Decade of North American Geology. However, the GGT emphasizes the continental crust, which contains 95% of the preserved record of earth history, rather than the transition between oceans and continents.

The term 'transect', as used in the GGT Project, refers to a cross-section showing the composition and structure of the entire crust of the Earth and, where possible, the lower lithosphere. It incorporates and integrates all available geological, geochemical and geophysical data. Transects lie along corridors 100 km wide and up to a few thousand kilometers long, positioned by regional experts to cross major crustal features. Ideally, they are a type of geological strip map in the vertical plane that can be used to show how the crust in that region has formed.

A major purpose of GGT is to encourage preparation of transect displays in a common format, so that crust in different parts of the world can be compared directly. The project is intended to utilize the vast amount of geological and geophysical information, collected partly for economic reasons, that already exists, mainly in national surveys. The quality and availability of these data are best known and evaluated by local experts. Thus, for the project to be viable, it must involve scientists at the "grass-roots" level, and give them the opportunity to share in a project of global scope and at the forefront of earth science research.

WHAT IS A CRUSTAL TRANSECT?

Geological cross-sections, and the maps they commonly accompany, are a necessary part of the geological analysis of any area. They are based on direct observation and current interpretation of composition, age and structure of surface rocks. The regional structural style obtained from the map is important, so that large-scale structures along trend can be projected down the plunge to define subsurface structures in the line of section. In places, surface data are supplemented directly by subsurface information from drillholes and tunnels, and indirectly, by geophysical data particularly in regions of interest to the petroleum industry. Typically, such geological cross-sections extend no more than a few thousand metres below the earth's surface.

In contrast to these relatively shallow and local cross-sections are interpretative sections based on regional geology, which have for many years been part of tectonic syntheses of regions such as the European Alps. The intensive work on the well-exposed rocks of the Alps with their distinctive and contrasting sequences led to the early recognition that very different stratigraphic facies were superimposed on thrust faults of great displacement. The magnificent serial cross-sections drawn to depths of 10 km by Staub (1924) led directly to the first transects: namely, cross-sections embracing much of the crust and showing the origin and probable present disposition of the major crustal elements.

Deeper parts of the crust and the subcrustal lithosphere are generally investigated by geophysical techniques. Geological observations on these inaccessible regions are limited to the few places where former deep crust, recognized by its metamorphic mineralogy, is exposed as the result of tectonic uplift, and where rocks from great depth are brought to the surface in volcanic necks and diapirs. The geochemistry of magmatic rocks yields somewhat equivocal clues about the composition of deeper regions. The cost of direct access by deep drilling is too great for more than a few deep holes globally, and with present technology these are unlikely to penetrate far into lower parts of the continental crust.

Geophysical techniques are widely applicable and relatively cheap, but gravity, seismic refraction, magnetic and heat flow studies provide very different kinds of data from geological sampling, and these may be difficult to interpret in geological terms. On the one hand, crustal sections derived from geophysics are typically simple and quantitative, but rather poorly constrained, as they generally represent only one of a possible range of options. On the other hand, sections derived from geological studies tend to be complex and qualitative, due to the widespread nature of surface sampling.

Geophysics typically measures today's world, whereas geology is very concerned with changes through time. For example, a representative section through the Alps (see figure 7, Mueller & others, 1980) shows a simple density layering, whereas a geological section shows the possible crustal disposition of rock units, based on the concept that rocks once widely distributed now are stacked vertically to make up the crust. Cross-sections of the crust in regions where there is deep geophysical information commonly exhibit a dichotomy in which the upper few kilometres show a great complexity based on geology, and in which the lower section is more simplistic with divisions based on densities.

The integration of these different geological and geophysical perspectives is perhaps the major challenge of GGT. This integration has been done in the upper crust for many years, with considerable commercial success, by the petroleum industry. Seismic reflection images, particularly the multi-channel computer-processed variety, resemble geological cross-sections in that they show the geometry of reflectors within the crust. These images may be readily related to geology where the reflector coincides with a lithological unit or structure that can be directly observed at the surface or in drill holes.

The first cross-sections of the Canadian Cordillera that compared in scope with the classic Alpine sections were drawn by petroleum explorationists. Cross-sections in the petroliferous Rocky Mountain fold and thrust belt, well constrained at depth by drill-holes and seismic reflection lines, were extended westward by the technique of balancing cross-sections into regions of granitic and high-grade metamorphic rocks.

More recently, the COCORP program in the United States has explored the crust to great depths using sophisticated, commercially available seismic reflection techniques pioneered by the petroleum industry. The spectacular results of this program, such as tracing of the Wind River Thrust in Wyoming to a depth of 24 km (Smithson & others, 1978), and the raising of the possibility that crystalline rocks of the Appalachians form part of an enormous thrust sheet (Cook & others, 1979), have substantially modified our concepts of crustal structure.

THE NORTH AMERICAN TRANSECTS PROGRAM

In 1978, the United States Geodynamics Committee initiated the continent-ocean transects project. This was joined shortly afterwards by the Canadian Committee on the Lithosphere and the Institute of Geology, University of Mexico, to become the North America transects program. This was designed to incorporate and integrate the considerable amount of mainly seismic reflection data from the offshore regions of North America, with existing geological and geophysical information from contiguous parts of the continent. The purpose was to explore the transition between continental and ocean crust in marginal parts of the continent.

Prior to this time, most published cross-sections fell into one of two categories: classic on-land geological profiles, and newer off-shore sections based on seismic reflection data. The two were separated by the shoreline and rarely integrated into a single profile. The project resulted in a series of cross-sections of the entire crust from the stable continental interior, through the marginal Phanerozoic mobile belts, into the ocean basins. The transects are now being published by the Geological Society of America as a product of the Decade of North American Geology (e.g. Monger & others, 1985).

Annual workshops for transect compilers reviewed progress, discussed problems, and modified as necessary the guidelines for the format of the transect displays. The workshops were a critical and personally rewarding part of the project in that they brought compilers together, made each aware that the transect problems wrestled with in isolation were not unique, and instilled a strong sense of common purpose. For GGT to become viable, it is essential that similar workshops be held.

After much discussion, it finally became mandatory for each North American transect display to consist of the following items:

1. A geological strip map (1:500 000 or 1:1 000 000) of a corridor 100 km wide containing the line of the transect, with rock units coloured according to age.

2. A geological cross-section with rock units coloured according to age, and at 1:500 000, with vertical equal to horizontal scales. These include, where available, gravity and magnetic profiles, deep seismic reflection line diagrams, velocity/depth curves and seismic refraction models, a magnetotelluric model, and ancillary data such as heat flow, hypocentres, epicentres, and selected isotope geochemistry.

3. A diagram showing the distribution of stratigraphic, structural, intrusive and metamorphic relationships of rock units along the line of the transect with space-time coordinates.

4. The transect proper, accompanied by a pamphlet giving brief descriptions of major units, sources of data and the rational for constructing it.

The transect proper is a crustal cross-section that integrates all of the above geological and geophysical information in an interpretation of the origin and disposition of crustal components and, thus, the origin of the crust in that region. Rock units are coloured according to the tectonic settings in which they are inferred to have formed: continental platform/shelf, slope/rise, magmatic arc, oceanic, successor and foreland basins and basins related to rifting, and generalized continental crust for old cratonic rocks. Compilers were allowed considerable latitude in making the interpretation; they could be conservative and merely newer offshore data in the Pacific Ocean west of Vancouver Island.

The problem faced by the compiler, and generally by others with similarly limited data, was how to bridge the gap between the geological detail in the upper 5-10 km and the Moho. The latter varies from 45-55 km under the continental interior and easternmost Cordillera, to 20-25 km under Vancouver Island (Fig. 1). A hypothesis based on geology and geochemistry was used, according to which the present crust of the western Cordillera is largely derived from two large composite suspect terranes, each composed of rocks originating in ocean basins. The eastern of these (Composite Terrane I of Figure 1) accreted to the ancient continental margin in Jurassic time, and the western (Composite Terrane II) in

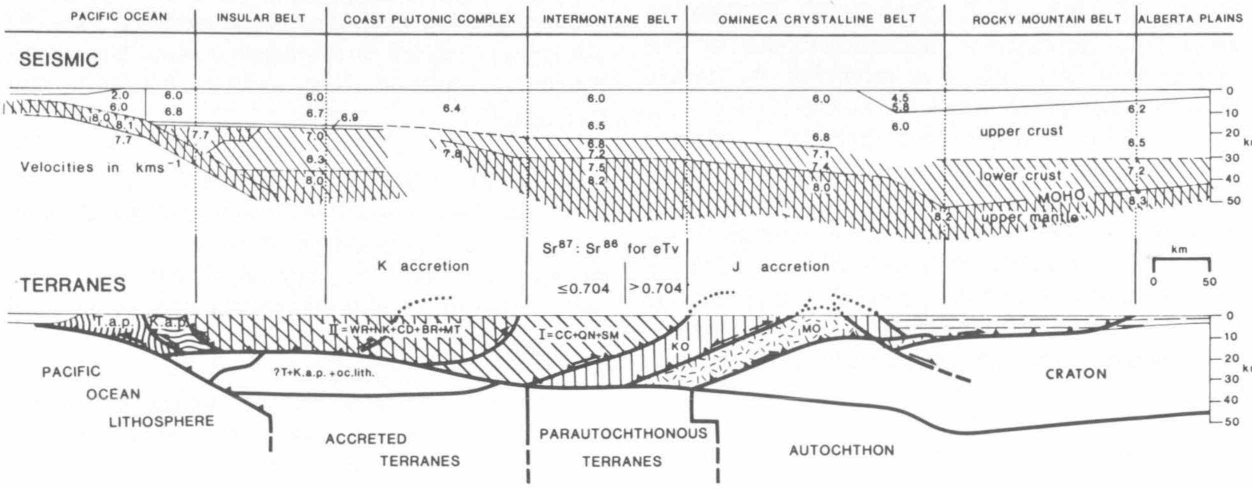

Fig. 1 The Southern Canadian Cordilleran Transect (from Monger & others, 1985), showing morpho-structural belts: (a) seismic refraction model; and (b) accreted terranes. Composite terranes I and II, comprising smaller terranes amalgamated prior to accretion, are shown as giant nappes. T.a.p. - Tertiary(T) to Cretaceous(K) accretionary prisms. e.T.v. - early Tertiary volcanics.

show generalized lower crust, or else be creative and show composition and relationships of units making up the entire crust.

AN EXAMPLE FROM THE CANADIAN CORDILLERA

To illustrate the problems involved, it may be useful to recall the steps followed in compiling Transect B2 across the southern Cordillera of western Canada (Monger & others, 1985). At the time of compilation (1982), the database included complete coverage by 1:250 000 scale geological maps, local detailed maps, complete gravity and sea/airborne magnetic coverage, and a seismic refraction profile, which incorporated data of varying quality obtained in the previous 20 years. At that time the only available seismic reflection data had been collected from relatively shallow depths by oil companies in the easternmost part of the transect, and Cretaceous time (Monger & others, 1982). In the accretion process, the ensimatic terranes were thrust over or beneath North American rocks.

As seen from the refraction model of Figure 1, the eastern step in the Moho coincides with the western edge of the little-deformed part of the Precambrian craton and the western step is located beneath the Coast Plutonic Complex. The terranes were treated as giant nappes, on the basis of structural relationships observed at the surface; these, with magmatic additions, make up the crust and lower lithosphere of the western Cordillera. Presumably the lower crust and lower lithosphere of the accreted terranes were detached and became incorporated in subcrustal mantle.

This interpretation is no different in princple from the early Alpine cross-sections. Both are speculative because the following features could not be resolved:

- the vertical distribution of the various lithological packages through the crust, and their attitudes;

- the relative contributions to the fabric of the present crust by various structural events seen at surface;

- the physical and chemical nature of the lower crust;

and

- the volume and composition of magmatic rocks in the entire crust and their vertical distribution.

The existence of this transect, as a model to be tested, was partly responsible for the choice of location of two deep seismic reflection lines run as part of LITHOPROBE, the Canadian multidisciplinary research program to investigate the nature and evolution of the lithosphere. A recent deep seismic reflection study across Vancouver Island shows Pacific Ocean crust going down beneath the outermost accreted terrane (Yorath & others, 1985). A profile in the eastern Cordillera shows reflections interpreted as late Mesozoic to earliest Tertiary compressional structures offset across Tertiary extension faults (Cook & others, 1987). The new results confirm the general crustal pattern shown for those parts of transect B2, but constrain the location and orientation of rock packages and structures at depth. Eventually it is hoped to complete a deep seismic reflection profile across the entire southern Cordillera. The transect will then be redrawn, possibly as a contribution to GGT.

Two lessons are to be learnt from this experience. First, given reasonable control of the surface geology and a concept of its evolution, transects can be drawn with limited geophysical data, and they can provide models to be tested by further geophysical work. Second, the approach to the problem, in particular the scale of lithological units displayed on the transect, breaks down many of the barriers between geologists and geophysicists; both are working on comparable scales.

CURRENT TRANSECT PROGRAMS ELSEWHERE

Other nations have proposed and implemented similar programs. For example, Australia has produced profiles of orogenic belts that are similar in many ways to the North American transects (Duff & others, 1985). Many countries are now actively promoting data-collecting along transect lines, rather than just using existing material as in the North American project. The European Geotraverse is a 4000 km long transect extending from the Precambrian Fennoscandian Shield in the north, across the Variscan realm, to the Alpine-Mediterranean region in the south (Mueller & Banda, 1983). The Soviet Union has an ambitious and exciting program of long seismic lines that are linked to, and presumably controlled by, super-deep boreholes (Kozlovsky & Yanshin, 1984). The working group on Studies in East Asian Tectonics and Resources (SEATAR) is preparing transects using extensive marine seismic lines. Encouraging responses have been received from these sources on the possiblity of final publication in GGT format.

THE GLOBAL GEOSCIENCE TRANSECTS PROJECT

The GGT project was conceived in August 1985 in Tokyo by the Inter-Union Commission on the Lithosphere (ICL), which runs the ILP. Letters announcing the project were sent out to ILP National Committees and drew a widespread and favourable response. At the request of ICL, Muawia Barazangi of Cornell University prepared a global map of over 150 possible transect lines. Locations of these transects were determined by reference to major geological structures and to previously implemented and proposed transects. They were designed to emphasize continental regions, though some marginal seas and island arcs are also included.

GGT was launched in 1986 and the most critical task at that time was to encourage scientists to start drawing transects. Initially, guidelines developed for the North American transects were followed, although these were subsequently modified as the project progressed and new problems were encountered.

In August 1986, ICL met in conjunction with the Fourth Circum-Pacific Energy and Mineral Resources Conference in Singapore. Representatives of several ILP National Committees brought proposed transect lines to this meeting. At the Singapore meeting, ICL established a new Coordinating Committee on Global Transects (CC-7), which will be made up of about 10 regional coordinators, plus the writer as chairman. The functions of CC-7 are to encourage compilation of transects within and between interested countries, to ensure that data and interpretation are presented uniformly so as to permit direct comparison of crust in different parts of the world, and ultimately to edit the transects for publication. Following the Singapore meeting, requests for numeric transect location data (to locate transects accurately on maps of various projections), and brief descriptions were sent out to ILP National Committees, various international groups, and interested individuals. This material is compiled as a GGT catalogue that provides an accurate overview of the entire project.

A key part of GGT is the direct and personal exchange of information and ideas and discussion of problems by the scientists making transect compilations. New problems will certainly arise that were not addressed in the North American program, which was concerned largely with Phanerozoic rocks of the continental margins. For example, how are Precambrian rocks in extensive platform areas with Phanerozoic cover to be treated, or what tectonic environments are to be resolved after scientists have wrestled with them and discussed their findings with others working on similar problems. For GGT to be viable and truely "global", as many scientists as possible who are compiling transects must attend project workshops and contribute directly.

The digitization of transect data for GGT is being explored with ILP Coordinating Committee 5, concerned with data centres and data exchange. W. Hinze of Purdue University is attempting to digitize one of the existing North American transects. Digitization of transects will permit rapid and accurate testing of the feasibility of transect crustal models; for example, is the crustal section as drawn compatible with the observed gravity profile? Digitization will allow rapid incorporation of new material and modification of earlier line work, and will speed up editing and publishing.

CONCLUSIONS

Data on the nature and composition of the earth's crust and mantle are rapidly being collected by many countries. The Global Geoscience Transects project encourages all countries to use this information in a systematic manner and present it in such a way that the crust in all parts of the world can be directly compared.

We appear to be at the beginning of a new approach to an understanding of our planet: namely, the systematic mapping in three dimensions of major features of the crust of the continents. We have seismic techniques that make routine exploration of the deep crust possible. During the 1950s, discoveries were made in the ocean basins (magnetic stripes, transform faults, etc.) whose interpretation profoundly influenced the whole of geological science. It is not possible to forecast the outcome of any scientific endeavour, but discoveries of at least comparable magnitude may well result from global studies of the continental crust. The enthusiastic response by many countries to GGT suggests that it is the ideal vehicle for encouraging a new scientific endeavour.

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SUMMARY OF GEOLOGICAL DEVELOPMENTS ALONG THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT

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ABSTRACT

Beneath the Mesozoic platform cover of the Great Artesian Basin, the Eromanga-Brisbane Geoscience Transect crosses, from west to east, the early to middle Palaeozoic Thomson Fold Belt, the late Palaeozoic to early Mesozoic Bowen Basin, and the middle Palaeozoic to Mesozoic New England Fold Belt. In general, orogenesis becomes younger from west to east along the transect, possibly reflecting progressive continental accretion since the inception of the Thomson Fold Belt at the end of the Precambrian.

The Thomson Fold Belt is believed to comprise mainly Cambrian and Ordovician quartzose turbidites deposited on thinned continental crust southeast of prominent gravity and aeromagnetic lineaments which mark the boundary with the thick Proterozoic crust of the North Australian Craton. The fold belt was deformed and uplifted close to sea level by the beginning of Devonian time, when the Adavale and Drummond Basins developed, probably as extensional basins. Its orogenic history was closed by east-west compression in the mid-Carboniferous.

The New England Fold Belt was a convergent plate margin related to a west-dipping subduction zone in Devonian-Carboniferous time, with a western volcanic arc, a central fore-arc basin, and an eastern accretionary wedge. A substantial component of strike-slip faulting in the Late Carboniferous or Early Permian led to displacement of stratotectonic units and formation of a large double orocline or megafold. Subduction-related volcanism resumed in the northern part of the fold belt in Early Permian time, but not in the south, which was in an extensional back-arc setting. Voluminous silicic magmatic activity in the Permian and Triassic produced granitoids and comagmatic continental volcanics. The final phase of orogenesis, which folded strata of the Maryborough Basin along the eastern extremity of the fold belt, was of Cretaceous age.

Although the most widely accepted origin for the Bowen Basin (and its southern extension, the Gunnedah-Sydney Basin) is as a foreland basin to the New England Fold Belt, recent interpretations support an extensional origin, possibly associated with strike-slip faulting in a transtensional environment. The basin consisted of a series of depocentres in which subsidence varied with time, and which received thick sequences of continental and marine sediments, including enormous coal resources, from the beginning of the Permian to Middle Triassic time. Intensity of deformation throughout the Bowen Basin is very variable, but in general increases from west to east, probably reflecting basement geology, sediment thickness, and application of stress.

INTRODUCTION

This paper presents a broad overview of the geological evolution of tectonic units along the Eromanga-Brisbane Geoscience Transect, concentrating on events which are believed to have been of particular relevance to the development of continental crust. The treatment largely follows a chronological succession from the early to middle Palaeozoic Thomson Fold Belt, progressing to the middle Palaeozoic to Mesozoic New England Fold Belt, and then to the intervening late Palaeozoic-early Mesozoic Bowen Basin (Fig. 1). Because these basement rocks are concealed by Mesozoic strata of the Eromanga, Surat and Clarence-Moreton Basins along the entire transect, it is necessary to take account of data from exposed areas to north and south, covering an area across eastern Australia in southern Queensland and northern New South Wales. The geological development of this region has been described by several authors including Marsden (1972), Scheibner (1973, 1976), Harrington (1974), Kirkegaard (1974), Day & others (1983), Veevers (1984), Degeling & others (1986), and Murray (1986). Figure 2 gives a general impression of known developments in the transect region throughout geological time.

THOMSON FOLD BELT

Definition and Limits

The name Thomson Fold Belt was introduced by Kirkegaard (1974) for a largely concealed belt of rocks which he considered to be of Cambrian to Carboniferous age and broadly correlative with the Lachlan Fold Belt of New South Wales. He included the exposed strata of the Anakie Inlier, Drummond Basin, Ravenswood Block, and Broken River Embayment at the northern end of the fold belt. However, subsequent authors have made it

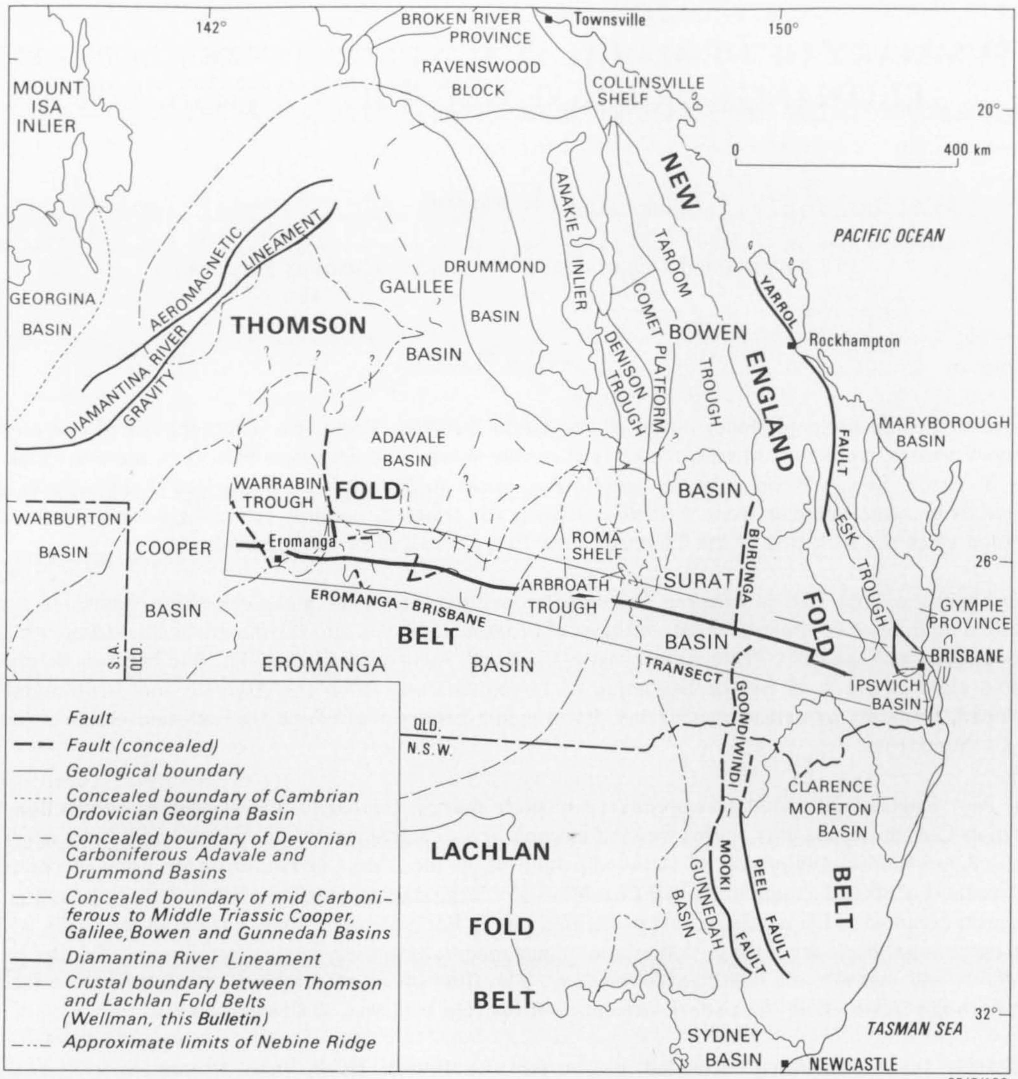

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Fig. 1 Simplified geological province boundaries and structural features in southern Queensland and northern New South Wales. The boxed area is the location of the Eromanga-Brisbane Geoscience Transect: the principal east-west deep seismic reflection traverse is shown within the boxed area.

clear that the Broken River Embayment (Day & others, 1978; Withnall & others, 1987) and the Ravenswood Block (Henderson, 1980, 1986) should be regarded as separate tectonic units.

The boundary of the Thomson Fold Belt with the Proterozoic Mount Isa Inlier or Block to the northwest is well defined by very prominent northeast-trending gravity and magnetic lineaments (Wellman, 1988; Murray & others, 1989), which mark the Diamantina River Lineament of Scheibner (1974) (Fig. 1). To the east, its contact with the New England Fold Belt is concealed beneath the Bowen Basin. The boundary with the Lachlan Fold Belt to the south is more controversial. Some authors (e.g. Powell, 1984a) consider that the Thomson Fold Belt is merely a northern continuation of the Lachlan Fold Belt. However, the dominant northeasterly trend of the Thomson Fold Belt (Kirkegaard, 1974; Murray & Kirkegaard, 1978) differs from that of the Lachlan Fold Belt, and there are differences in crustal structure as well. Wellman (this Bulletin) has defined a boundary between these two fold belts based on gravity and magnetic trends. To the west, the boundary between the Thomson Fold Belt and the subsurface Warburton Basin (Gatehouse, 1986) appears to be gradational, with increasing deformation and metamorphic grade eastwards.

Pre-Devonian Geology

Examination of basement cores of petroleum exploration boreholes has shown that the dominant rocks of the Thomson Fold Belt are steeply-dipping, low-grade metasediments intruded by Silurian and Devonian granites (Murray, 1986). The sediments are quartzose turbidites (P.J. Conaghan, pers. comm.) derived from a metamorphic-plutonic source area. They have generally been assigned an early Palaeozoic (Cambrian and Ordovician) age through correlation with strata of the Warburton Basin to the west (Murray & Kirkegaard, 1978; Gatehouse, 1986). Limited isotopic dates on metamorphic basement rocks are consistent with this interpretation (Murray, 1986).

The most probable surface correlatives of these presumed early Palaeozoic rocks occur in the northern part of the Anakie Inlier (Fig. 1), and comprise folded micaceous quartzite, quartz-rich sandstone, siltstone, slate, phyllite and minor schist. These rocks may be unconformable on a basal sequence of metamorphosed mafic volcanics, muscovite-quartz schist, calc-silicates, recrystallised limestone and serpentinite which forms the southern end of the inlier. The subsurface basement sediments may also be equivalent to the very extensive Late Ordovician quartz-rich turbidites of the Lachlan Fold Belt (Cas, 1983; Powell, 1984a).

Pre-Devonian Tectonic Evolution

It is postulated that the Thomson Fold Belt formed by rifting and extension of the Proterozoic craton southeast of the prominent gravity and aeromagnetic gradients which mark the Diamantina River Lineament (Fig. 1). Three different models can be envisaged:

1. Complete removal of the Precambrian craton by rifting and sea floor spreading. This requires that the entire fold belt was an oceanic area. There is evidence for an oceanic setting for the older basal sequence of the Anakie Inlier. However, it is considered unlikely that this was the case for the entire fold belt because much of it was uplifted above sea level by the beginning of Devonian time, requiring an unbelievably large rate of continental accretion in the early Palaeozoic.

2. Splitting of a continental margin arc and creation of a marginal sea. This model has been proposed for the Lachlan Fold Belt by numerous authors (see discussion in Powell, 1984a, p. 305), and has been applied specifically to the Thomson Fold Belt by Harrington (1974). Harrington proposed separation of the Nebine Volcanic Arc (which he considered to be represented by the Nebine Ridge, a broad, poorly defined basement ridge, see Fig. 1) from the Proterozoic craton to form the Barcoo Marginal Sea. This model has not been confirmed by seismic data, which have failed to reveal anomalous zones of crustal structure that could be equated unequivocally with Precambrian terranes such as the thick Proterozoic crust of the northern Australian craton (Finlayson, 1982; Collins, 1983). However, Finlayson & Collins (1987) have pointed out that there are differences in the seismic characteristics of the crust under the Nebine Ridge from those under the central Eromanga Basin to the west. An additional problem with this model is that boreholes into basement along the Nebine Ridge have not penetrated volcanic rocks.

3. The model currently favoured by most workers on the Thomson Fold Belt is that it is largely floored by extended and thinned Proterozoic continental crust. It is possible that the stretched Precambrian crust is now represented by the layered lower crust evident on seismic reflection profiles (Finlayson & others, this Bulletin). The layering may be due either to intrusion of sill-like mafic bodies (underplating) or to horizontal shearing of contrasting lithologies.

Whatever its origin and the nature of its crust, it appears that in the early Palaeozoic the Thomson Fold Belt was a relatively deep sea area, east of shallow-marine shelf deposits of the Warburton and Georgina Basins at a passive continental margin. As yet there is no confirmation of any volcanic arc deposits which may have marked the limits of a marginal sea, but the supposed older sequence at the southern end of the Anakie Inlier is a possible example. Whether the quartzose turbidites of the Thomson Fold Belt were sourced from the Proterozoic craton to the west, or were mainly transported northwards as in the Lachlan Fold Belt (Cas & others, 1980), is unknown.

The early Palaeozoic strata of the Thomson Fold Belt were folded and uplifted above sea level by the beginning of Devonian time. It is doubtful if this was a single orogenic event, as suggested by Murray & Kirkegaard (1978), because of the range of isotopic dates obtained from metamorphic basement cores and from muscovite schist of the Anakie Inlier (536 to 416 Ma, or Middle Cambrian to Late Silurian), and their uncertain significance (Murray, 1986). Intrusive granites from basement cores are Silurian to Early Devonian (426 to 405 Ma), but a Middle to Late Ordovician age (460 to 452 Ma) has been obtained from granite intruding schist which forms a small inlier just south of the Anakie Inlier (Murray, 1986). The relationships of the granitoids are shown in Figure 2.

Adavale and Drummond Basins

The Adavale Basin (Fig. 1) developed within the Thomson Fold Belt basement rocks during Early Devonian time (Evans & others, this Bulletin). The earliest sequences were continental volcanics and red beds (Gumbardo Formation). Early and Middle Devonian deltaic and shallow-marine carbonate sequences gave way to evaporitic and red-bed units in the Late Devonian and possibly the Early Carboniferous, when the Adavale Basin must have been continuous with the Drummond Basin to the northeast.

Most interpretations of the tectonic setting of the Adavale Basin have favoured a rifted back-arc basin or foreland basin model west of a volcanic arc - subduction zone complex (Passmore & Sexton, 1984; Remus & Tindale, 1988). However, the basement geology east of the Adavale Basin, which is relatively well known through numerous drillholes, provides no evidence to support such a model. The Nebine Ridge, which has been proposed as the site of the volcanic arc - subduction complex (Remus & Tindale, 1988) is composed of sediments which were metamorphosed in the Late Silurian (Murray, 1986), and therefore pre-dates the Adavale Basin. East of the Nebine Ridge, basement rocks of the Roma Shelf (Fig. 1) are tightly folded, steeply dipping metasediments of the Timbury Hills Formation. Although probably of Devonian age, and therefore coeval with the Adavale Basin sequence, these sediments are quartz-rich (Murray, 1986), and therefore unlikely to represent a fore-arc sequence.

An extensional model for the Adavale Basin was proposed by Veevers & others (1982) and Powell (1984b) who compared the tectonic setting to that of the Basin and Range Province in the southwestern U.S.A. This model accounts for the fact that Early Devonian volcanics, with a thickness of more than 750 m, are confined to the basin itself. One possible problem is that, although the volcanics range from basalt to rhyolitic ignimbrite, petrographic examination suggests that they are dominantly of intermediate composition (Murray, 1986). No geochemical data are yet available.

The strata of the Adavale Basin were deformed during two Carboniferous events, the first with crustal shortening in a north-south direction and the second in an east-west direction (Leven & Finlayson, 1987; Finlayson & others, 1988). The present extent of the basin is only the erosional remnant of a more extensive depositional area (Passmore & Sexton, 1984).

During Late Devonian - Early Carboniferous time, the Drummond Basin (Fig. 1) developed mainly to the west of the Anakie Inlier, but also with a small section to the east (Day & others, 1983). This basin has been described as a classic foreland basin west of a continental margin volcanic arc along the Anakie Inlier (Veevers & others, 1982; Powell, 1984c) or within the New England Fold Belt (Murray, 1986). This interpretation does not accord with the presence of a thick basal unit of dominantly silicic volcanics throughout the basin postulated from seismic reflection results (Pinchin, 1978) and penetrated by some boreholes, and possibly an extensional origin may be more applicable (Murray & others, 1989). In the mid-Carboniferous, the Drummond Basin was folded by the same east-west compressional event which affected the Adavale Basin.

Two mica granites, possibly of S-type, were emplaced during Early Carboniferous time into the folded Timbury Hills Formation on the Roma Shelf, a platform area south of the Denison Trough (Fig. 1) (Houston, 1964). Following folding of the Adavale and Drummond Basin sequences in mid-Carboniferous time, two large intracratonic sag basins (Galilee and Cooper Basins) developed over the Thomson Fold Belt, Warburton Basin and the southeastern margin of the Mount Isa Inlier (Fig. 1). They continued to receive extensive but thin continental sediments until the Middle Triassic, and for much of their history were connected to and probably drained into the Bowen Basin to the east.

NEW ENGLAND FOLD BELT

The New England Fold Belt (Korsch & others, this Bulletin) is the easternmost and youngest part of the Tasman Fold Belt System, and is separated from the older Thomson and Lachlan Fold Belts to the west by the Permian - Middle Triassic Bowen-Gunnedah-Sydney Basin (Fig. 1).

There is general consensus that the palaeogeographic setting of the New England Fold Belt in Devonian - Carboniferous time was a convergent plate margin related to a west dipping subduction zone (Day & others, 1978; Powell 1984c; Cawood & Leitch, 1985; Murray & others, 1987; Korsch & Harrington, 1987). Three parallel north - northwest trending stratotectonic elements can be recognised: a western volcanic arc (exposed in the north, but not in the south); a central fore-arc basin; and an eastern accretionary wedge that grew oceanwards by accreting trench-fill volcaniclastic turbidites and minor amounts of oceanic crust. The accretionary wedge is now separated from the fore-arc basin to the west by major fault zones (Yarrol and Peel Fault systems) marked by serpentinite lenses (Fig. 1).

The palaeogeographic relationship between the New England Fold Belt and the remainder of the Tasman Fold Belt System at this time is uncertain. It is usually assumed that from Late Devonian time, the New England Fold Belt developed essentially in its present position at the eastern edge of the Australian continent; the Late Devonian - Carboniferous volcanic arc was of continental margin (Andean-) type (Marsden, 1972), and appears to have contributed detritus to basins along the eastern side of the Lachlan Fold Belt (Powell, 1984c). The proposition, based on palaeomagnetic data, that the New England Fold Belt had moved southwards by a large amount relative to the Lachlan Fold Belt in mid-Carboniferous time (Klootwijk, 1985) has now been questioned in view of an alternative apparent polar wander path for Australia (Klootwijk & Giddings, 1988).

The possibility of large-scale movements of smaller areas within the New England Fold Belt cannot be discounted. The Late Silurian to Middle Devonian volcanic arc in the northern part of the fold belt may have been an island arc (Marsden, 1972; Day & others, 1978) separated from the Australian continent by a marginal sea or ocean basin of uncertain extent (Veevers & others, 1982). This postulated island arc must have reached its present position relative to the remainder of the fold belt by the end of the Middle Devonian, as it is overlain by Upper Devonian fore-arc basin strata. Coastal regions of the Devonian - Carboniferous accretionary wedge near Rockhampton and south of Brisbane contain quartz-rich sandstones which contrast markedly with the more characteristic volcaniclastic turbidites (Kirkegaard & others, 1970; Korsch, 1977; Murray, 1988). These anomalously quartz-rich units have been proposed as possible exotic terranes by Korsch & Harrington (1987). The best candidate for an exotic

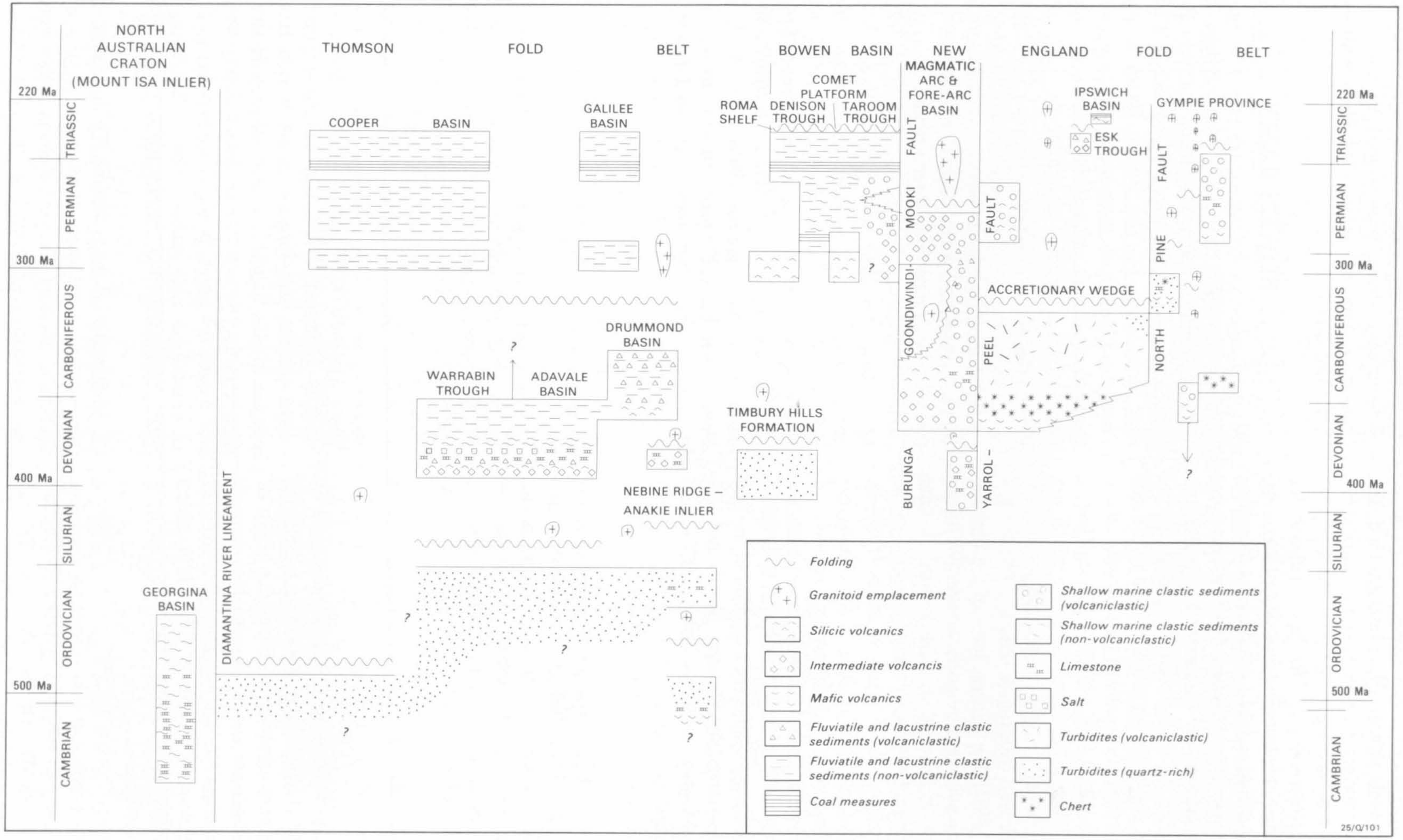

Fig. 2 Space-time diagram for the southern Queensland region indicating the principal geological developments based on data from Day & others (1983) and Murray (1986). Ages of folding events are derived from metamorphic isotopic dates and from observed angular unconformities.

terrane is the Gympie Province (Fig. 1), a unique stratotectonic unit which does not fit into the overall palaeogeographic pattern of the New England Fold Belt (Day & others, 1978).

It is highly probable that the New England Fold Belt was not a purely convergent plate boundary throughout its history, but that varying proportions of strike-slip motion were accommodated. The best evidence for strike-slip faulting is displacement of stratotectonic units. The New England Fold Belt is divided into two sections by Mesozoic cover of the Clarence-Moreton and Surat Basins in the region of the Queensland - New South Wales border (Fig. 1). Stratotectonic elements in the northern section are displaced about 200 km to the east of similar elements in the southern part (Bryan, 1925). This offset is associated with a large double orocline or megafold with a dextral sense of vergence which has folded the accretionary wedge sequence and possibly the fore-arc basin as well (Murray & others, 1987; Korsch & Harrington, 1987). The general form of the orocline is indicated by gravity and magnetic trends (Wellman, this Bulletin). It is believed to have formed in Late Carboniferous or Early Permian time. Formation of the orocline was attributed by Murray & others (1987) to large-scale dextral transform faulting of the eastern part of the New England Fold Belt following conversion of a purely convergent plate boundary to a combined convergent-transform boundary. This model is based on an analogy with the late Cainozoic evolution of the San Andreas Fault system in western North America (McKenzie & Morgan, 1969; Atwater, 1970), and involves interaction between an offshore trench and a mid-ocean ridge - transform fault system.

Voluminous calc-alkaline volcanism resumed in the northern New England Fold Belt in Early Permian time, coinciding with development of the Bowen Basin. Widespread volcanic and sedimentary rocks of this age appear to represent arc and fore-arc basin assemblages, but no Permian accretionary wedge sequences have been recognised. In contrast, the southern part of the New England Fold Belt was in a back-arc setting, and several relatively small, deep extensional basins developed.

Throughout Late Permian and Triassic time, most of the New England Fold Belt was uplifted as voluminous granite plutons were emplaced in the region south of Rockhampton. Plutonism was accompanied by eruption of widespread comagmatic silicic continental volcanics. Although most workers favour a subduction-related origin for the magmatism, this has not yet been conclusively demonstrated.

Transtensional events in the Triassic, possibly commencing as early as the Late Permian, formed the Middle Triassic Esk Trough and the Late Triassic Ipswich Basin (Fig. 1) in the central part of the New England Fold Belt (Korsch & others, 1989).

Following a period of cratonic stability in the Jurassic, silicic volcanism and plutonism resumed in the New England Fold Belt in Early Cretaceous time in the Maryborough Basin (Fig. 1) and at the northern end of the fold belt southeast of Bowen. Strata of the Maryborough Basin were folded in mid- to Late Cretaceous time into asymmetric folds with steep to overturned eastern limbs.

BOWEN BASIN

The Bowen-Gunnedah-Sydney Basin system extends along the entire western margin of the New England Fold Belt, separating it from the Thomson and Lachlan Fold Belts (Fig. 1). The Bowen Basin in southern Queensland comprises three main structural elements: the Denison Trough and associated Arbroath Trough in the west, the Taroom Trough in the east, and the intervening Comet Platform.

Several tectonic models have been proposed for the Bowen-Gunnedah-Sydney Basin (Harrington, 1982), but these can be grouped into three categories:

1. Probably the most widely accepted origin for the basin is that it is a foreland basin to the New England Fold Belt (Flood, 1983; Jones & others, 1984; Murray, 1985). Almost all features of the Bowen Basin are compatible with, and explicable by, a retroarc foreland basin model with initial subsidence being due to excess mass of the Early Permian volcanic arc to the east (Murray, 1985). This retroarc foreland basin model cannot be applied to the Gunnedah-Sydney Basin, because Early Permian volcanics along its eastern margin are bimodal rift type, not calc-alkaline arc type (Harrington & Korsch, 1985).

2. An extensional origin for the Gunnedah-Sydney Basin, which envisaged its formation as a volcanic rift, was proposed by Scheibner (1973, 1976). He considered that this model could not be extended into the Bowen Basin, and that the Gunnedah-Bowen Basin evolved into a foredeep (foreland basin) following mid-Permian orogenies in the New England Fold Belt, including major westward thrusting of fold belt sequences. More recently, Hammond (1987) has outlined an extensional model for the early history of the Bowen Basin. In support of this hypothesis he cited seismic reflection data from the Denison Trough, which clearly show that this feature developed in the Early Permian as a series of extensional half grabens (Paten & others, 1979; Brown & others, 1983; Ziolkowksi & Taylor, 1985).

3. Some authors have related the formation of the Bowen-Gunnedah-Sydney Basin to major transcurrent faulting or shearing. Evans & Roberts (1980) suggested that the basin was created as one of a series of en echelon troughs produced by a long-lived dextral rotational force couple with a northwest orientation, whereas Harrington (1982) proposed that it was formed as a pull-apart basin by large-scale dextral transform

faulting along the line of the Burunga-Goondiwindi-Mooki Fault (Fig. 1) (Harrington & Korsch, 1979, 1985). However, the validity of such large-scale movement was questioned by Cherry (1989). A possible resolution of this problem was provided by Korsch & others (1988) who combined an extensional and strike-slip model. They noted that deep reflection seismic data (Finlayson & others, this Bulletin) are not compatible with a purely extensional model of basin formation and that there must have been a component of strike-slip movement (amount not specified) along the Burunga-Goondiwindi-Mooki Fault. They concluded that the basin-forming mechanism was transtension.

The available data certainly lend strong support to the argument that extension and rifting, with or without strike-slip faulting, were involved in the initiation of the Bowen-Gunnedah-Sydney Basin. Further support is provided by the presence of the Meandarra Gravity Ridge (Wellman, this Bulletin), which extends for 1200 km along the axis of the basin, and coincides with the site of maximum sedimentation in the Taroom Trough. In the Sydney Basin, the southern end of this gravity ridge was modelled by Qureshi (1984), who attributed it to a mafic body up to 12 km thick within the crust. Whether this represents volcanics in the basement or intrusives deeper in the crust, or a combination, it can most readily be explained as the product of mafic igneous activity associated with rifting. Previous explanations of the Meandarra Gravity Ridge - as the gravity expression of a buried Devonian-Carboniferous volcanic arc that supplied detritus to the southern part of the New England Fold Belt (e.g. Day & others, 1978) - must be incorrect, because the northern end of the ridge is parallel to and separated from the exposed volcanic arc in the northern part of the fold belt (Murray & others, 1989).

As far as sedimentation is concerned, the Bowen Basin should be treated as a series of troughs or depocentres in which subsidence varied with time (Malone, 1964; Dickins & Malone, 1973; Flood, 1983). Deposition commenced in the Early Permian (Dickins & Malone, 1973; Fielding & others, this Bulletin). Thick continental sediments, including coal, were deposited in the elongate, rapidly subsiding Denison Trough. A subsequent marine transgression extended over much of the basin. Along its eastern margin, marine sediments, including limestone, were laid down over calc-alkaline volcanics of the Early Permian arc in the New England Fold Belt to the east. Marine conditions prevailed over the eastern part of the basin for much of mid-Permian time, while several regressive-transgressive cycles produced intertonguing marine, deltaic and fluviatile sequences in the west and north. A brief, widespread marine transgression was followed by final withdrawal of the sea and deposition of Late Permian coal measures throughout the basin. Triassic sedimentation was dominated by continental red-bed conditions.

Intensity of deformation in the Bowen Basin is very variable, but in general increases from west to east, probably reflecting basement geology, sediment thickness, and application of stress. It has long been recognised that the eastern margin of the basin is overthrust from the east by rocks of the New England Fold Belt (see Murray, 1985), and it is now known that this style of thin-skinned thrust tectonics extends much farther west (Hobbs, 1985). Thrusting was active from mid-Permian to Middle Triassic time, when deposition ceased in the Bowen Basin. The major large wavelength folds of the Taroom Trough must have formed in Late Triassic time, as they are truncated by basal Jurassic sediments of the Surat Basin (Fig. 1). An anomalous area of tight, large amplitude folds with steep axial planes in the central eastern part of the basin (the Folded Zone of Dickins & Malone, 1973) may have formed by gravity sliding following uplift of the New England Fold Belt to the east (Malone & others, 1969). In the Denison Trough, the Early Permian tensional phase was succeeded by compressional events, which began at the end of the Early Permian and reached a peak in Middle and Late Triassic time (Paten & others, 1979; Bauer & Nelson, 1980; Nelson & Bauer, 1980; Ziolkowski & Taylor, 1985). At least in part, the compressional forces have been attributed to transpression associated with left-lateral shearing along the major intra-trough fault systems (Paten & others, 1979; Brown & others, 1983).

MESOZOIC AND TERTIARY COVER

With the exception of the Maryborough Basin and the northern extremity of the New England Fold Belt, eastern Australia has been a stable cratonic region since the end of the Triassic. At this time the Eromanga, Surat, and Clarence-Moreton Basins developed as broad intra-cratonic sags. Strata of these basins are essentially undeformed and cover basement sequences along all but the easternmost extremity of the transect (Fig. 1). Mild Tertiary reactivation of older fault systems has been described by Finlayson & others (1988), and is evident in the Mesozoic sequences. The most significant Tertiary tectonism was the formation of narrow, relatively deep grabens within the area of the New England Fold Belt along the eastern margin of the continent during opening of the Coral Sea basin in the Paleocene and Eocene (Grimes, 1980; Shaw, this Bulletin).

Large volumes of continental flood basalts were erupted over the New England Fold Belt and extended as far west as the Anakie Inlier in mid- to late Cainozoic time. Eruption of these basalts has been attributed to passage of the Australian continent over hot spots in the mantle during northward movement away from Antarctica (Wellman & McDougall, 1974).

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A TECTONIC INTERPRETATION OF THE GRAVITY AND MAGNETIC ANOMALIES IN SOUTHERN QUEENSLAND

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ABSTRACT

Gravity and magnetic anomalies are used to divide the northern Tasman Orogen into crustal blocks, each with a different primary structural trend, and map bands of reworking at the boundaries of these blocks. The boundaries define the margins of the Thomson, Lachlan and New England Orogens in southern Queensland and northern New South Wales. In general, changes in the character of seismic reflections occur near these crustal block or reworking boundaries. Within the Thomson Orogen, the change in thickness of the mid-crustal seismic non-reflective zone near Quilpie corresponds to an inferred minor crustal block boundary. On the southeast boundary of the Thomson Orogen, along the Nebine Ridge, potential field data do not identify any sedimentary troughs, but do indicate that the relatively-thick, less-reflective upper crust is consistent with the ridge having been a reworked, ductile band. Seismic reflections within basement under the western Surat Basin area probably indicate early cratonization, but some are also probably related to the crustal extension events during the formation of the Bowen Basin.

In the eastern part of the transect, mafic crustal underplating of late Cretaceous to mid-Cainozoic age, that was previously inferred to occur due to Tasman Sea extension together with highland uplift and associated mid-Cainozoic intraplate volcanism, is not evident in the expected thickness on the seismic reflection profile.

During the early stages of crustal formation, upper and mid-crustal horizontal density changes were probably accommodated in a thin, weak, lithosphere with local isostatic compensation. At the present time, however, the comparative uniformity in depth and thickness of the reflective lower crust, and the small variation in depth to the base of the crust, shows that the isostatic compensation for these horizontal density changes must be regional.

INTRODUCTION

The Eromanga-Brisbane Geoscience Transect is, for its whole length, situated on cover rocks overlying the northern Tasman Orogen. The geology of the upper-crustal basement rocks is only known with certainty from the few drill cores of wells that penetrated the cover (Murray, 1986). This cover is generally 1 to 3 km in thickness. In the eastern part of the transect, basement outcrops are only 20 to 200 km from the traverse; however, according to interpretations presented in this paper, structures change between the outcropping rocks and the seismic reflection traverse, so that the type of basement under the seismic reflection traverse cannot be inferred directly from the outcropping basement. There are two previous studies inferring the structure of basement from gravity anomalies (Wellman, 1976; Murray & others, 1989). In the interpretations presented in this paper both gravity and magnetic anomalies are used to infer the structure of basement in the region of the geoscience transect. The interpretations are consistent with the two data sets, giving one confidence that features are correctly correlated, and allowing a more complicated model of basement structure.

GRAVITY MAPS

Gravity data in the region of the transect are presented at 1:1 000 000 scale as a mapsheet in this Bulletin (Map 2), and a larger area is presented at a much smaller scale in Figure 1. The sources of the critical gravity data are, (1) Bureau of Mineral Resources Geology & Geophysics (BMR) surveys covering the whole area at 11 km spacing, (2) observations along the 1100 km BMR deep seismic traverse at 0.5 km spacing, and (3) surveys covering the Clarence-Moreton Basin at the eastern end of the traverse at 4 km spacing by Alan Maher of University of Queensland and John Williams of BMR. All the digital survey information is held in the Australian National Gravity Databank, in BMR. The information has been compiled and plotted by BMR computer programs (Murray, 1974, 1977). Observed gravity values were used to calculate a 3 minute grid, and contours are derived from that grid.

MAGNETIC MAPS

Aeromagnetic data in the region of the transect are of two types: (a) New South Wales and Australian

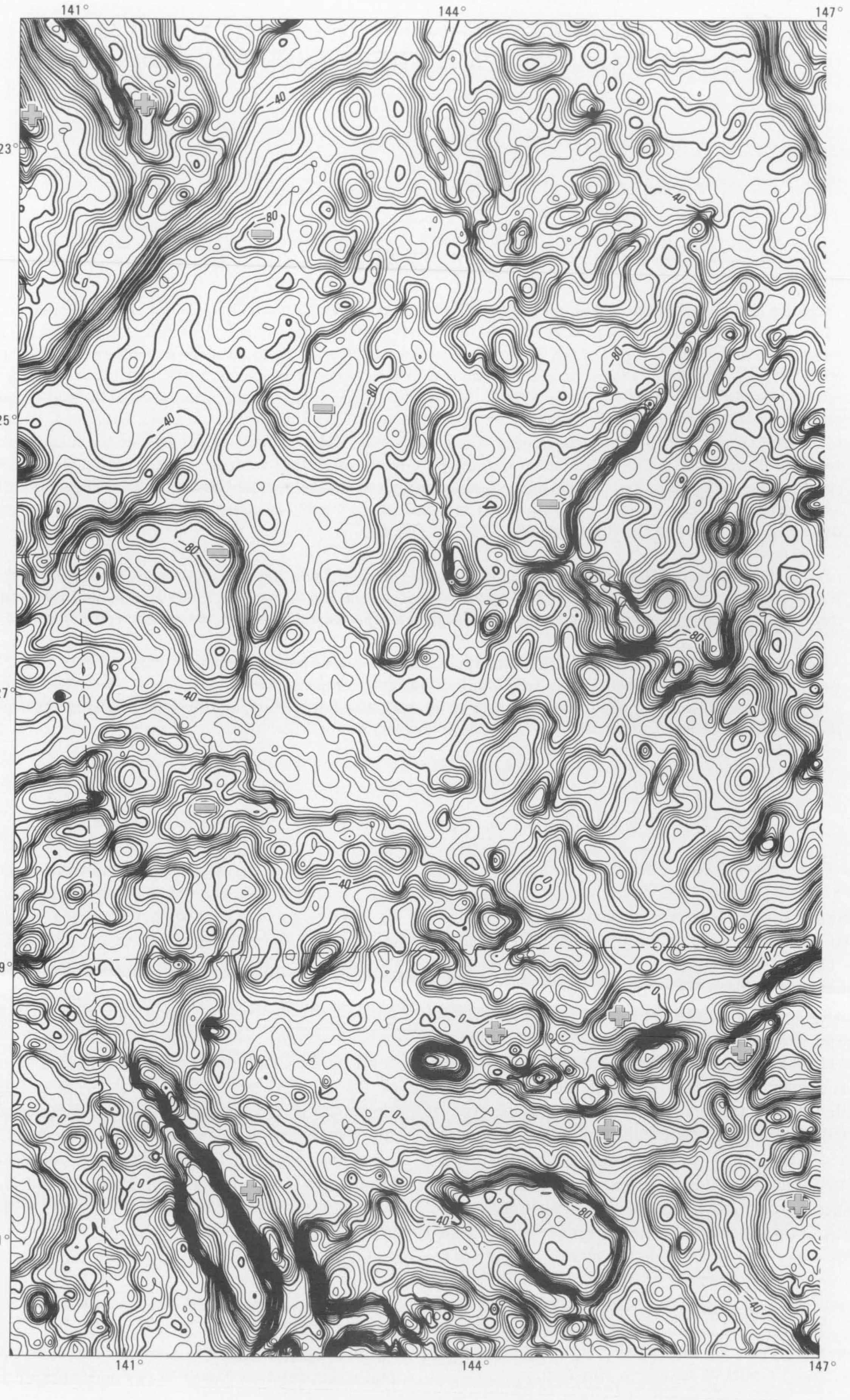
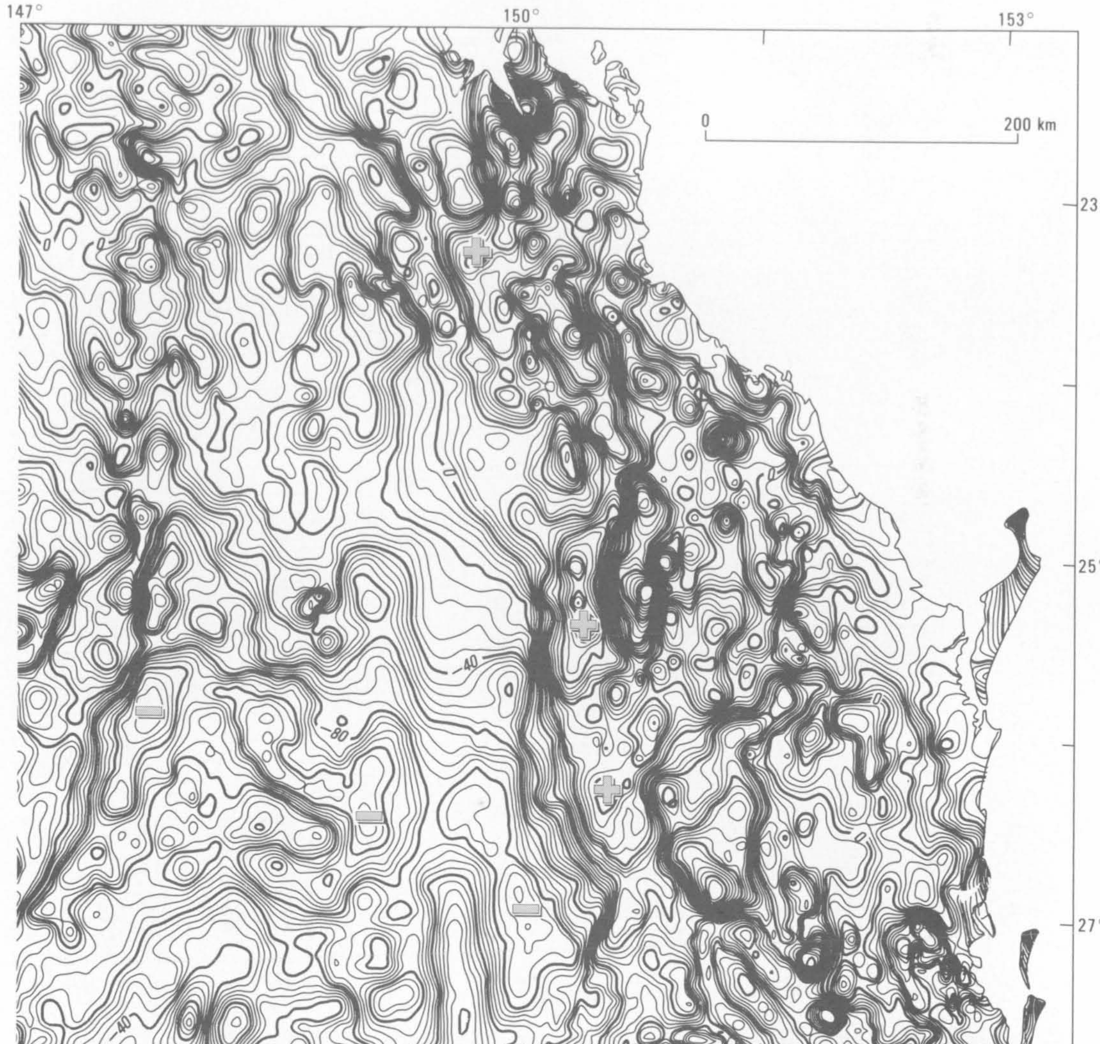
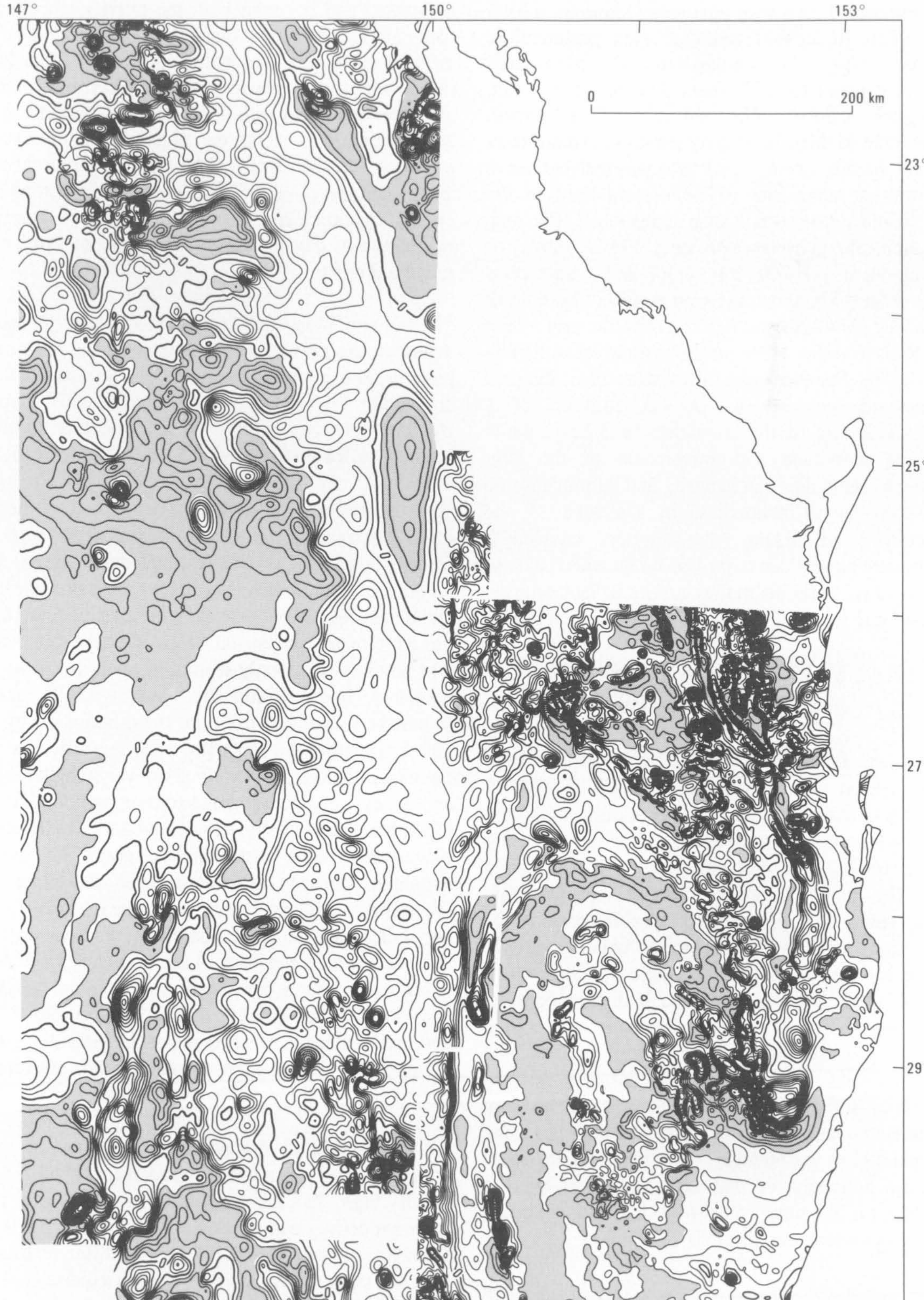


Fig. 1 Bouguer gravity anomalies. Major contours 100 $\mu m.s^{-2}$, others 20 $\mu m.s^{-2}$.



147°
150°
153°
0
200 km
23°
25°
27°
29°
31°
-40
-80
-120
0
25/A/214

Fig. 2 Magnetic anomalies. Total magnetic intensity, upward continued to 3 km. Contour interval 25 nT.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Government surveys with digital recording and 1.5 km flight-line spacing; and (b) older oil company surveys inthe Eromanga Basin with analogue recording, wider-spaced flight lines, and analogue data presentation, deposited with the Australian Commonwealth Government under the Petroleum Search Subsidy Act. An integrated map at flight-level cannot be presented, because some of the oil company surveys have not been digitized in this form, and because the extensive high-amplitude anomalies in the eastern third of the profile would make such a map unreadable. The map shown along the geoscience transect (Fig. 2, Map 3 of this Bulletin at 1:1 000 000 scale) is for anomalies calculated for 3 km above the land surface. For most of the area, the contours were derived from the grid which is the basis for the 1976 magnetic map of Australia (BMR, 1976). For the areas flown since then, the grid was upward-continued to 3 km altitude. This upward-continuing of the anomalies to 3 km altitude leaves the amplitude and wavelength of the long wavelength anomalies unchanged, and eliminates the shorter wavelength anomalies. In the area of the Clarence-Moreton Basin, the shorter wavelength anomalies are largely due to post-basin igneous rocks, so the removal of these anomalies assists in isolating the anomalies due to pre-Clarence-Moreton Basin rocks.

ANALYSIS OF GRAVITY AND MAGNETIC MAPS

Gravity and magnetic anomally trend patterns have long been used to subdivide continents into areas of different deformation direction. Major studies employing this method include those of Provodnikov (1975), using magnetic trends, and Wellman (1976), using gravity trends. The major trends in gravity and magnetic anomaly data can be determined by either tracing the axes of highs and lows or tracing the major gradients. The crust can then be divided into regions, within which the trends are sub-parallel. At the boundary between regions, where trends change direction, one region generally has trends oblique to the boundary, and the other has trends parallel to it. It is inferred that the structures causing the trends oblique to the boundary are older than the boundary, and that those causing the trends parallel to the boundary may be either the same age as the boundary, or may be younger, or may be older than the boundary and moved to the boundary when the boundary was formed.

In this paper, the Tasman orogenic system has been divided into crustal blocks with different deformation direction, and different age structures. The orogenic system is subdivided by analysis of gravity and magnetic data using three techniques that have previously been useful overseas and in Australia: (1) analysis of the trend distribution, (2) identification of major anomalies following the block boundaries, and (3) identification of magnetically quiet zones at crustal block boundaries. Any one of these features on their own in the gravity or magnetic data would not provide sufficient strong evidence for dividing the crust into blocks with a different history. However, where the features are associated and occur in both the gravity and magnetic data, a model of crustal blocks and their boundaries can be formulated with reasonable confidence. The changes in anomaly trends provide evidence that the boundaries are between blocks with different deformation histories. Major anomalies along the block boundaries provide evidence that the identified boundaries represent places were there are changes in the physical properties of the crustal section, and magnetically quiet zones are evidence that the boundaries are adjacent to a zone of reworked crust.

In the area along the Eromanga-Brisbane Geoscience Transect, the gravity and magnetic trends have similar patterns, and, in particular, the regional trend boundaries are similar. Figure 3 shows the gravity trends as mapped by the relative gravity highs, together with the boundaries for the gravity and magnetic trend regions.

At the boundaries between crustal blocks there are often conspicuous gravity and magnetic anomalies. These boundary anomalies generally have larger amplitude, longer wavelength, and are more linear than anomalies elsewhere. However, because of reworking at the edge of an old crustal block, the true boundary of that block is commonly displaced away from the trend boundary, being within the block with the younger trends, and separated from the trend boundary by 20 to 100 km. Within the reworked zone near the true block boundary, there are generally short wavelength magnetic anomalies due to faulting and folding parallel to the true block boundary. In some reworked zones there also are longer-wavelength, small-amplitude gravity and magnetic anomalies caused by structures of the older block (Wellman, 1988a; Rivers & others, 1989). These anomalies, and the corresponding geological structures, die out towards the true block boundary, the decline being caused by a combination of erosion, burial and reworking. The gravity trends of the older block generally extend closest to the block boundary; this is thought to be because the magnetic trends are primarily caused by structures closer to the surface.

At crustal block boundaries two types of gravity anomaly occur: (a) a gravity gradient, and (b) a narrow gravity high. The gravity gradient is thought to reflect different density and seismic profiles, because it is at the boundary of crustal units either with different origin, or with the same origin but different subsequent geological histories. Crustal boundaries A, B, C, F, G, and H of Figure 3 are associated with outstanding gravity gradients. Crustal boundaries A and H, between Proterozoic and Phanerozoic crust, have a gravity low on the side with younger trends, while crustal boundaries B, C, F, and G between Phanerozoic crust with different ages, have a gravity high on the side with older trends. Crustal boundaries D and E are along a narrow gravity high caused by a serpentine belt, with some outcrop. The serpentine is at the geologically defined boundary of a Devonian-Carboniferous accretionary prism to the east.

Magnetic anomalies occurring along crustal boundaries are generally more linear, and of greater amplitude and wavelength than those away from the boundaries. The magnetic anomalies at the crustal boundaries given in Figure 3 have the following three causes, the effects of which are superimposed.

(i) A change in crustal magnetization occurs at all the crustal boundaries. A magnetic low on the side with younger trends occurs at crustal boundaries A, B and H; at boundaries A and H there seems to be no volcanic rock associated with the boundary. A magnetic low on the side with older trends occurs at boundaries C, F and G.

(ii) Between the regional trend boundary and the true crustal block boundary there is generally a band 20 to 100 km wide that is both a regional magnetic low and without large-amplitude, short-wavelength magnetic anomalies (the magnetically quiet band of Kasner & King, 1986). At boundaries where trends are not sub-parallel, this magnetically quiet band/magnetic low is interpreted as an area of low magnetization due to reworking of the edge of the block with older trends, the low magnetization being due to erosion, mechanical deformation, or metamorphism. It cannot caused by to lithological changes because the magnetically quiet band cuts across lithology. These magnetic quiet zones/magnetic lows are well developed at boundaries A, F and G. At boundary B there are two magnetic lows, but neither is in the expected positon; however a magnetic quiet zone is also present. At trend boundaries C, D, and E, where trends are subparallel, there is a magnetic quiet zone, and a magnetic low on the older side of the block boundary. The magnetic low is largely due to lithology differences, but it may also be caused, in part, by demagnetization at lower crustal levels.

(iii) Crustal boundaries D and E have a narrow magnetic high along the boundary, that is known to be due to serpentine intruded along the boundary. A narrow, elongate, magnetic high, extending 200 km along crustal boundary G under cover sediments, is interpreted as a possible serpentine belt, and/or a mafic intrusion.

At most boundaries, the gravity and magnetic anomalies are consistent with the geological structures on the side of the boundary with younger trends being semi-continuous, elongate and gently curved sub-parallel to the boundary. This crust appears to have been deformed plastically against the crust with older trends. However at boundary F south of 28.5°S, the crust with younger trends is broken into short segments (50 to 100 km long) separated by northeast to east-trending strike-slip faults with displacements of 10 to 30 km. The faults do not cross the boundary into the crust with older trends, so the faulting is inferred to date from the time the two crusts were either first deformed jointly, or earlier. The crust must have been brittle at this time, attributable to prior cratonization.

NAMING AND ORIGIN OF CRUSTAL BLOCKS

The crustal block boundaries, defined in this paper by the above means, are similar to those recognized previously on the basis of geological data, and are similar to those in earlier geophysical studies (Wellman, 1976; Murray & others, 1989). In detail, the separate boundaries, defined solely by gravity, magnetics or geology, will show disagreements, because these boundaries refer, on average, to different depths. For example, a boundary that is mapped geologically at the surface as a low-angle thrust may be displaced considerably from the corresponding boundary inferred from gravity data.

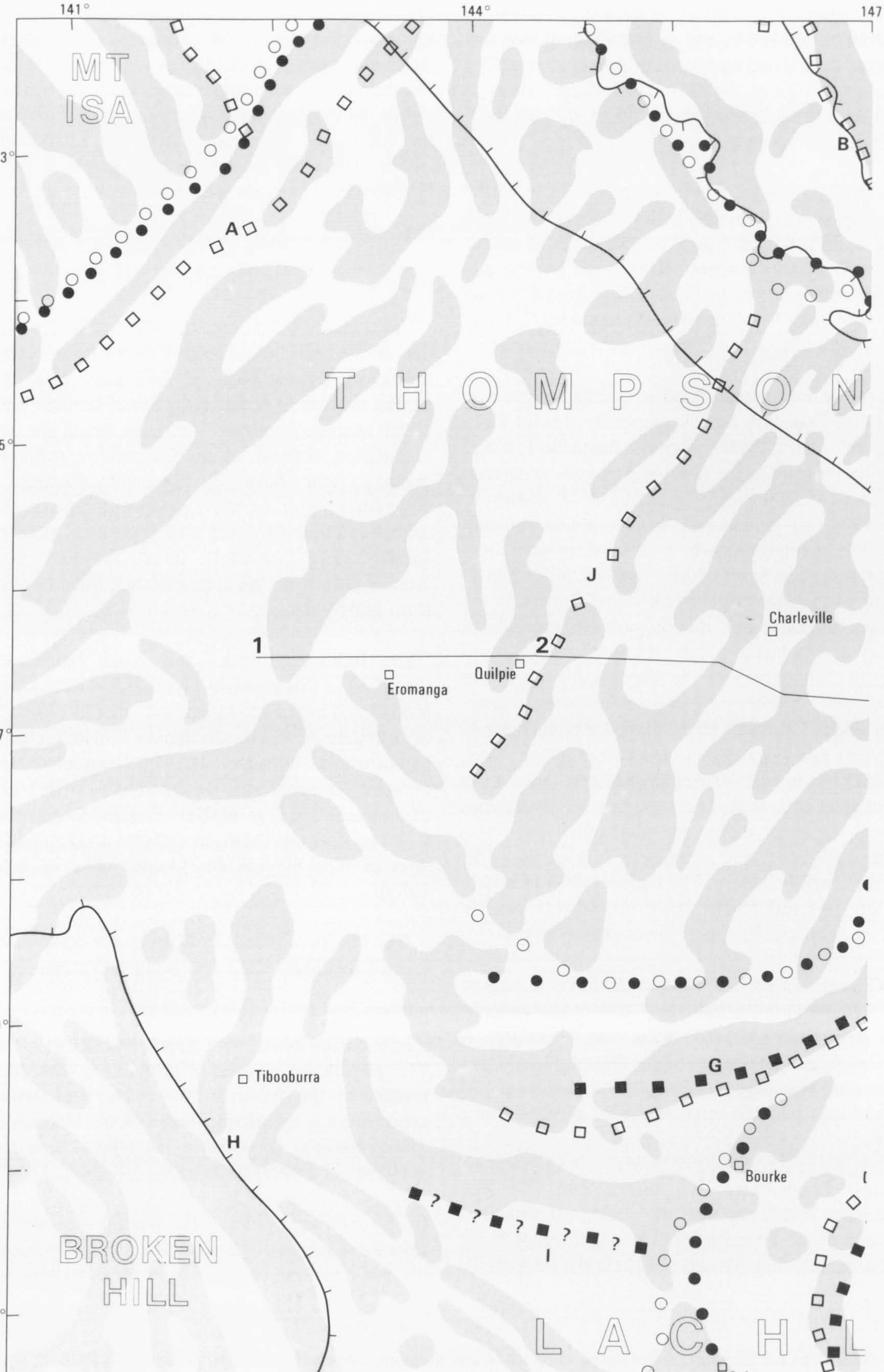
The following blocks and block boundaries are identified in this paper in Figure 3. Proterozoic crust is represented by the Mount Isa and other blocks northwest of boundary A; and by the Broken Hill and other blocks southwest of boundary H. The Tasman Orogen, of Phanerozoic age, forms the remainder of the crust. East of the boundary C is the New England Orogen; west of C, F and G is the Thomson Orogen; and south of G and west of E is the Lachlan Orogen (Murray & others, 1989).

The blocks of crust identified above could have been formed by either of the following two methods:

(a) by total reworking of part of a continent to produce a new crustal block with trends parallel to its margin. The reworking may have been along the continetal margin of the time, or may have been within the continent, e.g. the Mount Isa Block of Wellman (1986). The proportion of old crust incorporated in the new block may be small or large.

(b) by removing crust at the margin of a continent to truncate trends with the older direction, then the addition of new material to the continent. The new material

Fig. 3 (next page) Crustal model based on gravity and magnetic anomalies. Thin lines give the extent of gravity anomaly highs. Rectangles give inferred block boundaries (open rectangles from gravity data, solid rectangles from magnetic data), circles give inferred margins of reworking (open circles form gravity data, solid circles from magnetic data). Medium thickness lines with ticks give the position of prominent magnetic gradients, or magnetic lows. The block boundaries are given letters, that are referred to in the text. Numbers are points on the main east-west seismic reflection profile referred to in the text. The dotted line is the BMR seismic reflection traverse.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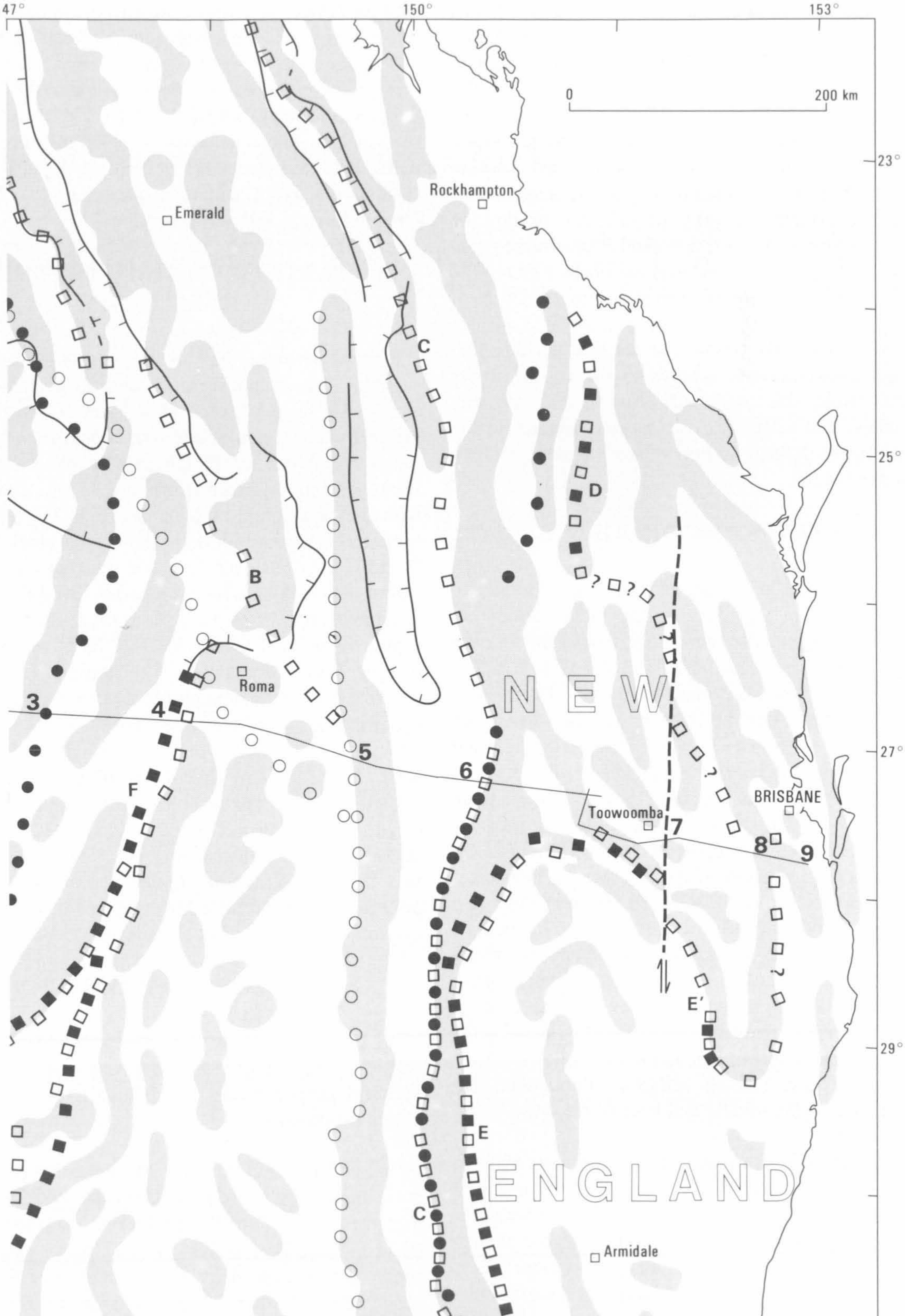
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would have trends sub-parallel to that margin, because of structures formed at the time of accretion and/or (for some crustal blocks) prior to accretion. The eastern part of the New England Orogen, east of boundaries D and E of Fig. 3, is generally agreed to be an accretionary wedge of new material, with major structures formed during, and subsequent to, accretion. Between boundaries I and G the trends appear to be younger than the surrounding areas, possibly, in part, due to reworking within the continent. For the Thomson and Lachlan Orogens there is no conclusive gravity, magnetic or geological evidence bearing on whether or not the western portion is, in part, reworked Precambrian crust, and also no conclusive evidence on whether or not the eastern part is a reworking of the western portion.

Figure 3 shows the position of the seismic reflection and geoscience traverse relative to the crustal blocks inferred from this study of gravity and magnetic anomalies. The relation of the seismic features to the crustal blocks defined by gravity and magnetic data are discussed below.

THOMSON OROGEN

The western half of the seismic traverse from 142° to 148°E (1 to 4 in Fig. 3) is in the older part of the Thomson Orogen with northeast trends; the eastern portion of this area, from 147.2° to 148°E (3 to 4 in Fig. 3), is in the reworked margin of this block. The Thomson Orogen can be divided into two halves by boundary J; a northwest half where northeast gravity and magnetic trends are poorly defined, and a southeast half where the gravity and magnetic trends are relatively straight and regular. On the seismic reflection traverse, the boundary is about Quilpie at 144.2°E (2 in Fig. 3). This boundary correlates with a change in the character of the deep seismic reflections. To the west there is a conspicuous non-reflective zone with an interval of about 6 sec two-way time in the middle/upper crust, and a well-defined band of short, sub-horizontal reflectors in the lower crust. In contrast, to the east the crust is thicker, the non-reflective zone is thinner (about 4 sec two way time at most), and the thicker mid- to lower crust has longer, sloping reflectors throughout, but concentrated in the middle and base of the crust.

The reworked margin to this block, coinciding with the Nebine Ridge (between 3 and 4 in Fig. 3), has a similar structural position to the reworked margin farther south within the Lachlan Orogen outcropping as the Wagga Metamorphic Belt (Wellman, 1988b). This latter belt is a structural high comprising high-grade metamorphic rocks. It appears to have deformed plastically compared with the terrane immediately to the east, and, possibly because of this, it is not cut by sedimentary troughs. The Nebine Ridge has a similar metamorphic grade, being amphibolite grade in places (Murray, 1986), with the grade dropping off to the east and west; it too is a structural high and it is not cut by sedimentary troughs. The seismic reflection profile across the eastern margin of the Thomson Orogen (3 - 4, Fig. 3) has some similarities to that of section 2 - 3 (Fig. 3) to the west, but differs in that the crust thins to the east from 14 to 12.5 sec TWT, and, below the cover, the less reflective upper crust is relatively thick, consistent with deformation extending to greater depth. The gradual change in crustal thickness between that of the Thomson Orogen and that of the Lachlan Orogen, the high metamorphic grade, the absence of sedimentary troughs, and the thicker sequence of deformed rocks, are all features consistent with this margin of the Thomson Orogen being a zone of reworking.

TAROOM TROUGH OF THE BOWEN BASIN

The next portion of the traverse, from 148° to 150°E (4 - 6, Fig. 3) crosses a region between the boundaries C and F. The area between boundaries B-F and C was originally interpreted as one long block. However, near the northern end of F, the gravity pattern becomes irregular, and there are major cross-cutting structures with the same trend as boundary B to the north. The preferred interpretation is for the crustal block between B and C to be younger than, and truncate, the northern end of the block between boundaries C and F. With this interpretation most of the crust in the seismic reflection profile between 4 and 6 (of Fig. 3) is the reworked margin of two crustal blocks. This portion of the transect gave a seismic reflection profile with no clear non-cover upper-crustal non-reflective zone, with reflecting horizons that are shorter and more horizontal than to the west. In the eastern two thirds of this segment the reflections are shorter and more evenly distributed. Two origins for this seismic reflection character are thought possible: (1) the structure may be original layering that was never deformed when the adjacent blocks were deformed, or (2) the structures may be due to the crustal extension associated with forming the Bowen Basin.

NEW ENGLAND OROGEN

The eastern quarter of the transect between 151° and 153.5°E (6 to 9, Fig. 3) is in the New England Orogen. Figure 4 gives a model of this area from Wellman & others (in press).

The orogen is divided into two contrasting sections by major faults marked by serpentine belts: the Yarrol Fault in the north (D in Fig. 3), and the Peel Fault in the south (E in Fig. 3). To the west of these major sutures, the rocks are stratigraphically coherent fore-arc basin sequences, with some volcanic components especially in the north. To the east are coeval strongly-deformed and disrupted accretionary wedge assemblages. Stratotectonic units in the northern part of the orogen (Yarrol Province) are displaced about 200 km east of equivalent units in the south (New England Province). This displacement is believed to be due to large-scale oroclinal bending, possibly associated with major strike-slip faulting (Murray & others, 1987; Korsch & Harrington, 1987).

The oroclinal bending has destroyed the typical palaeogeographic pattern of the orogen (fore-arc basin in the west, and accretionary prism in the east) by repeating forearc basin and accretionary wedge strata, and has almost doubled the width of the orogen.

The seismic reflection profile crosses the New England Orogen just north of the Texas Orocline or megafold, which has bent rocks of the accretionary wedge from their characteristic north-northwest trend to northeast, east and southeast orientations. The gravity and magnetic trends outline the general form of the oroclinal bending (Fig. 3). In particular, the prominent aeromagnetic anomaly associated with serpentinites along the Peel Fault (Fig. 2) can be traced beneath cover rocks part way around the Texas Orocline, and occurs also as the Baryulgil serpentinite on boundary E'. This suggests that at least part of the fore-arc basin sequence should also be bent around the Texas Orocline, and that beneath the Mesozoic platform cover the seismic profile runs entirely through fore-arc strata from 6 to 8 (of Fig. 3), where accretionary wedge rocks crop out. This is consistent with the occurrence of small areas of fore-arc basin strata both to the north and south of the seismic line, at Alice Creek, Mount Barney, and forming the Emu Creek Block. Outcrops of Emu Creek Block extend slightly over the boundary E'; this latter boundary is based on geophysical anomaly data and is thought to exist at depth.

Some sedimentary troughs under the Clarence-Moreton Basin are imaged in seismic reflection profiles. These troughs correlate with gravity lows, so their basin fill is of lower density than basement. The troughs also correlate with magnetic highs, consistent with the interpretation that the fill is, in part, magnetic volcanics similar to Triassic Esk Trough rocks. Magnetic anomalies over the basement between troughs have trends similar to those in the troughs; hence these trends are thought to also reflect pre-Triassic structure. The distribution of these troughs throughout the Clarence-Moreton Basin (Fig. 4) is such that the whole of the Clarence-Moreton Basin can be explained by thermal relaxation following the formation of these troughs by transtensional crustal deformation.

At 152.2°E (7 on Fig. 3), the traverse crosses the proposed northern continuation of the Demon Fault (Shaw, 1969). The gravity and magnetic anomalies show that this fault, and its northern continuation, has a nearly northern strike, and 22 km of dextral displacement along at least 350 km of its length. The movement must have occurred when the igneous rocks in the south were displaced by about 23 km, i.e. after 222 Ma ago (Korsch & others, 1978; McPhie & Fergusson, 1983). The gravity and magnetic data show that the troughs below the Clarence-Moreton Basin sequences were displaced by the fault, so the movement was likely to be after the Esk Trough activity and before the Clarence-Moreton Basin sedimentation; that is from the middle to the end of the Triassic. The fault is not imaged in the seismic

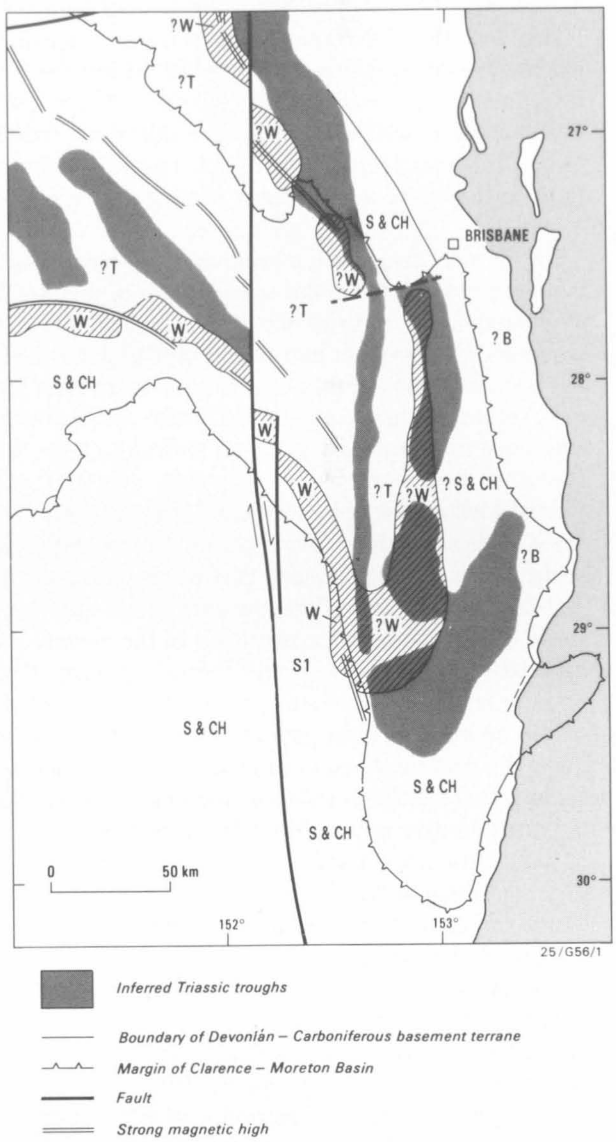

Fig. 4 Geological model of the central part of the New England Orogen. Letters refer to the following tectono-stratigraphic associations and terranes (after Korsch, 1977): S & CH, Sandon and Coffs Harbour; W, Woolomin; T1, non-volcanic older part of Tamworth; B, Beenleigh. The major fault is the Demon Fault (from Wellman, Williams and Maher, submitted).

reflection section; this part of the section is under the Main Range volcanics which created poor recording conditions.

East of 152.8°E (8 to 9 in Fig. 3), the seismic traverse is over the Beenleigh Block, interpreted variously as part of the accretionay wedge, or an exotic terrane unrelated to the other parts of the New England Orogen.

CRUSTAL THICKNESS, UNDERPLATING, AND UPLIFT

This analysis of the gravity and magnetic anomalies has subdivided the region crossed by the seismic traverse into three major crustal blocks which differ in their evolution. The seismic reflections within these crustal blocks differ in character, and this character is attributed to the differing tectonic histories of each crustal block.

The seismic data show a relatively flat crust/mantle boundary (Moho) along the whole length of the profile at approximately 12 to 14 sec TWT, and also, for long sections of the western part of the profile, a relatively flat boundary between the non-reflective upper crust and more reflective lower crust. The Moho has a lower-amplitude relief than the sediment thickness of the late Mesozoic basins, so it is likely that, either (a) the present Moho topography is a post-Mesozoic feature, or (b) it is detached from the upper crustal processes. In some places, in the western part of the traverse the mid-crustal boundary reflects the base of sub-Eromanga Basin troughs, so in the western part of the traverse the boundary must have been formed earlier than Carboniferous when these troughs were deformed. Many of the structures in the reflective lower crust are of lower dip than the structures in the deformed basement, so, in part, the lower crust is inferred to be younger than the non-reflective upper crust, or to be detached from it at some mid-crustal level.

Late Cretaceous-Cainozic underplating of the eastern quarter of the profile has been inferred from (a) the petrology of the mid-Cainozoic shield volcanoes (Ewart & others, 1980), (b) from nodules with high-pressure mineralogy included in the Cainozoic volcanics (Wass & Hollis, 1983; Knutson & others, 1984; Rudnick & others, 1986), (c) from consideration of late Cretaceous to Cainozoic uplift and crustal strength (Wellman, 1979), and, (d) from consideration of the consequences of forming the Tasman Sea 80 to 60 Ma ago by detachment faulting, with the Australian margin forming the upper plate (Lister & others, 1986). This underplating, if it exists, should be thickest either under the larger volcanoes, or under the areas of greatest Late Cretaceous - Cainozoic uplift. If the underplated material was above the Moho, then the added material would depress the Moho from its earlier position. The density contrast of upper crustal rocks with air is about 2.7 $t.m^{-3}$, and that of crust with mantle is 0.4 $t.m^{-3}$ or less, so the amount of depression of the Moho would be more than seven times the regional surface uplift. In the Toowoomba area, the uplift of the land surface, and the base of the Clarence-Moreton Basin, is about 0.4 km, corresponding to an expected Moho depression of over 2.8 km. The seismic reflection data in the Toowoomba area, although relatively poor because of the surface volcanics, does not indicate a depression of the Moho of this magnitude. In the area of the New England highlands, to the south of the seismic traverse, there is uplift of about 0.8 km, corresponding to an expected Moho depression of over 5.6 km. Seismic refraction crustal thicknesses in the New England region (Finlayson, Collins & Wright, this Bulletin) are not more than 5.6 km thicker than in adjacent regions. Hence, either the expected amount of underplated material does not exist in the crust, or it has always been below the Moho.

The presence of layered lower crust, and a horizontal base to the crust, are of interest in terms of isostatic compensation of the variations in density within the upper crust. In the Roma area, gravity anomalies have an amplitude of 100 to 200 $\mu m.s^{-2}$, and a wavelength of 25-35 km. The gravity lows are Roma granite in part, and the gravity highs are Timbury Hills Formation in part. When the granite were emplaced, the lithosphere would have been relatively weak, and the upper crustal horizontal density variation would have associated isostatically-compensating lower-crustal structures, either in or out of phase depending on the type of structural deformation. The original wavelength of the upper and lower crustal structure may have been originally the present wavelength - 25 to 35 km. The horizontal layering in parts of the lower crust, and the almost horizontal base to the crust, probably mean that the original lower crustal structures have been remobilized, over-printed or eliminated since the early Palaeozoic. The present support for the density variation in the upper crust is the relatively strong lower crust and lithosphere with few lateral density variations; this supports the upper crustal inhomogeneities by regional isostatic compensation.

CONCLUSIONS

(1) Using gravity and magnetic anomalies, the Tasman Orogen can be divided into crustal blocks. Each block has a similar tectonic grain, and the blocks with inferred younger trends truncate and rework the margins of the blocks with inferred older trends. The block boundaries, subdivision boundaries, and boundaries of reworking, correspond to changes in the character of the seismic reflection images. Using these crustal block subdivisions, the Eromanga-Brisbane Geoscience Transect is shown to cross the Thomson, Lachlan, and New England orogens.

(2) Between the western and eastern halves of the Thomson Orogen, there is a thinning of the seismically non-reflective middle crust, corresponding to a change in character of the gravity and magnetic anomalies from irregular, to relatively straight and regular. The Nebine Ridge area, forming the deformed southeastern margin of the Thomson Orogen, and the band to the south in the same structural position, are inferred to have been relatively ductile, on the basis of their relatively high metamorphic grade, the semicontinuous gravity and magnetic anomalies, and the absence of fault troughs in section. In the Bowin Basin area, the seismic reflections in the middle and lower crust are more uniformly distributed, shorter and more horizontal. These features may be due to deformation during crustal extension to form the Bowen Basin, or reflect a lack of crustal

shortening in a strip of crust that is inferred, from the pattern of gravity and magnetic anomalies to the south, to have been cratonized early.

(3) In the New England Orogen, interpretations of geology, gravity and magnetic anomalies, are consistent with most of the seismic reflection traverse being within the fore-arc basin sequence. Sedimentary troughs, seen in the seismic reflection profiles below the Clarence-Moreton Basin sediments, are associated with magnetic highs and gravity lows, consistent with the troughs containing some magnetic volcanics, which are correlatives of the Esk Trough rocks of Triassic age.

(4) Many authors infer that underplated material was intruded in association with Cainozoic intra-plate volcanoes and late-Cretaceous to Cainozoic detachment faulting to form the Tasman Sea. This underplating is inferred to be the cause of the uplift of the Eastern Highlands. The thickness of lower crustal underplating inferred from the amount of uplift, is not found in the seismic reflection and refraction results.

(5) Horizontal density variations in the upper and middle crust must have originally been isostatically balanced by local compensation at the base of the crust. The existence now of a horizontally layered lower crust, and a near-horizontal base to the crust, mean that these density variations must now be balanced by regional isostatic compensation.

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TECTONICS OF THE NEW ENGLAND OROGEN

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ABSTRACT

The New England Orogen of eastern Australia evolved during the Devonian and Carboniferous in a convergent plate margin tectonic setting related to a west-dipping subduction system. Parts of the volcanic arc, fore-arc basin and accretionary wedge are still preserved in the orogen. The earlier history of the orogen is fragmentary, but the geochemistry of ophiolitic units and the petrography of sandstones suggest that, from the Cambrian to Silurian, the orogen was probably island arc-related in origin. The later history of the orogen has involved strike-slip faulting and major oroclinal bending, with about 450-500 km of displacement being involved. This was followed by massive amounts of volcanism and plutonism in the Late Permian and Early Triassic when the orogen was in an extensional, back-arc setting.

INTRODUCTION

The easternmost geological unit on the Eromanga - Brisbane Geoscience Transect is the middle Palaeozoic to Mesozoic New England Orogen (Fig. 1). Comprehensive summaries on the geology of the Queensland and New South Wales sectors of the Orogen were included in Hill & Denmead (1960) and Packham (1969) respectively. More recent articles, including those by Marsden (1972), Harrington (1974), Leitch (1974), Korsch (1977), Day & others (1978, 1983), Korsch & Harrington (1981), Murray (1986), Degeling & others (1986), Murray & others (1987), and Flood (1988), highlight the complicated nature of the orogen. This paper provides a summary of the geology of the New England Orogen by outlining current interpretations, reviewing recent developments, and highlighting some of the unsolved problems. Data that reflect on the nature of the crust beneath the orogen, and how it has developed through time, are also discussed.

Most recent interpretations of the orogen prefer a convergent plate margin model associated with a west-dipping subduction zone, with refinements due to concepts of terrane analysis (e.g. Scheibner, 1973, 1976, 1985; Leitch, 1975a; Korsch, 1978; Day & others, 1978; Crook, 1980; Cawood, 1982, 1984a; Fergusson, 1984a; Cawood & Leitch, 1985; Murray, 1986; Murray & others, 1987; and several papers in the following volumes: Flood & Runnegar, 1982; Leitch & Scheibner, 1987a; Kleeman, 1988).

In the northern New England Orogen, a tripartite subdivision of the Devonian and Carboniferous into an arc/fore-arc/accretionary wedge assemblage has been proposed, and consists of the Connors and Auburn arches in the west (magmatic arc), the Yarrol Belt in the centre (fore-arc basin) and the Coastal Block and its southern equivalents in the east (accretionary wedge) (Fig. 1) (for full details see Day & others, 1978, 1983; Murray, 1986). In the New South Wales sector, the arc is less well developed, but the fore-arc basin is represented by the Tamworth Belt, and possibly the Hastings Block (Fig. 2), and the accretionary wedge by the Tablelands Complex comprising the Woolomin-Texas and Coffs Harbour blocks (Fig. 1).

In temporal terms, the subduction-related model is applicable for at least the Devonian and Carboniferous, but there are hints of an earlier history to the orogen that extend back to at least the Cambrian. Rutland (1976) even suggested that a lower crust of continental Precambrian material extended under the New England Orogen.

EARLY HISTORY OF THE NEW ENGLAND OROGEN

Evidence for the early history of the orogen is mostly confined to a narrow zone associated with the Peel Fault and its southeastern extension towards the coast (Fig. 2). Cambrian (Cawood, 1976), Ordovician (Philip, 1966; Packham, 1969; Hall, 1975; Wass & Dennis, 1977), and Silurian (Hall, 1970) fossils have been reported, but most of these are probably from fault slivers or allochthonous blocks and may not date the age of the confining rocks.

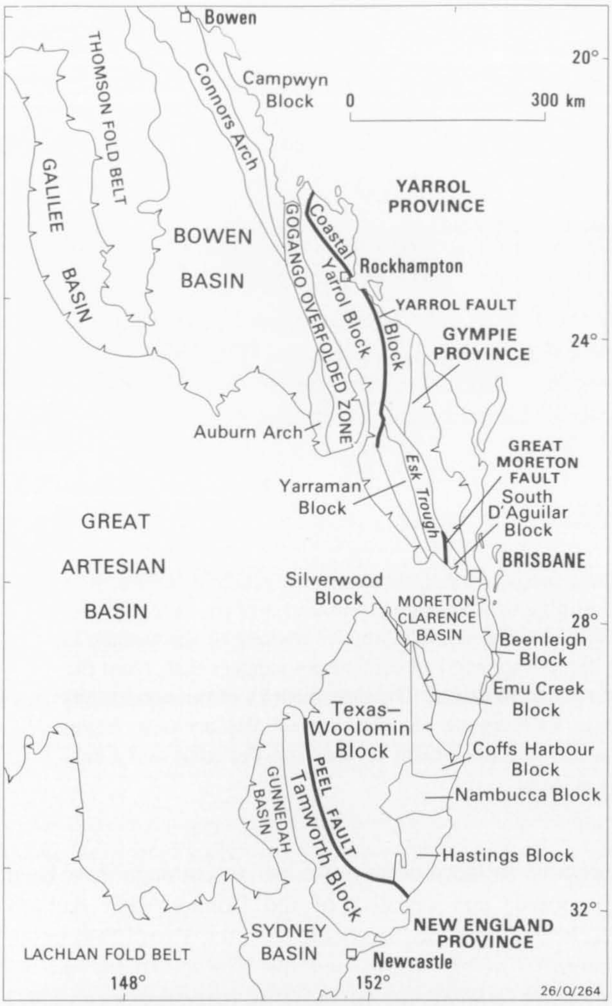

Fig. 1 General map of eastern Australia showing the present structural units in the New England Orogen, including the locations of the New England, Yarrol and Gympie provinces.

The Cambrian and Ordovician fossils reported by Cawood (1976) occur in a sequence in the eastern part of the Tamworth Belt, which was volcanic-derived (Leitch & Cawood, 1987) and was probably deposited close to an island arc system. Cross (1983) considered that the Pigna Barney Ophiolitic Complex in the same area was part of an island arc - ophiolitic association. A Pb-Pb zircon age of 436 $\pm$ 9 Ma for a tonalite of the Pola Fogal suite (in spatial association with the Pigna Barney Ophiolitic Complex) was obtained by D.L. Kimbrough, K.C. Cross and R.J. Korsch (unpublished data), which they interpreted as the minimum age for crystallisation of the suite. Thus, rocks of the Tamworth Belt in the Pigna Barney area were deposited on island-arc related rocks of probable early Silurian age. Hence, the eastern margin of the Tamworth Belt in the vicinity of the present Peel Fault system may have had a very complicated evolutionary history from the Cambrian to the Early Devonian, related to the presence of a nearby island arc (or arcs) and island-arc type ophiolitic rocks.

Other isotopic dates also provide some constraints on the early history of the New England Orogen. Glaucophanes from blueschists in melange of the Port Macquarie Block have K-Ar isotopic ages of 383 $\pm$ 5 Ma (Webb, in Pogson & Hilyard, 1981) and 444 Ma (Lanphere, in Scheibner, 1985).

DEVONIAN-CARBONIFEROUS MAGMATIC ARC

In New England, the volcanic arc, poorly exposed as minor Devonian breccias and Carboniferous felsic volcanic centres near Boggabri and Gunnedah and in the Carboniferous of the Hunter Valley, is inferred to have been located to the west of the Tamworth Belt. Day & others (1978) suggested that the position of the arc, buried beneath the Sydney-Gunnedah Basin to the west of the New England Orogen, was indicated by a positive gravity ridge, i.e. the Meandarra Gravity Ridge (see Wellman, this Bulletin). Recent interpretation of a filtered gravity image, however, clearly indicates that the Meandarra Gravity Ridge is parallel to the gravity ridge defining the exposed Devonian-Carboniferous volcanic arc of the northern New England Orogen (Murray & others, 1989b), so the suggestion of Day & others is incorrect.

Strong evidence for the existence of an arc is provided by the composition of the sandstones from the Tamworth Belt and Tablelands Complex, with virtually all of them being arc-derived volcaniclastics (Crook, 1960a, 1960b; Korsch 1977, 1984; Cawood, 1983a). Using the geochemistry of sandstones from the Late Devonian Baldwin Formation, Chappell (1968) considered that they were derived from tholeiitic andesites. Cawood (1983a) used detrital clinopyroxene geochemistry to show that sandstones of principally Devonian age from the Tamworth Belt came mainly from subalkaline volcanic arc material, and Morris (1988a) has extended this study by the application of discriminant function analysis.

Vallance (1969), Offler (1979, 1982) and Morris (1988b) have examined the geochemistry of altered basalts (spilites) and albite dolerites that are intercalated with the Devonian Tamworth Group sediments on the eastern side of the Tamworth Belt. Island arc tholeiitic, alkaline, and calcalkaline compositions indicate significant heterogeneity in the source regions (or multiple source regions); the rocks are probably not associated with the western magmatic arc, but possibly are related to rifting in the outer fore-arc region (Morris, 1988b). The pre-Late Devonian sediments of the Tamworth Belt are volcaniclastic and obviously arc-related; insufficient of this ancient system remains to confidently identify the arc polarity (see also Cross, 1983).

The geochemistry of volcanic clasts from conglomerate lenses in the Late Devonian - Early Carboniferous Goonoo Goonoo Mudstone indicate that they were

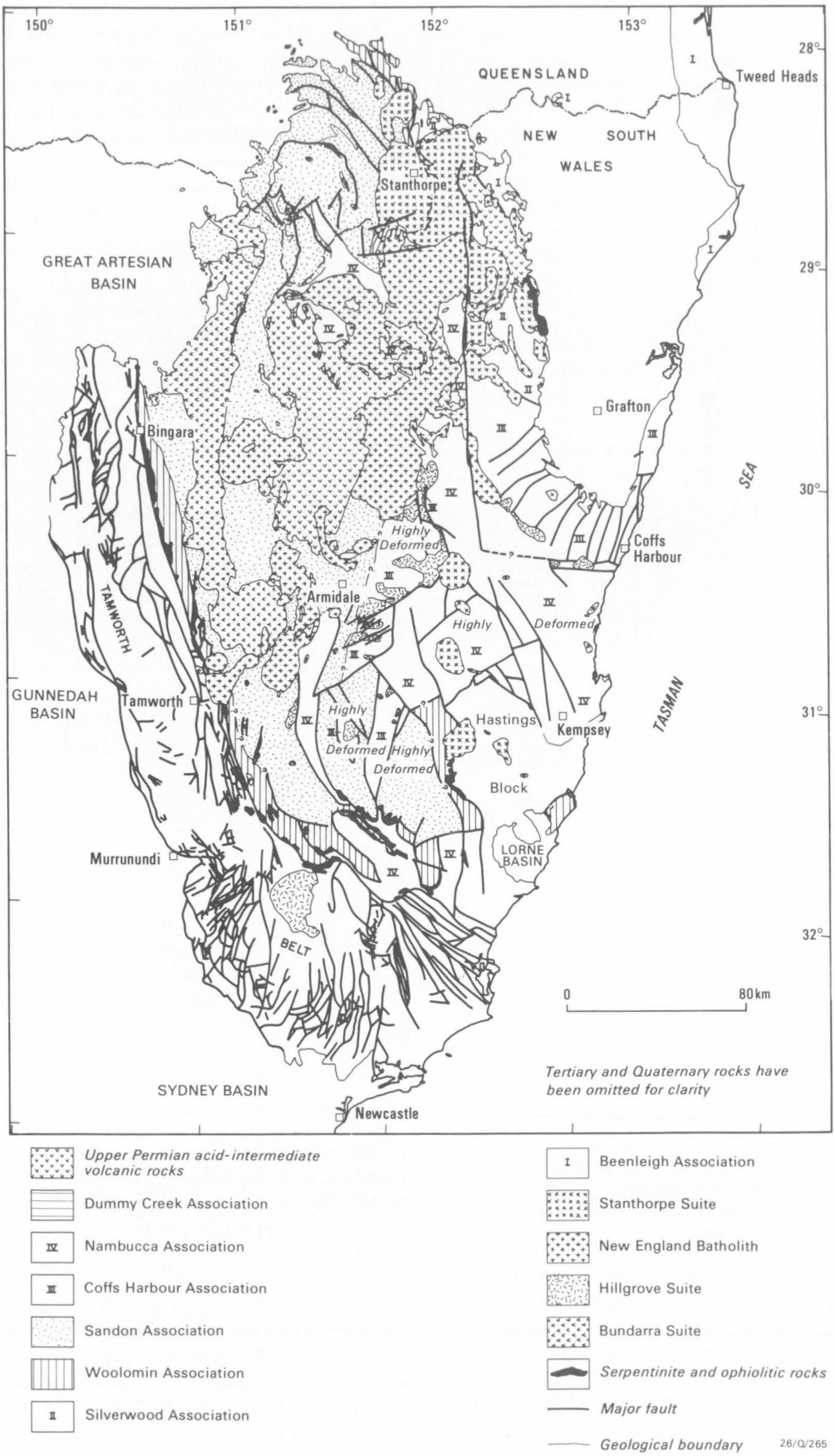

Fig. 2 Map showing the present-day distribution of lithological associations recognised by Korsch (1977) in the New England Province of the New England Orogen. The Woolomin, Sandon and Coffs Harbour associations represent the Devonian-Carboniferous accretionary wedge. The Beenleigh association is a suspect terrane consisting of accretionary wedge material.

derived from calcalkaline volcanic arc rocks that were erupted through thinned continental crust (Morris, 1988c). Late Early and Late Carboniferous dacites and andesites are also calcalkaline in character (Wilkinson, 1971; Cook & Dawson, 1988). The Late Carboniferous lavas and pyroclastics are markedly different in chemistry to Devonian volcanics and volcaniclastic sediments. The volcanology of some of these ignimbrite sheets has been described in detail by McPhie (1983, 1986), who proposed an Andean continental margin analogue for their setting (McPhie, 1987; see also Leitch, 1974).

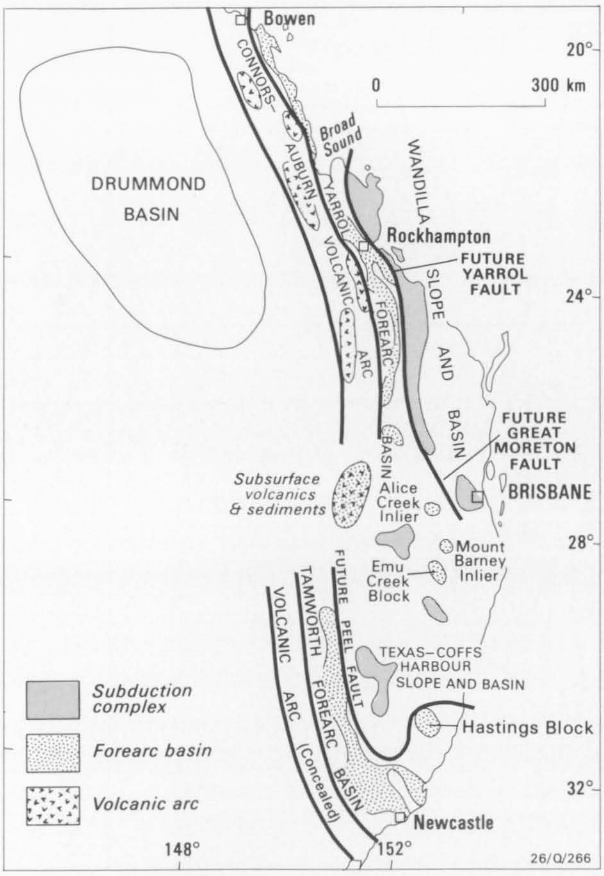

Fig. 3 Map (after Murray & others, 1987) of the Late Devonian - Early Carboniferous palaeogeography (not palinspastically restored) showing present-day distribution of the volcanic arc, fore-arc basin and accretionary wedge (subduction complex).

In the northern part of the New England Orogen (Yarrol Province, Fig. 1), the oldest volcanic arc rocks are exposed in isolated fault blocks and inliers within the Yarrol Belt. The strata range in age from Late Silurian to Middle Devonian. The precise nature and composition of the volcanics are not well known. Andesitic flows and pyroclastics are probably dominant, accompanied by dacitic pyroclastics and tholeiitic basalt flows. Associated sediments are medium to fine-grained clastics of volcanic provenance, coralline limestone and minor chert, and conglomerate (Malone & others, 1969; Kirkegaard & others, 1970; Dear & others, 1971; Whitaker & others, 1974; Olgers & others, 1974). The assemblage was compared with that of modern island arcs by Marsden (1972); Day & others (1978) grouped all the Late Silurian - Middle Devonian fault blocks and inliers together as the Calliope Island Arc. The arc, or arcs, must have been separated from the Australian continent by an ocean basin of uncertain extent (Veevers & others, 1982), and therefore represent an exotic terrane or terranes.

An alternative interpretation is that the late Silurian - Middle Devonian sequences were part of a continuous belt along the northeastern edge of the Australian continent, which was an active, Andean-type continental margin (Henderson, 1980). This view was supported by Morand (1989), who pointed out that the presence of locally abundant silicic volcanics and detrital quartz in the associated sediments is characteristic of continental margin arcs rather than island arcs (cf. Baker, 1982). However, the most voluminous silicic volcanics in the area southwest of Rockhampton have unusual compositions with remarkably high Na_2O/K_2O ratios (7.1 - 19.4; Cornelius, 1969), which are probably more compatible with an island arc setting. Additional factors, which support the island arc hypothesis, are: the lack of evidence for the existence of early Palaeozoic basement except for the presence of clasts of Late Ordovician limestone in conglomerate of the Silverwood Group near the Queensland - New South Wales border (Wass & Dennis, 1977); and the local juxtaposition of volcanic rocks and coralline limestones.

The relationship of the Calliope Island Arc or continental margin arc to the concealed coeval arc west of the Tamworth Belt is unknown. Some authors (e.g. Bryan, 1925; Cawood & Leitch, 1985) have regarded the Silverwood Group, which was included in the Calliope Island Arc by Day & others (1978), as a correlative of the Early-Middle Devonian Tamworth Group of the Tamworth Belt.

The rocks of the Calliope Island Arc are overlain by Late Devonian and younger fore-arc basin strata of the Yarrol Belt. Therefore, if the arc was an exotic terrane, it had reached its present position relative to the remainder of the New England Orogen by the end of the Middle Devonian, with the accretion event being represented by an unconformity with the overlying Late Devonian volcaniclastic rocks.

Thick sequences of volcanics assumed to be representative of the Late Devonian - Carboniferous arc form a belt to the west. This arc was of continental margin type (Marsden, 1972). In the north (Connors Arch), andesite flows are the main rock type. Basalt is abundant locally and porphyritic rhyolitic and dacitic lavas are common. Andesitic and silicic pyroclastics constitute up to half the unit (Malone & others, 1966, 1969; Jensen & others, 1966). In the south (Auburn Arch), dacitic and rhyolitic pyroclastics are dominant, and only locally are andesites as abundant as the silicic volcanics (Dear & others, 1971; Whitaker & others, 1974).

The age limits of the volcanics are not well defined. They have generally been assigned a Late Devonian - Early Carboniferous age because, (1) they are similar in lithology and stratigraphic position to volcanics interbedded with fossiliferous marine sequences along the western margin of the fore-arc basin (Jensen & others, 1966; Malone & others, 1969; Dear & others, 1971); (2) they are intruded by Middle Carboniferous granitoids (Rb-Sr isochron of 316 $\pm$ 15 Ma; Webb & McDougall, 1968); and (3) an Early Carboniferous K-Ar date of 343 Ma has been obtained from a hornblende-bearing dacitic tuff at the southern end of the exposed arc (Whitaker & others, 1974). As in New England, the arc has provided voluminous clastic material that has been deposited in the fore-arc basin and deeper settings to the east.

DEVONIAN - CARBONIFEROUS FORE-ARC BASIN

To the west of the Peel and Yarrol faults (Figs. 1, 3), the Tamworth Belt in New England and the Yarrol Belt in the Yarrol Province are usually interpreted to be parts of a once-continuous fore-arc basin, located between the volcanic arc to the west and the accretionary wedge to the east (Fig. 4).

The Devonian to Carboniferous history is well documented, with stratigraphic and palaeogeographic aspects summarised by Marsden (1972), Day & others (1978), Roberts & Engel (1980), Mory (1982), Roberts (1985), Murray (1986) and Murray & others (1987). As such, the fore-arc basin is probably the best understood unit in the subduction-related system. Essentially, the basin deepened towards the east and consists predominantly of continental to shallow marine (shelf) clastic sediments that were later deformed into shallowly plunging, upright regional folds and associated thrusts. As discussed above, the provenance of the clastic sediments was dominated by the volcanic arc to the west. Early Carboniferous sedimentation was characterised by the repeated development of calcareous ooliths and oolitic limestones on shallow banks during periods of decreased terrigenous deposition (Roberts & Engel, 1980). Ooliths were transported eastwards to deeper parts of the basin, where oolitic greywackes were deposited, and down the continental slope beyond.

DEVONIAN - CARBONIFEROUS ACCRETIONARY WEDGE

Despite reservations expressed by Flood (1988), there is now consensus that, during the Devonian and Carboniferous, the New England Orogen developed at a convergent plate margin related to a west-dipping subduction zone (see references listed in the Introduction). In the east, the Tablelands Complex (Woolomin slope and basin) in New England and the Coastal Block and South D'Aguilar Block (Wandilla slope and basin) in the Yarrol Province (Figs. 1, 3) are interpreted as a once-continuous accretionary wedge (Fig. 4) that grew oceanwards by accreting trench-fill volcaniclastic turbidites (derived from a magmatic arc) and minor amounts of oceanic crust (basalt, chert, pelagic mudstone). Until recently, age constraints were poor for the deposition of the clastic sediments and growth of the wedge (see discussions in Korsch, 1977). An almost continuous belt of oolitic greywackes in the central part of the accretionary wedge is assumed to have been sourced from the fore-arc basins to the west, and therefore to be of Early Carboniferous age (Korsch, 1984; Murray & others, 1987). This age is supported by the occurrence in the oolitic greywackes of elliptical crinoid stems (Fleming & others, 1974, 1975), which are restricted to a single Early Carboniferous (Late Tournaisian - Early Visean) brachiopod zone in the fore-arc basin strata (Roberts, 1987).

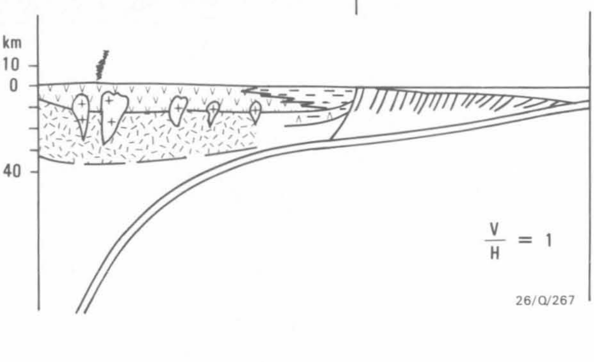

Fig. 4 Cartoon-cross section (after Fergusson & others, 1988) across the Yarrol Province to illustrate the convergent margin model for the Late Devonian and Carboniferous. The Yarrol terrane is a volcanic arc and fore-arc basin and the Wandilla terrane is an accretionary wedge. The word terrane is used to indicate that the two sub-provinces are suspect terranes that could have moved relative to one another by strike-slip faulting.

Several recent studies have been very successful in extracting radiolarians, and sometimes conodonts, particularly from cherts and siltstones (e.g. Ishiga & others, 1987, 1988a, 1988b; Ishiga, 1988; Aitchison, 1988a, 1988b, 1988c; Blake & Murchey, 1988a, 1988b). These studies place considerable constraints on the age of formation of the accretionary wedge, and suggest that most chert has an age range of Late Devonian to Early Carboniferous implying substantial accretion in the mid-Carboniferous, which is more limited than previously thought. An Rb-Sr whole rock isochron of 318 $\pm$ 8 Ma for low-grade regional metamorphism of accretionary wedge rocks in the Coffs Harbour Block (Graham & Korsch, 1985) places an upper limit on the formation of the wedge. This age was confirmed by K-Ar ages of 312$\pm$6 and 318$\pm$9 Ma from Oxley Metamorphics in the central Tablelands Complex (Watanabe & others, 1988a).

Provenance studies on sandstones from the accretionary wedge (e.g. Korsch, 1977, 1984; Cawood, 1983a; Fergusson, 1984a) all show that they are volcaniclastic in character. Cawood (1983a) and Korsch (1984) suggested that, in the New England area, there was a provenance linkage between the Tamworth Belt and the Tablelands Complex, i.e. detritus in both were derived from the same volcanic arc. On the other hand, Flood (1988) raises the possibility that they are not related. Nevertheless, the Early Carboniferous calcareous ooliths and associated fossil debris (McKellar, 1967; Fleming & others, 1974, 1975) provide an indisputable provenance linkage between the fore-arc basin and the accretionary wedge. Although a considerable amount of strike-slip faulting (e.g. Offler & Williams, 1987) at the present position of the Peel Fault has probably occurred (and hence rocks now adjacent across the Peel Fault were originally deposited some distance apart), we consider that the evidence is sufficiently strong to invoke the existence of only a single volcanic arc in the Late Devonian and Carboniferous providing detritus to both the fore arc and accretionary wedge.

The timing of the intense deformation(s) observed within the accretionary wedge (e.g. Korsch, 1981a,

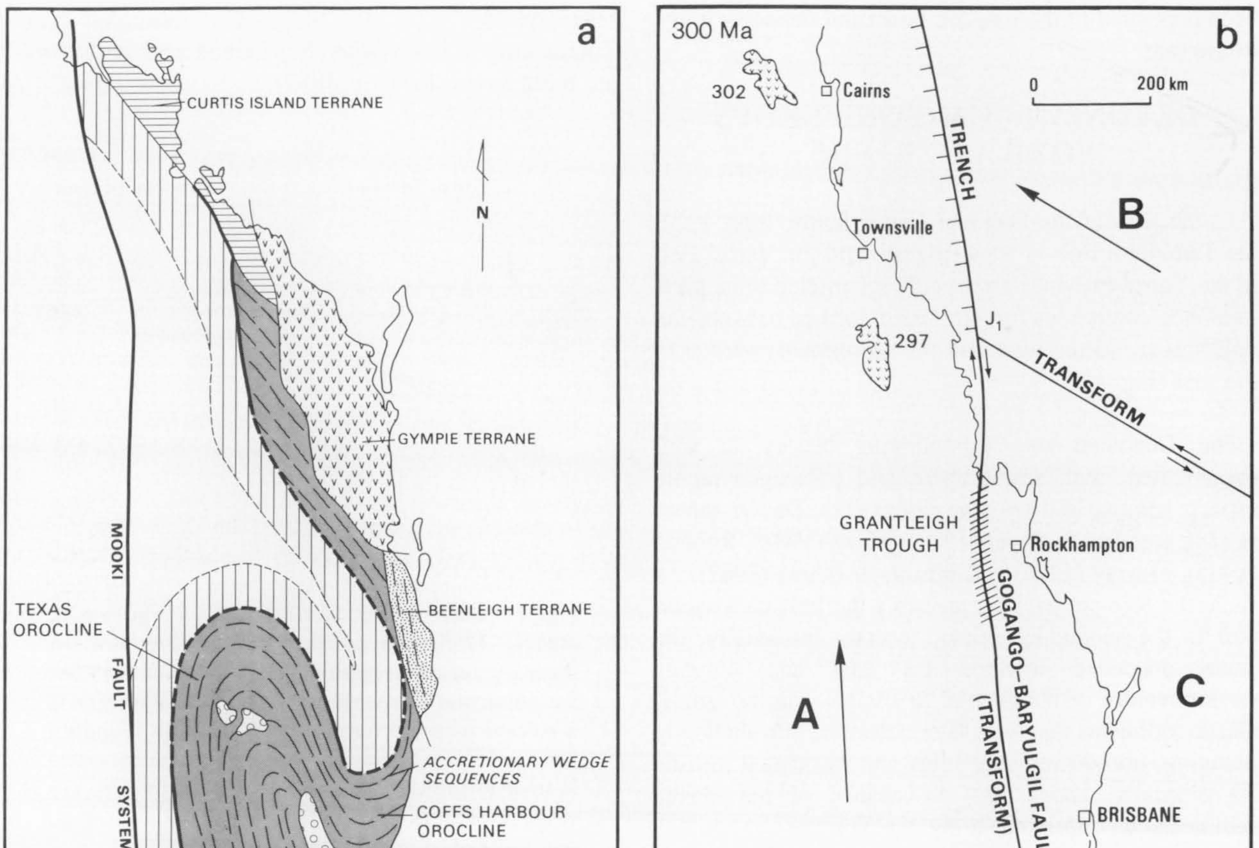

Fig. 5 Orocline models of (a) Korsch & Harrington (1987) and (b) Murray & others (1987). The major difference is in the location of the controlling strike-slip fault zone, which Korsch & Harrington infer to be their Mooki Fault System, whereas Murray & others interpret it to be along their postulated Gogango-Baryulgil transform fault zone.

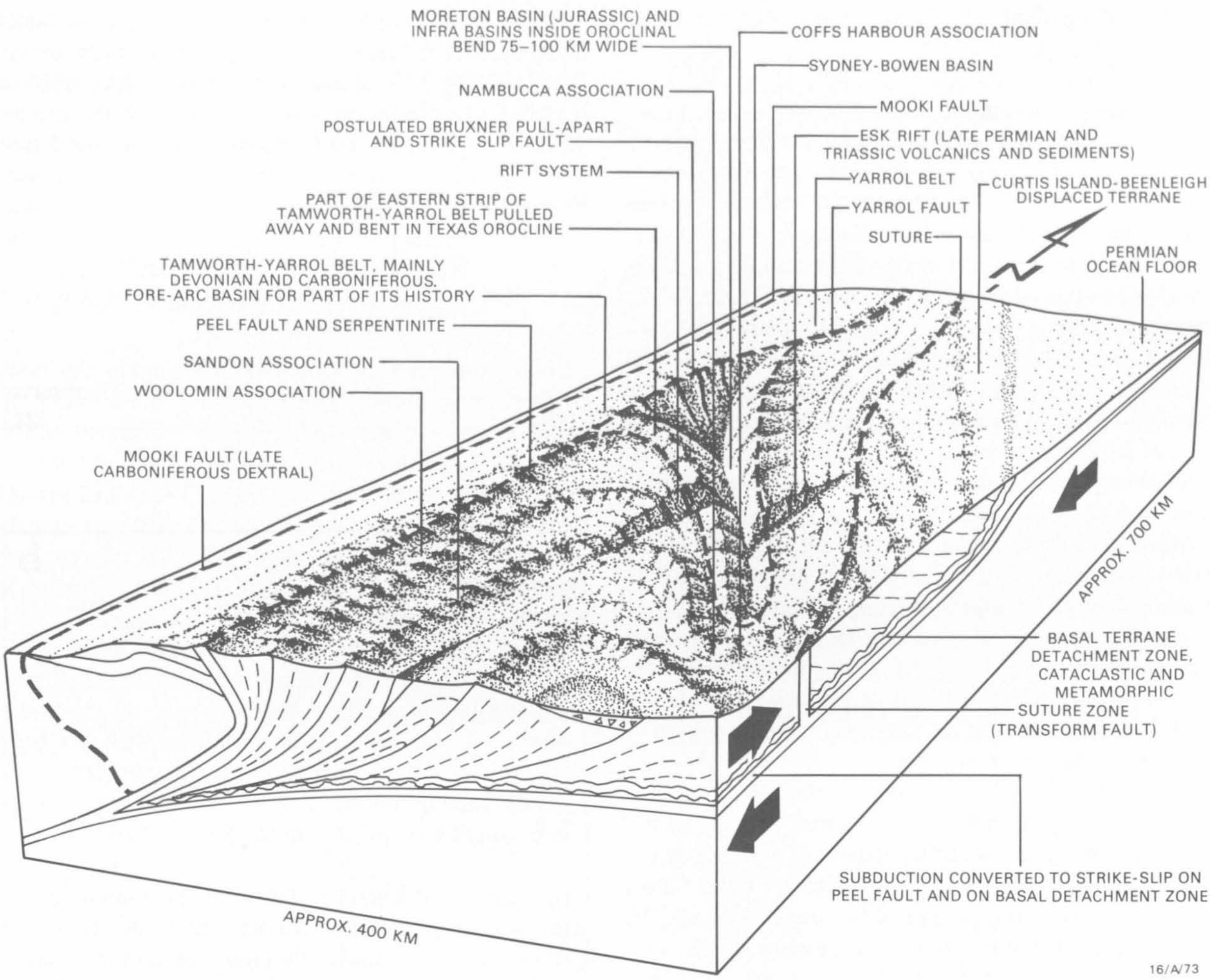

Fig. 6 Block diagram interpretation of the oroclinal bending by Harrington & Korsch (1987) showing sliding and bending of the accretionary wedge above a detachment zone which is postulated to be the top of the subducted oceanic plate. Note that the Peel Fault and the adjoining serpentine belt are bent in the orocline. The Mooki Fault is postulated in this diagram to be the main western boundary of the orocline region, but the attempt to draw it on this diagram showed the need for the deep seismic line that is described in this Bulletin. The seismic line was roughly across the centre of the diagram from west to east. On Fig. 5a the eastern part of the seismic line passed through the word 'Mooki' to the word 'Beenleigh'.

1981b; Fergusson, 1982, 1984b; Cawood, 1982) is poorly constrained but, if the accretionary wedge model is correct, the deformation should be diachronous and also migrate oceanwards with time. Melange zones are a common product of this deformation (e.g. Fergusson, 1984b; Fergusson & others, 1985; Cross & others, 1987; Fergusson & others, 1988). In the Coffs Harbour Block, the major deformation was synchronous with the mid to Late Carboniferous low-grade regional metamorphism (Korsch, 1978; Graham & Korsch, 1985). An older history for parts of the accretionary wedge is suggested by a K-Ar whole rock age of 389±18 Ma for a schist from the upper Manning area (Watanabe & others, 1988b) and an age of 387±12 Ma for an Rb-Sr whole rock isochron on metasedimentary rocks from the Wongwibinda metamorphic complex (Hensel & others, 1985).

OROCLINAL BENDING

The northern part of the accretionary wedge in New England has been involved in oroclinal bending which has produced mega-folds in the Texas and Coffs Harbour areas and possibly elsewhere, and has led to a widening of the orogen in this area with apparent repetition of some internal elements (Fig. 5). Lucas (1960) recognised that the rocks in the Texas area were folded into an arc, and Korsch (1975) postulated oroclinal bends in rocks of the Coffs Harbour Block and southern part of the Tamworth Belt and Hastings Block. Flood & Fergusson (1982) and Fergusson & Flood (1984) considered that the Texas and Coffs Harbour oroclines were part of the same Z-shaped megafold because regional units could be traced around the fold. Recently, interpretation of aeromagnetic and gravity anomalies for the region by Wellman & Korsch (1988, in preparation) and Wellman (this Bulletin) has confirmed the oroclinal bending in the Texas and Coffs Harbour regions.

The timing of formation of the orocline is contentious. Korsch & Harrington (1987) prefer an Early to mid-Permian age (ca. 280-265 Ma), because they related strong deformation of Permian slivers in the Texas area and intense deformation (Leitch, 1978) and metamorphism (Leitch, 1975b) of the Nambucca Block in the Permian (Leitch & McDougall, 1979) to the oroclinal bending (Fig. 5a). Murray & others (1987; see also Lennox & Roberts, 1988, 1989) prefer a Late Carboniferous age (ca. 310-300 Ma) to account for the timing of subduction-related magmatism in the northern part of the orogen, which appears to have ceased in mid-Carboniferous time and resumed in the Early Permian. The model presented by Murray & others (1987) involves a change from convergent plate motion in the Early Carboniferous to strike-slip movement in the Late Carboniferous. Oroclinal bending accompanied about 450-500 km of dextral transform faulting of the eastern part of the orogen (Fig. 5b). One problem with this model is that the postulated transform fault can no longer be recognised. If the magnetic anomaly pattern of Wellman (this Bulletin) does in fact represent the continuation of geological units around the orocline, there is no displacement at the inferred position of the fault. The magnetic anomaly pattern suggests that the Baryulgil (Gordonbrook) Serpentinite, rather than delineating the southern part of the transform fault, is, instead, a continuation around the orocline of the zone of serpentinites adjacent to the Peel Fault. This is a modification of an idea by Runnegar (1974) who proposed that the Peel Fault was a spoon-shaped thrust that dipped eastward under the Tablelands Complex to eventually resurface at the position of the Bayugil Serpentinite. The magnetic anomaly pattern also implies that rocks of the fore-arc basin sequence (Tamworth and Yarrol belts) have been involved in the oroclinal bending (Fig. 5a), as previously inferred by Korsch & Harrington (1987).

Harrington & Korsch (1987) considered that the oroclinal bending was confined to the upper part of the crust, with movement taking place on a gently-dipping detachment that allowed sliding at shallow to mid-crustal levels. They considered also that this detachment was the top of the old subduction zone (Fig. 6). If this were the case, the contrast between the relatively coarse-grained trench-fill turbidites and the hemipelagic sediments of the ocean floor would result in shearing and formation of the detachment (cf. Karig, 1982). We consider it significant that on BMR deep seismic line 16 across the orogen, there appears to be a mid-crustal, subhorizontal detachment present (e.g. Finlayson & others, 1990). The above model has considerable implications for the current composition of the entire crust in easternmost Australia (see below). In their model, Harrington & Korsch (1987) inferred that the steeply-dipping bounding fault was the Peel Fault, but the involvement of fore-arc sediments in the bending suggests that the Mooki Fault (*sensu lato*, Harrington & Korsch, 1985a) is more likely to bound the orocline to the west. The shape of the orocline (Fig. 5) indicates that there has been a horizontal displacement towards the south of at least 450-500 km of the northern extension of the rocks involved in the oroclinal bending, i.e. the Yarrol Province.

ROLE OF STRIKE-SLIP FAULTING

Limited amounts of strike-slip faulting in the New England Orogen have been known for some time (e.g. about 22 km of post-Middle Triassic movement on the Demon Fault: Shaw, in Packham, 1969; Korsch & others, 1978; McPhie & Fergusson, 1983) and several authors have proposed large amounts of displacements on faults in plate tectonic models for the orogen (e.g. Scheibner & Glen, 1972; Leitch, 1975a; Harrington & Korsch, 1979, 1985a, 1985b; Evans & Roberts, 1980; Cawood, 1982; Murray & others, 1987). Because the orogen has had a long history related to a convergent plate margin, it is likely that the role of strike-slip faulting has been much more important than has been recognised even to date, with large displacements possibly having occurred (e.g. the 450-500 km of displacement associated with the oroclinal bending).

Harrington and Korsch (1985a) proposed major strike-slip movement on the Mooki Fault in the latest Carboniferous to Early Permian. However, Cherry (1989) has recently suggested that large-scale horizontal displacements were unlikely, based on a clast in the Currabubula Formation which he considered was derived from the Lachlan orogen. An alternative interpretation is that the clast was derived from the Drummond Basin, and, if so, 500+ km of displacement on the Mooki Fault system is required to move the Currabubula Formation to its present position. In yet another interpretation Thomson & Flood (1984) proposed that the Mooki Fault was a Late Permian - Early Triassic thrust feature. To the east, there is controversy over movement directions on the Peel Fault system, particularly in the Permian. Evidence for sinistral strike-slip has been presented by Corbett (1976) and Offler & Williams (1985) which conflicts with dextral transtension determined by Katz (1986). Recent work shows that an even more complicated movement history for the Peel Fault system is starting to emerge (e.g. Blake & Murchey, 1988a, 1988b; Offler & others, 1989).

Thrust sheets, interpreted as gravity glides by Roberts & Engel (1987) and Engel & Morris (1987), resulted in overthrusting over Permian sediments of the northern Sydney Basin by structural blocks of the southern Tamworth Belt, i.e. a north over south movement. Bounding faults for these thrust sheets would most likely have a strike-slip character, with opposite senses of movement on the eastern and western sides of the thrust sheets.

Recently, Korsch & others (1988, 1989) have suggested that Early Permian to Cretaceous sedimentary basins peripheral to the New England Orogen were initiated during transtensional events. Both the Esk Trough and Taroom Trough are asymmetric in shape, being bounded to one side by very steeply dipping faults that are considered to have had a strong strike-slip component to their movement during basin initiation. O'Brien & others (in press; this Bulletin) showed that beneath the Clarence-Moreton Basin there are several strike-slip faults that controlled the geometry of the basin. These faults, and those bounding the Esk and Taroom Troughs, were active, at least intermittently, throughout the depositional history of the basins. Even the youngest sediments in the vicinity of the faults have been folded or thrusted; these structures have been interpreted as positive flower structures above the more deeply rooted strike-slip faults (Korsch & others, 1989; O'Brien & others, in press). Thus, strike-slip faulting has exerted a considerable influence on the New England Orogen, at least from the latest Carboniferous or even earlier, until the Late Cretaceous or Early Tertiary.

TERRANE RECOGNITION

Several different schemes have been proposed for the recognition and definition of terranes in the New England Orogen (e.g. Cawood, 1983b; Cawood & Leitch, 1985; Scheibner, 1983, 1985; Leitch & Scheibner, 1985, 1987b). Flood & Aitchison (1988) modified the lithological associations of Korsch (1977) to define a number of terranes in the Tablelands Complex and they provided a synonymy for the previous terrane nomenclature in New England. Most of the schemes subdivided the accretionary wedge described above into several terranes, but the uniformity in sandstone compositions and provenance indicates that there is a lack of clearly defined, exotic terranes for much of the New England Orogen.

As outlined above, the late Silurian to Middle Devonian volcanic arc sequences of the northern New England Orogen, if an island arc, almost certainly represent an exotic terrane (or terranes) which docked at the end of the Middle Devonian.

The most consensus occurs on the Gympie Province (Fig. 1, see also Fig. 5a), which is considered to be allochthonous to the rest of New England (Harrington, 1983) and is thus either a suspect or exotic terrane. The Gympie Province has been inferred to be related to rocks of the same age in New Zealand (Harrington, 1983, 1987; Cawood, 1984b; Waterhouse & Sivell, 1987; Sivell & others, this Bulletin). The main problems centre on recognition of the boundaries to the Gympie Province and whether it comprises more than one terrane (see Murray, 1988). Murray & others (1989a) suggested that the Gympie Province consists of at least three suspect terranes: (1) the western terrane comprising slate, phyllite and sandstone with mafic metavolcanics and Middle Carboniferous limestone lenses (Roberts, 1987); (2) the central terrane composed of the greenschist facies metasediments and mafic metavolcanics of the North D'Aguilar Block of uncertain age (Murray, 1988; Sivell & others, this Bulletin); and (3) the eastern terrane (the Gympie terrane of Harrington, 1983; Harrington & Korsch, 1985c) consists mainly of Permian to Early Triassic shallow marine sediments and island arc volcanics (Waterhouse & Sivell, 1987). This eastern terrane itself, however, is almost certainly composite, because it contains mafic volcanics of different tectonic settings (Murray, 1987; Sivell & Waterhouse, 1988) and also Early Carboniferous radiolarian chert lenses (Murray & others, 1989). Harrington & Korsch (1985c) proposed that the Gympie terrane docked with the rest of New England in the Middle Triassic, and that this event had a marked consequence on the geological history of eastern Australia, with sedimentation ceasing in the major coal-bearing basins.

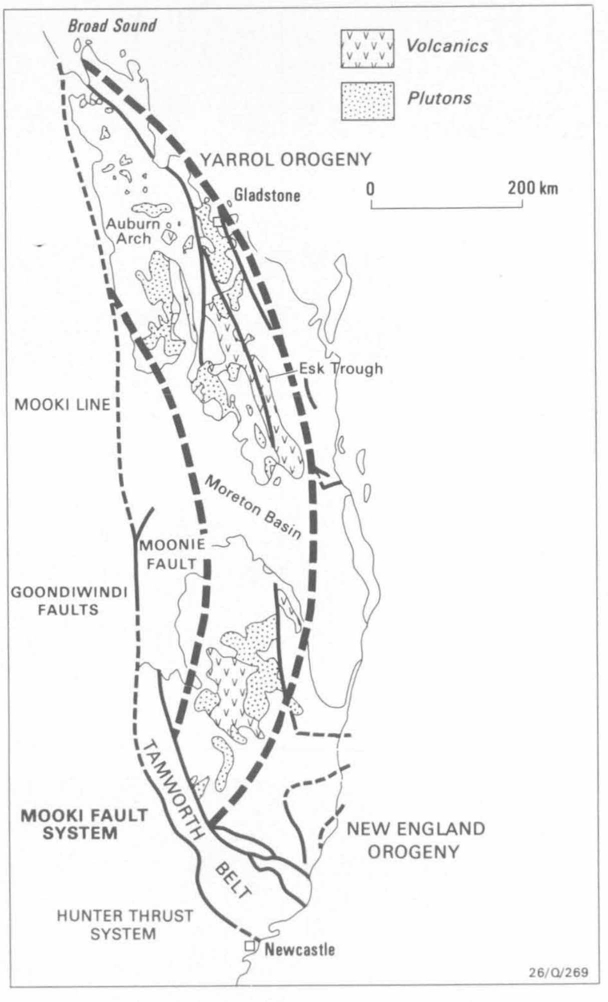

Fig. 7 Major granitoid belt of New England suite and associated volcanics (Late Permian to Early Triassic) after Harrington & Korsch (1985b). At this time, the volcanic arc and subduction system were located to the east of the orogen, probably in palaeo-New Zealand and the orogen was in a back-arc position. Thus, the granitoid belt represents a major area of back-arc extension, volcanism and plutonism.

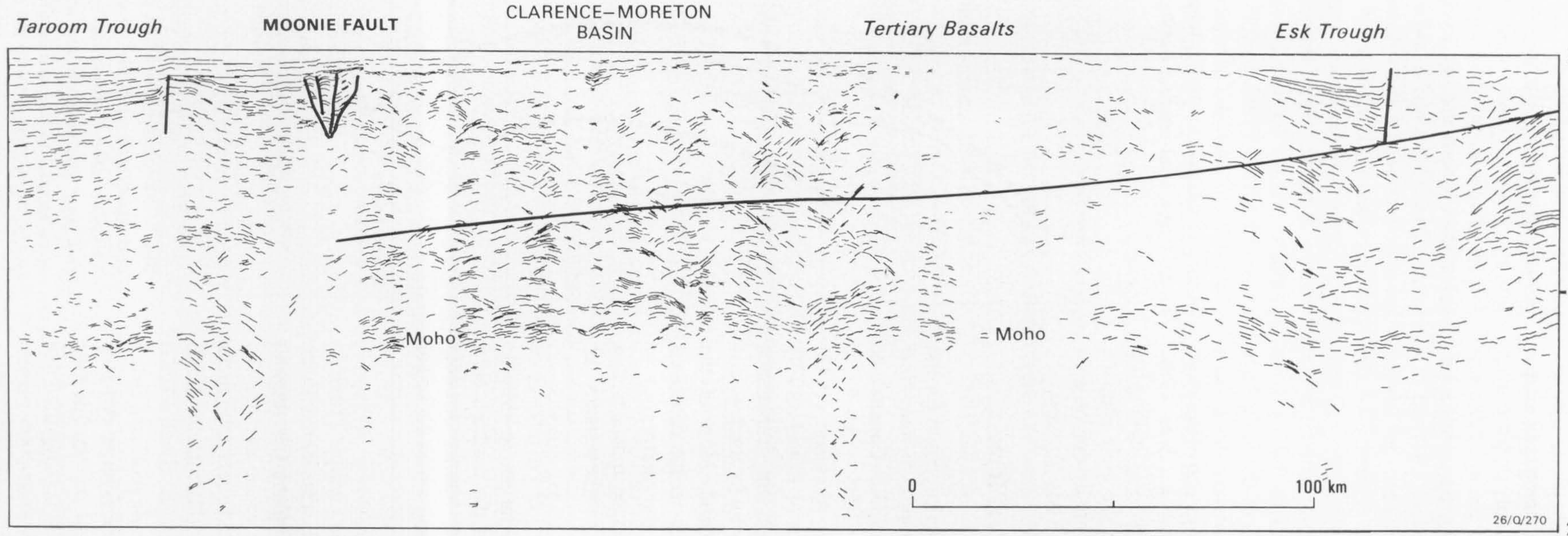

Fig. 8 Line drawing of BMR deep seismic data from part of traverse 14 and traverse 16 across the New England Orogen. The mid-crustal detachment separates more-reflective lower crust below it from less-reflective crust above it, and is inferred to be the detachment on which oroclinal bending took place. Also shown are the deep, layered sequence in the easternmost part of the line and the peripheral sedimentary basins: Taroom Trough, Esk Trough, and Clarence-Moreton Basin.

Other possible suspect or exotic terranes in the New England Orogen are anomalous quartz-rich sandstone sequences in the eastern parts of the Beenleigh Block (Fig. 5a) south of Brisbane (Korsch, 1977; Lohe, 1980; Korsch & Harrington, 1987; Murray, 1988) and in the Coastal Block near Rockhampton. Korsch & Harrington (1987) considered the whole of the Coastal Block to be a suspect terrane (their Curtis Island terrane, see Fig. 5a) whereas Ferguson & others (1988) considered only the eastern part of the block (their Shoalwater terrane) to be suspect.

EARLY PERMIAN EVENTS

After an apparent break in the Late Carboniferous, calcalkaline volcanism resumed in the Early Permian in the northern New England Orogen along the site of the Late Devonian - Early Carboniferous volcanic arc (Connors and Auburn Arches), and also extended over much of the area of the former fore-arc basin (Yarrol Belt). The volcanics range from basalt to rhyolite, and are dominantly andesitic, with pyroclastics slightly more abundant than flows (Malone & others, 1964, 1969; Jensen & others, 1966; Clarke & others, 1971; Dear & others, 1971; Paine & others, 1974; Whitaker & others, 1974). Preliminary geochemical data (L.J. Hutton, pers. comm., 1988) are consistent with a subduction-related arc origin.

This arc did not continue southwards into the New England area along the site of the Devonian - Carboniferous arc, but instead trended southeast, and if projected, would meet the present coastline just south of the Queensland - New South Wales border. A possible explanation, if arc formation followed oroclinal bending, was the doubling of the width of the orogen in New England by the megafolding event. Alternatively, if oroclinal bending was later than the arc formation, the arc would have been transposed to its present position during the bending. Cessation of subduction and related arc volcanism in southern New England possibly was related to major strike-slip movement on the Mooki Fault System (Harrington & Korsch, 1985a). In either case, the northern part of the New England Orogen (Yarrol Province) was in an arc or fore-arc position, whereas the southern part (New England Province) was in a back-arc, extensional setting. No accretionary wedge sequence of Early Permian age has been recognised in the Yarrol Province. It may be located offshore or have been subsequently removed. As the locus of subduction moved east, a thick accretionary wedge built up in New Zealand in the Permian and Mesozoic (e.g. Korsch & Wellman, 1988).

Latest Carboniferous - Early Permian volcanics on the western margin of the Tamworth Belt in New England (e.g. Werrie Basalt, Boggabri Volcanics) were interpreted by Scheibner (1973), Korsch (1982), Harrington (1982), Harrington & Korsch (1985a) and Korsch & others (1988) as extension-related volcanics associated with transtension on the precursor of the Mooki Fault and formation of the Gunnedah-Bowen Basin. On the other hand, McPhie (1984) considered the volcanics to be the final stages of the Devonian - Carboniferous subduction-related, continental margin volcanic arc. Leitch & others (1988) suggested that the compositions of the volcanics reflect the influence of lithosphere previously subducted during the Carboniferous, and that the eruption of the large volume of lava was due to an extensional environment during the Early Permian. Chemical and isotopic data on volcanic centres of the Werrie Basalt indicate that the rocks are significantly different from the Late Carboniferous lavas and ignimbrites (Flood & others, 1988).

Large areas of New England consist of deep-water marine diamictites and fine-grained sediments that contain Permian fossils and are interpreted to have formed in an extensional basin or basins (Korsch, 1982; Leitch 1988). These rocks are usually intensely deformed and in places such as the Nambucca slate belt are among the most severely deformed rocks that occur in the orogen. This deformation also occurred in the Permian because the rocks are then intruded by post-orogenic Late Permian to Early Triassic plutons and are overlain by Middle Triassic basin sediments that are essentially undeformed.

The Early Permian represented a time of significant change in the New England Orogen. While the northern part of the orogen (Yarrol Province) remained in an arc and fore-arc setting, an eastward jump of the subduction zone and associated arc in New England resulted in a change for the first time to a back-arc environment. There is no record anywhere in the exposed orogen of trench-fill sedimentation of Permian age. Several models (e.g. Korsch, 1982; Cawood, 1982, Harrington & Korsch, 1985a, 1985b; Murray & others, 1987; Leitch, 1988) have been proposed to explain the change in tectonism, but none of them is completely satisfactory. Problems still to be fully addressed include the nature of the mechanisms that led to development of the localised, relatively deep, sedimentary basins and also the process for the deformation following closely after sedimentation.

TECTONIC SIGNIFICANCE OF THE GRANITES

Trenches and accretionary wedges are areas of low heat flow, in comparison with the volcanic arc and back-arc spreading regions. The New England Orogen, in comparison with some other ancient accretionary wedges (e.g. Franciscan Complex of California, Torlesse Complex of New Zealand), has a significant (enormous) volume of granitoid plutons and related volcanics. Both S-type and I-type plutons have been described and have been divided into several suites and/or belts (e.g. Shaw & Flood, 1981; Hensel & others, 1985). These workers found that, in general, the S-type suites were older than the I-type suites and suggested that the plutons were

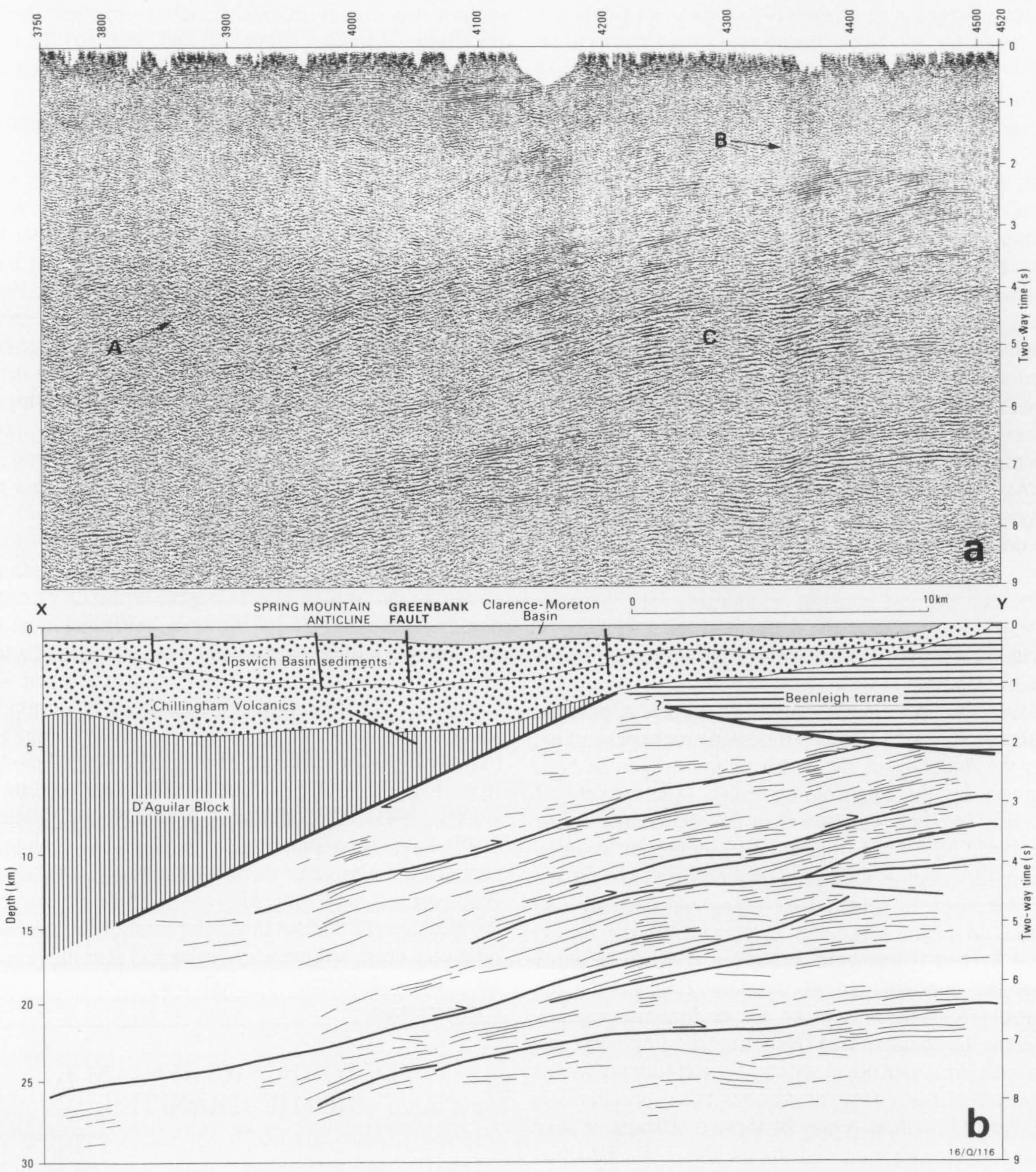

Fig. 9 **(a) Unmigrated seismic reflection profile of the easternmost part of BMR traverse 16; (b) an interpretation of the profile integrated with surface data (after Korsch & others, 1986). The character of the layered sequence from 2 to 8 s in the easternmost part is considerably different from that recorded beneath the forearc region to the west (see Fig. 8).**

derived from felsic sandstones and mudstones of the accretionary wedge. Two mica granites of similar (Early Permian) age intrude the accretionary wedge in the Yarrol Province, but may not be true S-type granites.

The I-type plutonism occurred in the Late Permian to Triassic, occupying a curved belt over 1000 km long (Fig. 7). At this time, the orogen was in a back-arc, presumably extensional, setting behind the arc, remnants of which are now located in New Zealand (see Korsch & Wellman, 1988). Isotopic dating of the plutons, mainly using mineral separates for K-Ar and whole rock samples for Rb-Sr, is sporadic. There has been almost no U-Pb zircon work attempted, and hence the ages of initial magma generation and crystallisation are not clear. Thus, there has been no satisfactory explanation as to how the plutonics relate to the overall tectonic development of the orogen. Nevertheless, significant tectonic events altered the heat flow regime and caused the generation of the different magma types and emplacement of the plutons, along with associated thermal metamorphism, into the former accretionary wedge.

The formation of the orocline, and consequent widening of the orogen, may be related to the generation of the I-type magmas, but the bending may have occurred after the emplacement of the S-type plutons, which would therefore require another explanation. In New England, the plutons are abundant in the accretionary wedge, but extremely rare in the fore-arc basin. This difference, however, is not at all apparent in the Yarrol Province to the north (Murray, 1986). It is unlikely that the plutons were derived by melting of the accretionary wedge, because insufficient thickness of wedge material would have developed to provide the depths needed for magma generation.

NATURE OF THE CRUST IN THE NEW ENGLAND OROGEN

Rutland (1976) suggested that a lower crust of continental Precambrian material extended under the New England Orogen. Based on Pb-Pb isotopic data from the Drake volcanics and associated granites, Gulson & Bottomer (1984) concluded that the granites and volcanics were derived from a Proterozoic source. On the other hand, Nd and Sr isotopic data from New England granitoid and sedimentary rocks severely limit the involvement of old continental crust in the granitoid source regions (Shaw & Flood, 1981; Hensel & others, 1985). O'Reilly & others (1988) showed that granulite xenoliths brought to the surface near Gloucester by Mesozoic pipes have an isotopic signature consistent with the involvement of subducted continental and ocean-floor or island-arc material.

The early history of the orogen suggests that it was related to an island arc-type convergent margin; hence the existence of Proterozoic crust beneath the orogen is unlikely. Nevertheless, one possible explanation for the Proterozoic signature at Drake is that Proterozoic continental crust could extend as far east as beneath at least part of the fore-arc basin, i.e. to the position of the continent-ocean transition. Rocks of the fore-arc sequence have been involved in the oroclinal bending, and the Drake region is possibly a displaced portion of this fore arc (see Fig. 5a). Thus, a Proterozoic signature in the Drake volcanics could reflect Proterozoic continental crust beneath the fore arc. The limited data of Gulson & Bottomer (1984) warrant further investigations to confirm their results.

The accretionary wedge in the orogen is intensely deformed, both as a result of accretion processes and later oroclinal bending which also widened it. On BMR deep seismic line 16 (Fig. 8), the seismic records show the existence of a subhorizontal, mid-crustal detachment, required at some crustal level because the orocline cannot be expected to continue to an indefinite depth. With regard to the nature of the lower crust, the model of Harrington & Korsch (1987) suggests that the lower crust under the oroclinally-bent accretionary wedge consists of a "frozen" part of the subducted slab of oceanic crust, with or without pelagic or trench-fill sediments (Fig. 6). To date, isotopic and geochemical evidence cited above are compatible with this interpretation.

The easternmost part of BMR seismic line 16, across the eastern Clarence-Moreton Basin and Beenleigh Block (Fig. 9), revealed a thick, deeply buried, layered sequence that was interpreted by Korsch & others (1986) to have been sedimentary rocks telescoped by thrusting into a stack over 20 km thick. The stack is overlain by the Beenleigh Block, which is seismically non-reflective and has a maximum thickness of about 4.6 km. The layered sequence is noticeably different in character to rocks of the Beenleigh Terrane and the inferred fore-arc region to the west (cf. Fig. 8).

CONCLUSIONS

The New England Orogen in eastern Australia has had a complicated evolutionary history that stretches from the Cambrian to the Late Mesozoic. The major component of the orogen evolved in the Devonian and Carboniferous in a convergent plate margin tectonic setting related to a west-dipping subduction system. Parts of the volcanic arc, fore-arc basin and accretionary wedge are still preserved in the orogen. The earlier history of the orogen is fragmentary, but from the Cambrian to Silurian, the orogen was probably island arc-related. The later history of the orogen has involved strike-slip faulting and major oroclinal bending, with about 450-500 km of displacement being involved. This was followed by massive amounts of volcanism and plutonism in the Late Permian and Early Triassic when the orogen was in an extensional, back-arc setting. One technique that, in the future, might help to decipher the complex history recorded in the rocks of the New

England Orogen is a rigorous terrane analysis approach coupled with palaeomagnetic studies.

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DEVELOPMENT OF THE TASMAN SEA AND EASTERNMOST AUSTRALIAN CONTINENTAL MARGIN - A REVIEW

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ABSTRACT

The eastern end of the Eromanga-Brisbane Geoscience Transect is located in the hinterland adjacent to the eastern Australian continental margin. This margin delineates the boundary between the eastern extent of the Australian continental mass and oceanic crust of the adjacent Tasman Basin. The development of this margin is best described in four phases, similar to those of the more generalised models of passive margin development; Pre-rift, Rift, Post-Breakup (Active Seafloor Spreading Phase) and Post-Breakup (Quiescent Phase). Sediments of the Upper Triassic to Late Jurassic Clarence-Moreton Basin were deposited during the pre-rift stage. Rifting commenced along the southeastern Australian margin in Early Cretaceous times and progressively extended northwards reaching the vicinity of the transect during the Late Cretaceous. Seafloor spreading in the adjacent basin was active from approximately anomaly 28 (64 Ma b.p.) to anomaly 24 (about 55 Ma b.p.).

Features of the margin near the eastern end of the transect, including the virtual absence of rift associated sedimentation, steep continental slope, and narrow width, can be reconciled with a model of passive margin development by invoking an initial amount of strike-slip motion coupled with asymmetric rupture of the rift system. Erosion associated with thermal uplift prior to the onset of seafloor spreading is attributed to having truncated up to 2 km off the eastern flank of the Clarence-Moreton Basin.

Uplift and volcanism responsible for the formation of much of the present day hinterland relates to a more recent period of tectonic activity. Commencing in the early Tertiary, much of this activity is of Neogene age and associated with compressional underplating of the eastern Australian continent.

INTRODUCTION

The eastern end of the Eromanga-Brisbane Geoscience Transect is located in the hinterland adjacent to the eastern Australian continental margin (Fig. 1) (Johnstone & others, 1985; Wake-Dyster & others, 1985; Finlayson, this Bulletin). This paper describes the development of the Tasman Sea and adjacent continental margin in the vicinity of this eastern end of the transect.

MORPHOLOGY

The morphology in the vicinity of the eastern end of the Eromanga-Brisbane Geoscience Transect is dominated by a conspicuously elevated hinterland adjacent to a narrow, steep continental margin. The hinterland forms part of the eastern highlands, a broad uplifted region with a steep eastern, coastal, escarpment and gradual western slope (Ollier, 1982). Along much of the eastern Australian margin the crest of the highlands runs north-northeast, parallel to the coast and adjacent continental margin. The highlands, which are well delineated by the 400 m elevation contour, attain elevations in excess of 2 km in the south. North of Brisbane the highlands swing north-northwest in a broad arc extending inland to a distance of over 400 km from the continental margin (Fig. 1). Here the highlands are less pronounced, lying at elevations generally less than 400 m. The crest of the highlands forms a major watershed separating the dendritic drainage of the inland and tablelands from the complex river capture drainage patterns of the more intensely eroded coastal zone (Ollier, 1982).

Figure 2 shows a portion of a BMR seismic traverse which highlights the morphology of the offshore portion of the margin east of the Eromanga-Brisbane Geoscience Transect. This seismic traverse reveals a morphology typical of much of the eastern Australian continental margin, but atypical of that of a classic passive or Atlantic-type margin. The continental shelf is only approximately 40 km wide and lies adjacent to a continental slope which is also narrow. As shown on Figure 2 the slope is dominated by a single steep (6°) fault escarpment. The presence on this profile of a lower slope platform and continental rise is atypical; most profiles across the eastern Australian margin reveal the outer slope to terminate abruptly onto the adjacent abyssal plain.

The abyssal plain, lying at depths exceeding 4 km, forms part of the Tasman Basin. This oceanic basin separates eastern Australia from the Lord Howe Rise, a submerged continental mass of 1800 km length and 400 km width, whose crest lies in water depths of 1000 -

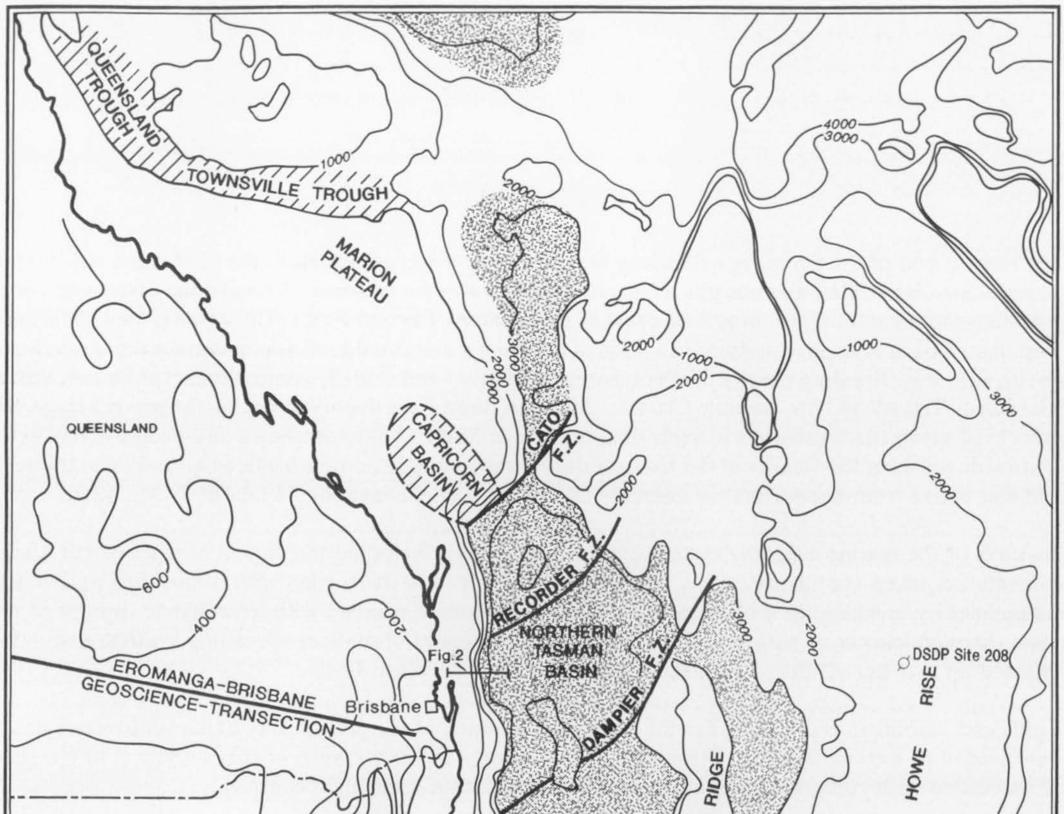

Fig. 1 Eastern Australia regional setting, showing major structural and tectonic elements; topographic contours from Wellman (1987).

2000 m. By world standards, the Tasman Basin is narrow, the bounding continental margins converging rapidly, such that in the north and adjacent to the Eromanga-Brisbane Geoscience Transect, its width is only 250 km.

The northern Tasman Basin (Fig. 1) is bounded to the north by the Marion Plateau across the Cato Fracture Zone. To the south, the Dampier Fracture Zone separates the basin from the wider central Tasman Basin. This latter part of the basin is flanked to the east by the Dampier Ridge, another submerged rifted continental fragment lying in water depths of 2000 - 3000 m.

GEOLOGICAL & TECTONIC SETTING

Figure 3 shows the geology near the eastern end of the transect to be dominated by fault-bounded outcrops of Palaeozoic rocks, of the Yarrol and New England Orogens, which constitute basement throughout the region. Surrounding these are the Mesozoic sedimentary sequences of the Ipswich, Esk and Clarence-Moreton Basins, which are locally overlain by surficial Cainozoic sedimentary deposits and volcanics.

The Beenleigh and D'Aguilar Blocks comprise Late Palaeozoic deep water marine sequences subsequently deformed during incorporation into subduction related accretionary wedges. The Beenleigh Block is interpreted as a thin skin exotic terrane accreted onto the New England orogen prior to the Middle Triassic (Korsch & others, 1986).

Early Mesozoic sediments are confined to fault-bounded depressions between these Palaeozoic blocks in the Esk Trough and in the Ipswich Basin, east of the West Ipswich Fault. To the south these sequences are unconformably overlain by Late Triassic to Late Jurassic sediments of the more extensive Clarence-Moreton Basin (Fig. 3). The Ipswich Basin, Esk Trough and Clarence-Moreton Basin are representative of a number of basins located along the eastern Australian margin with tectonic affinities that remain unclear (Korsch & others, 1989; O'Brien & others, this Bulletin).

Harrington & Korsch (1985) proposed that the Esk Trough originated as a pull-apart basin in response to dextral strike-slip motion. More recently Korsch & others (1989) proposed that the Ipswich Basin, Esk Trough and Clarence-Moreton Basin are generically related to the same period of oblique extension. Both the Esk and Ipswich Basins are interpreted by Korsch and others (1989) to have formed during successive rifting phases of oblique extension, active between 240-260 and 225-240 Ma, respectively. The much more extensive Clarence-Moreton Basin sequences, which unconformably overly these rift-related sequences and Palaeozoics rocks, formed during the subsequent associated lithospheric cool-down phase.

The continental margin in southern Queensland and northern New South Wales delineates the easternmost extent of the Australian continental landmass. Seismic studies (e.g. Doyle & others, 1966; Cleary, 1973) reveal this margin to coincide with a zone of crustal transition - a shallowing of the crust from depths of approximately 40 km or more beneath the continental mainland interior, to less than 25 km on the continental shelf, as true oceanic crustal thicknesses are attained beneath the adjacent Tasman Basin.

Seismic data over the continental shelf near the eastern end of the Eromanga-Brisbane Geoscience Transect reveals an absence of any thick layered sedimentary sequences. As shown in Figure 2 the thin sedimentary section (less than 0.5 km) overlying Palaeozoic basement is interpreted to be Neogene to Recent in age, consistent with adjacent onshore geology which indicates depositional thinning and erosional truncation of the Clarence-Moreton Basin sequences towards the coast.

Conspicuously absent across the margin is the block faulted half-graben, taphrogenic basement structure and rift related sedimentary infill typically observed along passive continental margins. However, such features do occur along the western flank of the Lord Howe Rise (BMR, 1987), which was juxtaposed to the eastern Australian margin prior to Tasman Sea opening.

Crustal thicknesses of 18-29 km derived from seismic refraction studies (Shor & others, 1971) confirm the continental affinities of the Lord Howe Rise. Data from Deep Sea Drilling Project (DSDP) Sites 207 and 208 located on the Lord Howe Rise document its subsidence history. Shallow marine or subaerial conditions prevailing during the Late Cretaceous gave way to upper bathyal depths during the Paleocene or early Eocene, these depths having persisted through to the present day (Burns & others, 1973).

That the Tasman Sea formed by processes of seafloor spreading has now been well documented, particularly for the wider, and more southerly portion of the basin (Hayes & Ringis, 1973; Weissel & Hayes, 1977; Shaw, 1978). The narrower, more northerly portion of the basin, including the area adjacent to the Eromanga-Brisbane Geoscience Transect, is also considered to have formed as a result of this same episode of seafloor spreading, although the exact nature of opening and its subsequent history is less well understood.

Figure 4 shows the seafloor magnetic anomaly pattern identifed by Shaw (1979) in the northern Tasman Basin. The presence of numerous seamounts, fracture zones, and the orientation of the recording ship's traverses, conspire to make identifications difficult. Nevertheless, the individual identifications give rise to a consistent pattern of magnetic lineations which, as shown in Figure 4, are distinguished by the central anomaly, 24, and a slightly asymmetric spreading sequence, which can be traced out from the central anomaly to anomaly 31. As the width of the basin decreases north of the Dampier Fracture Zone the range of identifications also decreases, extending out to only anomaly 28 or 29. North of the Recorder Fracture Zone, the sequence is further truncated and identifications range from anomaly 24 to only anomaly 26 or 27.

In keeping with the absence of active, present-day, seafloor spreading, the Tasman Basin is aseismic except for two small epicentre groups located in the southern Tasman near latitude 40°S and scattered minor seismicity on the continental shelf off southeast Queensland. Low-level earthquake activity is also an attribute of eastern mainland Australia where a diffuse pattern of shallow, small magnitude earthquake epicentres occurs along the uplifted highlands adjacent to the eastern margin. Earthquakes have been recorded in the vicinity of the Eromanga-Brisbane Geoscience Transect (Cuthbertson, 1989) and also in the Southern Highlands located some 1000 km to the south. Stress and earthquake focal mechanism solutions of these earthquakes reveal a largely pan-continental contemporary stress regime which, although locally variable, is dominated by east-west, nearly horizontal compression (Veevers, 1984).

Fault and fold orientations within Neogene sediments in the Gippsland Basin imply that this contemporary stress regime has been operative for some time. Veevers & Powell (1984) suggest that this compression reflects an overall dextral or anticlockwise motion of mainland Australia, relative to the rest of the enclosing plate, caused by the sinistral shear of the Pacific and Indo-Australia plates in the New Guinea region.

MARGIN EVOLUTION

Generalised models of passive margin evolution (e.g. Falvey, 1974) propose rifted margin development under a regime of high heatflow initially associated with uplift, increased volcanism, erosion and subsequently rift basin formation. Following crustal rupture (breakup) and the onset of seafloor spreading the post-rift phase is characterised by continental lithospheric subsidence of the rift system in a declining heatflow regime as the spreading heat source retreats away from the margins.

Although the Australian margin adjacent to the Eromanga-Brisbane Geoscience Transect shares much in common with this generalised scenario of passive margin evolution, it also has attributes which imply significant differences, not the least of which are, as already noted, the presence of an elevated hinterland and narrow, steep continental margin with a virtual absence of rift related sediment cover. In describing the evolution of this region it is convenient to consider four phases of margin

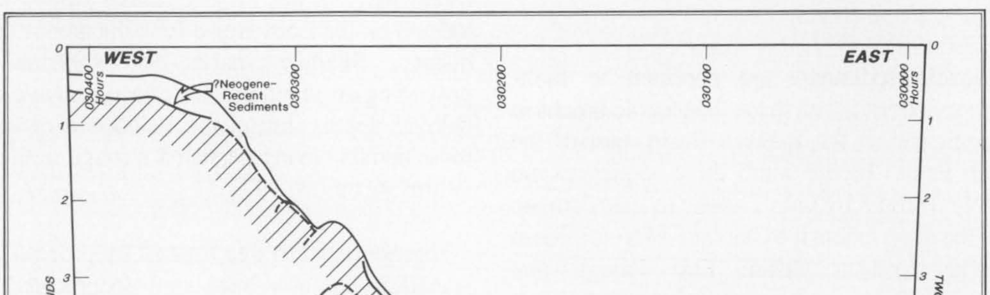

Fig. 2 Interpreted portion of BMR seismic profile 15/45 off southeastern Queensland.

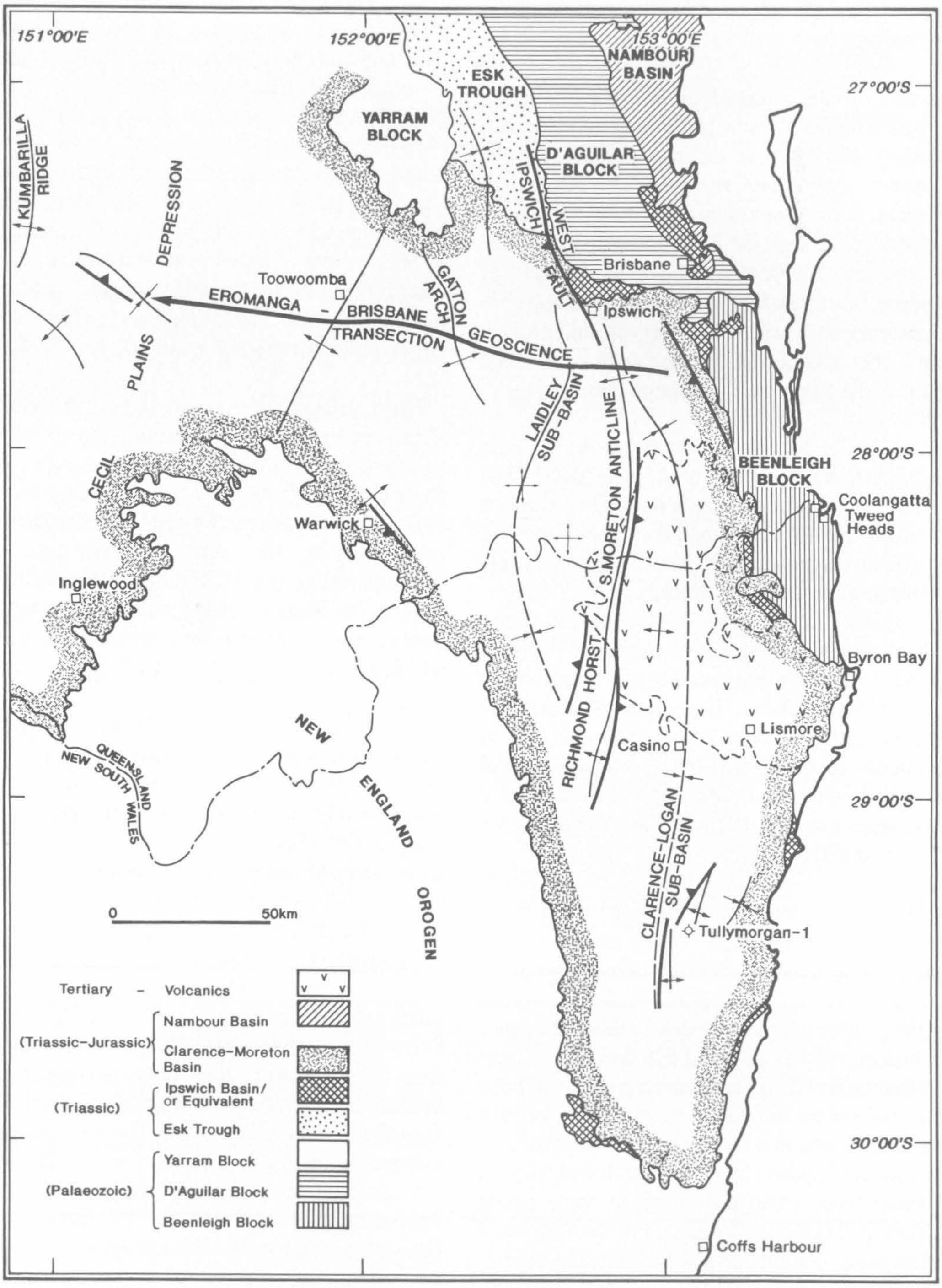

Fig. 3 Regional geological framework, eastern end Eromanga - Brisbane Geoscience Transect.

development: (1) pre-rift, (2) rift, (3) post-breakup, (active seafloor spreading phase), and (4) post-breakup (quiescent phase). These phases are discussed below.

PRE-RIFT PHASE

Deposition of rift related sedimentation, which accompanied the down faulting of the central graben of the Gippsland Basin, place the age for the onset of rift development along the eastern Australian margin at about 140 Ma ago.

Prior to this, in the Jurassic pre-rift stage, eastern Australia was dominated by a proto-Pacific-Australian plate boundary located east of, and parallel to, the present eastern Australian margin. This boundary extended from New Guinea in the north, south of Tasmania, and on through the Transantarcic Mountains (Jones & Veevers, 1983). Plate convergence north of 30°S latitude is interpreted by Jones & Veevers (1983) to have been associated with an active volcanic orogen, which was the source of labile sediments shed westwards into the Surat and Eromanga Basins.

During the Early Jurassic, sedimentation in these basins highlights the presence of a widespread, and shallow-marine-dominated, depositional regime across a large part of eastern interior Australia, even though, at that time, southeastern Australia was, by contrast, the site of non-deposition and volcanism.

Along the Eromanga-Brisbane Geoscience Transect, pre-rift sedimentation is represented by the sequences of the Clarence-Moreton Basin. Deposition in that basin commenced in the Upper Triassic and continued until at least the Middle to Late Jurassic. The Walloon Coal Measures, which are one of the highest stratigraphic units in this sequence, are dated as Callovian (160 Ma b.p.) (Helby & others, 1987).

Whereas initial sedimentation in the Gippsland Basin was rapid and controlled by down-faulting, sedimentation in the Clarence-Moreton Basin during the Late Jurassic, was more uniform and widespread reflecting the greater maturity of that basins subsidence and burial history. Notwithstanding that deposition in the Clarence-Moreton and Gippsland Basins partially overlap in time, the differences in deposition and basin-forming style serve to highlight that the genesis of the Clarence-Moreton Basin, and underlying Esk Trough and Ipswich Basin, is interpreted not to be related to those rifting processes responsible for the formation of the Gippsland Basin and adjacent margins of southeastern Australia.

RIFT PHASE

Rift development, as a precursor to the Tasman Sea spreading episode, is preserved along the eastern Australian margin only in the Gippsland and Capricorn Basins. These basins are separated by approximately 1750 km of margin, which, as already indicated, is conspicuously devoid of substantial late Mesozoic or early Tertiary sedimentation. Both basins occur near major offsets in the adjacent Tasman Basin magnetic lineations pattern, the Gippsland across the Bass Fracture Zone, and the Capricorn across the Cato Fracture Zone (Fig. 1). Both probably represent "failed-arms" in the Burke & Dewey (1973) triple-arm aulocogen model.

Rift basin development in the Gippsland Basin coincided with the onset of Strzelecki Group deposition. This commenced approximately 140 Ma ago (Jones & Veevers, 1983) when sediments comprising non-marine, greywackes, arkoses, mudstones and coals were deposited into a rapidly subsiding, fluvial system located within a developing central graben complex. In the Capricorn Basin initial rift development is interpreted to have commenced during the Late Cretaceous (Ericson, 1976). Here initial sedimentation, comprising conglomerates, lithic and arkosic sandstones with red-bed affinities, were deposited within a developing rift complex measuring some 100 km by 300 km.

Rift development recorded in both the Gippsland and Capricorn Basins reflects the time-transgressive nature of rifting along the eastern Australian margin, as predicted from the identification of seafloor spreading anomalies and reconstruction history of the Tasman Sea. It is analogous to the time of transgressive margin development proposed for the western Australian margin (Falvey & Mutter, 1981) and is consistent with the overall time-transgressive, anticlockwise rifting pattern which ultimately defined Australia's continental margins.

It is proposed that rifting began during the earliest Cretaceous, along the southern Tasman margin in the vicinity of the Gippsland Basin, and extended progressively northwards along the eastern Australian margin. As indicated in Figure 3, there is no evidence of rift-associated sedimentation onshore adjacent to the transect. Offshore there are also no known remnants of rift-associated sedimentation preserved; however, such sedimentation is anticipated to have occurred in association with widespread rift development across much of what is now the western flank of the Lord Howe Rise and Dampier Ridge. Rift development and associated sedimentation only impinged beyond the present-day eastern Australian continental slope in the Gippsland and Capricorn Basins. Based on the interpreted timing for rift onset in the Capricorn Basin, rift development in the northern Tasman Basin (including the area east of the Eromanga-Brisbane Geoscience Transect) commenced sometime during the Late Cretaceous, shortly before the thermal event which heralded the emplacement of the first oceanic crust.

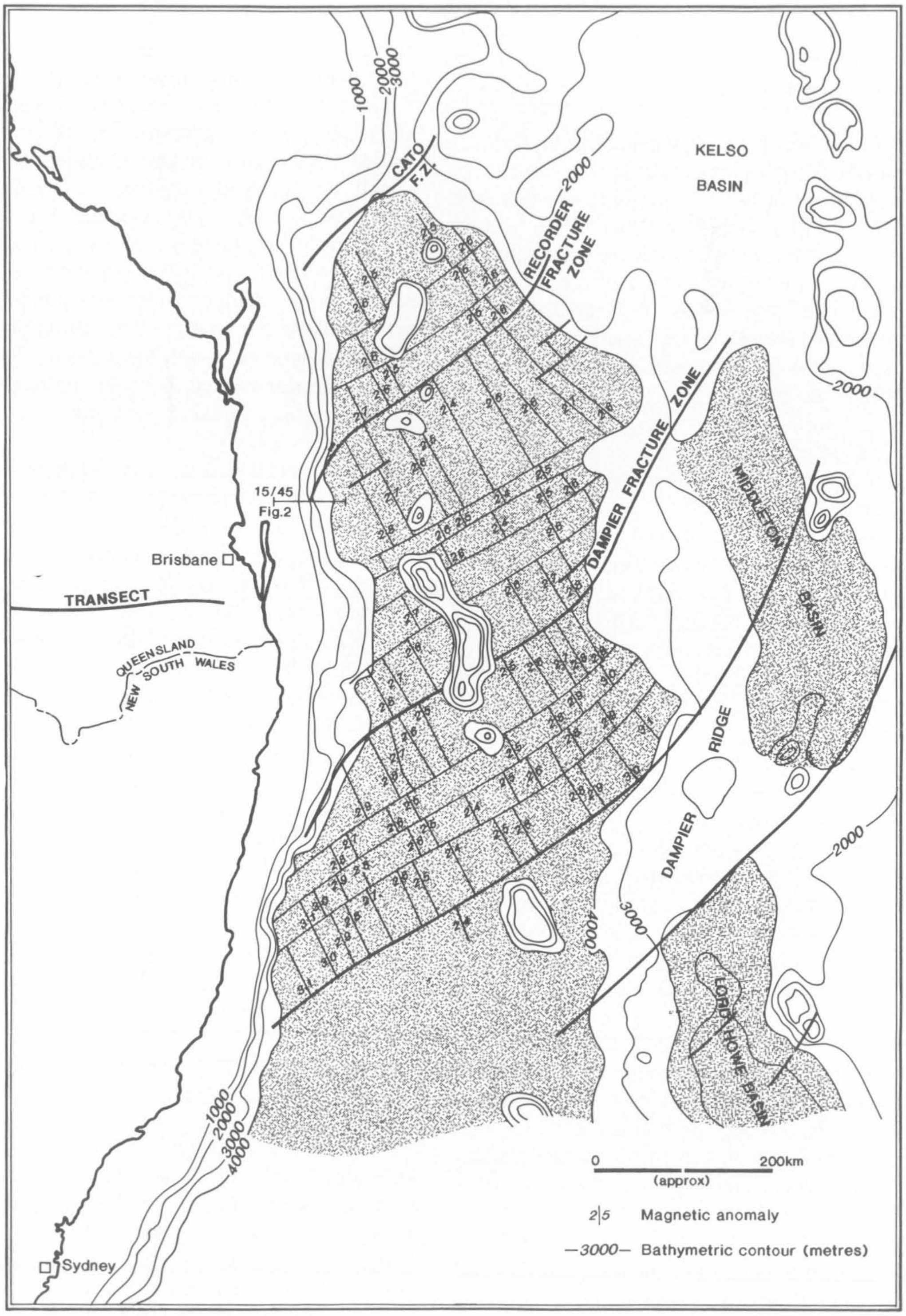

Fig. 4 Magnetic lineations and fracture zone locations, northern Tasman Basin (after Shaw, 1979).

POST-BREAKUP (ACTIVE SEAFLOOR SPREADING) PHASE

The Tasman Sea underwent breakup with the inception of active seafloor spreading initially in the south, prior to anomaly 33 time (>85 Ma b.p.). At this time, and during the preceding 10 Ma, substantial changes occurred in the geology of eastern Australia, which are interpreted to reflect a fundamental change in the pre-existing tectonic setting.

During the Cenomanian, sedimentation in both the Eromanga and Surat Basins ceased, as did convergence along the plate boundary off the Queensland margin (Jones & Veevers, 1983). Although magmatism occurred within the Gippsland Basin Strzelecki Group prior to 100 Ma, it became more widespread during the interval 90-100 Ma. This period was also marked by the emplacement of several small, mafic intermediate intrusions along the incipient eastern Australian margin (Jones & Veevers, 1983).

Apatite fission track analyses (Moore & others, 1986) document the period 80-100 Ma as corresponding with increased heatflow and uplift along the southeastern Australian margin. Contours of maximum heating based on apparent apatite ages lie parallel to, and form a maximum at, the present-day coastline. They also form an embayment around the Gippsland Basin, which is consistent with its origins as a failed arm of an aulocogen within the Tasman rift system. Heating appears to have affected the margin up to approximately 100 km landward of the present coastline in southeast Australia (Moore & others, 1986), and a similar pattern of coastal zone heating is predicted along the entire eastern Australian margin.

This increased heatflow was also apparently responsible for overprint magnetization recorded over a wide area of southeast Australia and dated independently by Schmidt & Embleton (1981) as 70-100 Ma ago. Schmidt & Embleton (1981) estimate that temperatures of 100°- 200°C were reached at the surface in order to achieve this overprint. Such temperatures are consistent with the 100°C temperatures necessary to anneal the apatite fission tracks in the coastal region (Moore & others, 1986). This heating event is interpreted to have been associated with the initial intrusion of accretionary material prior to continental rupture and active seafloor spreading.

The apatite fission track results indicate that the uplift associated with this heating event presumably produced a rim, with flanks extending landward to at least some 100 km beyond the present coastline. That is, only partially coincident with the present distribution of the eastern highlands. Moore & others (1986) suggested that up to 2-2.5 km of erosion may have occurred within the coastal zone during this period of high heat-flow and uplift. Such magnitudes of uplift are compatible with results obtained elsewhere along the eastern Australian margin. For example, in the Sydney Basin, Middleton & Schmidt (1982) estimated 1 km, to in excess of 2 km, of erosion on the basis of present-day surface maturities derived from vitrinite reflectance gradients.

By linking this uplift with the onset of seafloor spreading, it follows that similar uplift and erosion would have been experienced progressively northwards along the eastern Australian margin, reaching the vicinity of the northern Tasman Basin and the region adjacent to the Eromanga-Brisbane Geoscience Transect during the Campanian or Maastrichtian. In the absence of any evidence for significant Late Cretaceous sedimentation inland, or outside the confines of the Gippsland and Capricorn Basins, it is assumed that the erosional products of this uplifted rim were deposited eastwards to form the syn-rift and initial post-breakup depositional sequences on the adjacent developing margin. Based on seismic data from northern New South Wales, up to 2 km of sediments could have been shed from the developing rim located along the eastern flank of the Clarence-Moreton Basin at this time.

The magnetic lineations of the adjacent Tasman Basin (Fig. 4), when incorporated with those from the southern Tasman Basin, provide the basis for reconstructions of the Tasman Sea opening episode. Figure 5 shows reconstructions of the developing Tasman Basin at several specific times, using identified magnetic anomalies. A key feature of these reconstructions is the recognition that each of the Dampier, Recorder and Cato Fracture Zones acted as major transforms at specific periods during the opening of the central and northern Tasman Sea (Shaw, 1978). Strike-slip motion along these fracture zones is interpreted to have occurred prior to successive, northward extensions of the spreading centre. The presence of numerous smaller transforms, which offset the spreading axis, coupled with the restricted spreading history, gives rise to an illusion of discordance between the magnetic lineations and the adjacent margins. However, there is no evidence of anomaly truncation along either the eastern or western margins of the Tasman Basin and thus of subduction, as has been previously proposed (Hayes & Ringis, 1973) or implied in some reconstructions (Weissel & Hayes, 1977).

The inception of active seafloor spreading, prior to anomaly 33 time (>85 Ma.B.P), was confined to the southern Tasman Basin, and the Middleton and Lord Howe Basins to the east of the Dampier Ridge. Subsequent divergence of the spreading centre northwards occurred during a series of discrete plate boundary relocations. Spreading in the central Tasman Basin commenced following migration of the spreading centre out of the Middleton and Lord Howe Basins coupled with the fragmentation of the Dampier Ridge, during the anomaly 31 to 33 period. Spreading in the northern Tasman Basin was established later, during a phase of divergence which saw extension of the spreading centre northwards, between anomaly 30 to anomaly 27 times (Figs. 5a, 5b). Seafloor spreading was active throughout the entire Tasman Basin by approximately anomaly 27 time (63 Ma b.p.) and ceased throughout the basin at about anomaly 24 time (~55 Ma b.p.), immediately prior to the increase in spreading rate between Australia and Antarctica (Cande & Mutter, 1982).

The inception of seafloor spreading throughout the Tasman Basin at anomaly 27 time was coincident with the inception of seafloor spreading in both the Cato Trough and Coral Sea (Shaw, 1979); that is, it formed part of a major increase in plate divergence between the Indo-Australia and the Pacific plate boundary. Poles of rotation calculated from the identified magnetic anomalies imply a breakup and dispersal history in the northern Tasman Sea, including the margin adjacent to the eastern end of the Eromanga-Brisbane Geoscience transect, involving strong oblique, sinistral pull-apart motion. Relative changes in the pole positions during the anomaly 32 to 27 period imply that this was a time when plate kinematics were changing rapidly.

During the period from anomaly 32 to 30, the Dampier Ridge slid past the northern New South Wales coastline in a north-northeast, and later, north-east trajectory along an oblique shear boundary, the Dampier Fracture Zone (Fig. 5a). Increased relative divergence in the plate movements was reflected by the progressive northward propagation of the spreading centre into the northern Tasman Sea, first from the Dampier to the Recorder Fracture Zone (Fig. 5b) and then later to the Cato Fracture Zone (Fig. 5c).

This breakup scenario, in which initial margin movements involved substantial strike-slip rather than extensional movements, accounts for the margin morphology which is atypical of a passive margin. The steep continental slope observed on the margin adjacent of the Eromanga-Brisbane Geoscience Transect is interpreted as an artifact of the original shear escarpment along which segments of the diverging margins separated. Jongsma & Mutter (1978) proposed that the initial strong strike-slip component led to asymmetric breakup, rupture occurring adjacent to the eastern Australian margin so that almost the entire rift system was inherited by the eastern side of the diverging Tasman Sea margin. This proposal explains both the observations of an almost complete lack of rift related sediment cover on the margin, including that adjacent to the Eromanga-Brisbane Geoscience Transect, and the origin of the thick sedimentary cover on the western flank of the Lord Howe Rise (BMR, 1987).

Lister & others (1986) state that such asymmetric development of passive margins may be more the norm, proposing that major detachment faulting plays a key role in passive continental margin development. Depending on the original dip of this detachment surface Lister & others (1986), recognise two broad classes of passive margins. Upper plate margins are those above the detachment surface, whereas lower plate margins, those below the detachment surface, are commonly the more highly structured containing most of the typical rift phase development. Applying these principles to the case of the Tasman Sea, the eastern Australian margin would be seen to be on the upper plate with the detachment fault dipping westwards beneath eastern Australia (Etheridge & others, 1990). The relevance of such a detachment surface to the westward dipping thrust surface intepreted by Korsch & others (1986), and separating an accretionary wedge beneath the New England Orogen, remains to be seen.

POST-BREAKUP (QUIESCENT PHASE).

The quiescent phase of margin development commenced with the cessation of seafloor spreading at approximately anomaly 24, about 55 Ma b.p., in the early Eocene. Overall rapid cooling of the entire basin probably enhanced subsidence rates compared to those normally anticipated for passive continental margins (e.g. Sleep, 1971) and DSDP Sites 207 & 208 on the Lord Howe Rise record current depths being reached during the Eocene. Along the eastern Australian margin, subsidence during this phase is documented only from petroleum exploration wells in the Gippsland and Capricorn Basins.

In the Gippsland Basin, a marine transgression established during the Maastrichtian continued and culminated in the Oligocene with deposition of a marine-dominated sequence, the Seaspray Group, at a time when palaeo-water depths increased from littoral to outershelf. However, deposition was interrupted for approximately 9 Ma commencing at about 35 Ma ago, with an erosional and non-depositional hiatus known as the Cobia event. Similar events have been identified in wells throughout the southwestern Pacific, including the DSDP sites on the Lord Howe Rise, where the breaks are of longer duration. The origins of this event are thought to be in part related to breakthrough of the circumpolar currents following the subsidence of the Tasman Basin to bathyal depths. As such, erosion and non-deposition would have dominated the entire eastern Australian margin for much of the Oligocene and perhaps early Miocene.

In the Capricorn Basin, cessation of seafloor spreading was followed initially by continuing terrestrially dominated sedimentation, with marginal marine conditions becoming established during the Oligocene. The delay in marine transgression across the Capricorn Basin is probably only apparent for two reasons. Firstly, the wells which provide control are located in relatively structurally high shelfal positions, and secondly, the marine transgression itself at these well locations may have coincided with the period of non-deposition and erosion - the Cobia event correlative of the Gippsland Basin. Subsequently, however, over 1000 m of calcareous claystones, marls and limestones were deposited within the Capricorn Basin during the Neogene-Recent period.

Correlation of burial histories in the Gippsland and Capricorn Basins suggests that the offshore portion of the margin east of the Eromanga-Brisbane Geoscience Transect became established at depths similar to those of the present day, during the Oligocene to early Miocene. However, non-deposition and erosion would have been expected until probably the Miocene, when shelfal, carbonate-dominated deposition would have become established. Consistent with this interpretation it is predicted that the sediment cover of the shelf adjacent to the Eromanga-Brisbane Geoscience Transect is of Neogene-Recent age.

Concurrent with overall margin subsidence, the post-breakup period also witnessed the uplift of the eastern highlands and establishment of the coastal hinterland, which, with its associated volcanism, dominates the onshore morphology of the eastern end of the Eromanga-Brisbane Geoscience transect. Opinions vary greatly as to when uplift commenced, estimates ranging from pre-Mesozoic (Stephenson & Lambeck, 1985) to pre-Cainozoic (Jones & Veevers, 1983), Early Cainozoic (Jones & Veevers, 1982), Mid-Cainozoic (Young & McDougall, 1982) and Late Cainozoic (Browne, 1969).

Wellman (1987), in a review of current evidence, favoured semi-continuous tectonic uplift throughout the Cainozoic. Previously, most workers placed strong emphasis on the apparent association of the eastern highlands with the adjacent continental margin. Propositions for a causal link between the two inexorably led to conclusions that the eastern highlands were uplifted during the rift or active seafloor spreading phase; that is, that uplift occurred during the Late Cretaceous and Early Tertiary. Corroboration for this was seen in the absence of any Late Cretaceous sedimentation covering the highlands, and that this time corresponded to a cessation of sedimentation in both the Surat and Eromanga Basins. For example, models by Karner & Weissel (1984) favoured lithospheric heating, as a precursor of Tasman Sea opening, as the responsible mechanism, whereas Lister & others (1986) proposed lithospheric underplating during the same period. However, uplift at that time on a scale which would have encompassed the eastern highlands should have been followed subsequently by subsidence, which is not observed.

In evaluating causal uplift mechanisms, Wellman (1987) placed more weight on the coincidence of the distribution of volcanism and extent of the eastern highlands, noting that the close association between the margin and highlands breaks down north of Brisbane, where the highlands swing inland extending over 400 km from the coastline (Fig. 1). By contrast, Cainozoic volcanism occurs along the entire highlands, including the inland Queensland areas where it has been active up to within the past 1 Ma (Coventry & others, 1985). In addition, the morphologic position of this volcanism relative to datable peneplanation surfaces, provided a means of dating successive episodes of uplift and established that relief developed at several times during the Cainozoic. For example, basalt flows dated as late Eocene (40 Ma) (Wellman 1974) infer 600 m of relief around the Koscuisko region, and at Barrington Tops basalts outcroping from 900 m to 1500 m elevation, indicate that similar relief was established some 50 Ma ago (Wellman, 1987).

The coincidence of volcanism and uplift supports an underplating mechanism, as do gravitational studies, which indicate that the eastern highlands are virtually in isostatic equilibrium (Wellman, 1987). As speculated by Wellman (1986, 1987), it is likely that the uplift of the eastern highlands was by a combination of mechanisms: heating and perhaps underplating during the mid-Cretaceous and underplating and some heating and erosion in the Cainozoic. Evidence in support of this is seen within the vicinity of the transect, especially in the Clarence-Moreton basin in northern New South Wales. Along the eastern coastal margin, sediments of the Clarence-Moreton Basin are truncated by surface erosion. Cross-section reconstructions indicate at least 1-2 km of erosion along a coastal zone (Ties & others, 1985). The interpretation of petroleum exploration seismic sections reveals that some of this truncation occurs across, post-depositional, reverse faulted structures. It is proposed that this erosional truncation occurred mainly as the net result of margin uplift associated with the rift onset during the heating stage and prior to the onset of seafloor spreading. To-day, these structures having been extensively eroded and show little, if any, obvious surface expression.

In contrast, some structures such as the Tullymorgan Anticline, also located relatively close to the coast, exhibit conspicuous topographic relief. Similar relief is observed across the Richmond Horst, a southern extension of the Moreton Anticline (Ties & others, 1985), where Wellman (1987) noted a step of 200-300 m in a basalt flow dated as mid-Cainozoic (20-24 Ma). Petroleum exploration seismic coverage across both features reveals their association with post-depositional compressional structuring and the step in elevation of the basalt flow indicating movement on the Richmond Horst since the early Miocene. Lying on the same trend as the Richmond Horst, structural reactivation also probably occurred at this time, and gave rise to much of the present-day relief, along the Moreton Anticline and West Ipswich Fault, which is traversed by the eastern end of the Eromanga-Brisbane Geoscience Transect (Fig. 3).

Further afield, two episodes of compression are documented in the Gippsland Basin (Esso, 1988). The older, Eocene, episode, being responsible for the creation of anticlines which trap the major petroleum accumulations. The nature of the link between

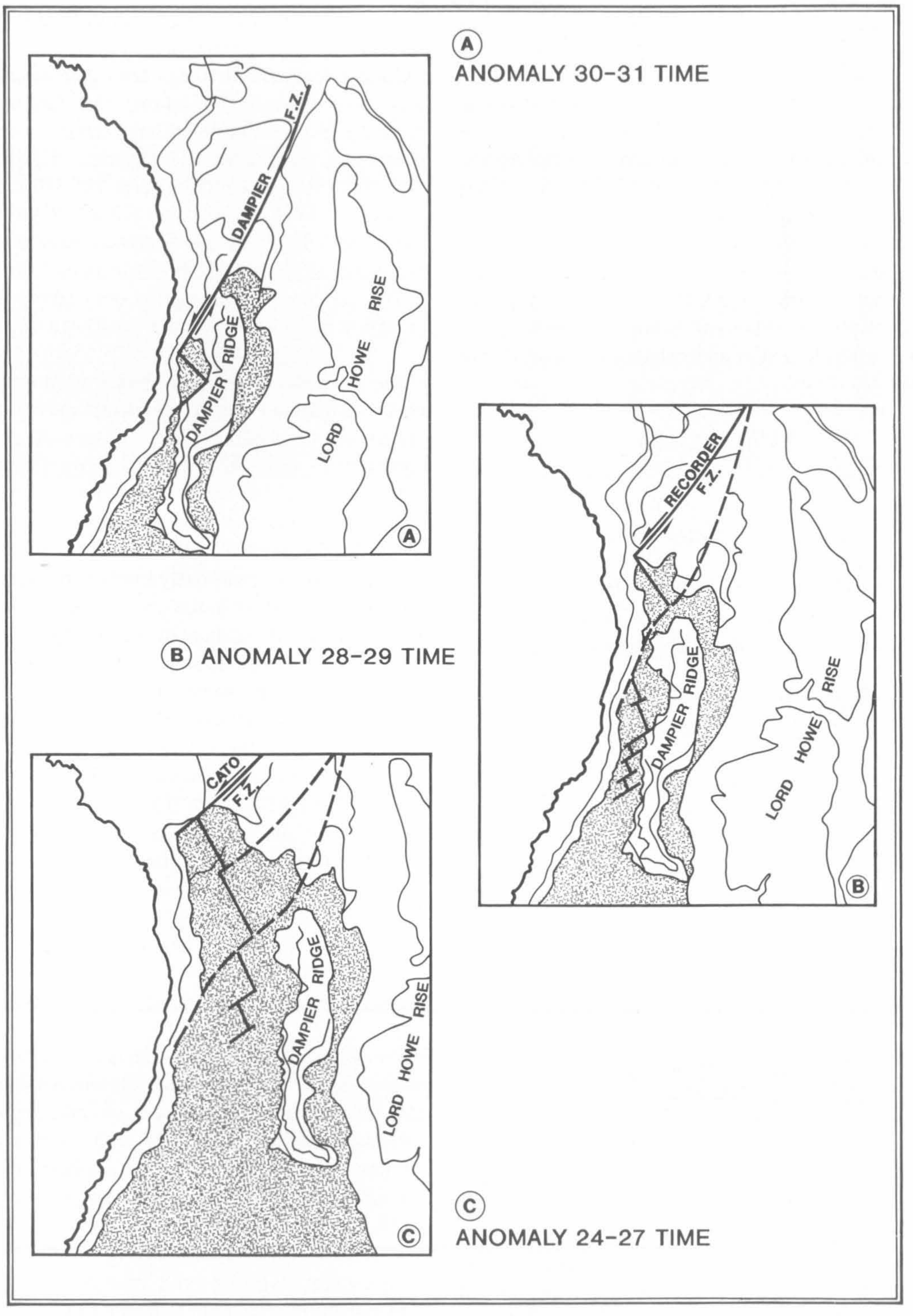

Fig. 5 Northern Tasman Basin reconstructions: A - Anomaly 30-31 time; B - Anomaly 28-29 time; C - Anomaly 24-27 time (after Shaw, 1979).

compression, volcanism and uplift in the Cainozoic in eastern Australia is unclear, although Jones & Veevers (1983) noted that in the Murray Basin recurrent cycles of transgression followed by regression could be associated with phases of more, then less, intense volcanic activity in the adjacent highlands. The most recent phase commenced about 7 Ma following a dormant interval from 7-10 Ma, with volcanism reaching a peak at 2-3 Ma. In the Gippsland Basin, this corresponds to the second phase of compression which is interpreted to be still active. Late Miocene to Pliocene uplift is also implied from discontinuities in the structural attitudes of basaltic flows across peneplanation surfaces in central Queensland (Coventry & others, 1985). Although volcanism of such recent age is not present in the immediate vicinity of the transect, many of the Clarence-Moreton Basin structures, especially those associated with topographic expression, may well have contemporary structural histories reflecting the effects of currently active underplating beneath eastern Australia.

SUMMARY

The post-Triassic development of the region adjacent to the eastern end of the Eromanga-Brisbane Geoscience Transect can be described with respect to the development of the adjacent Tasman Basin, in terms of a pre-rift phase, a rift phase, post-breakup (seafloor spreading phase) and a post-breakup (quiescent) phase.

During the pre-rift phase, eastern Australian was dominated during the Jurassic by a proto Pacific-Australian plate boundary located east of, and parallel to, the present eastern Australian margin. In the vicinity of the Eromanga-Brisbane Geoscience Transect, pre-rift sedimentation is represented by sequences of the Ipswich Basin, Esk Trough and Clarence-Moreton Basin. Deposition of these sequences commenced in the Upper Triassic and continued until at least the Late Jurassic. The Clarence-Moreton sequences were deposited during the final cooling subsidence phase following transtensional rifting, which earlier had produced the Esk Trough and Ipswich Basins. As such, their formation is unrelated to the Tasman Sea rifting processes which commenced about 140 Ma and led to the formation of the eastern Australian continental margin.

The Tasman Sea underwent breakup with the inception of active seafloor spreading, initially in the south, prior to anomaly 33 time (>85 Ma). Syn-rift sedimentation, and basin down-faulting in the central graben of the Gippsland Basin, place the age for the onset of rift development along the eastern Australian margin at about 140 Ma.

Rifting in the Gippsland Basin commenced in the earliest Cretaceous with the deposition of the Strzelecki Group. In the Capricorn Basin, rift development commenced in the Late Cretaceous, implying that rift development along the intervening margin, including that adjacent ot the Eromanga-Brisbane Geoscience transect, occurred time transgressively during the Cretaceous period. Remnants of the Tasman Sea rift development along the eastern Australian margin now are recognised definitively only in the Gippsland and Capricorn Basins. These basins are separated by approximately 1750 km of margin, which is conspicuously devoid of late Mesozoic or Early Tertiary rift-related sedimentation.

Commencing about 10 Ma prior to breakup, heat flow increased dramatically along the eastern coastline resulting in the resetting of apatite ages, magnetization overprints, and increased volcanism. Uplift associated with this event extended at least 100 km landward and resulted in erosion of perhaps 1-2 km of sediments from the eastern flank of the Clarence-Moreton Basin and enhanced exposure of the Palaeozoic Beenleigh and D'Aguilar blocks. This period also coincided with the cessation of Eromanga and Surat Basin deposition.

The inception of seafloor spreading in the central and northern Tasman Basins occurred during a series of northward propagations by the spreading centre. By anomaly 27 time (63 Ma b.p. Early Palaeocene) the spreading axis was continuous throughout the Tasman Basin, including the area adjacent to the Eromanga-Brisbane Geoscience Transect. Spreading ceased at anomaly 24 time in the early Eocene (55 Ma.b.p.). Initial breakup movements in the central and northern Tasman Basins involved substantial strike-slip motion (Shaw, 1978, 1979). This motion is reflected in the margin morphology, the steep slopes representing shear escarpments. The absence of rift sediments off most of the eastern Australian margin is attributed to this strike-slip motion and asymmetric breakup, the preferential rupture along the western side of the rift complex which formed over an interpreted westward-dipping crustal detachment surface (Lister & others, 1986; Etheridge & others, 1990). As a consequence, most of this rift complex was subsequently inherited by the Lord Howe plate following breakup and seafloor spreading.

During the Cainozoic, the margin near the eastern end of the Eromanga-Brisbane Geoscience Transect, as with the remainder of the margin, experienced subsidence and overall marine transgression. This was interrupted during the Oligocene by a period of non-deposition and submarine erosion, associated with the breakthrough of circumpolar currents. Present margin depths were probably established during the Oligocene in the south, and Miocene in the north.

Concurrent with overall margin subsidence, the post-breakup period also witnessed uplift of the eastern highlands and associated volcanism. Notwithstanding the appeal of mechanisms, which link the apparent close field association of the continental margin and adjacent highlands, evidence now favours more the interpretation that uplift occurred mostly during the Cainozoic, associated with episodic volcanism and underplating. Evidence of multiple staged uplift in the vicinity of the

transect is seen from structural deformation of basalt surfaces contrasted with other post-depositional structures showing no apparent topographic expression. The most recent movements date from the early Miocene and involve compressional reactivation of structures, such as the South Moreton Anticline and West Ipswich Fault. Uplift during these periods of reactivation have enhanced differential erosion and led to the development of coastal river capture systems as distinct from the interior dendritic drainage patterns.

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THE SEISMO-TECTONICS OF SOUTHEAST QUEENSLAND

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ABSTRACT

The relationship between seismicity and tectonics in southeast Queensland and northeast New South Wales has been investigated using a database of over 1300 earthquakes. Most of these events were located using a seismograph network that was installed primarily to monitor seismicity associated with several large reservoirs, and the configuration of the network creates a non-uniform detection level over the study area. However, careful consideration has been given to the accuracy of the computed locations, and events that occurred prior to the installation of the sensitive network have been excluded from the analysis because they have large uncertainties in location.

Earthquake activity is mainly concentrated in the onshore coastal regions, but also extends off the coast southeast from Fraser Island, beyond the continental shelf. The coastal activity is mainly within the New England Belt. The central area of the fold belt overlain by the Clarence-Moreton Basin is also active. Inland there is an area of activity near St. George that correlates with a basement high, the Wunger Ridge. Fraser Island and the adjacent onshore area, near Maryborough, are notably aseismic.

Within the New England Fold Belt is a zone of activity that includes the South D'Aguilar Block and the Esk Trough. This zone can be extended northward to include the concentrated activity in the Gayndah region. The Nambour Basin, and the North D'Aguilar, Gympie, and Yarraman Blocks have been significantly less active. Within the New England Block in the south of the study area the earthquake activity is concentrated between the Peel and Demon Faults in the Armidale-Inverell-Texas region.

Accurate hypocentre locations have enabled a detailed study of the seismicity in the Wivenhoe Dam area. Activity is concentrated within several kilometres of the southern end of the eastern bounding fault of the Esk Trough. Farther north, the activity is up to 10 km to the west of the mapped outcrop of the fault. Seismicity in this area defines a northeast-dipping structure to at least 10 km depth that closely parallels a gravity anomaly modelled by Leven (1977) as a shallow dipping reverse fault. Composite focal mechanisms of several groups of accurately located earthquakes in southeast Queensland indicate predominately reverse faulting. This faulting is the result of the crust being compressed horizontally in a northeast-southwest direction.

INTRODUCTION

The results of 12 years of seismic monitoring (1977-1989) in southeast Queensland are presented in this paper. The study area includes southeast Queensland and northeast New South Wales from 22.5° to 31° South and from 148.5° to 156° East (Fig. 1). A summary of the tectonic evolution of this area is given by Murray (this Bulletin) and by Korsch & others (this Bulletin). Particular emphasis is given to the Wivenhoe Dam area where the seismograph network provides sensitive earthquake detection and accurate location. The configuration of the seismic network, and data processing techniques are described. Results of focal mechanism analyses and correlations between seismicity and mapped geological structures are presented.

The Queensland Department of Mines (now Department of Resource Industries) began seismicity studies in 1977 with the primary aim of investigating seismicity associated with the filling of large reservoirs. The first installation was the Wivenhoe Dam network in southeast Queensland. This was followed by the installation of single seismographs at Boondooma and Awoonga Dams in 1980 and 1981 (Fig. 1). In addition to monitoring specific reservoirs, these seismographs form a regional network that monitors seismicity in southeast Queensland and northeast New South Wales. However, the positioning of the seismographs in the primary study areas around each reservoir has created a network that is not ideal for monitoring on a regional scale. The linearity of the network also increases errors in some earthquake locations.

Despite these limitations, the data provided by the network have enabled a better understanding of the seismo-tectonics of southeast Queensland and northeast New South Wales. Studies can be undertaken to delineate zones of seismic activity and the correlation of these zones with mapped geological structures provides the basis for understanding the current tectonic regime. Accurate earthquake locations are required for meaningful correlations. In this paper, analysis is restricted to those earthquakes with location errors smaller than the dimensions of the geological structures being studied. Earthquakes located solely from arrival times at distant seismographs, or from intensity reports, are rejected.

SEISMIC INSTRUMENTATION

The Wivenhoe Dam network commenced operation in 1977 with the installation of four recorders in a small network measuring 20 km on each side. This network was expanded both in size and number of recorders to the current configuration of seven instruments forming a network 60 km by 40 km. Improved timing accuracy and a decrease in recorded noise levels now enables micro-earthquakes inside the Wivenhoe Dam network with magnitudes as low as -0.6 to be accurately located.

The installation of single seismographs at Boondooma Dam (BDM) in 1980 and at Awoonga High Dam (AWG/AWA/AWD) in 1981 improved earthquake detection levels and provided additional control to earthquake locations in southeast Queensland.

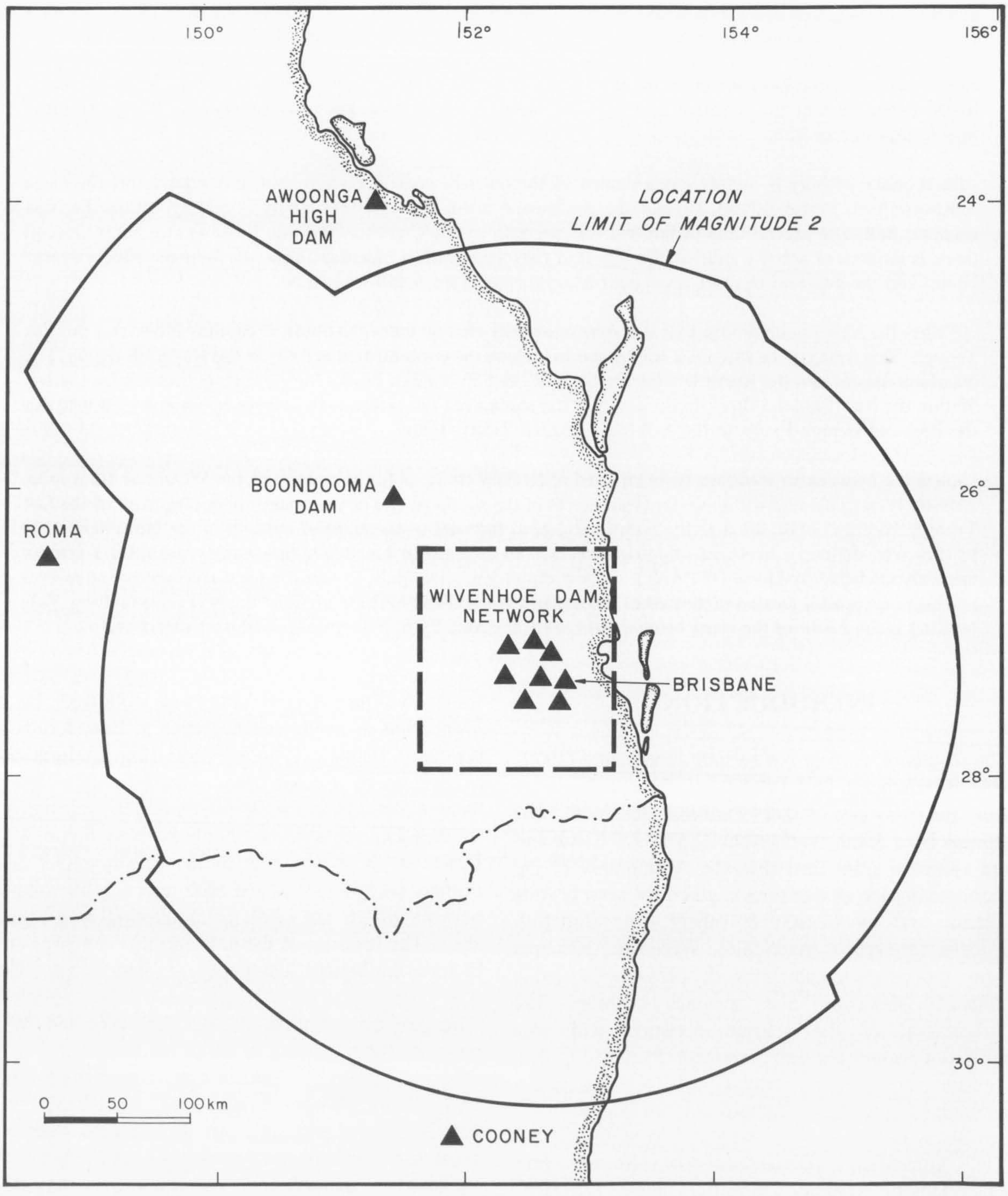

Fig. 1 Seismographs in the study area (as at 1/12/89) and location limit for magnitude 2 earthquakes. Inset shows area of Figures 3 and 7.

TABLE 1

Seismographs in southeast Queensland and northeast New South Wales.

CODE	LONGITUDE°E	LATITUDE°S	START	FINISH	NAME
Currently operating					
BRS	152.7750	-27.3917	63-06-00	- - -	Brisbane (Mt. Nebo)
COO	151.8916	-30.5777	74-04-00	- - -	Cooney, NSW
WPM	152.7355	-27.5357	78-08-04	- - -	Pine Mountain
WTG	152.3333	-27.1458	79-04-08	- - -	Toogoolawah
WMB	152.5502	-27.1155	79-05-28	- - -	Mount Brisbane
BDQ	153.0417	-27.8817	80-04-25	intermit.	Beaudesert
BDM	151.4444	-26.1123	80-07-29	- - -	Boondooma Dam
WBA	152.3082	-27.3527	84-07-06	- - -	Buaraba
WRC	152.6631	-27.1874	84-07-11	- - -	Reedy Creek
WWH	152.5872	-27.3702	84-07-12	- - -	Wivenhoe Hill
RMQ	148.755	-26.489	84-08-11	- - -	Roma
AWD	151.3157	-24.0780	87-07-01	- - -	Awoonga Dam #3
WCR	152.455	-27.520	89-02-27	- - -	Cricket Road
Discontinued sites					
CODE	**LONGITUDE**	**LATITUDE**	**START**	**FINISH**	**NAME**
BRS	153.0311	-27.4778	37-09-00	51-04-00	Brisbane (George St.)
BRS	153.018	-27.502	51-03-00	63-06-00	Brisbane (St. Lucia)
WDC	152.6717	-27.4318	77-03-18	78-12-01	Deep Creek
WLR	152.5438	-27.4607	77-03-24	78-12-13	Lowood Range
WKC	152.6438	-27.3448	77-03-30	78-02-10	Kipper Creek
WLC	152.5243	-27.3485	77-04-05	78-02-10	Logan Creek
WMC	152.6203	-27.2305	77-08-24	78-06-17	Middle Creek
WME	152.4712	-27.2347	77-08-31	78-07-03	Mt. Esk
WBG	152.6542	-27.1158	78-03-03	78-07-17	Byron Gully
WSC	152.4177	-27.0180	78-03-03	78-07-17	Scrub Creek
WPK	152.6018	-27.6035	78-08-07	78-12-13	Perry's Knob
WSD	152.5540	-27.1137	78-12-22	79-05-28	Somerset Dam
WLL	152.5615	-27.4258	78-12-22	79-03-15	Lowood
WSH	152.4418	-27.6047	79-04-09	79-05-27	Summerholm
WGR	152.4457	-27.6262	79-05-28	79-09-09	Grandchester
WPL	152.4168	-27.6058	79-10-10	84-05-23	Plainland
AWG	151.3157	-24.0462	81-09-22	87-04-14	Awoonga Dam #1
RTQ	148.8514	-26.5678	83-11-16	84-08-10	Roma (Temp)
WTR	152.4645	-27.5286	84-01-31	89-02--26	Thallon Road
AWA	151.3017	-24.0681	87-04-14	87-06-30	Awoonga Dam #2

Only those sites that have operated for more than 2 months are listed.
Start and finish dates are listed as year, month, day.

The seismographs at Wivenhoe Dam, Boondooma Dam, and Awoonga High Dam consist of a single, short-period (1 Hz), vertical seismometer recording onto a smoked-paper recorder. Careful preparation of the smoked-paper records and the application to the records of timing marks every second has ensured that arrival time errors are kept below 0.1 s. Local micro-earthquakes produce seismic waves with a dominant frequency range of 4 to 20 Hz. Electronic filters are used to reject seismic signals outside this range. Signals from distant earthquakes are attenuated by this filtering.

The seismographs operated by the Bureau of Mineral Resources at Roma (RMQ) and Cooney (COO), and by the University of Queensland at Brisbane (BRS), have a lower overall magnification and a response in a lower frequency band. These sites are better equipped to record distant events. Photographic or pen-and-ink recording, coupled with the lack of one-second timing marks, limit timing accuracy and resolution.

Details of seismographs in southeast Queensland and northeast New South Wales are given in Table 1. Figure 1 shows the area within which earthquakes with magnitude ≥2 can be located with the present operating network. This area covers most of southeast Queensland and northeast New South Wales.

DATA PROCESSING

Event Identification

A major practical problem in any seismic monitoring in an urbanised area, such as southeast Queensland, involves differentiating between natural earthquakes and man-made explosions. It is imperative that no earthquakes are assumed to be explosions and, conversely, it is important to ensure that blasts are not incorrectly identified as natural earthquakes. Various criteria, such as the sense of the first motion, presence or absence of surface waves, and P and S wave amplitudes, are studied to ensure the correct identification of seismic events.

Earthquakes from within the study area account for less than 5% of the seismic events recorded. Large distant earthquakes account for approximately 10%, with the remainder (over 85%) being blasts from coal mines, hard-rock quarries, or construction sites. Of the 800 events recorded at Boondooma Dam seismograph during 1988, less than 40 were earthquakes from within the study area. The recording of underground collapses at coal mines around Ipswich (Rynn & others, 1984-1985) is an additional problem in the Wivenhoe area.

Earthquake Location

Event location is done using a program that allows easy editing of arrival times (Gibson & others, 1981). The interactive nature of the program allows fast and accurate earthquake location. Events recorded by three or more seismographs are usually located with an accuracy of less than 10 km in a horizontal direction.

Information for each earthquake, including arrival times and location data is stored on computer files for subsequent output or further analysis. The files also include earthquakes which occurred prior to the installation of the Wivenhoe Dam network, such as those listed in Jones (1958) and Rynn (1987).

The database now contains information on over 1300 earthquakes in eastern Queensland and northeast New South Wales, from the first recorded event in 1866 to the end of 1988. Over 1000 of the events on file have occurred since 1977, reflecting the increased coverage and sensitivity of the seismograph network in recent years.

Location Accuracy

Of particular interest to this study is the earthquake location accuracy. The accuracy code entered into the computer database for each earthquake is based initially on the standard deviations that are computed by the

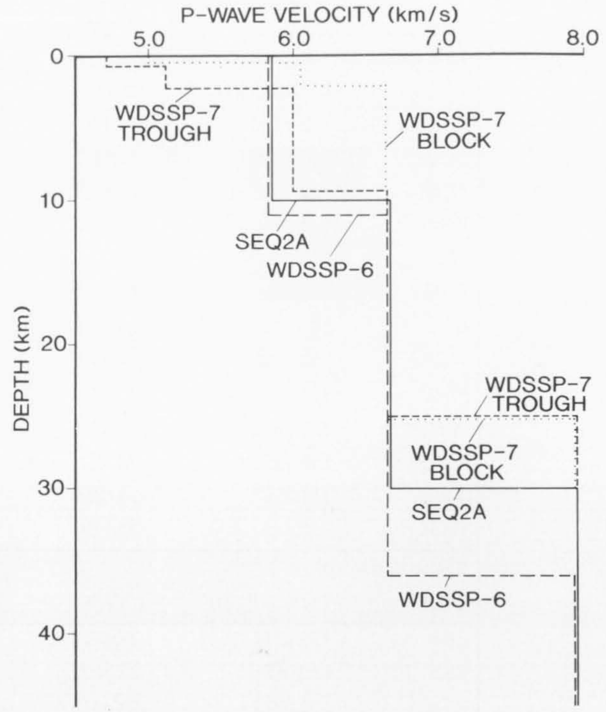

Fig. 2 Crustal velocity models for southeast Queensland. Model WDSSP-6 from Rynn (1984a), WDSSP-7-Trough and WDSSP-7-Block from Rynn (1984c), and model SEQ2A by author.

earthquake location program. The computed errors may be reassessed if the operator considers they are not a true indication of the real error in location.

Location errors that are unrealistically low can occur from the use of numerous seismic phases in the computer location routine. Arrival times for a phase arriving at several seismographs may be used even when the identification of the phase is uncertain. The proper identification of these phases is not taken into account by the computer program, with the result that the mathematically determined location error will be artificially low. In this case the operator may choose to decrease the computed accuracy of the location.

Unrealistically high mathematical location errors can result from an event that is close to one seismograph. The event location may be well constrained by the P to S time interval at the seismograph. The computed location error in this instance may be much larger than this constraint warrants. Again, in this instance the operator would override the computed location error and increase the accuracy code of the location.

Appropriate accuracy codes are also given to earthquakes that have not been instrumentally located. Most of the earthquakes that occurred prior to the commencement of the Wivenhoe Dam network were located using a calculated distance to one or perhaps two seismographs, and intensity reports. The accuracy code indicates the level of confidence that can be placed on the calculated locations of these earthquakes.

Crustal Models

A model of seismic velocities in the crust is a prerequisite for the location of an earthquake using seismic arrival times. The model used in this study (denoted SEQ2A in Figure 2) was obtained by the author in 1979 using small timed blasts from local quarries and coal mines, as well as larger shots from the Bowen Basin coal mines to the northwest. It represents the average structure of the crust in southeast Queensland. Using this model in a location program will give earthquake hypocentres that do not differ significantly from those obtained with a more detailed model. This model compares well with a preliminary model obtained subsequently by Rynn (1984b) for the Wivenhoe Dam area (WDSSP-6 in Figure 2), and with a model for the Esk Trough (WDSSP-7 Trough in Figure 2) (Rynn, 1984c).

Magnitudes

Two methods of magnitude calculation have been used in this study. Local Richter magnitudes (ML) have been determined from the formula of Richter (1935) with appropriate corrections to allow for instrument response (Gutenberg & Richter, 1956). Simple tests have shown that this method provides magnitude estimates that have little or no distance dependence.

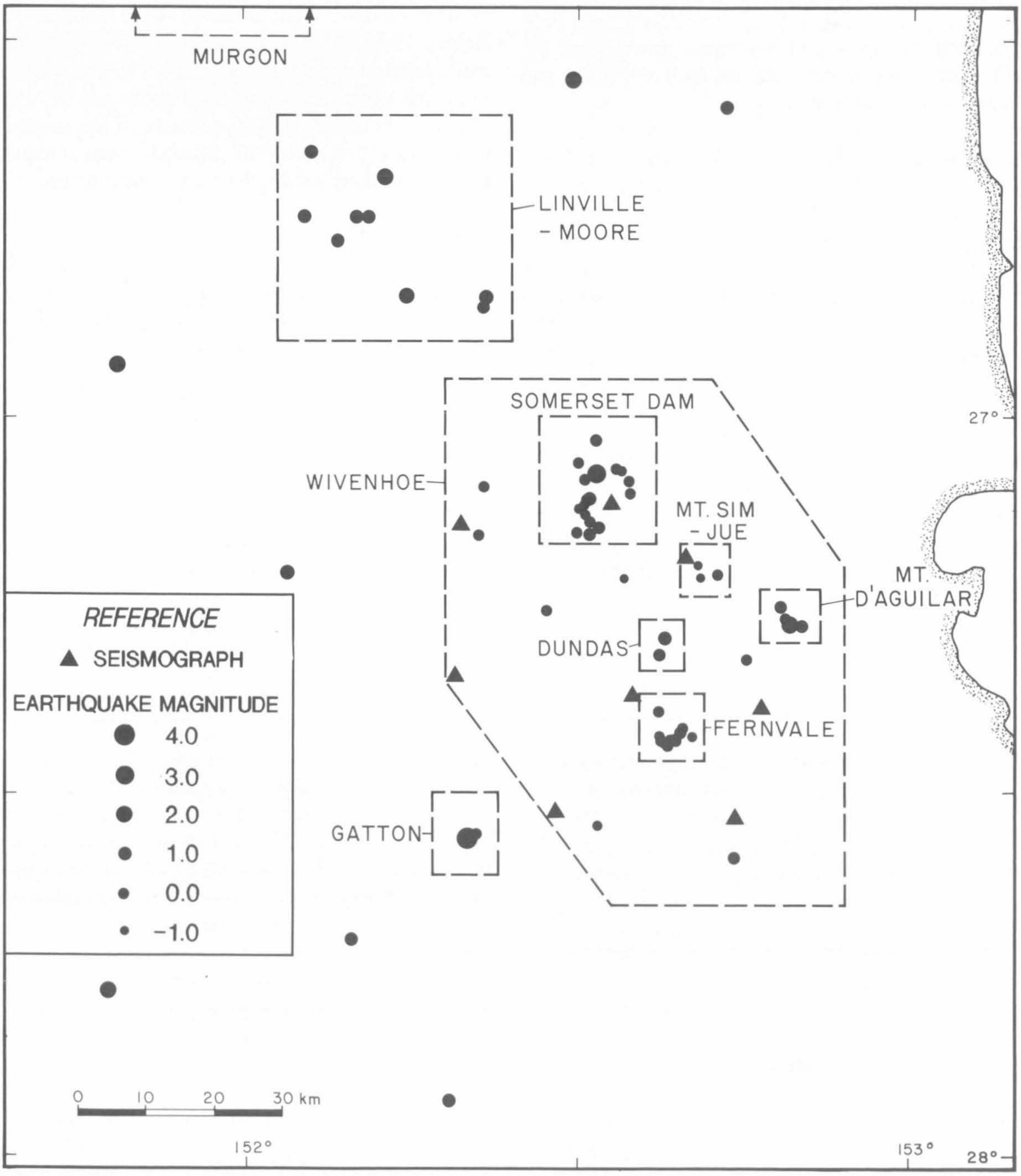

Fig. 3 Earthquakes and zones used in focal mechanism analysis. See Table 2 for details on zones and solutions.

Duration magnitudes (Md) have been calculated for only a small proportion of the events in the database using a formula determined by the author (Cuthbertson, 1977) for similar instrumentation in Victoria. The direct application of this formula to Queensland produces magnitudes that are overestimated by an average of 0.3 units. The slight overestimation in the small number of Md's in the database is of little consequence, especially for this study where the earthquake location is of prime importance.

FOCAL MECHANISMS

Theory

The sense and direction of motion at the focus of an earthquake can be derived from the distribution of first motions of arrivals. These focal mechanisms have been used since the 1920's to determine more about the earthquake source; in particular the fault orientation and the causative stress field.

A knowledge of the earthquake location and the velocity structure of the crust is required to obtain a focal mechanism. If the earthquake location and crustal structure are known then the initial direction at which the rays left the hypocentre can be determined. The sense of first motions is plotted on an imaginary unit-sphere surrounding the earthquake focus. The distribution of compressions and dilatations on the focal sphere can generally be divided with two orthogonal planes. One of these planes represents the fault plane, upon which the earthquake motion occurred. The second plane, termed the auxiliary plane, is perpendicular to both the fault plane and the direction of motion on the fault plane. Distinguishing between the fault plane and the auxiliary plane is generally not possible using first motion data. For the analysis in this paper the distinction is not relevant.

Uncertainties in the crustal model can introduce errors into the analysis. The distance from an earthquake at which the first arrival changes from a direct arrival to a refracted arrival is particularly sensitive to the crustal model. This distance is termed the crossover distance. An arrival close to the crossover distance may be incorrectly assumed to be a refracted arrival when it is in fact a direct arrival. These phases have markedly different angles of departure from the focus. The inclusion in the analysis of an arrival close to the cross-over distance can lead to substantial errors. The departure angles of other arrivals, in contrast, are not critically dependent on the crustal model. The departure angle of first arrivals at a seismograph directly above an earthquake will only change slightly if the earthquake location and crustal model are altered.

The crustal stress field can be inferred if the orientation of the fault plane and auxiliary plane are known. Anderson (1951) showed that for the fracture of homogeneous material with no prior faulting, the principal compressive stress (S_1) must lie:

- in the plane that is perpendicular to both the fault and auxiliary planes (third orthogonal plane),
- in the dilatational quadrant,
- at an angle to the fault plane of 45° or less.

The internal coefficient of friction accounts for deviations from 45°.

The required condition of homogeneous material with no prior faulting is seldom observed. Low stress-drops measured in shallow earthquakes, being too small by an order of magnitude to produce fracture in homogeneous material, indicate that earthquakes occur on pre-existing faults (Chinnery, 1964; Brune & Allen, 1967; Wyss & Brune, 1968). McKenzie (1969) showed that for an earthquake on a pre-existing fracture, in a triaxial stress field, the principal compressive stress can be located anywhere in the dilatational quadrant. If the stress field is uniaxial ($S_2=S_3$), then the principal stress is restricted to lie somewhere on the third orthogonal plane.

The focal mechanism for an earthquake is controlled by the orientation of the causative stress field and the pre-existing fracture. Over an area of uniform stress the variations in focal mechanisms will be due to different orientations of the pre-existing fractures. It is possible to estimate the stress field by comparing the focal mechanisms of several earthquakes. Various statistical techniques have been used to obtain stress field orientations from a family of focal mechanisms (Gephart & Forsyth, 1984; Angelier & others, 1982), but the simplest method is to find the range of directions which is common to all dilatational quadrants. The principal stress, S_1, must lie within this range (Carey-Gailhardis & Mercier, 1987).

Focal Mechanisms in southeast Queensland

Application of focal mechanism theory to southeastern Queensland involved selecting a set of well-located earthquakes. The only area in which earthquakes have been located with sufficient accuracy is in the vicinity of the Wivenhoe Dam network. From a set of 68 accurately located earthquakes, 226 first motions were measured. Removal of arrivals at distances close to the crossover distance between direct and refracted phases, reduced the number of first motions to 208.

Individual focal mechanism solutions for each event were not possible because of the small number of observations and poor azimuthal distribution. The earthquakes were therefore grouped into zones based on their geographical locations, and composite focal mechanisms were attempted for each zone. This method assumes the stress field is invariant within each area being considered. Furthermore, it is assumed that the earthquakes are either located on a single fault plane or

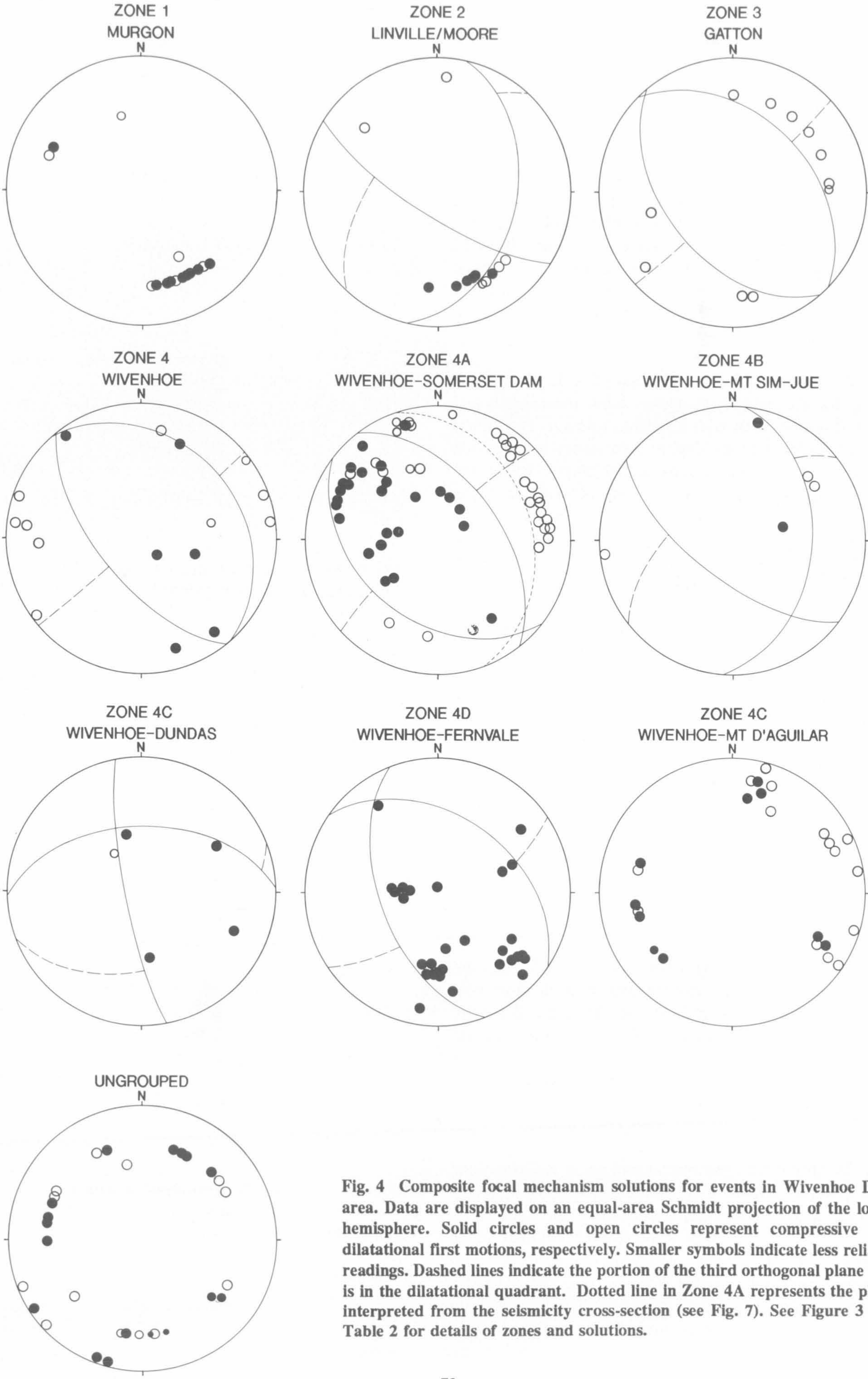

Fig. 4 Composite focal mechanism solutions for events in Wivenhoe Dam area. Data are displayed on an equal-area Schmidt projection of the lower hemisphere. Solid circles and open circles represent compressive and dilatational first motions, respectively. Smaller symbols indicate less reliable readings. Dashed lines indicate the portion of the third orthogonal plane that is in the dilatational quadrant. Dotted line in Zone 4A represents the plane interpreted from the seismicity cross-section (see Fig. 7). See Figure 3 and Table 2 for details of zones and solutions.

on multiple fault planes with identical orientations. If these two requirements are met, then all earthquakes in the zone will have identical focal mechanisms which can be combined to obtain a more constrained solution.

Figure 3 shows the locations of earthquakes used in this analysis. The earthquakes were grouped into several geographical zones: Gatton, Murgon, Linville-Moore and Wivenhoe Dam. Concentrations of activity in the Wivenhoe Dam zone suggested it could be subdivided into smaller zones: Somerset Dam, Mt. Sim-Jue, Dundas, Fernvale, and Mt. D'Aguilar (Table 2). Events that were not in any of the above zones were classified as "ungrouped". Figure 4 shows the data for each zone and the fitted focal planes. Less reliable data are indicated by a smaller symbol. Less weight was given to these arrivals.

Data for zones 1 (Murgon) and 4E (Mt. D'Aguilar), and for the ungrouped events, were inconsistent and could not be fitted with a reliable solution. The failure to fit solutions to the data in these zones indicates the failure of the assumptions stated above. Either the individual earthquakes in the zone occurred on faults with varying orientations or the stress field in the zone was not constant. It is possible that errors in the earthquake locations or inaccuracies in the crustal model contributed to the inconsistent data.

Data for zones 3 (Gatton) and 4A (Somerset Dam) provided reliable solutions that indicated reverse faulting. The predominance of compressive first arrivals in the centre of zone 4D (Fernvale) also indicated reverse faulting. The limited data in zones 2 (Linville-Moore), 4B (Mt. Sim-Jue), 4C (Dundas), did not constrain the focal planes. These zones were fitted with solutions indicative of predominately reverse faulting so as to agree with the solutions of the other zones. The final solutions were based on agreement of the third orthogonal plane (see following). The solutions for zones 2, 3, 4, 4A, 4B, and 4D all had a focal plane striking northwest.

The dilatational quadrants of all the focal mechanism solutions (Table 2) had a common range of directions. The interpretation is that there is a uniform stress field over southeast Queensland. In addition, the third orthogonal plane for these solutions converges at a point (Fig. 5) which indicates the stress field is also uniaxial. The point of convergence indicates that the stress field in southeast Queensland has a principal stress acting horizontally in a northeast-southwest direction.

The only other focal mechanism study in Queensland is by Rynn (1984a) who studied earthquakes in an area close to the Wivenhoe Dam. There are differences between the data used by Rynn (Fig. 6) and those in the equivalent zone for this study (zone 4D in Fig. 4). Rynn incorporated all located earthquakes, regardless of location accuracy, and all arrivals, regardless of sensitivity to the crustal model. Some differences are

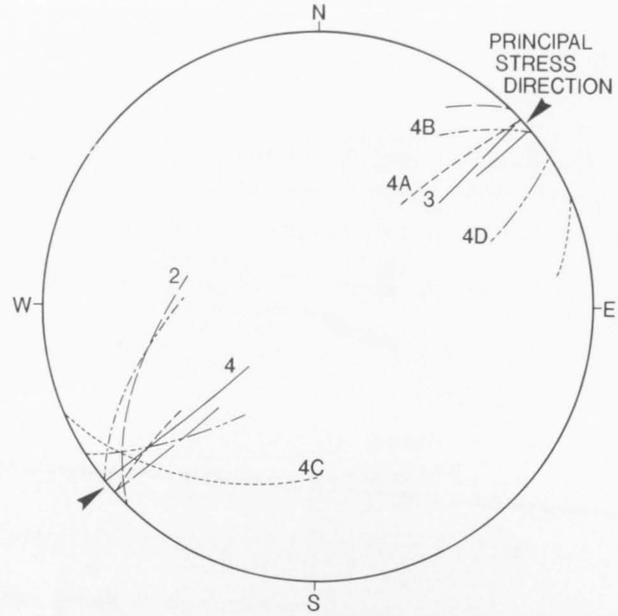

Fig. 5 Portions of third planes in the dilatational quadrants. Taken from individual solutions in Figure 4.

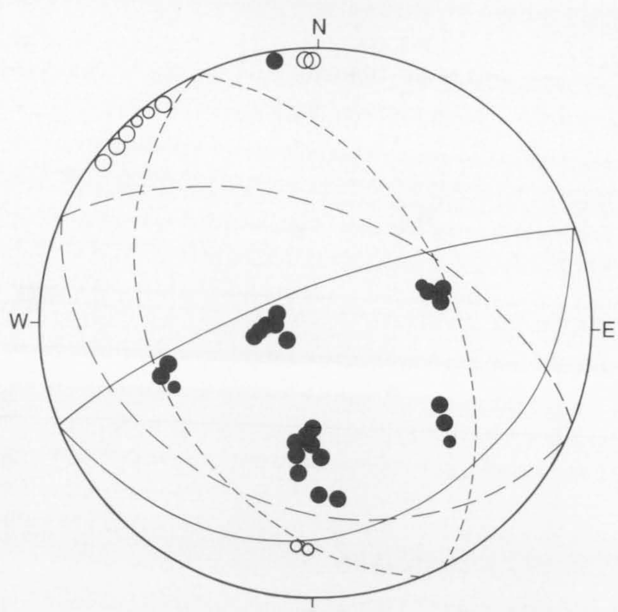

Fig. 6 Composite focal mechanism data for the Fernvale Bridge area (modified from Rynn, 1984a). Solid lines represent original solution from Rynn (1984a). Dashed lines represent alternative solutions by author.

also due to the different location programs used to locate the events.

The interpretation by Rynn (1984a) is for reverse faulting with a strike of 68°. Two alternative reverse-faulting focal mechanism solutions, with strikes of 110° and 155° have been drawn by the author to fit the original data of Rynn (Fig. 6). These solutions fit the data as reliably as the original solution of Rynn. The range of possible strike directions indicates that the focal planes are not well constrained.

Rynn (1984a) used the result of the focal mechanism to relate the seismicity in the Fernvale Bridge area to a small inferred fault on the Caboolture 1:100 000 map (Murphy & others, 1979). The additional data used in this paper, obtained from a number of geographical locations, has provided an alternative solution with a stress direction perpendicular to that of Rynn.

SEISMICITY

Wivenhoe Dam Area

The Wivenhoe Dam network has provided 12 years of earthquake monitoring in southeast Queensland. Over 200 earthquakes, with magnitudes from -2.0 to 4.0, have been located in the Wivenhoe Dam area. Of these, 68 have been located with sufficient accuracy to allow correlation between seismicity and geological structure.

Earthquakes located in the Wivenhoe Dam area since 1977 with error bounds of less than 50 km are plotted in Figure 7, together with the gross geological structure. Most activity has been in a belt trending north-northwest (shaded in Fig. 7) that includes the South D'Aguilar Block and the Esk Trough. Within this belt there are several areas where activity is concentrated. The Nambour Basin, the Gympie Province, and the Yarraman Block are outside the main seismicity belt and are significantly less active.

An additional line, that corresponds in part to the exposed margin of the Yarraman Block, has been drawn in Figure 7. This lineament is based on limited data, but parallels the main belt of activity.

Several of the areas of concentrated seismic activity within the main seismic belt are associated with the eastern bounding fault of the Esk Trough (Eastern Border Fault in the north, Great Moreton Fault in the south) (Rynn & others, 1984-1985). Activity is concentrated within several kilometres of the bounding fault at its southern end. Farther north, the activity is up to 10 km to the west of the mapped outcrop of the fault.

As has been observed elsewhere in Australia (Lambeck & others, 1984) the majority of epicentres had focal depths in the middle to upper crust, at depths of approximately 10 km. A regular variation of focal depths was observed at only one locality in the Wivenhoe Dam area (see following). Earthquakes from the remainder of the region showed no regular variation. This may be due to inaccuracies in the depth estimates for earthquakes away from the Wivenhoe Dam network.

Figure 8 is the cross-section, A-B, of earthquakes in the area shown as the inset in Figure 7, and includes the gravity model of Leven (1977) (see following). This plot shows a deepening of earthquake foci towards the east. All events in this figure have accurate epicentres, but have varying depth errors. The majority of the events lie within 1 km of the fitted line. This line represents a plane striking approximately 160° and dipping 30° to the east.

The accuracy of the strike and dip is limited because of the errors in the depth estimates. However, there is

TABLE 2

Focal planes for earthquake zones

ZONE		NAME	EVENTS	ARRIVALS	STRIKE	DIP	STRIKE	DIP
1		Murgon	2	17	-	-	-	-
2		Linville-Moore	9	12	13	40	122	75
3		Gatton	2	11	140	48	314	42
4		Wivenhoe	7	16	141	64	318	26
4	A	- Somerset Dam	17	58	128	39	322	52
4	B	- Mt. Sim-Jue	3	6	127	66	5	40
4	C	- Dundas	2	5	169	80	268	50
4	D	- Fernvale	3	28	159	54	304	42
4	E	- Mt. D'Aguilar	4	25	-	-	-	-
----		"Ungrouped"	9	30	-	-	-	-

Note :- Dips are measured to the right in the direction of the strike.

little doubt that the events in this area are associated with a shallow east-dipping plane with a northwesterly strike.

The plane determined from Figure 8 is also shown on the focal mechanism plot for the same area (Fig. 4, zone 4a). The dip and strike of the plane determined from the cross-section and one of the planes from the focal mechanism plot are in good agreement.

Modelling of Bouguer gravity anomalies from a profile across the Esk Trough and Northbrook Block revealed a structure interpreted as a low-angle thrust fault dipping to the east beneath the Northbrook Block (Leven, 1977). The profile was within the inset area of Figure 7. The location of the modelled fault was several kilometres to the west of the mapped location of the eastern bounding fault.

Gravity maps (Wellman, this Bulletin) show the gravity contours in this area to be aligned north-northwest. This direction may be taken to represent the strike of the fault modelled by Leven. Cross-section A-B, which is perpendicular to the assumed strike of the fault, shows the fault modelled from gravity (Fig. 8).

Alternative models for the gravity anomaly were tried in an attempt to better match the fault delineated by the seismicity. The dip and throw of the modelled fault can be increased if the density contrast used by Leven is

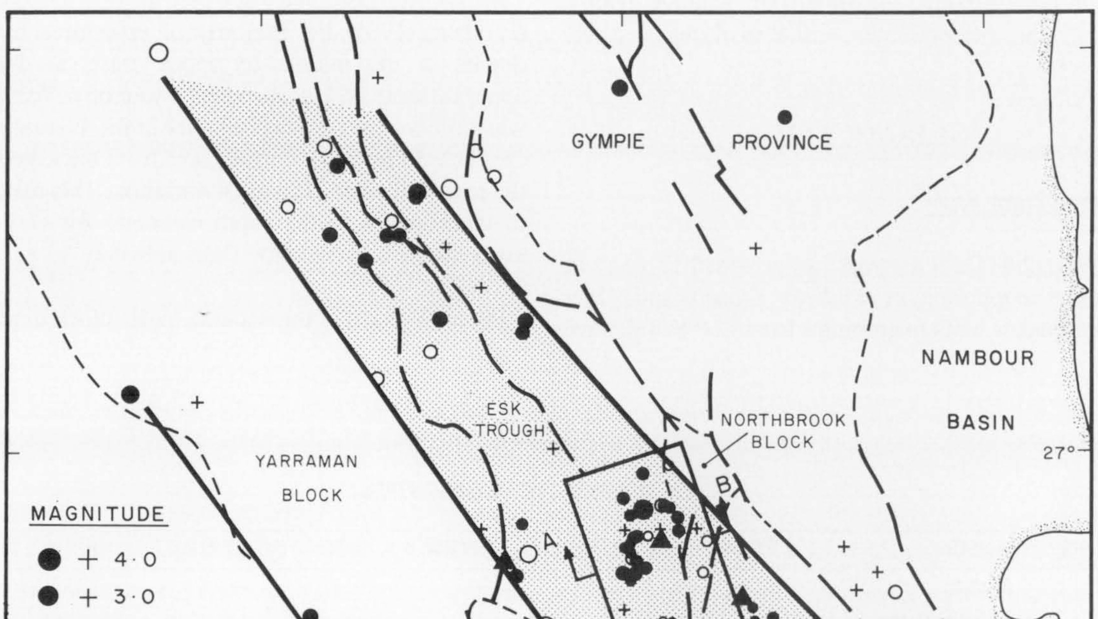

Fig. 7 Earthquakes, seismicity lineaments, and generalised geological structure in Wivenhoe Dam area. Closed circles, open circles and crosses represent earthquakes located to within 5, 10 and 50 km, respectively. Inset indicates area of section A-B (Figure 8). Generalised geology from Whitaker & Green (1980).

reduced. However, the gravity anomaly restricts the possible models so that the best agreement that can be achieved is to have the seismicity fault and the gravity model to be parallel but separated vertically by 7 km.

Southeast Queensland and Northeast New South Wales

Seismicity of southeast Queensland and northeast New South Wales is shown in Figure 9. Earthquakes since 1977, with magnitudes greater than 1, are plotted in the same area as depicted in Figure 1. Only earthquakes located with an accuracy of 50 km or better are plotted. This restriction is acceptable for correlations at this scale. There is a tendency for the location uncertainties to increase towards the edge of the figure, due to the increased distance from the majority of the seismographs.

Any plots of recorded seismicity in the study area will be affected by variations in earthquake detection levels. If this effect is to be minimised, then only those events with magnitudes greater than 2 should be plotted (see location limit in Fig. 1). However, this would severely reduce the number of events for analysis. A compromise between uniform data coverage, and acceptable levels of activity had to be reached. Plotting earthquakes with magnitudes greater than 1 will introduce a bias, due to the large number of small magnitude events detected close to the Wivenhoe network. However, this is offset by the larger number of events available for analysis.

Most of the epicentres depicted in Figure 9 are concentrated in the coastal and offshore regions. Activity in the inland region, with the exception of the border area near St. George, is markedly lower than that in the coastal areas.

Rynn (1987), using earthquake data to the end of 1984, noted a "pocket" of activity in the region of the Tasmantid Guyots of the North Tasman Basin (Vogt & Conolly, 1971) (indicated by the isolated 4000 m water depth contours). This activity is now seen as a continuous zone, extending from near Fraser Island, southeast over the edge of the continental shelf.

The true level of activity in the St. George area, and also off the Queensland coast, is probably much higher than indicated by the number of recorded events. Many small events in these areas will have gone undetected because of the poor detection capabilities of the existing network. This will also be true of the activity near Armidale, in New South Wales. In contrast, the concentrated activity near Gayndah is due in part to the improved detection capability in this area.

The level of seismic activity near Lismore and offshore is low. Detection capability in this area is similar to that around Armidale in northeast New South Wales, which displays a considerably higher level of activity. The low level of activity in the southeast and northwest of the study region is not an effect of poor detection capabilities; it is considered real. These areas have a similar detection level to the area off the coast of

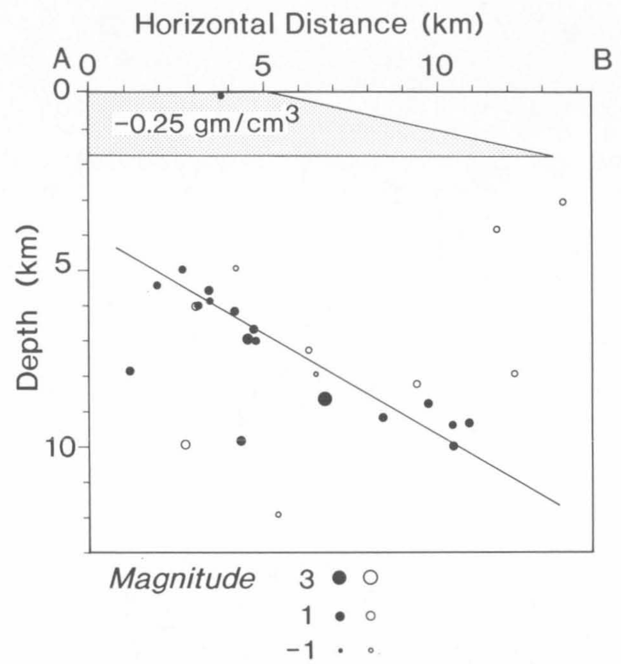

Fig. 8 Cross-section A-B of earthquakes in Figure 7. Closed circles indicate earthquakes with uncertainty in depth of less than 3 km. Earthquakes with no significant depth control are indicated by open circles. Shaded portion represents gravity fault model (from Leven, 1977).

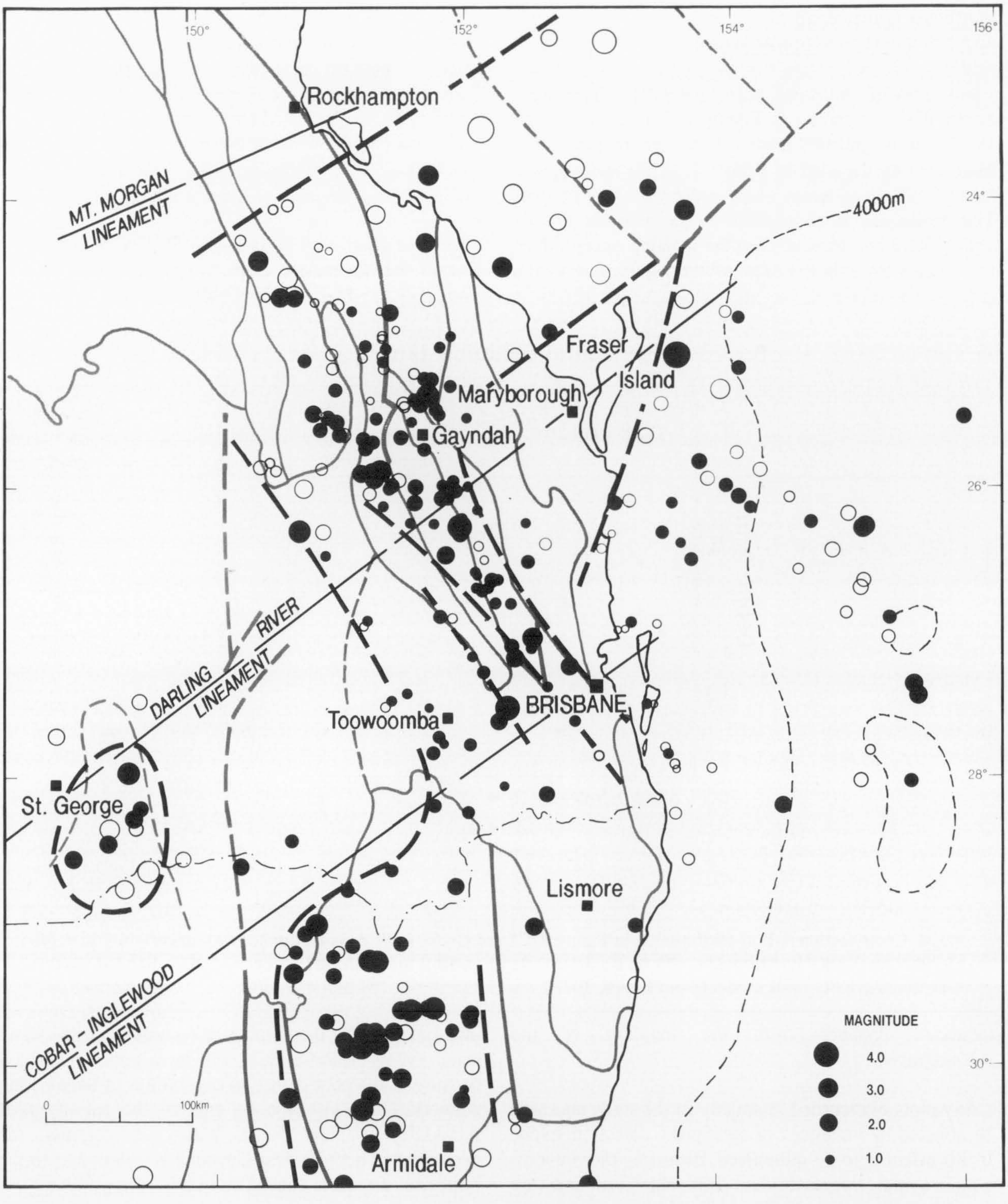

Fig. 9 Earthquakes, seismicity lineaments, tectonic lineaments, and generalised geology in study area. Closed circles and open circles represent earthquakes located to within 10 and 50 km, respectively. Small dashed line is the 4000 m water depth contour (from Division of National Mapping, 1969).

Gladstone and Bundaberg, within which several large earthquakes have been located.

Fraser Island and the adjacent onshore area, near Maryborough, are notably aseismic. The two earthquakes of 1947 and 1952 that were located near Maryborough by Jones (1948,1958) could have occurred farther east, off Fraser Island, in the area of activity found in this study. The published locations of Jones are based on distance from Brisbane (calculated from P and S-wave arrival times), and intensity reports in the Maryborough region. A location off Fraser Island is the same distance from Brisbane and would produce similar felt effects in Maryborough.

The seismicity lineaments in the Wivenhoe Dam area (Fig. 7) are also shown in the central portion of Figure 9. The belt of activity in the Wivenhoe Dam area has been extended further north-northwest to include the concentrated activity in the Gayndah area. Increasing errors in location accuracy away from the Wivenhoe Dam network prevent further extrapolation. Other seismicity lineaments reflecting the features noted above have been added to Figure 9.

Correlation between several large tectonic lineaments and observed seismicity lineaments is poor. The Darling River Lineament (Scheibner, 1974) is parallel to one side of the aseismic zone around Fraser Island. As suggested by Rynn (1987), the decrease in activity near Gladstone may be associated with the Mt Morgan Lineament (Horton, 1978). The correlations, noted by Rynn (1987), between the Darling River Lineament (inland from Maryborough), the Cobar-Inglewood Lineament, and seismicity are not verified in this study.

Tectonic units in the study area have been plotted in Figure 10. All these units are related to crustal evolution of Palaeozoic and Mesozoic age and therefore may have no relation to present day seismicity. However, because of their ancient lineage compared with the oceans, continental regions usually have zones of weakness persisting from earlier orogenic episodes, which are susceptible to reactivation (Watterson, 1975; Sykes, 1978).

Most of the seismicity is restricted to the New England Fold Belt. The area of the fold belt that is overlain by the Clarence-Moreton Basin is also active. The western limit of the seismicity in this region does not coincide with the Moonie-Goondiwindi Thrust or the Gogango-Baryulgil Fault Zone. The western limit does, however, coincide with a large-scale fault postulated by Murray & others (1987) to account for the structure of the New England Fold Belt.

The aseismic area near Maryborough corresponds largely to the Gympie Province and the Maryborough Basin, and may be related to the Gympie Terrane described by Harrington (1983). The activity near St.George correlates in part with the Wunger Ridge

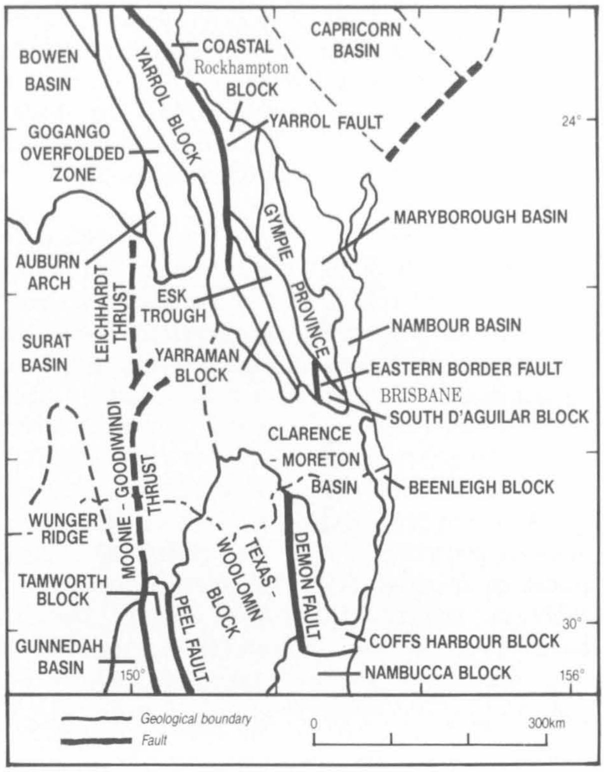

Fig. 10 Generalised geological structure and mapped faults in study area. Geology adapted from Geological Survey of Queensland (1975), Davies & others (1989), and Power & Devine (1970).

(Power & Devine, 1970). The Demon Fault appears to separate the activity in the Armidale area from the coastal area, which has considerably less activity (Rynn & Lynam, 1984).

DISCUSSION

Attempts to directly associate seismicity with geological structure are often confused by the need to correlate mapped structure at the surface with tectonic processes at depth. Mapped structures represent the surface expression of a multitude of past tectonic events, whereas observed seismicity represents the current tectonic process occurring at depth within the crust.

Inaccuracies in earthquake locations will also hinder seismicity studies. The activity depicted in Figure 8 extends for over 10 km horizontally and from depths of 5 to 10 km. Observing this seismicity without considering the focal depths of the events would lead to the conclusion that the activity is occurring in a diffuse zone.

If the plane observed in the Wivenhoe Dam area represents an active fault then the absence of hypocentres at depths less than 5 km indicates that stress is being released at shallower depths via aseismic slip. Laboratory experiments and studies of the depth distributions of earthquakes have shown that low temperatures and high stress are required for earthquake generation (Brace & Byerlee, 1970; Stesky & others, 1974). These conditions are met in the mid-crustal region at depths of the order of 10 km (Brace & Byerlee, 1970; Stesky & others, 1974). At shallow depths the stress is too low for earthquake generation, and in the deep crust the temperature is too high.

Focal mechanism studies of well-located events in the Wivenhoe area have shown the crust to be currently under horizontal compression in a northeast-southwest direction. All other stress measurements in Australia, whether from earthquake focal mechanisms, or by direct measurement in shallow boreholes, have found horizontal compression of the crust (Lambeck & others, 1984). The principal stress directions determined from these measurements vary widely within short distances, but there is a tendency for the principal stress to be aligned perpendicular to the continental margin (Lambeck & others, 1984). The direction found in this study in southeast Queensland and the northerly direction found by Bock & others (1987) at the Burdekin Falls damsite in northeast Queensland, also fit this pattern.

The results presented in this paper represent only 12 years monitoring. This is a relatively short period from which to draw firm conclusions regarding the presence or absence of seismicity in particular areas. With continued and improved monitoring, the seismicity lineaments observed will undoubtably be modified and refined. Further monitoring at a more detailed level is required to confirm the aseismic nature of the Maryborough area and to investigate the focal depth distribution in the Wivenhoe area.

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GEOLOGY OF THE EROMANGA SECTOR OF THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT

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ABSTRACT

The western third of the Eromanga-Brisbane Geoscience Transect, extending for 500 km from the Nebine Ridge to the Mt Howitt Anticline in southwestern Queensland, provides critical evidence of the history of the foreland sector of the Tasman Fold Belt. The stratigraphic sequence within or related to the transect above Lower Palaeozoic basement consists of three mega-sequences: Devonian in the Adavale Basin and Warrabin Trough; Upper Carboniferous to Late Triassic in the Cooper and Galilee Basins; and Jurassic to Cretaceous in the Eromanga Basin. Each mega-sequence is separated from the next by a regional unconformity linked to a period of widespread tectonism and erosion.

The foundations of all structures were set during the (?)Proterozoic-Early Palaeozoic, when crustal extension created a series of NE-SW trending deep crustal fractures, followed by the Thomson Orogeny. Intra-cratonic extension along the foreland margin of the Tasman Fold Belt established a lesser series of basin-forming fractures during the Early Devonian. Three phases of foreland thrusting during the remainder of the Devonian and culminating in a major event in the Middle Carboniferous firstly loaded then deformed the crust. Both the extensional phase and the subsequent compressional phases of tectonism are viewed as responses to N-S sinistral shear stress along the length of the Tasman Fold Belt.

The Early Devonian tectonic regime is compared with the Cainozoic of western Southeast Asia, where oblique slip between the Indian plate and the Southeastern Asian plate has created a rift corridor. The later Devonian-Carboniferous phases of foreland thrusting bear resemblances to local effects of the Laramide Orogeny upon the North American craton.

The Late Carboniferous - Triassic Cooper and Galilee Basins formed from downwarps to the west and to the north and east of the Adavale Basin, respectively, probably under the influence of NNW-SSE dextral strike-slip stress related to events along the palaeo-Pacific margin farther east. The rest of the western sector of the transect was generally stable throughout this period. Australia-wide tectonism towards the end of the Triassic ended subsidence of the Cooper and Galilee Basins. In the vicinity of the transect, however, the surface of the crust was close to the base level of erosion and remained stable from the end of the Permian until late in the Early Jurassic. Renewed subsidence later in the Jurassic was centred upon the Cooper Basin. Infill of the basin during the Jurassic was controlled as much by repeated uplift of the hinterland as by sea level changes and intra-cratonic sag. Fluctuating sea levels were major influences upon Cretaceous sedimentation. Neogene stresses reactivated pre-existing fractures and flexures to warp the terrane into its present form.

INTRODUCTION

The western third of the Eromanga-Brisbane Geoscience Transect, in southwest Queensland, parallels the rail and road link from Charleville to Eromanga, across the eastern flank of the Mesozoic Eromanga Basin (Figs 1, 2). The terrane is very subdued, averaging 60 m above sea level, but reflects structures in the Eromanga Basin that to a large extent mirror the geometry of underlying basins and basement ridges (Senior & others, 1978; Hoffmann, 1989b). The Mount Howitt Anticline at the western end of the transect is one such structure, between the Cooper and Wilson Rivers.

The aim of this paper is to outline the evidence within the sedimentary sequences upon which our present understanding of geology of the transect is based, and to identify the inferences and constraints this evidence places upon models of the region's evolution.

The Eromanga Basin (Mott, 1952) occupies an area of about 1 200 000 km^2 of Queensland, South Australia, New South Wales and the Northern Territory. It is part of the Great Artesian Basin, which overlies the Late Proterozoic-Early Mesozoic Tasman Fold Belt (Fig. 3).

The section of the Tasman Fold Belt underlying the Eromanga Basin is referred to as the Thomson Fold Belt (Murray, this Bulletin). The Eromanga Basin contains up to 3000 m of Jurassic and Cretaceous strata (Exon & Senior, 1976; Armstrong & Barr, 1986), distributed between two main depressions. The Poolowanna Trough in the west is separated from the Cooper Syncline or Barrolka-Jundah Depression (Hawkins & others, 1989) and other depocentres in the east by the Birdsville Track Ridge (Moore, 1986; Passmore, 1989). The southeast flank of the basin is referred to as the Thargomindah Shelf, and gently rises eastwards across the Cunnamulla Shelf towards the Nebine Ridge.

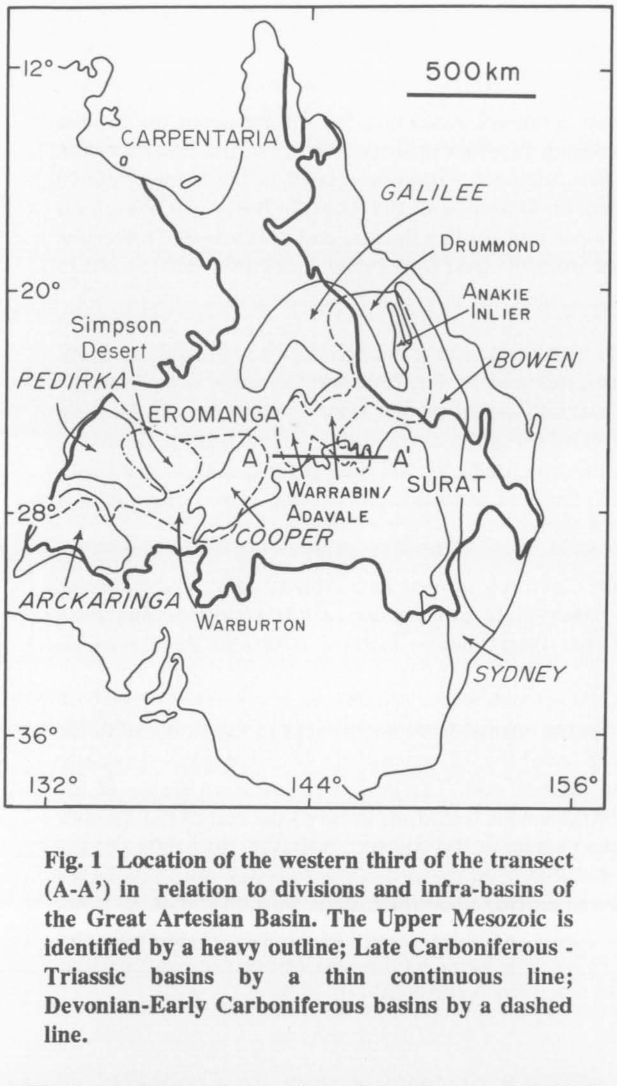


Fig. 1 Location of the western third of the transect (A-A') in relation to divisions and infra-basins of the Great Artesian Basin. The Upper Mesozoic is identified by a heavy outline; Late Carboniferous - Triassic basins by a thin continuous line; Devonian-Early Carboniferous basins by a dashed line.

The later phases of tectonism of the Tasman Fold Belt, during the Middle - Late Palaeozoic, led to the development of a series of basins which are overlain by Eromanga Basin rocks (Fig. 1). The present-day Adavale Basin and its associated troughs (Fig. 4) are structural remnants of the originally much broader Adavale Depression after it underwent Carboniferous deformation and erosion. The Canaway Ridge is a major basement high separating the Warrabin Trough in the west from the remnant Devonian sequences to the east (Fig. 4). In this paper, the term Adavale Basin is used to refer to the Devonian sequences east of the Canaway Ridge, including those in the Quilpie, Cooladdi, and Westgate Troughs. Sequences of the succeeding Late Carboniferous - Triassic Cooper Basin in southwestern Queensland and South Australia, and the Galilee Basin in central Queensland (Fig. 5), are also represented within the transect area.

The Adavale Basin occupies about 60 000 km^2 to the east of the Canaway Ridge and contains over 8000 m of Devonian sediments (Table 1) (Passmore & Sexton, 1984). The Warrabin Trough occupies about 35 000 km^2 and has up to 3000 m of Devonian section (Pinchin & Senior, 1982; Hoffmann, 1989b). It should be noted that Pinchin & Senior (1982) used the term Barcoo Trough to denote another depression to the southwest of Longreach, that is regarded by Hoffmann (1988, 1989b) as part of the Warrabin Trough. In this paper, we refer to all the residual Devonian sequences west of the Canaway Ridge as the Warrabin Trough sequences.

The Cooper Basin is completely covered by the Eromanga Basin. It is an elongate, partly faulted, complex depression in an area of 127 000 km^2, and contains over 2000 m of entirely terrestrial Late Carboniferous, Permian and Triassic strata (Battersby, 1976) (Table 2). Its limits are defined by the extent of the Permian sediments. The basin consists of several deep troughs separated by basement highs that are reflected in the overlying Mesozoic sequences (Fig. 3). South of the Naccowlah Trend (Kuang, 1985) are fault-controlled troughs with thick sequences of Lower Permian sediments. North of the trend the basin contains mainly Triassic sediments.

The Galilee Basin occupies an area of 250 000 km^2 and contains up to 2500 m of Late Carboniferous to Middle Triassic sediments (Fig. 5; Table 1) (Evans, 1980). The southern sector of the Galilee Basin extends as far as the transect, but is only represented by a veneer of Late Carboniferous and Permian sediments near BMR Line 11 (Fig. 6). Most of the Galilee Basin is concealed by the Eromanga Basin (Table 2); only its northeastern margin is exposed.

Only outcrops of the youngest Mesozoic and Cainozoic sequences occur within the immediate neighbourhood of the transect (Map 1, this Bulletin). Upper Palaeozoic and older Mesozoic rocks crop out 300 km to the north of the transect, and there are limited exposures of older Mesozoic around the Eromanga Basin's southern and western borders in South Australia and northern New South Wales. Interpretations of the geological history of the transect rely entirely upon data from deep wells, seismic traverses, and other geophysical data. The presence of ground water and commercial quantities of hydrocarbons in the Jurassic and Permian of the Eromanga and Cooper Basins, respectively, and of non-commercial reserves of gas in the Adavale Basin, has ensured widespread exploration of the region over the past thirty years (Sprigg, 1986; Finlayson & others, 1988; Wilkinson, 1988; Passmore, 1989).

The geological history of the region is described below in terms of three mega-sequences which were deposited during, (1) the Devonian-Middle Carboniferous, (2) the Late Carboniferous-Triassic, and (3) the Jurassic-Cretaceous. Widespread Cainozoic sediments are of considerable interest in that they are still accumulating within structural downwarps. In a general way, these sediments delineate the positions of concealed infrabasins, and perhaps represent the ealiest phase of a future mega-sequence. However, they are not considered in detail here.

DEVONIAN - MIDDLE CARBONIFEROUS

A number of major structural units and provinces are identified within and near the Adavale Basin and Warrabin Trough (Fig. 4). The Pleasant Creek Arch is regarded as part of the pre-Late Carboniferous "basement" to the east of the Adavale Basin, referred to here as the Westbourne Block. The Block is characterised by a series of elongate anticlines bounded by reverse or thrust faults that extend NNE-SSW. Some N- and NNW-trending faults also intersect the area, but there is insufficient seismic coverage to determine the fault pattern in detail. Strata of presumed Devonian or Early Carboniferous age are preserved between the anticlines. No wells yielding age-indicative fossils have intersected any of these sections. The Westbourne Block is defined as the region extending eastwards to the Nebine Ridge (Fig. 4), or possibly the Foyleview Geosuture. Its northern boundary is placed at the change of structural style from NE-trending horsts and graben to the NW-trending *en echelon* anticlines at the southern end of the Drummond Basin. The southern boundary of the Westbourne Block is within the transect and is marked by the Langlo and Wanka Embayments and the Westgate Trough that open into the Cooladdi Trough. At least 2000 m of Devonian sediments are preserved in these synforms (Pearce, 1980).

Southwest of the Westbourne Block and Cooladdi Trough are the Cheepie Shelf and Quilpie Trough. The Quilpie Trough comprises presumed Devonian sediments preserved in a post-depositional asymmetric synform, closely comparable in shape and dimensions to the Westgate Trough. To the west of the Pleasant Creek Arch is a series of uplifted crustal blocks, including the Gumbardo Anticline, the Cothalow Arch, the triangular-shaped Gilmore structure, and the Carlow and Bonnie Anticlines. The margins of these uplifts are defined by reverse or thrust faults at depth (Fig. 6). Between the Cothalow Arch and the Canaway Ridge is the relatively stable Powell Block that merges to the northwest with the Yaraka Shelf.

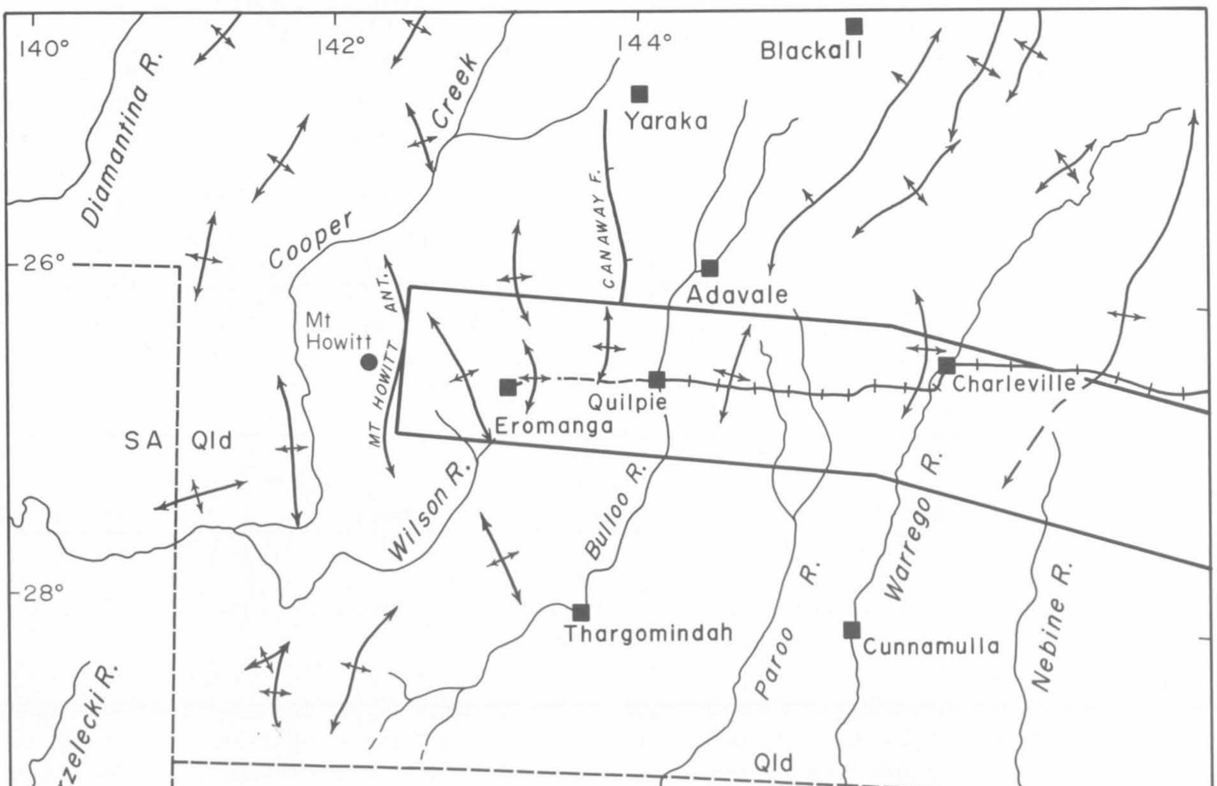


Fig. 2 Relation of the western third of the transect (boxed) to surface structures, topography and principal settlements (after Senior & others, 1978).

In contrast to the NNE-SSW trend of structures across the Westbourne Block and the dominant NE-SW trend of structures within the Adavale Basin, the Canaway Ridge and associated Canaway Fault complex strike N-S (Pinchin & Anfiloff, 1986) and are traceable over at least 150 km. The faults are relatively straight and steep to reverse (Figs 4, 7). The overall thickness of the Devonian in the Adavale Basin increases from the Yaraka Shelf and Powell Block in the northwest towards the southeast (Slanis & Netzel, 1967; Paten 1977; Passmore & Sexton, 1984; Finlayson & others, 1988). a net downthrow to the south and a long history of movement.

Structures within the southern Warrabin Trough (Fig. 4) trend NNW-SSE, curving at their southern ends to WNW-ESE (Map 1, this Bulletin). The NNW-SSE trend includes faults which had been down-thrown to the east during deposition of the Early Devonian sequence. The WNW-ESE trending faults acted as reverse faults during the Carboniferous (Pinchin & Senior, 1982).

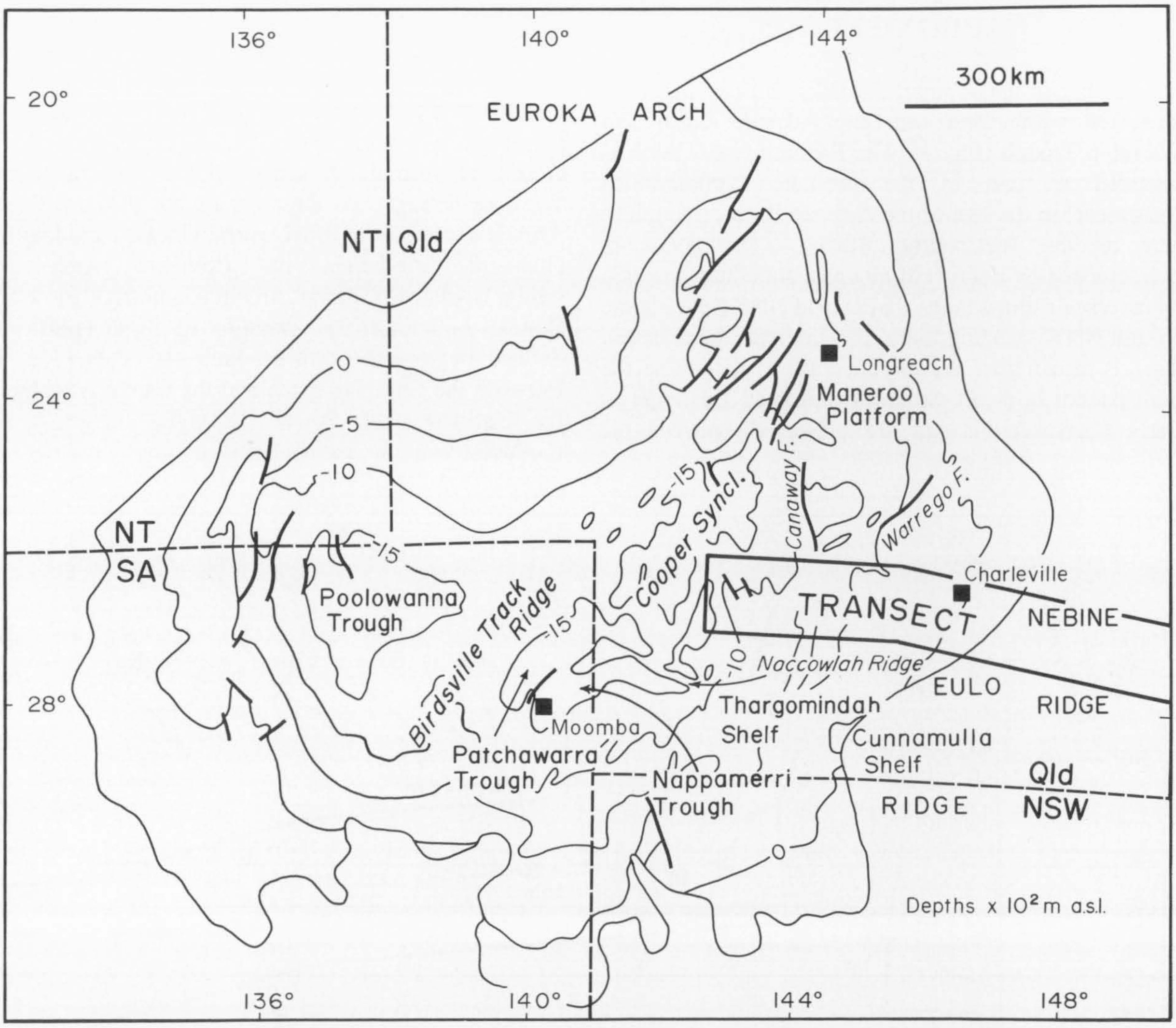


Fig. 3 Structure of the Eromanga Basin at the level of the Cadna-owie Formation, at the junction between Cretaceous non-marine and marine facies (after Moore, 1986, Figs 3 & 4).

The Early Devonian sequence onlaps the Canaway Ridge from both the Adavale Basin and Warrabin Trough, implying the ridge was a central basement high at that time (Hoffman, 1989b).

The Blackall Ridge (Fig. 4) is a complex basement uplift that trends to the northwest along the northern flank of the Adavale Basin between the Pleasant Creek Arch in the east and the Tara Fault in the west (Robson, 1986). The ridge marks a major step in basement with

The Devonian in the Adavale Basin and Warrabin Trough (Table 1) may be subdivided into four sequences that reflect phases of evolution of the Adavale Depression. The first is the product of an initial stage of rifting and accompanying volcanism. The other three phases are interpreted as the result of episodic loading of the crust to the east of the Adavale Depression that converted the depression into a foreland basin.

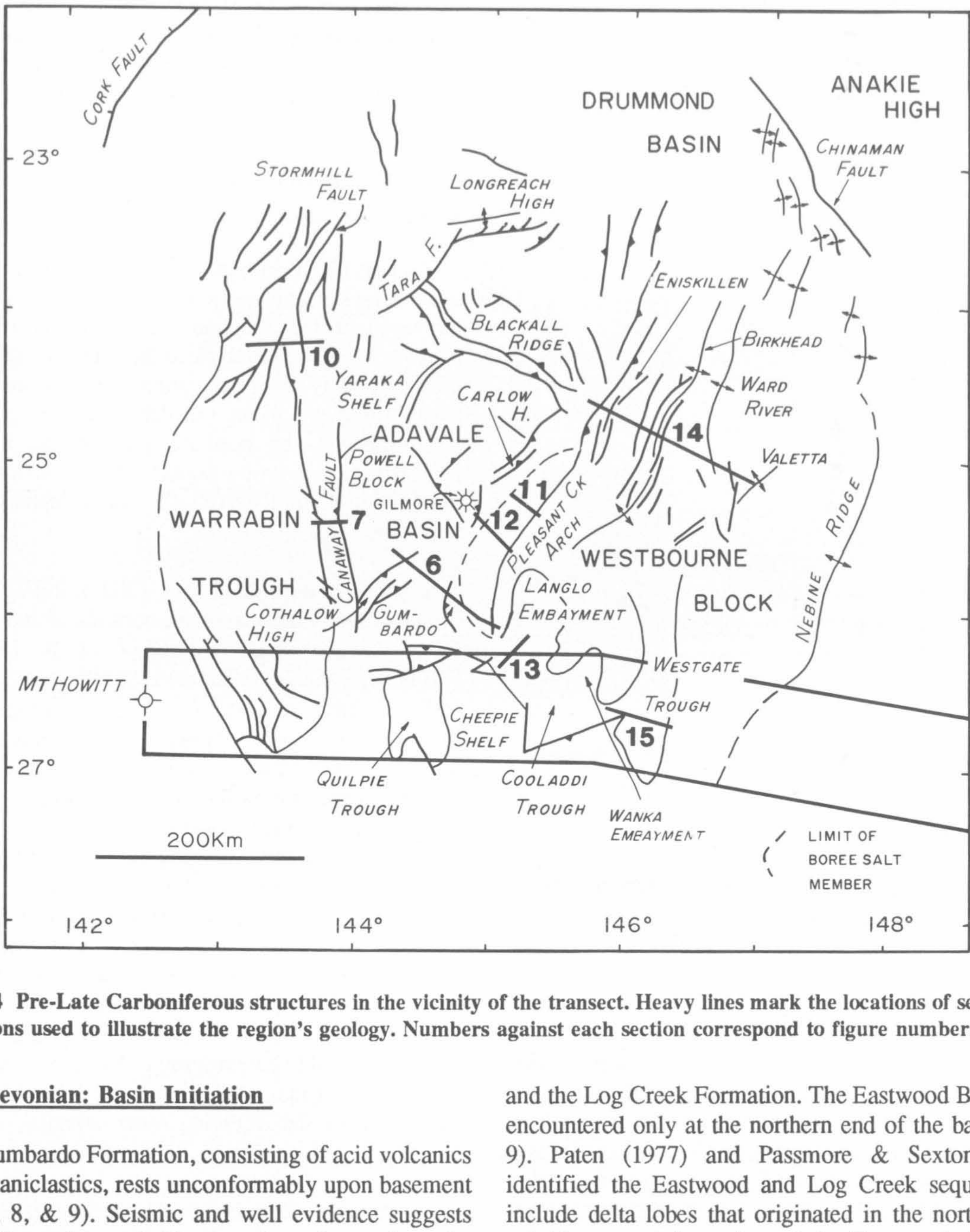


Fig. 4 Pre-Late Carboniferous structures in the vicinity of the transect. Heavy lines mark the locations of seismic sections used to illustrate the region's geology. Numbers against each section correspond to figure numbers.

Early Devonian: Basin Initiation

The Gumbardo Formation, consisting of acid volcanics and volcaniclastics, rests unconformably upon basement (Figs. 6, 8, & 9). Seismic and well evidence suggests that the formation probably extends as far north as the flank of the Blackall Ridge (Robson, 1986) and across the northwest flank of the Westbourne Block (Fig. 14). The presence of volcanics in the Warrabin Trough is not proven. The Gumbardo Formation is below drilled depths in the east of the basin (Fig. 6) (Auchincloss, 1976).

The deepest, readily identifiable seismic marker in the sedimentary succession in the Adavale Basin and Warrabin Trough is created by the Cooladdi Dolomite in the north and west and by the top of the Bury Limestone in the east (Fig. 8). The thickness and lithofacies of the section between the carbonates and basement varies greatly (Figs 6-9), due, at least in the early stages, to rift development.

Paten (1977) divided the sequence underlying the carbonates between the Early Devonian Eastwood Beds and the Log Creek Formation. The Eastwood Beds were encountered only at the northern end of the basin (Fig. 9). Paten (1977) and Passmore & Sexton (1984) identified the Eastwood and Log Creek sequences to include delta lobes that originated in the north. Initial subsidence and formation of the Eastwood delta was confined to the northeast of the basin. The Log Creek delta facies is more widespread, merging in the south of the basin with shaley formations and minor carbonates that are regarded as the marine shelf and slope facies of the period (Figs 8, 9).

Seismic data indicate the Early-Middle Devonian rocks are restricted to half-graben and graben in the central and northern parts of the Adavale Basin, as well as the margins of the Warrabin Trough (Hoffmann, 1988, 1989b). In contrast, deposition in the southern part of the Adavale Basin and central Warrabin Trough occurred in broad depressions in which substantial thicknesses of sediment accumulated. In the Adavale Basin, the fault blocks are rotated from the east, northeast and southeast towards the Canaway Ridge, whereas the blocks within the Warrabin Trough show rotation towards the trough axis suggesting a rift corridor. The amount of

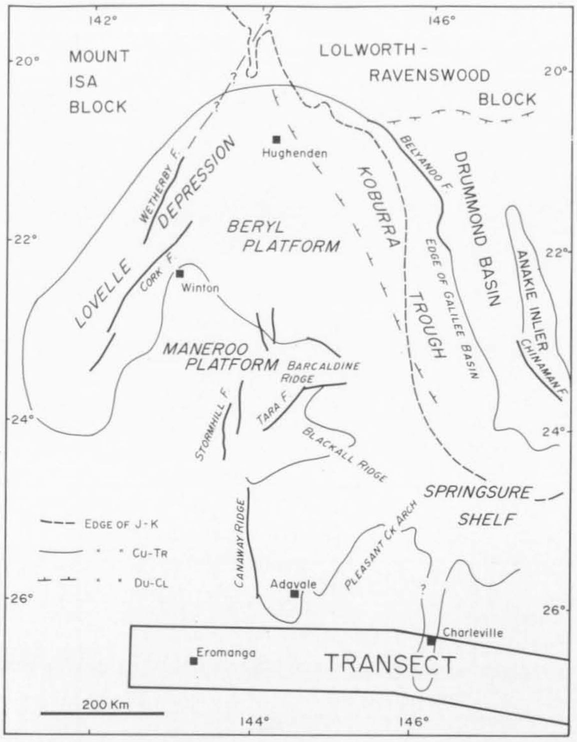


Fig. 5 Subdivisions and major structural features of the Late Carboniferous - Triassic Galilee Basin and adjacent crustal units to the north of the transect.

displacement on faults in the Warrabin Trough was less than movements observed within the Adavale Basin.

If these rifts involved simple extension, one of the two orthogonal fault trends would have been a set of normal faults, and the other a series of transfer or strike-slip faults (cf. Gibbs, 1984; Lowell, 1985). Determination of the original function of each set of faults is not conclusive. Hoffmann (1989a) postulated that the extensional terrane mapped in the Adavale Basin would require a detachment fault dipping to the west to accommodate the sense of movement on the extensional blocks.

Discounting the effects of post-Devonian movements, net regional extension during the Early Devonian appears to have been to the northeast. This is particularly evident in the Warrabin and Cooladdi Troughs and in the vicinity of the Blackall Ridge. Trends, such as the Cothalow-Carlow Arches the Tara, Stormhill and Cork Faults, and the Pleasant Creek Arch, would have acted as transfer or strike-slip faults. This suggestion depends on (1) identification of onlap onto basement by sediments infilling half graben, and (2) discounting such structures as ones accompanying the Stormhill Fault zone across the northern Warrabin Trough (Fig. 10), which appear to be largely post-depositional structures. Alternatively, if the NE-SW trending faults, such as those which cross the Westbourne Block, were initiated as normal faults as part of the extensional terrane and not transfer faults, net extension would have been to the southeast. The westward-dipping Westgate and Foyleview deep crustal ramps evident within the transect area farther east may have provided the detachment surfaces (Hoffmann, 1989a).

Early Middle Devonian: Initiation of the Foreland Basin

The Log Creek Formation is an extensive, generally coarsening upwards sequence of siliciclastic deposits in the north of the Adavale Basin and carbonate-shale facies at least 1000 m thick to the southeast (Fig. 8) that is succeeded by the cleaner, shallow-marine and possibly partly alluvial facies of the Lissoy Sandstone. Subsidence in the northern part of the basin was confined once more by faults of the Cothalow-Carlow trend (Fig. 9), but overall the basin deepened to the southeast.

An undrilled section, at least 1000 m thick, below the Cooladdi Dolomite marker in portions of the Warrabin Trough is a probable correlative of the Log Creek Formation (Pinchin & Senior, 1982; Pinchin & Anfiloff, 1986; Hoffmann, 1988, 1989b). The trough in Log Creek time was a broad, partly fault-controlled depression, such that the Log Creek Formation thinned by onlap towards the Canaway Ridge (Fig. 7). The lack of well intersections precludes comment on the extent of these Devonian strata in the north of the Warrabin Trough. Post-depositional half-graben adjacent to the Stormhill Fault trend at the northern end of the trough (Figs 4, 10) contain over 3000 m of Devonian section, a portion of which may be of Early Devonian age. Hoffmann (1988) proposed that these features developed as half-graben and were later uplifted and truncated during the mid-Carboniferous. Passmore & Sexton (1984) suggest that, at this time, the Warrabin Trough was emergent and receiving some terrestial sediments, possibly fluvial clastics.

The Adavale Depression during this phase of its evolution is interpreted as having assumed the characteristics of a foreland basin, deepening to the southeast, in response to crustal loading farther to the southeast (Fig. 16). Little sediment from the upthrusting zone entered the preserved part of the original depression: sediment was supplied from the craton. Either that part of the basin to receive sediment from the high to the east is no longer preserved, having been uplifted and eroded by later events, or the initial overthrust segment had insufficient elevation to contribute sediment to the foreland basin.

Middle Devonian: Tectonic Quiescence

The supply of clastics from the north and west diminished during the Middle Devonian. The Log Creek delta/shelf provided a very gently sloping substrate for accumulation of the Bury Limestone. Both the Canaway Ridge and Warrabin Trough were virtually level ground,

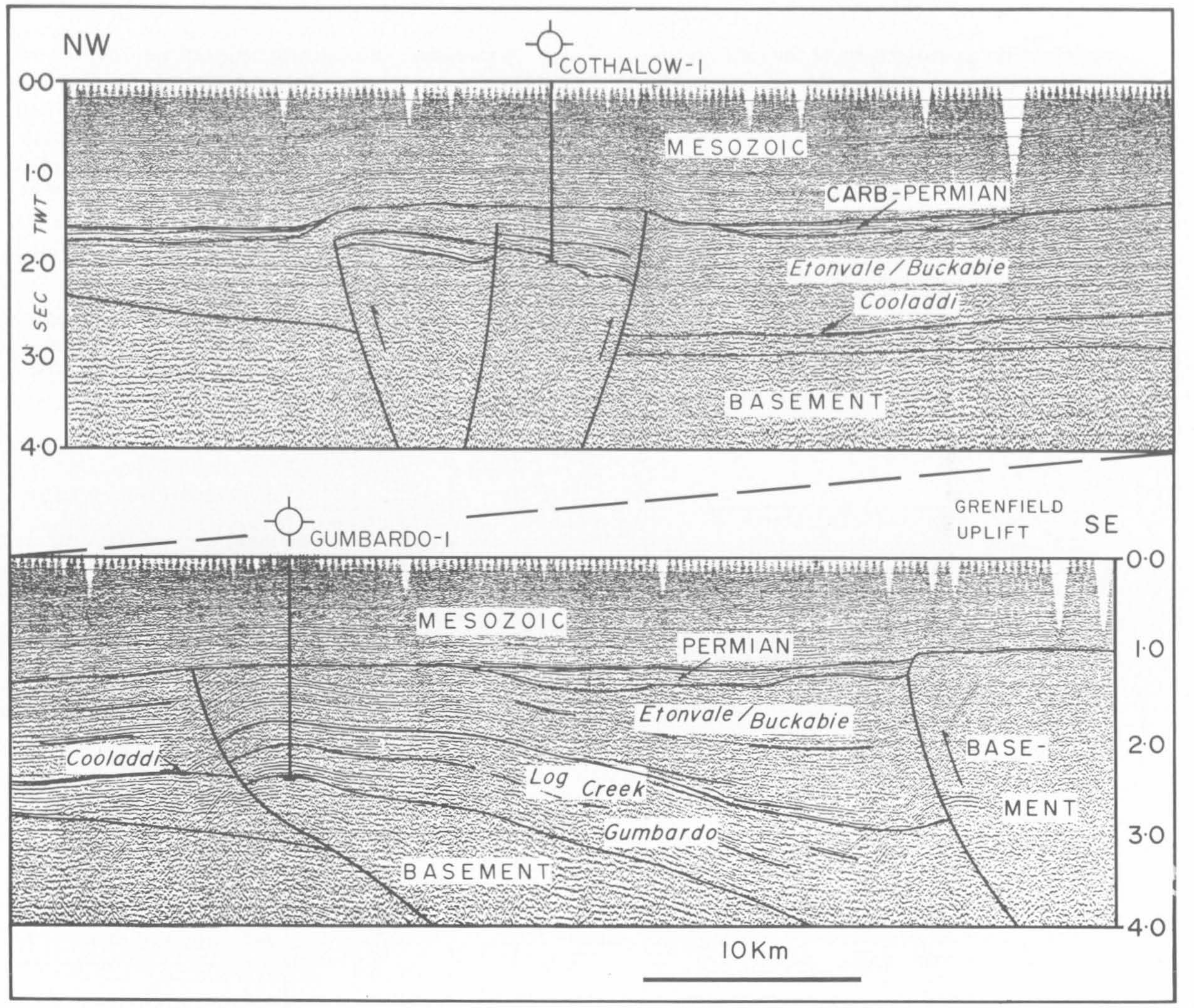


Fig. 6 Portion of BMR seismic section 11 (Finlayson & others, 1987) from the Grenfield Uplift, across the Gumbardo and Cothalow Anticlines. The vertical to horizontal scale ratio at the level of the Cooladdi Dolomite is about 2:1. The thrust front of the Grenfield Uplift ramps to the surface at about 45°. The Gumbardo Formation thickens to the SE. The Log Creek Formation also thickens to the SE, but to a lesser extent. The Gumbardo thrust is a fault that had little significance during deposition of the Lower Devonian Gumbardo/Log Creek section. Note the onlap of the Etonvale Formation to the west upon the Cooladdi Dolomite/Bury Limestone. The Cothalow structure was elevated principally during the Carboniferous and again during the Tertiary. The unconformity between the Devonian and the Lower Permian exhibits a distinctly uneven topography.

a vast sabkha in which carbonates and evaporites of the Cooladdi Dolomite accumulated.

Paten (1977) identified: (a) to the south and southeast, the limestone/shale facies of the Bury Limestone that were deposited below an open sea; (b) to the north, the limestone facies of the same formation where there was optimum carbonate deposition and reef development (Fig. 11); and (c) westward, the Cooladdi Dolomite that was formed in the landward zone of evaporitic lagoons and restricted water circulation. The disposition of these facies indicates open sea towards the east and southeast. This was the period of maximum marine transgression in the Adavale Basin. Interbedded arkosic and silicic clastics in the Bury Limestone (limestone facies) were possibly sourced from a rising thrust front to the east (discussed later in this paper).

The general lack of clastics from the west, the formation of such a gently sloping carbonate ramp, and the tectonically neutral role of the Canaway Ridge during this interval of the Middle Devonian indicate a period of overall tectonic quiescence.

Middle-Late Devonian: Second Phase of Foreland Thrusting

The Etonvale Formation succeeding the Bury Limestone and Cooladdi Dolomite (Fig. 9) consists of three lithofacies - dominant sandstone, shale, and salt (Galloway, 1970). The Boree Salt Member is confined to the east of the Adavale Basin, upon the Bury Limestone. It is thickest adjacent to the Pleasant Creek Arch, partly because of halokinesis. In places, the salt onlaps a flexed, uneroded surface of the Bury Limestone

TABLE 1

Generalised time stratigraphy of the Adavale and Galilee Basins. Palynostratigraphic divisions after Price & others (1985).

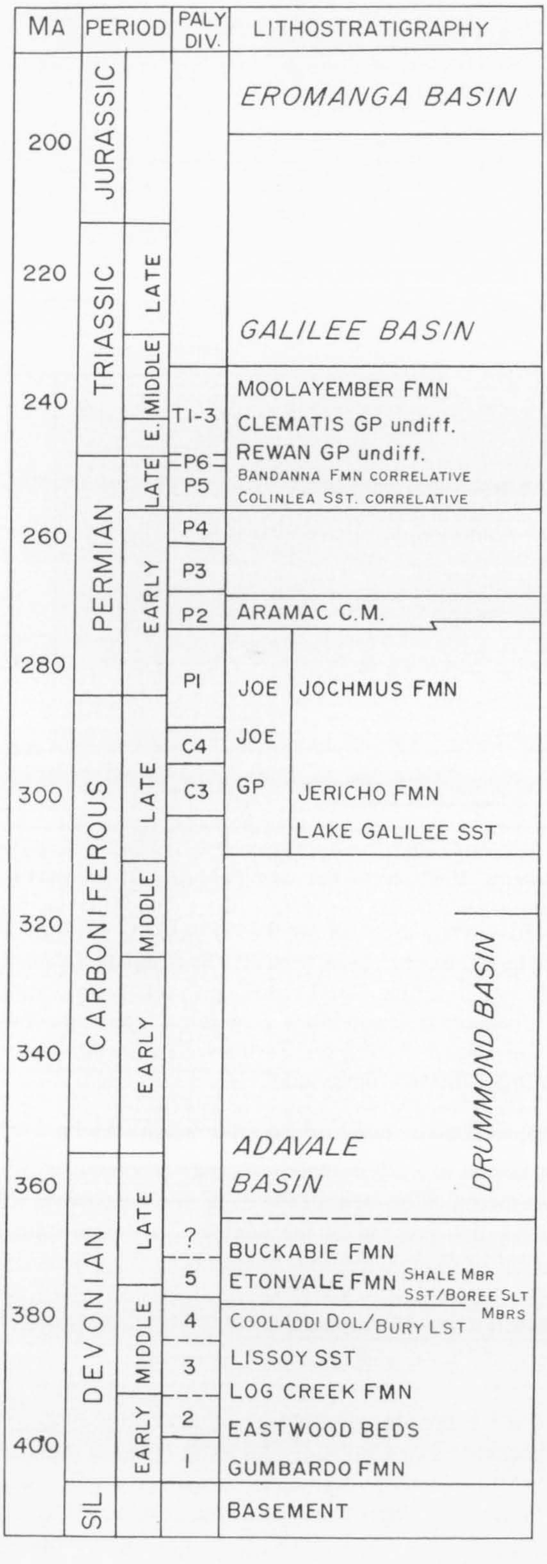
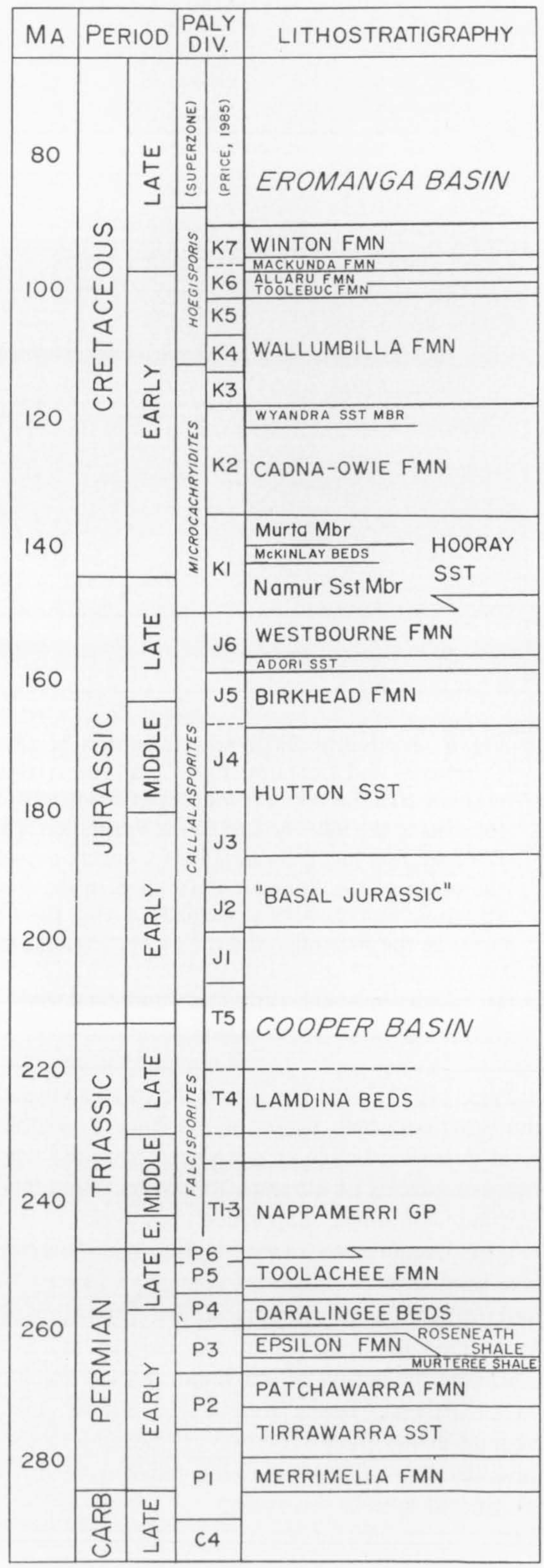


TABLE 2

Generalised time stratigraphy of the Cooper and Eromanga Basins. Palynostratigraphic divisions after Helby & others, 1987 (the superzones) and Price & others (1985). Other palynostratigraphic schemes have been devised, using Linnean binomial nomenclature, but the alphanumeric system used by Price & others is adopted mainly because of its simplicity and in order to show the extent to which subdivision is possible.



(Fig. 11). Tectonism had cut off the shelf from open marine circulation and deformed the substrate in water deep enough for the uplifted limestone not to be affected by subaerial or inter-tidal erosion. Deformation at this time was widespread, warping the sea floor, causing faulting of the Bury Limestone (Catuhe & Laws 1979), and also local uplift and erosion from the Cothalow Arch in the Adavale Basin (Passmore & Sexton, 1984; Hoffmann, 1989a).

The Etonvale Sandstone Member, a partially lateral correlative of the Boree Salt Member, onlaps the Cooladdi Dolomite to the west (Figs. 6 & 8). Dominantly continental red beds were deposited in the Warrabin Trough, while east of the Canaway Ridge, a fluctuating shoreline controlled more marine deposits. Restricted circulation returned periodically, as evidenced by horizons in the Etonvale Formation containing gypsum. The shale member of the Etonvale Formation was deposited during a transgressive phase that returned

regression and exclusion of the sea. Sediments probably entered the depression from both the rising ground to the east, as well as from the interior of the craton to the west.

The age of the Buckabie Formation is uncertain because of its lack of diagnostic fossils, but has been variously regarded as Late Devonian and possibly Early Carboniferous. The formation was regarded by Vine (1972) as no younger than Frasnian, because of the contrasting sequence in the Drummond Basin to the northeast of the Adavale Basin, where Late Devonian (Famennian) strata of limited distribution are the oldest of the basin fill, are volcanigenic, and are succeeded by up to 10 000 m of Lower Carboniferous, far in excess of the maximum thickness of the Buckabie Formation. It seems likely that deposition ceased in most parts of the basin at the end of the Devonian.

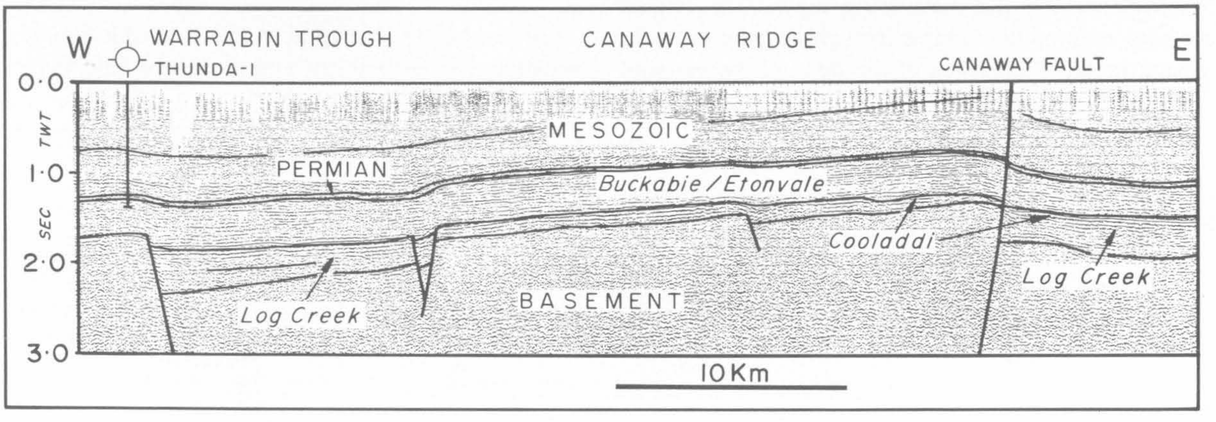


Fig. 7 Portion of BMR seismic section 8 (Finlayson & others, 1987) across the Canaway Ridge and eastern Warrabin Trough. The (?)Early Devonian Log Creek Formation in the Warrabin Trough thins towards the western side of the ridge. The Etonvale and Buckabie Formations extend across the ridge with little change in thickness. The Permian comprises a thin layer of the upper Gidgealpa Group (Toolachee Formation). The Canaway Fault was active during the (?)Early Devonian and reactivated mainly in the Tertiary.

normal marine conditions to much of the Adavale Basin.

These events are identified as responses to a second stage of foreland thrusting. The terrane rising to the east further loaded the crust, thereby deepening the basin, but simultaneously restricting access to the ocean, so that the Boree Salt Member was deposited in the deeper part of the basin. The open marine facies that were briefly re-established in the Adavale Basin near the top of the Etonvale Formation marked the final transgressive phase prior to final regression of the sea and deposition of non-marine shales and sandstones of the Buckabie Formation.

The Buckabie Formation of interbedded red, fine to medium-grained sandstone and shale was deposited in a fluvio-lacustrine environment and is the most widespread of the Devonian formations, extending from the Warrabin Trough, across the Canaway Ridge and throughout the Adavale Basin. It represents rapid

Late Devonian-Early Carboniferous: Migration of the Depocentre

A depocentre developed to the northeast in the Drummond Basin during the Late Devonian (Famennian), the time of major changes at the plate margin (Henderson, 1987). Pinchin (1978) showed that there was some fault control on deposition in the Drummond Basin. The NW-trending Chinaman Fault, ultimately an oblique slip thrust fault, appears to have had a significant role as a normal fault during deposition. The extent of this phase of faulting cannot be precisely determined.

Post-depositional faulting and collapse of the northern sector of the Warrabin Trough (Fig. 10) might also be related to this stage of tectonism, but the actual time of these Palaeozoic or Early Mesozoic movements has not yet been fully determined (Hoffmann, 1988).

(?)Carboniferous Deformation

Prior to the onset of deposition of the second mega-sequence during the Late Carboniferous, the entire Adavale Depression was deformed and deeply eroded by the third stage of foreland thrusting. In excess of 3000 m of sediments were eroded in places (Passmore & Sexton, 1984; Hoffmann, 1988). This compressive movement created all the major structures now present in the region, and has been referred to as the Quilpie Orogeny (Finlayson, this Bulletin). Movements were so severe that subsequent reactivation has done little to alter the essential characteristics of structures formed at that time.

The effect of this compressive movement was to change a number of existing normal faults to reverse or thrust faults and to create new ones. In the line of the Eromanga-Brisbane Transect, thrust movements created the remnant synforms of the Quilpie, Cooladdi and Westgate Troughs and the Langlo and Wanka Embayments and the intervening highs. The faulted northern edge of the Quilpie Trough appears to be an arcuate southern extension of the trend of the Pleasant Creek Arch.

The southern half of the Pleasant Creek Arch is a blind thrust duplex, where older formations have been thrust into the Boree Salt Member (Fig. 12) (Remus & Tindale, 1988). The southern end of the thrust curves into the northern margin of the Cooladdi Trough, paralleling the limits of the Boree Salt Member, but no longer acting as a blind thrust. It is presumably linked by a strike-slip or tear fault to the Grenfield upthrust, which ramps to the Carboniferous surface (Fig. 6). At the northern end of the Pleasant Creek Arch, beyond the limits of salt deposition, the up-thrust front is steep and had also ramped to the Carboniferous surface (Fig. 14). A dextral tear fault intersecting the front to the north of seismic line AAP 302 (Fig. 11), accommodates this change in structural style (Remus & Tindale, 1988).

To the northeast of the Langlo and Wanka Embayments, the elongate Birkhead and Ward River Anticlines within the Mesozoic cover reflect more accentuated highs and faulted troughs at depth that formed during the Middle Palaeozoic (Fig. 14). A number of the original (probably normal) faults were reactivated as reversed faults during these compressive movements.

The northeastern margin of the Cooladdi Trough is underlain by sheets thrust into the Boree Salt Member, whereas its southern margin shows a different seismic character and is underlain by faulted basement (Fig. 13), the Pingine Fault (Finlayson & others, 1988).

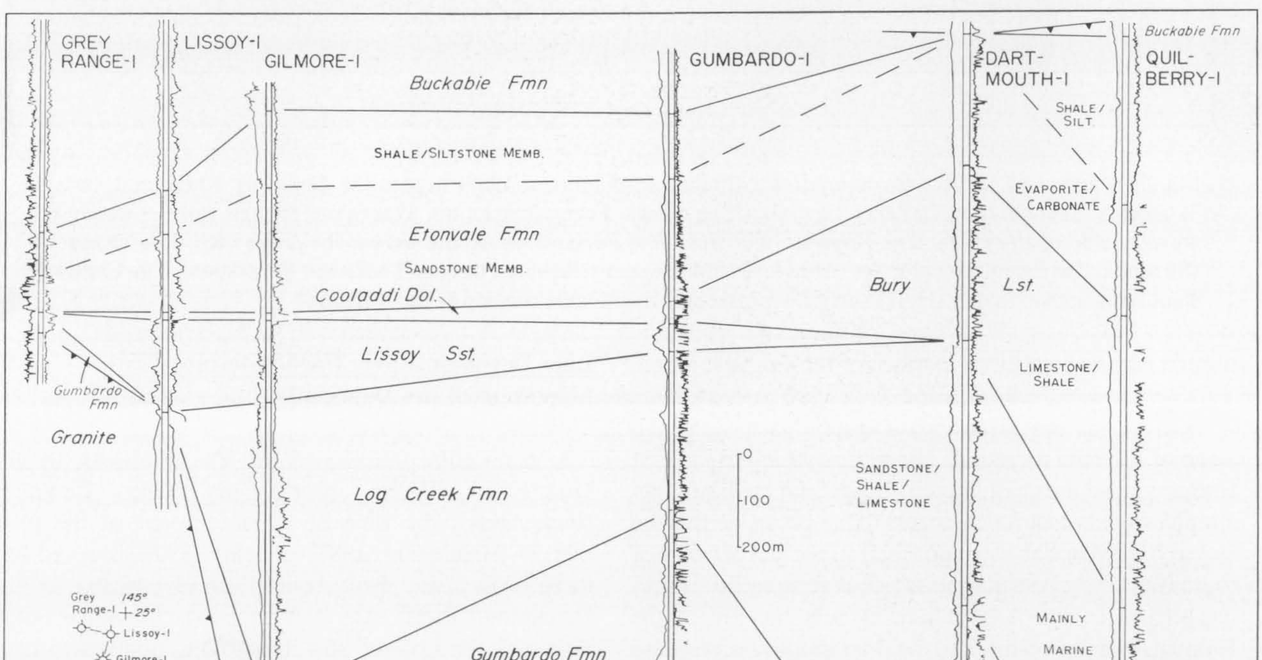


Fig. 8 Cross-section from Grey Range-1 in the margin of the Powell Block to Quilberry-1 at the junction of the Langlo Embayment and Cooladdi Trough, demonstrating variations in thickness and facies of the Early and Middle Devonian, based on SP and SN resistivity logs (after Price, 1980).

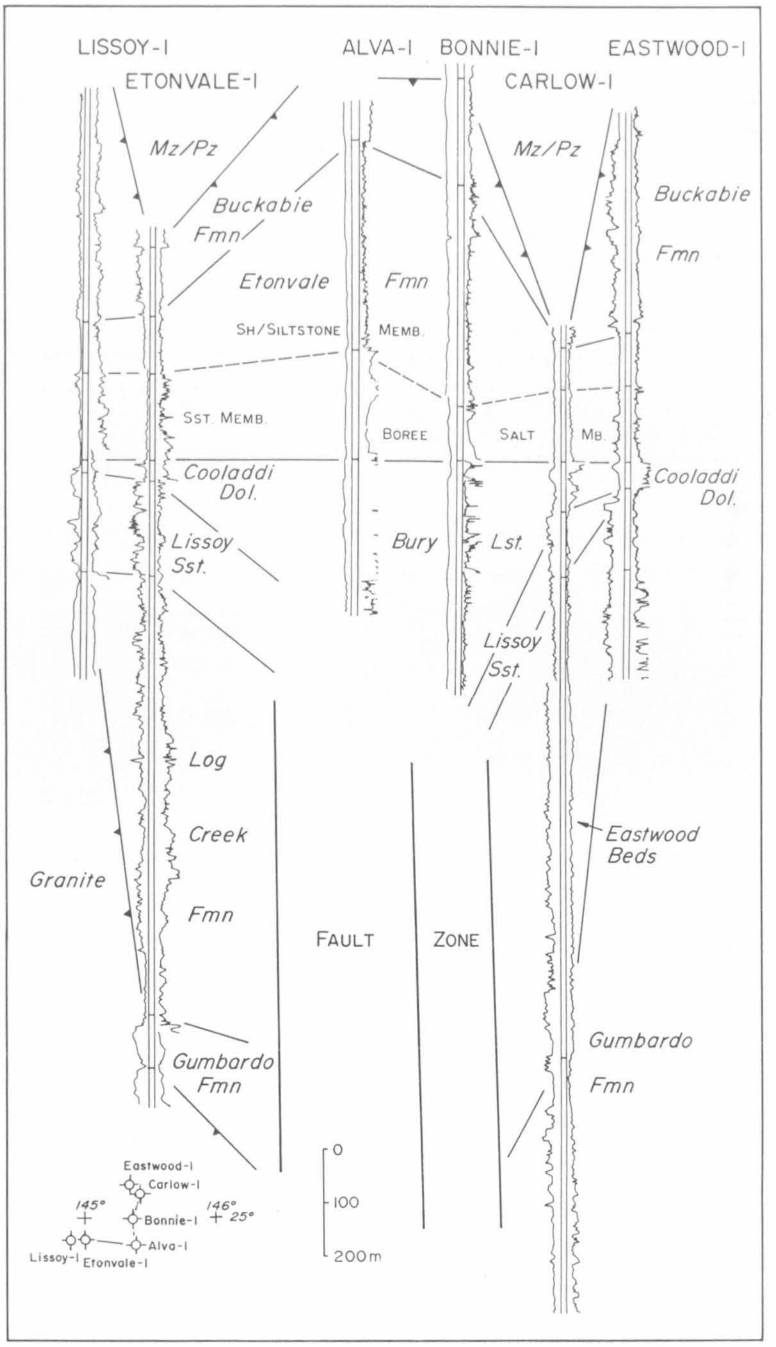


Fig. 9 Cross-section from Lissoy-1 (tie to Fig. 8) to Eastwood-1 via Alva-1, demonstrating known relationships between the Eastwood Beds and the Log Creek Formation, and facies variations in the Middle Devonian.

The Canaway Ridge, partly a horst (Fig. 7), was also uplifted and faulted during these movements, although reactivation of this complex continued sporadically until the late Middle Tertiary (Pinchin & Anfiloff, 1986; Hoffmann, 1989b). Basement at the southern end of the Canaway Fault includes a series of *en echelon* reverse faults stepping to the right, and upthrown to the south (Urschel & Griffiths, 1986). At the same time, most of the faults associated with the Blackall Ridge became reverse faults and continued to move periodically until after the Cretaceous.

The time of these movements is widely accepted as mid-Carboniferous, because they deformed Lower Carboniferous sediments in the southern Drummond Basin, south of the Chinaman Fault (Fig. 4), prior to deposition of the Upper Carboniferous Joe Joe Group (Table 1) (Olgers, 1973).

LATE CARBONIFEROUS - TRIASSIC

The second mega-sequence incorporates sediments of Late Carboniferous, Permian and Triassic age, which occupy the Cooper Basin to the west of the transect, and the Galilee Basin to the north (Tables 1, 2).

There are both similarities and differences between the sedimentary fill and structural histories of the two basins. Both contain a veneer of Late Permian sediments. The Galilee Basin has mainly Late Carboniferous and Early Permian formations; Triassic formations are largely confined to its northern sector. The Cooper Basin initially developed principally as a complex of rifts during the Early Permian and sagged again in the Triassic, although the Triassic is best developed in the northern half of the basin (Battersby, 1976). The sedimentary sequences in both basins are of terrestrial origin, with the exception of a tongue of Late Permian shallow marine facies in the east of the Galilee Basin (Evans, 1980). The contents of both basins are considered below in terms of three sequences, the distribution of which reflects the region's tectonic history.

Late Carboniferous - Early Permian

The Joe Joe Group in the Galilee Basin is an extensive sequence, some of which is glaciogenic, or of volcanic origin. The unconformity below the Joe Joe Group, particularly upon the Westbourne Block, is irregular, perhaps reflecting a glacially scoured surface (Figs 11, 14). In the Koburra Trough (Fig. 5), the Joe Joe Group is over 1500 m thick. Most of it is regarded as Late Carboniferous in age; however, the top of the group merges with the Aramac Coal Measures and is placed in the basal Permian (Kemp & others, 1977).

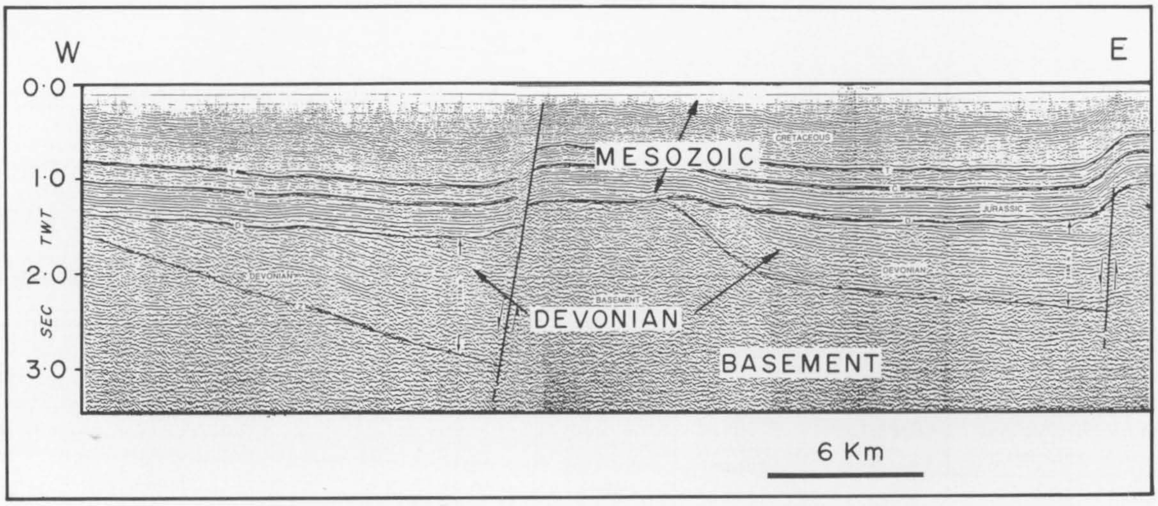


Fig. 10 Seismic section Esso Y81B-1225 across the Stormhill Fault complex, demonstrating the thick (?)Devonian sequences in half graben below the Mesozoic. Note there is no obvious onlap of the (?)Devonian onto basement and hence the half graben are likely to have been created by post-(?)Devonian rather than by syn-depositional movements (from Hoffmann, 1988). This interpretation emphasises possible Devonian structural style. Cainozoic movements have subsequently remobilised structures as reverse faults.

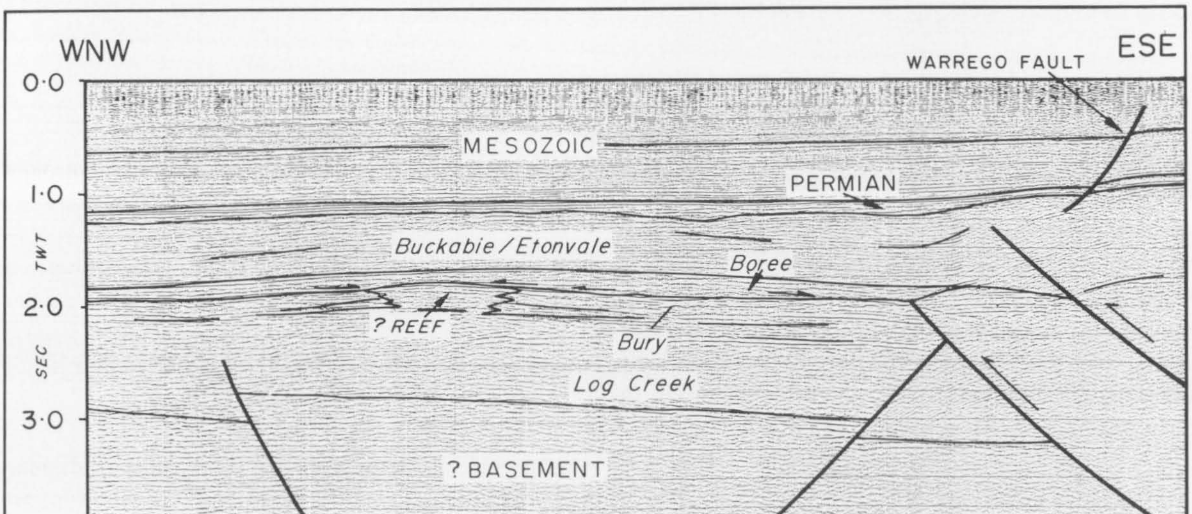


Fig. 11 Seismic section AAP 302 from the edge of the Pleasant Creek Arch to the north of Alva-1 across a possible reef in the Bury Limestone and a pillow in the Boree Salt Member. Note onlap of the reef crest by the salt. The thrust front of the Pleasant Creek Arch at this location was both a blind thrust into the Boree Salt Member and also a ramp to the Carboniferous surface (cf. Fig. 14). Vertical: horizontal scale ratio at the depth of the Bury

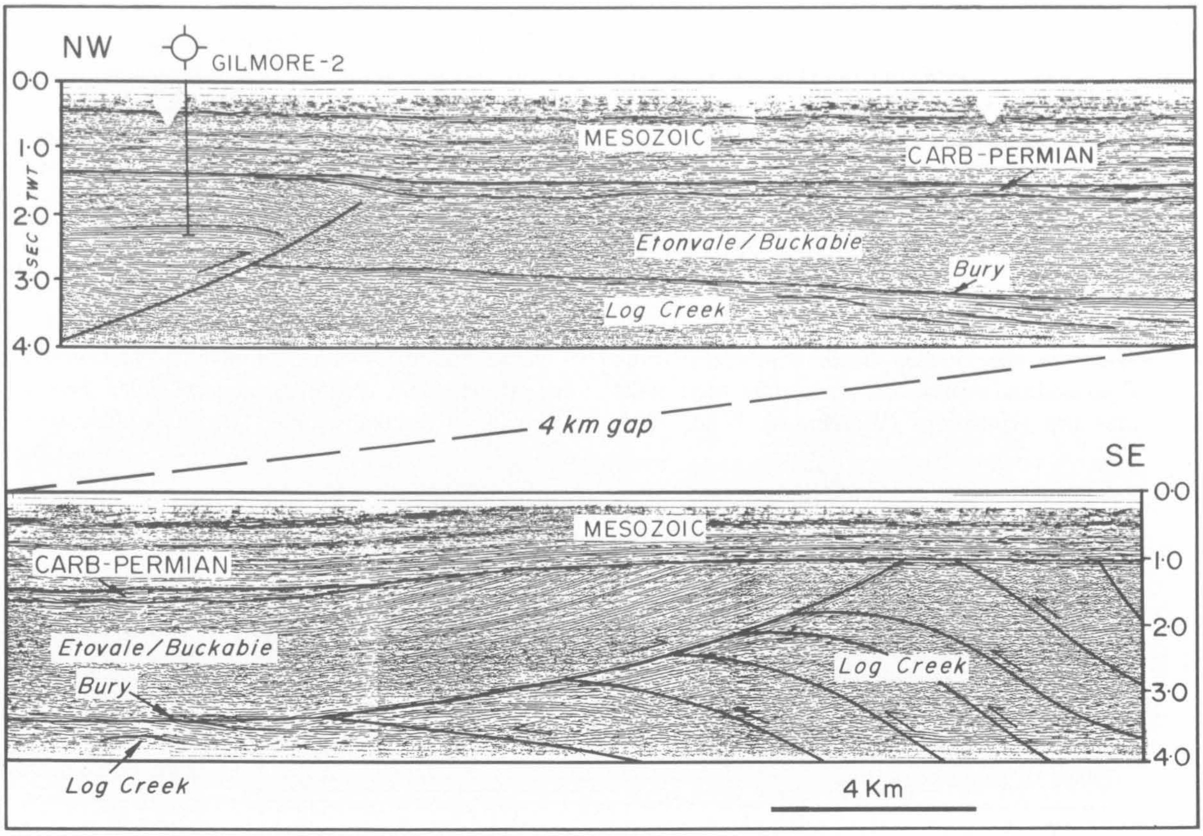


Fig. 12 Portion of seismic section AAP S484S-18 from the southern sector of the Pleasant Creek Arch to the Gilmore structure at a vertical to horizontal scale ratio of about 1:1. Thrusts below the western flank of the Pleasant Creek Arch constitute a blind thrust into the sedimentary column at the level of the Boree Salt Member. Note prograding of sediments below the Bury Limestone indicative of construction of the Log Creek delta.

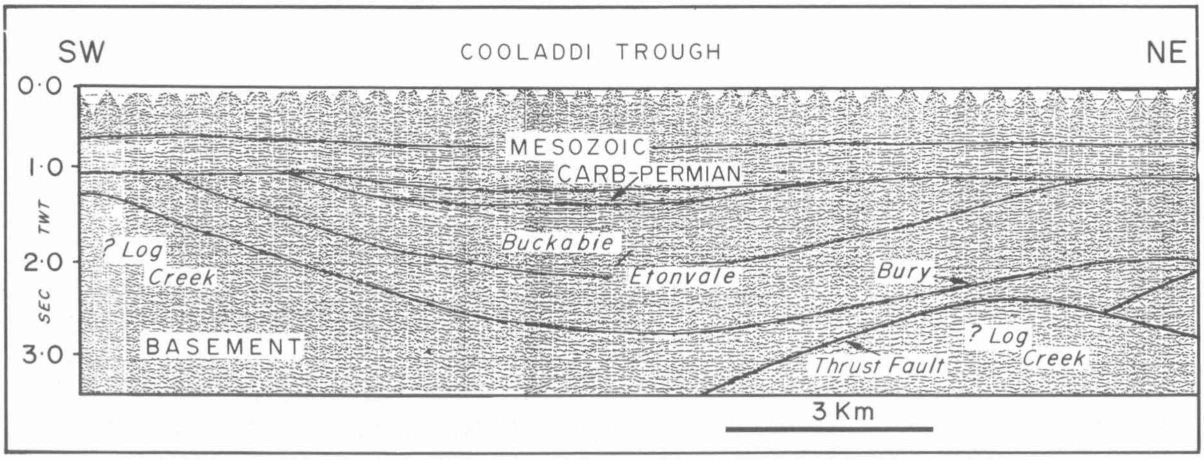


Fig. 13 Portion of Phillips seismic Line 164 (reprocessed by Hartogen) across the Cooladdi Trough at an approximate horizontal to vertical scale ratio of 1:1. Basement cannot be adequately distinguished from pre-Bury Devonian on the southern side of the trough. On the northern side, the line intersects a series of thrust faults that either ramp to the Carboniferous surface or are blind thrusts into the Boree Salt Member along the western front of the Pleasant Creek Arch (cf. Fig. 12). Note the net thickening of the Etonvale Formation to the NE.

The oldest formation in the group, the Late Carboniferous Lake Galilee Sandstone, is confined to the Koburra Trough. The succeeding Jericho Formation consists of mudstones and siltstones with subordinate volcaniclastic sandstones that are products of low energy stream and lacustrine environments. Correlatives of this facies have been recorded in all wells that have penetrated the southern Galilee Basin.

The Jochmus Formation and Aramac Coal Measures are confined to the Koburra Trough and the Lovelle Depression (Fig. 5). In places around the Blackall Ridge they unconformably succeed the Jericho Formation (Robson, 1986). The depositional environments remained glaciofluvial and glaciolacustrine and the sandstones were largely volcaniclastic. The Aramac Coal Measures, preserved mainly upon the Beryl Platform and within the Lovelle Depression, are products of a deltaic-lacustrine environment (Gray & Swarbrick, 1975). The coal measures in part correlate with the upper levels of the Jochmus Formation (Evans, 1980) and are envisaged as being created around the foot of a broad complex of humid alluvial fans that spread from the eastern and northern flanks of the Koburra Trough (Robson, 1986). No correlatives of this facies have been recognized in the southern Galilee Basin, which was a stable platform at that time.

Deposition in the Cooper Basin commenced with fluvio-(?)glacial and aeolian sediments of the Merrimelia and Tirrawarra Formations (Williams & Wild, 1984; Williams & others, 1985) and continued south of the Naccowlah Ridge with a series of fluvial and lacustrine formations up to 1200 m thick. Little of this series is preserved in the northern half of the basin. The distribution of the Lower Permian was controlled by active normal faults that defined a series of NE-SW trending troughs across the southern half of the basin.

Late Permian

The second major sequence, of Late Permian channel and flood plain deposits, covers older formations following a regional hiatus. The unconformity is of

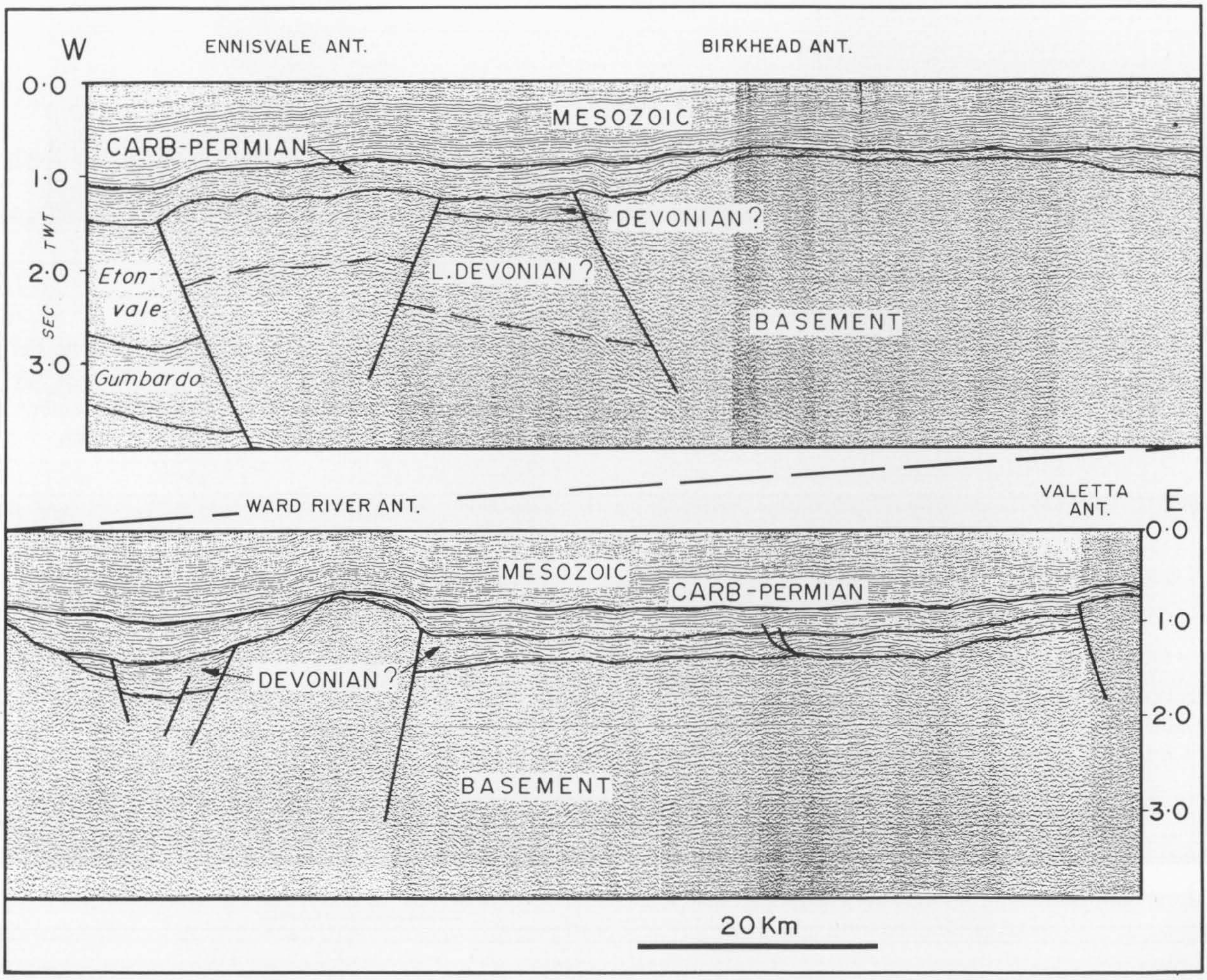


Fig. 14 Seismic sections Esso T81-1101 and T81-3003, extending for 150 km across the northern half of the Westbourne Block/Springsure Shelf, between the Eniskillen and Valetta Anticlines (Esso Australia Ltd, 1985). No well was directly located along this line, but its western end is correlatable to Fairlea-1 (Robson, 1986) and Valetta-1 was offset 2.5 km to the south of its eastern end. Birkhead-1, located on the northern extension of the Birkhead Anticline and 35 km to the north of the line of section, encountered marine Devonian (?)Bury Limestone. The horizontal: vertical scale ratio is about 4:1; faults flanking the Eniskillen and Birkhead Anticlines are therefore high-angle thrust faults. The lack of critical well control prevents positive identification of graben infill between the Eniskillen and Birkhead Anticlines and either side of the Ward River Anticline. Intermittent reflections below the Eniskillen Anticline are tentatively correlated to the Gumbardo Formation. The overlying sequence could be a correlative of the Etonvale Formation, or even as young as the Carboniferous of the Drummond Basin. The Permian includes the Joe Joe Group and younger formations. Basement forming the Eniskillen, Birkhead, Ward River, and possibly the Valetta uplifts was emergent during the Early Permian. Structuring was largely completed during the Carboniferous, but there was minor rejuvenation at the end of the Triassic and more substantial reactivation after the Cretaceous.

lesser duration in the Cooper Basin than in the Galilee Basin and follows basin-wide movements (Kuang, 1985). The Late Permian Bandanna Formation and Colinlea Sandstone equivalents in the Galilee Basin and the Toolachee Formation in the Cooper Basin form the most widespread Upper Palaeozoic unit; it extends over an area of at least 500 000 km^2, although it is only 100-150 m thick (Vine, 1976; Stuart, 1976). Accumulation of these sediments is attributed to back-up of the base level of erosion by a regional rise in sea level, rather than by tectonic subsidence (Evans, 1980). The limits of Permian deposition are shown in Map 1 of this Bulletin.

Triassic

The undifferentiated representatives of the Lower - Middle Triassic Rewan Group, Clematis Group and Moolayember Formation are mostly confined to the northern half of the Galilee Basin (Vine, 1976). Wells & Newhouse (1988) mapped Moolayember (Triassic age) in the eastern end of the southern Galilee Basin; however, there is unlikely to be Triassic along the line of the transect.

In the Cooper Basin, the fluvio-lacustrine sediments of the Triassic Nappamerri Group (Papalia, 1969; Powis, 1989) in most places conformably overlie the Permian. They comprise the bulk of basin fill to the north of the Naccowlah Ridge (Kuang, 1985), where they are up to 760 m thick (Battersby, 1976). Variations in thickness of the Triassic were controlled by flexure of the substrate and reactivation of basement faults.

The Cooper Basin was initiated as a series of tectonic rifts, the alignment of which corresponds to older trends within the Thomson Fold Belt. By the Triassic, the basin subsidence followed the typical development of an intra-cratonic sag basin. However, migration of the depocentre to the northeast of the main rift basin still requires explanation.

The absence of Triassic formations from most of the southern part of the Galilee Basin, where the underlying Upper Permian is little eroded, is further evidence of the prolonged stability of much of the western sector of the Eromanga-Brisbane Transect until the onset of deposition during the late Early Jurassic.

Much of this crust was stable for about 40 Ma. Tectonism towards the end of the Triassic that affected the entire continent (Evans, 1988) induced some minor wrenching along pre-existing fractures in the Cooper Basin along the Naccowlah trend (Kuang, 1985; Nelson, 1985), and there were major adjustments in the Bowen and Drummond Basins. Final deformation of the northern parts of the Drummond Basin was at the end of the Triassic (Evans & Roberts, 1980; Fenton & Jackson, 1989).

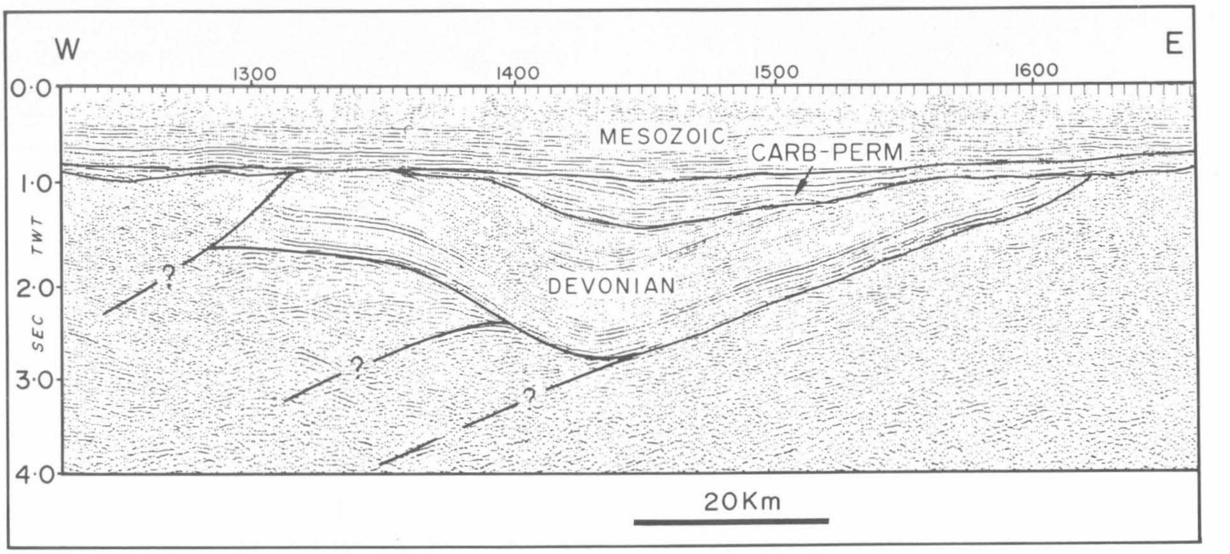


Fig. 15 Section from BMR traverse 14, across the Westgate Trough. The horizontal: vertical scale ratio is about 5:1, and hence the faults controlling uplift of the western flank of the trough are low-angled, to be compared with shallow crustal thrusts at the southern end of the Pleasant Creek Arch (Fig. 12). The southwestern flank of the Westbourne Block was evidently the source of much of the Permian following structuring during the Carboniferous and subsequent strong erosion.

JURASSIC-CRETACEOUS

The Jurassic-Cretaceous in the Eromanga Basin is the final and most widespread of the three mega-sequences that cover the western third of the transect, where it is generally less than 1000 m thick (Fig. 3). The sequence (Table 3) consists of a series of extensive cycles of dominant sandstone succeeded by dominant shale, that are traceable eastwards into the Surat Basin (Exon, 1976; Exon & Burger, 1981; Passmore & Burger, 1986).

Early Jurassic

The "basal Jurassic" and "basal Hutton" fluvial sandstones (Veevers & Evans, 1975; Wiltshire, 1982) comprising the first cycle, are centred above the Cooper Basin and extend eastwards across the Canaway Ridge and Springsure Shelf to join with the Precipice Sandstone-basal Hutton Sandstone of the Surat Basin. In the vicinity of the transect, this unit is over 60 m thick (John & Almond, 1987).

Early Jurassic-Early Cretaceous

The succeeding two cycles (upper Hutton Sandstone - Westbourne Formation) are of fluvial sandstones succeeded by lake and flood plain siltstones and shales with intermittent channel sands and fresh-water deltas, which are remarkably similar to each other in character. A major change from quartz lithic to volcanolithic provenance is recorded towards the top of the Hutton Sandstone (Watts, 1987).

The remaining depositional cycles are products of an overall marine transgressive - regressive episode, that left a widespread series of shales and sandstones with minor limestone across the basin. In the vicinity of the transect, this series amounts to about two-thirds of the thickness of the Jurassic-Cretaceous mega-sequence.

LATE CRETACEOUS - CAINOZOIC

Attaining equilibrium once more, the region was exposed to severe weathering for at least 40 Ma. Two major weathering episodes are recognised (Idnum & Senior, 1978) and span the Late Cretaceous to Paleocene and Late Oligocene to Early Miocene. During the Eocene, general uplift of the northern flanks of the Eromanga Basin led to deposition of quartzose sands, initially following synclinal valleys. The river systems that transported these sediments flowed towards the ancestral Lake Eyre Basin. Subsequent silicification during the Late Oligocene and Early Miocene created widespread silcrete profiles, which at present form cappings to erosional landforms carved in the Eyre and Glendower Formations. Along structural interfluves, where the sands were poorly represented or absent, the older weathered Cretaceous rocks were reweathered forming an unusual superposed profile. Gentle deformation and selective erosion of these indurated weathered profiles following the Early Miocene, mirrors the presence of faults and folds within the more-strongly deformed basin sequences and basement. A degree of strike-slip motion is registered along some of these fractures (Nelson, 1985). At present fluviatile and lacustrine sediments continue to accumulate within actively subsiding synforms and local basinal depressions, the largest of which is occupied by Lake Eyre.

The Thargomindah Shelf, to the south of the transect, is devoid of any obvious influential basement features and, therefore, deformation of the Eromanga Basin sequence upon it should reflect stress vectors during the late Cretaceous-Cainozoic. N-S and E-W oriented features are the dominant trends on the shelf and are believed to be the result of separate deformational events (Hoffmann, 1989b). There is difficulty in determining the relative timing of the two deformations. The E-W trending features have been attributed to N-S compression resulting from the tectonic activity between the Australian and SE Asian plates to the north of Australia (Hoffman, 1989b). The N-S trending structures are tentatively assigned a Miocene-Holocene age, correlating this E-W compressional phase of deformation with uplift and tilt of the Eromanga Basin during the late Tertiary. Other N-S trending structures, such as the Canaway Fault and the reverse faults in the southern Warrabin Trough, were preferentially reactivated during this final phase of deformation (Map 1, this Bulletin).

A rise in geothermal gradient that occurred in the Cooper Basin during the Tertiary is probably related to Tertiary deformation. The rise in temperature and the reactivation of the older structures was of major importance to the maturing and emplacement of hydrocarbons in the Eromanga Basin (Finlayson & others, 1988) and in other Mesozoic basins throughout Australia (Evans, 1988; Passmore, 1989).

EVOLUTIONARY MODELS

The evolution of central and southwestern Queensland during the Middle Palaeozoic has been speculatively related to plate convergence along a palaeo-Pacific margin. Marsden (1972) regarded the Adavale Basin as part of a mobile platform adjacent to a marginal sea. Day & others (1978) and Murray & others (1987) retained the idea that a marginal sea separated the Adavale region from an island arc at least until the Middle Devonian.

Evans & Roberts (1980) introduced the idea that formation of the Middle and Late Palaeozoic basins in northeastern Australia was controlled by a dextral shear couple that paralleled the craton margin. Evans (1982) suggested that the stress vector and sense of shear reversed during the Carboniferous from N-S sinistral to NNW-SSE dextral. Harrington & Korsch (1985) and Murray & others (1987) also recognized evidence of considerable shear forces affecting the palaeo-Pacific margin, but did not consider the implications of their models to the interior of the craton. Recent data and models of the New England Fold Belt are summarized by Korsch & others (this Bulletin).

Powell (1984) envisaged that eastern Australia during the Late Silurian and Early Devonian was part of a NW-SE dextral transtensional regime in which the Adavale Depression was located to the east of a volcanic arc. Powell regarded the volcanics of the Gumbardo Formation as part of a cratonic margin at the southern end of an arc that extended northwards to the Hodgkinson Basin, whereas they are now regarded as the fill of a rift complex.

Leven & Finlayson (1987), interpreting structures in the southern Adavale Basin, identified two sets of faults. They interpreted the dominantly E-W trending set as thrust faults that developed during the Early Devonian, and the other set striking N-S as the result of movements that commenced in the Late Devonian and continued into the Carboniferous. They attributed these events to thrusting within the lower crustal reflective zone (8-12 s TWT). In the present paper, both sets of structures are interpreted to be expressions of the northward migration of the thrust front that created the Pleasant Creek Arch. However, Hoffman (1988) interpreted the arch to have been initiated as the uptilted edge of an extensional block, which was later subjected to Middle Carboniferous compressional deformation.

Leven & others (1990) explained creation of the Quilpie and Westgate Troughs by westward and downward slippage upon mid-crust ramps. Recognition of these mid-crustal detachments under the transect is of major significance to the model presented here. However, the formation of the Quilpie and Westgate Troughs is interpreted as being caused by upper crustal megafolding (Fig. 15), rather than being formed by syndepositional subsidence along the ramps to produce ramp synclines and basins as suggested by Leven & others (1990).

The underlying tectonic causes of the observed structures cannot be determined solely from data directly within the transect, and a broader perspective must be taken. Dilation of the crust during the Early Devonian was characteristic of the entire foreland region of the Tasman Fold Belt. Not only the Adavale Depression, but also the Darling Depression (of which the Darling Basin in New South Wales is the result) were initiated at that time. Subsequent events during the Devonian and Carboniferous influenced an even greater part of this foreland region, from the Drummond Basin in the north to the Melbourne Trough in the south. (Fig. 17).

Powell (1984) compared the evolution of eastern Australia during the Late Silurian-Middle Devonian to the Cainozoic evolution of western North America. A more appropriate model for the corridor of Early Devonian rift basins is obtained by comparison with the series of Cainozoic basins in western Southeast Asia that were created during the oblique dextral convergence of the edge of the Indian plate with the Southeast Asian Plate (Fig. 17) (Tapponnier & others, 1986; Burri, 1989; Polachan & Sattayarak, 1989).

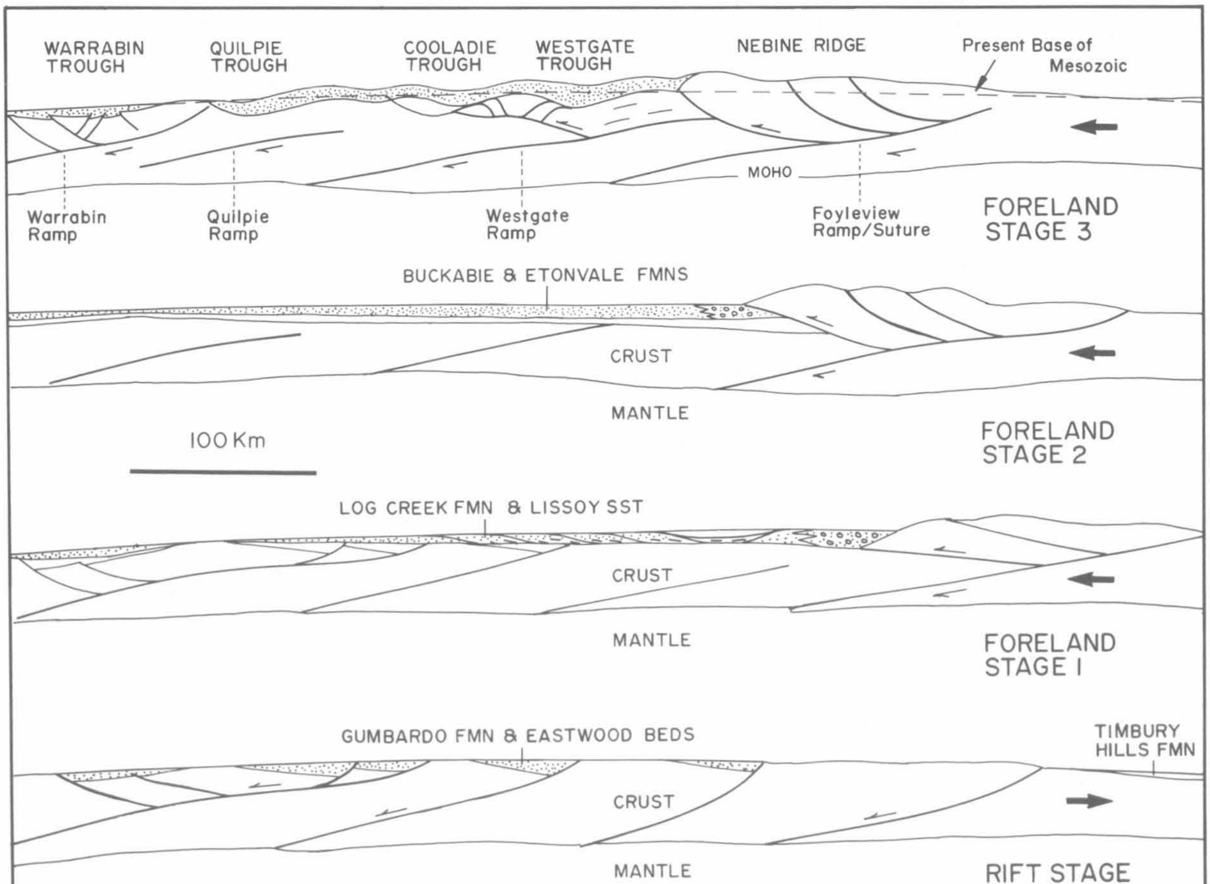


Fig. 16 Diagramatic sections depicting stages in the evolution of the region around the western third of the transect during the Devonian and Carboniferous.

Southeast Asia is a mosaic of welded, older Palaeozoic blocks (Bunopas, 1981), the shape of which influenced subsequent deformation. Tapponier & others (1986) demonstrated that Southeast Asia was "extruded" eastward during the collision of Greater India and Asia to form a series of wedges, movement of which was controlled by a series of strike-slip master faults. When this model is mirrored to accommodate a N-S shear along an eastern craton margin, as implied for the palaeo-Pacific margin of Australia, cross faults such as the Mae Ping and Three Pagodas Faults fit within the scale and trend, for example, of the Pleasant Creek Arch, and possibly the Westgate or Foyleview sutures. Within this model, the Phitsanulok, Petchabun and associated basins are analogues of the initial rifts into which the Gumbardo Formation was deposited. The complex of rift basins in the Gulf of Thailand has its parallels in the early development of the Darling Basin. The Andaman Sea would compare with the marginal sea that probably existed to the east of the Nebine Ridge during the Early Devonian. The Burma Ranges would compare with the Nebine Ridge.

There is general agreement that eastern Australia was fully cratonised by the mid-Carboniferous, when the palaeo-Pacific margin was of Andean type. Structures formed within the Adavale and adjacent domains during the Middle Devonian-Carboniferous bear comparison with structures in other foreland thrust zones. The blind thrust duplex at the southern end of the Pleasant Creek Arch and the Gumbardo structure, for example, have parallels in the front of the Canadian Rockies. Parallels with Appalachian and Laramide structures are limited, however. Ramps to the surface in the Adavale region are broadly spaced, and a distinct shear component seems to have been involved (Evans, 1982; Robson, 1986; Finlayson & others, 1988) (Fig. 17).

The Grenfield Uplift and the Cothalow-Carlow trend, that were previously normal, transfer or strike-slip faults, were approximately at right angles to the principle vector of net compression; they became upthrust reverse faults. The previously extensional fractures within basement below the Cooladdi Trough and along the Blackall Ridge became strike-slip faults in the Late Devonian - Carboniferous.

Movements within the upper crust could have been accommodated on the intra-crustal detachments zones, such as the Quilpie and Westgate, and the Foyleview suture (Fig. 16) in order to accommodate the degree of rotational shear required. Thrusts, such as at the front of the Pleasant Creek Arch in this model, would be back-thrusts accommodating the rising western block.

The change from transtensional to transcompressional regimes in the mid-Devonian does not require a major change in stress vector. However, a major change in stress field occurred in the Late Carboniferous or Early Permian, when a dextral shear vectored NNW-SSE affected eastern Australia (Evans & Roberts, 1980; Harrington & Korsch, 1985; Murray & others, 1987). The extensional regime under this shear couple, was confined to the north of the Adavale Basin, although fractures across the Westbourne Block were reactivated. A new fracture pattern was initiated to the west of the transect, forming the foundations of the Cooper Basin. A degree of shear motion controlling sedimentation in the Cooper Basin may also have been initiated at this time (Nelson, 1985).

The time of the change from the one stress field to the next is uncertain. It could have preceded, accompanied, or followed deposition of the Late Carboniferous sediments in the Koburra Trough, but the preferred interpretation is that a change occurred subsequent to deposition of the bulk of the Joe Joe Group. There was a shift in depocentre of the Galilee Basin during the Early Permian, which was also the time of initiation of the Cooper and Pedirka Basins (Fig. 1). Whatever the actual order of events, much of the landscape was eroded by the time that deposition recommenced in the Adavale region during the Late Carboniferous, when fans of sediments spread as a piedmont facies across the topographic low in front of a much reduced mountain chain and above the remains of the Adavale Depression (Figs 11-15).

The area centred on the transect was quiescent during the Late Permian. Even during the Triassic, the crust remained stable, neither receiving sediment nor being significantly eroded, although the northern part of the Cooper Basin and the Koburra Trough foundered at this time.

This period of stability along the transect continued until late in the Early Jurassic when depression of the Cooper Basin and the southern Galilee Basin resumed, perhaps initially due in part to sediment compaction as much as to tectonism, creating a broad riverine plain as large as the Amazon that debouched across the Springsure Shelf into the Surat Basin. The Canaway Ridge, which had been an important structure during the Palaeozoic, had no significant influence on deposition (Hoffmann, 1989b).

Thereafter, until the mid-Cretaceous, the entire region foundered in stages, each marked by the spread of firstly a widespread sand that was succeeded by a clay-rich facies. Burger (1986) and Passmore & Burger (1986) ascribed these cycles to the influence of global changes in sea level. Across the Eromanga Basin, the fluvially transported and deposited sand at the base of the Jurassic cycles is just as likely to be the result of changes in the elevation and rates of erosion of the sediment source areas, as to distant fluctuations in sea level (Evans, 1988). The increasing content of volcanolithic components from the Middle Jurassic onwards is a measure of the continuing influence of a volcanic arc along the palaeo-Pacific margin. However, depositional cycles during the Late Jurassic - Early Cretaceous were distinctly influenced by fluctuations in sea level. They

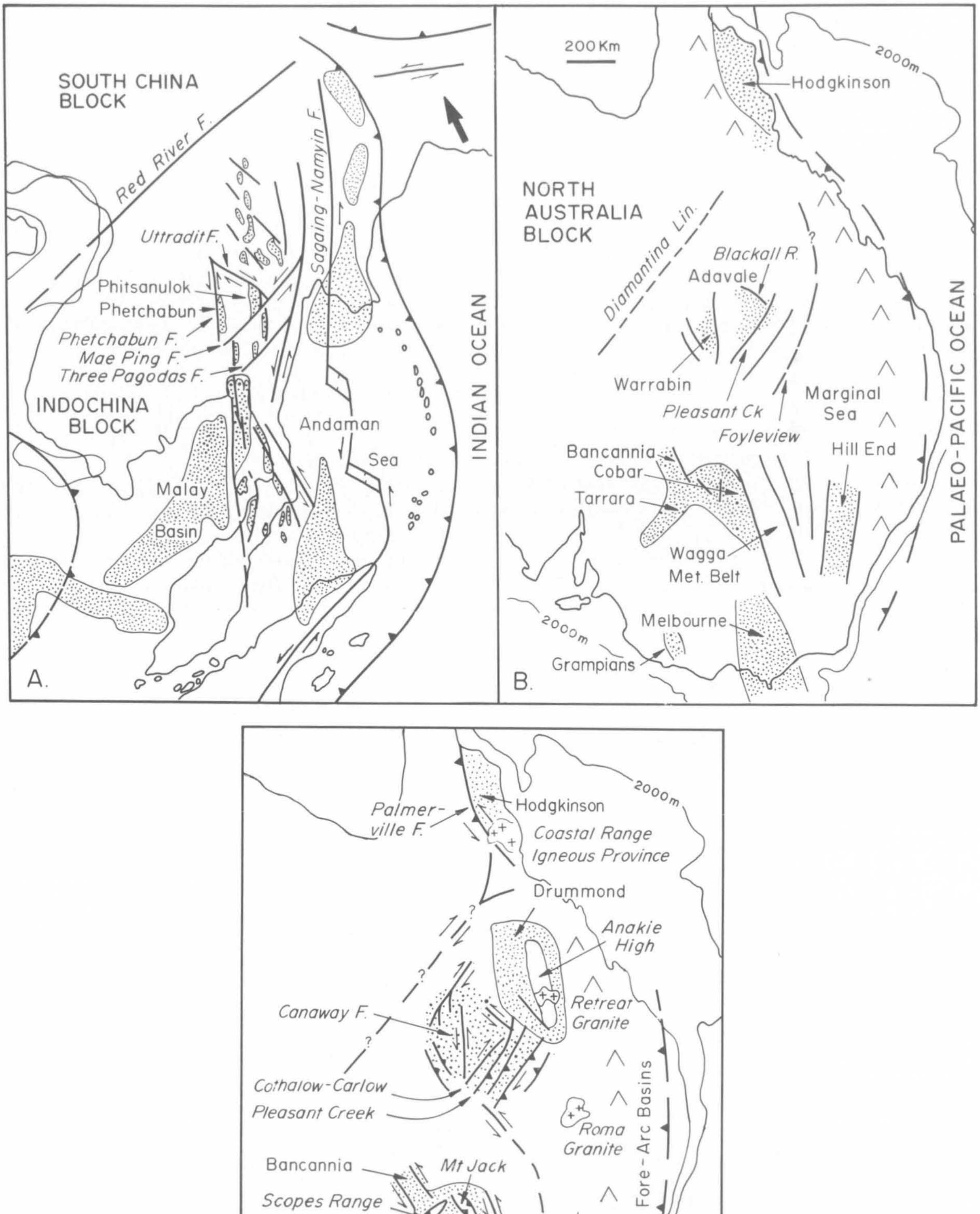


Fig. 17 Models of foreland basins to the Tasman Fold Belt during the Devonian and Carboniferous. (a) a mirror image of the tectonics of Southeast Asia, based on Burri (1989, fig.1); (b) the Tasman Fold Belt during the Early Devonian; and (c) the Tasman Fold Belt during the Late Devonian - Carboniferous. The three diagrams are at the same scale.

culminated with the transgression of marine environments across the Eromanga Basin. Broad, crustal flexure and uplift around the flanks of the basin restricted the final phases of crustal downwarp to the centre of the basin.

The crust entered yet another period of remarkable stability during the remainder of the Cretaceous and the Palaeogene. Rejuvenation of old structures during the Neogene created a series of anticlines, normal and reverse faults. The overall stress pattern of this period has not yet been definitively analysed. Finlayson & others (1988) noted the principal direction of crustal shortening was NE-SW. Evans (1982) and Veevers & Powell (1984) noted the sense of intra-continental and continental margin block rotations of the period. Veevers & Powell (1984, p. 102) attributed the Cainozoic movements of the continent to sinistral shear interaction between the Pacific and Indo-Australia Plates, drawing a mechanical analogue of "...anticlockwise rotation imparted by a moving slab to a mosaic of blocks set in a less rigid matrix".

The concept of a craton being a mosaic of crustal blocks and ramps fits well with the available evidence. The origins of the mosaic comprising eastern Australia are found in the Palaeozoic history of the underlying fold belt.

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THE BOWEN BASIN AND OVERLYING SURAT BASIN

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ABSTRACT

The Permo-Triassic Bowen Basin and overlying Jurassic-Cretaceous Surat Basin fills are intersected by the BMR Eromanga-Brisbane seismic transect between the Nebine Ridge and Kumbarilla Ridge structural highs to the west and east, respectively.

The Bowen Basin, which is of controversial origin, is a north-south trending, asymmetrical syncline bounded in places by faults. The basin is here interpreted as a retro-arc foreland basin, which developed on the landward (west) side of a continental arc associated with continent-ocean plate convergence. Initiation of the basin occurred via back-arc extension in Early Permian times, coincident with a phase of calc-alkaline arc volcanism. Early Permian, mostly marine sedimentation gave way to extensive coastal and alluvial plain depositional systems in Late Permian times, coincident with a resurgence of arc volcanism. Triassic sedimentation was almost entirely alluvial and coeval with westward propagation of thrust sheets from the eastern basin margin. Compressive tectonics climaxed in Middle to Late Triassic times, terminating sedimentation and uplifting the entire area.

Following a period of erosion, the Surat Basin was initiated by largely passive downwarping. The Surat Basin is part of the larger Great Artesian Basin, a major intracratonic basin complex which covers a substantial part of eastern Australia. A sequence of mainly alluvial and lacustrine Jurassic sediments was succeeded by a mainly marine Early Cretaceous interval. Mild rejuvenation of earlier faults occurred at intervals throughout this time.

Sediment accumulation in the Surat Basin was terminated in mid-Cretaceous times. Minor extensional fault movements characterised the Palaeocene period, associated with the opening of the Coral Sea, and mild compressive deformation in Oligocene-Miocene times coincided with a major phase of basaltic volcanism in eastern Australia. In late Tertiary and Quaternary times, the area evolved through erosion to its present topographic relief.

INTRODUCTION

The BMR Eromanga-Brisbane Seismic Transect intersects the Permo-Triassic Bowen Basin and the Jurassic-Cretaceous Surat Basin fills in central south Queensland. It is therefore appropriate that this Bulletin should contain a review of the geological evolution of these basins. This paper aims to gather available geological data on the Bowen and Surat Basins (Fig. 1), and to present an interpretation of the area's geological history since Late Carboniferous times.

A variety of geological data pertaining to the Bowen and Surat basins has been published in recent years, particularly as a result of subsurface exploration for oil and gas. These data include a considerable amount of seismic reflection records which yield information on the basins' structural and stratigraphic evolution. To date, however, this large volume of open file data has not yet been effectively synthesised into a unified geological model.

Several tectonic models have been proposed to explain the Late Paleozoic and later geological record of easternmost Australia. These are based largely upon the regional distribution of rock types and structural features, and have not for the most part employed modern techniques of basin analysis. In the present paper, regional correlation of major stratigraphic units in wells and from seismic records, and delineation of facies patterns, have allowed a more detailed examination of basin evolution. Research in progress is aimed at testing and refining interpretations presented herein.

STRUCTURE

Regional structure of the Bowen and Surat Basins is well known from the extensive exploration activities in the area. The major structural features of the two basins are illustrated in Figure 2.

The Bowen Basin is a narrow, north-south, elongate feature with an asymmetric, synclinal structure (Figs. 1,

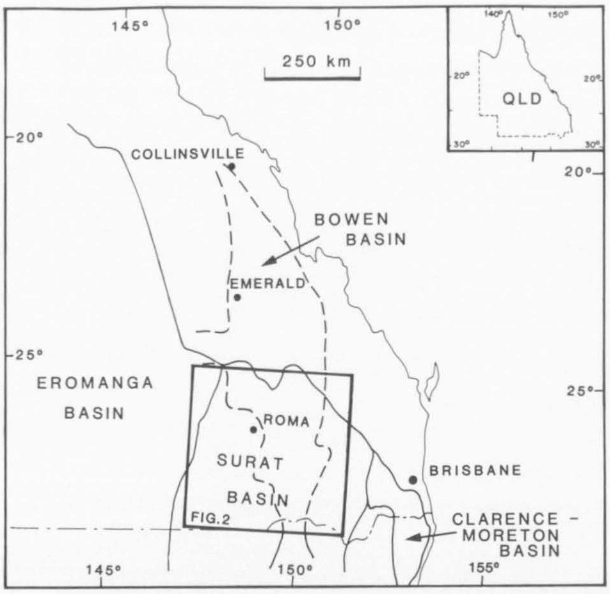

Fig. 1 Map showing the location of the Bowen and Surat Basins, and of Figure 2.

3, & 4). Along the basin axis, referred to as the Taroom Trough, up to 10 km of Permian and Triassic sedimentary rocks are preserved. The basin, which is contiguous with the Gunnedah and Sydney Basins to the south, was bounded to the east by an intermittently active volcanic chain, and to the west by cratonised Palaeozoic and Precambrian rocks. Permian and Triassic sedimentary rocks onlap the western basin margin and possibly parts of the eastern margin, although later faults complicate onlap patterns. Faults and fold axes associated with the Bowen Basin sequence are mostly north-south trending, with a few notable exceptions (Fig. 2). Faults are either normal, down-to-the-basin structures, or similar faults which were later reversed by compressional forces. Evidence of limited strike-slip movement is preserved in some areas (Fig. 2). Many faults are seen on seismic records to penetrate basement (Fig. 4).

The Surat Basin is a component of the much larger Great Artesian Basin system (Fig. 1). It is bounded to the east by the Kumbarilla Ridge and to the west by the Nebine Ridge, both basement highs which restricted sediment supply and accumulation at various times. The axial zone of the Mimosa Syncline is the major depocentre of the Surat Basin and is parallel to and slightly offset from the Taroom Trough axis. In this structural low, up to 2500 m of Jurassic and Early Cretaceous sedimentary rocks are preserved (Fig. 3).

Major faults and fold axes associated with the Surat Basin sequence may be divided into three groups; (1) north-south trending structures (which are mainly reactivated Bowen Basin structures), (2) north-north-west to north-west trending structures, and (3) north-east-trending structures.

The structural development of the area from Late Carboniferous time onward will be discussed in the context of basin evolution, following an outline of the stratigraphy of both basins.

STRATIGRAPHY

Bowen Basin

Stratigraphy of the Bowen Basin is presented in Figure 5. This represents a distillation of data from many sources, using the framework of Price & others (1985), QDM (1988) and the absolute time scale of Harland & others (1982). Emphasis is placed on the southern part of the basin, the area within the Geoscience Transect. For a review of stratigraphic correlations over the entire basin, see Draper (1985).

The basement to the Bowen Basin in the area of the transect comprises a complex mosaic of geological units. On the Roma Shelf, the basement is composed of Devonian metasedimentary rocks (the Timbury Hills Formation; see Murray, this volume) and Early Carboniferous acid intrusive masses (the Roma Granites). Isostatic buoyancy of the Roma Granites is thought to be responsible for maintaining the Roma Shelf as a structural high throughout Late Palaeozoic and Mesozoic times. Elsewhere, sedimentary and volcanic rocks including those of the Connors-Auburn arch and Yarrol Basin systems subcrop and outcrop. The tectonic context of these units is discussed by Murray (this volume).

The Late Carboniferous to Early Permian Combarngo Volcanics, and somewhat later Camboon and Lizzie Creek Volcanics represent the earliest recognised deposits of the Bowen Basin. These are acidic and intermediate lavas, volcaniclastics and sedimentary rocks, representing a major period of extrusive activity during Late Carboniferous and Early Permian times. Cosgrove & Mogg (1985) note that from the Roma Shelf the Combarngo Volcanics thicken northeastwards into the axis of the Taroom Trough. The Lizzie Creek and Camboon Volcanics are developed on the eastern margin of the basin.

The term "Kuttung Formation" has been somewhat loosely used to describe Late Carboniferous and Early Permian rocks under the Bowen Basin fill (Mack, 1963). This term, though never formally defined, has been adopted widely and indiscriminately to describe pre-Bowen Basin sedimentary and volcanic "basement". Because of the confusion this has created, it is here proposed that the term "Kuttung Formation" be discontinued with respect to the Bowen Basin area.

Unconformably overlying the early volcanics are localised sequences of Early Permian non-marine sedimentary rocks. These have to date been penetrated only on the western flank of the basin, although a more widespread distribution in the subsurface has been suggested. In the Roma Shelf area, at least five half-graben structures have been recognised, filled with at least 1 km of mostly fine-grained sedimentary rocks of possible alluvial and lacustrine origin.

Farther north, in the Denison Trough, several similar half-graben structures have been recognised (Bauer & Nelson, 1980; Nelson & Bauer, 1981; Ziolkowski & Taylor, 1985) which contain Early Permian coal-bearing sedimentary rocks of the Reids Dome beds (Draper & Beeston, 1985). The Reids Dome beds, totalling up to at least 2770 m in thickness, comprise deposits of alluvial fan, fan delta, alluvial valley/plain, lacustrine, and in the uppermost part of the unit, marginal marine environments, according to Draper & Beeston (1985). Price & others (1985) recognise an unconformity in the Reids Dome beds of the northern Denison Trough, as do Ziolkowski & Taylor (1985) in the southern Denison Trough.

The Camboon and Lizzie Creek Volcanics, and coastal and shallow-marine sediments of the Buffel and Tiverton Formations on the eastern margin of the basin, are at least partly contemporaneous with deposition of the Reids Dome Beds. Draper (1988) notes that the Buffel Formation of mixed carbonate and clastic lithologies accumulated on an initially uneven, perhaps fault-controlled topography.

Following deposition of the Reids Dome Beds and equivalents, the late Early to Late Permian Back Creek Group was deposited in marine to coastal environments

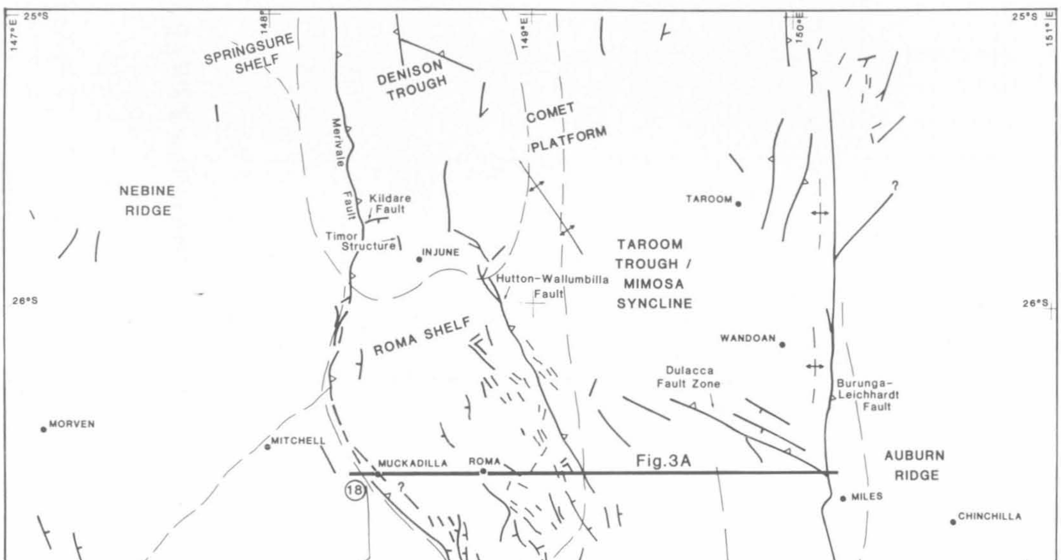

Fig. 2 Map showing major structural geological features of the Bowen and Surat Basins. Note the positions of cross-sections shown in Figures 3 and 4.

of deposition. In the Denison Trough, sediment accumulation was continuous, leading to a total sequence thickness in excess of 4.5 km, while in the east and south a substantial period of non- deposition was followed by Late Permian sediment accumulation, leading to more modest sequence thicknesses (Fig. 5). Near the top of the Back Creek Group sediments show a major change in mineralogy from quartzose to volcanic lithic.

The overlying Blackwater Group is present across the entire basin, and comprises tuffaceous and coal-bearing sedimentary rocks ranging up to 700 m in aggregate thickness. These were deposited in deltaic, coastal plain and alluvial plain environments of deposition (Dickens & Malone, 1973).

Above the Blackwater Group lies the Late Permian to Early Triassic Rewan Group, which is apparently conformable over much of the basin though unconformable in the Roma Shelf area (Price & others, 1985). The Rewan Group comprises reddened sandstones and siltstones of interpreted alluvial plain origin, and conglomerates of interpreted alluvial fan origin are known from the subsurface adjacent to the eastern margin of the basin (referred to as Cabawin Formation). This is succeeded by the alluvial but quartzose and more sandstone-dominated Clematis Group (late Early to Middle Triassic). The subsurface equivalent of the Clematis Group in the southwestern part of the basin is referred to as the Showgrounds Sandstone, and is interpreted to be of fluvial origin with some lacustrine or marine influences (Butcher, 1984; Schroder, 1988).

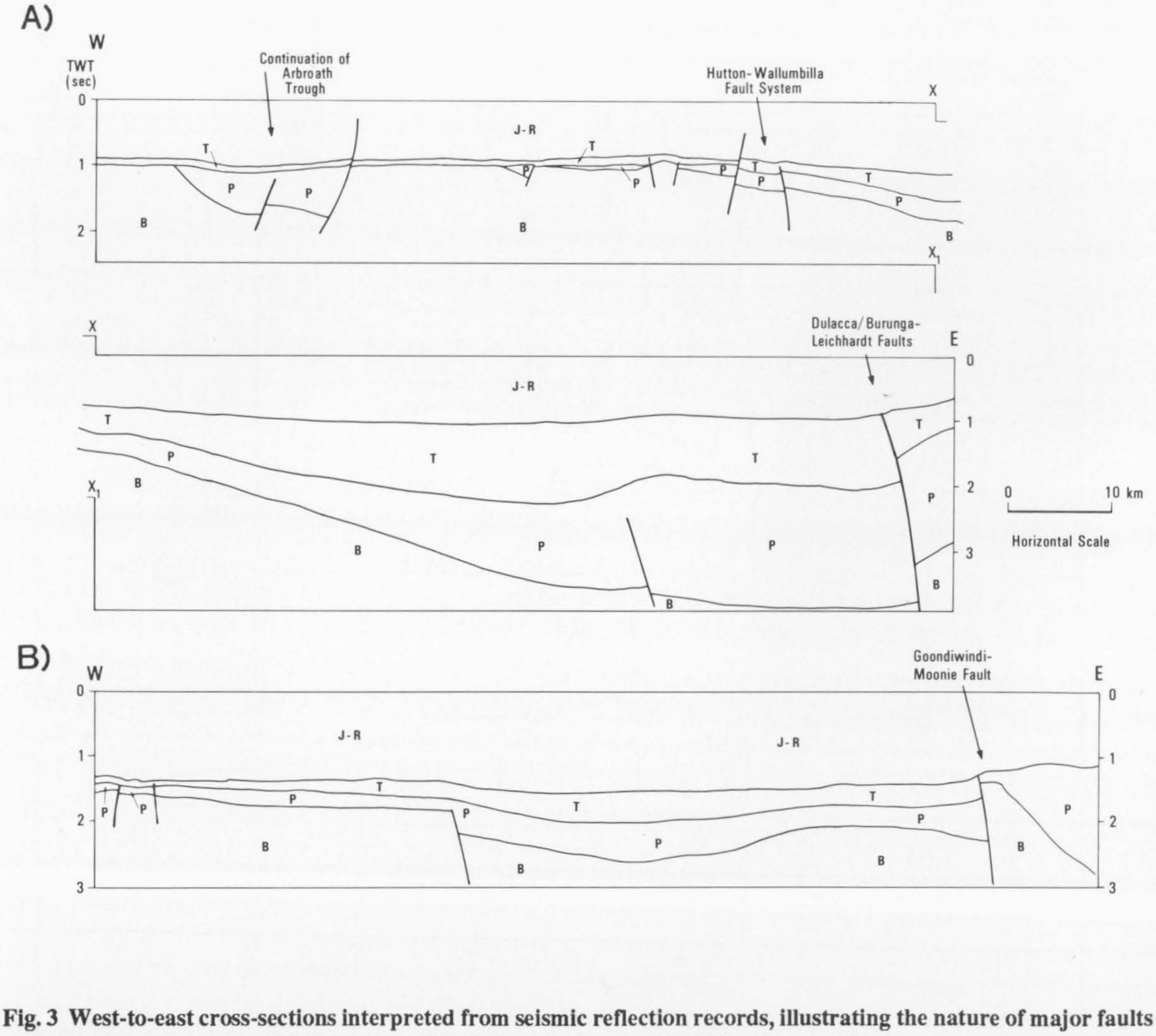

Fig. 3 West-to-east cross-sections interpreted from seismic reflection records, illustrating the nature of major faults and geometry of stratigraphic intervals (after Elliott & Brown, 1988). See Figure 2 for locations of cross-sections. B = Basement, P = Permian, T = Triassic, J-R = Jurassic-Recent.

The youngest unit preserved in the exposed part of the basin is the Middle to ?Late Triassic Moolayember Formation, which comprises mainly siltstones and sandstones of interpreted alluvial and lacustrine origin (Jensen, 1975). Laterally restricted conglomerates of interpreted alluvial fan origin outcrop and subcrop on the central eastern margin of the basin.

Isopachs of most stratigraphic intervals in the Bowen Basin show pronounced asymmetry in plan (Exon, 1976), with thickness maxima located along the axis of the Taroom Trough.

Surat Basin

The stratigraphy of the Surat Basin presented in Figure 5 is based primarily on the schemes of Exon (1976), Price & others (1985) and Elliott & Brown (1988).

The basement to the Surat Basin is partly the Palaeozoic units forming basement to the Bowen Basin, and the Bowen Basin fill itself (Exon, 1976). The Surat Basin has a considerably wider distribution than the underlying Bowen Basin fill which, following erosion, is largely confined to the Taroom Trough, Denison Trough and adjacent shelves. Surat Basin formations, therefore, have a more widespread and also more sheet-like distribution than their predecessors. The components of the Surat Basin fill have lateral equivalents across the entire Great Artesian Basin system, representing vast areas of broadly similar depositional conditions.

The earliest deposits of the Surat Basin are the Early Jurassic alluvial sandstones of the Precipice Sandstone (Martin, 1980). These are succeeded by probable lacustrine and coastal plain sediments of the Evergreen Formation (Golin & Smyth, 1986; Fielding, 1989c). In turn, overlying the Evergreen Formation is the Hutton Sandstone, interpreted as being deposits of an alluvial plain.

The basal Early Jurassic part of the basin fill is succeeded by the Middle to early Late Jurassic Injune Creek Group, comprising the Eurombah Formation, Walloon Coal Measures and its equivalent Birkhead Formation, Springbok Sandstone, and Westbourne Formation. These units are interpreted to have accumulated in alluvial and lacustrine environments (Exon, 1976).

The Blythesdale Group (Late Jurassic to Early Cretaceous), overlying the Injune Creek Group, again comprises sequences of mainly fluvial origin, in some cases associated with coastal plain environments. The uppermost component of the Blythesdale Group, the Bungil Formation, is a fine-grained interval of interpreted coastal plain and marine shelf origin. Succeeding the Blythesdale Group is the Rolling Downs Group (Early Cretaceous), comprising units of fluctuating marine shelf and coastal/alluvial plain origin.

Isopachs of Surat Basin sequences are more complex than those for the Bowen Basin, but are broadly concentric around several depocentres, including the Mimosa Syncline (Exon, 1976).

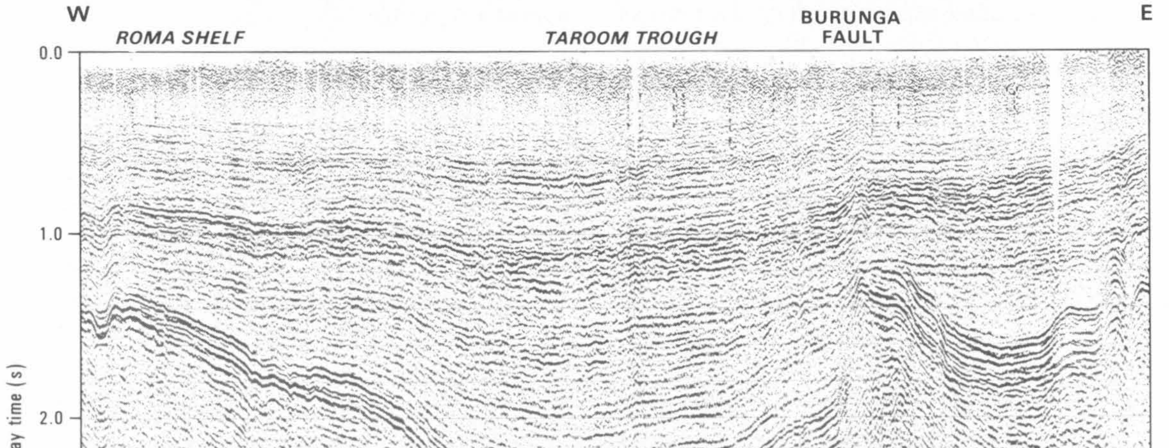

Fig. 4 Part of BMR Seismic Reflection Line 14, presented as an uninterpreted, vertically exaggerated section to show the asymmetrical nature of the Taroom Trough. Note the deep- seated nature of major faults. The Bowen and Surat Basin fills are separated by a regional unconformity, visible on the section at 1.3-1.5 secs. TWT. See Fig. 2 for location.

PUBLISHED TECTONIC MODELS

Before proceeding to interpret the geological development of the Bowen/Surat area, the various tectonic models already published will first be subjected to a brief critical review. The geological history of the area as a whole is reviewed in general terms by Murray (this Bulletin).

The Surat Basin is almost universally accepted as a part of an extensive intracratonic sag which subsided mainly passively, exploiting pre-existing crustal weaknesses (Exon, 1976; Veevers & others, 1982; Day & others, 1983). Veevers & others (1982) have drawn attention to the apparent timing of events within the Great Artesian Basin and possible coincidences with plate marginal events.

Murray (this Bulletin) sets the tectonic scene for the Bowen Basin. The New England Fold Belt, extending north-south along a large part of the present Great Dividing Range, is conventionally interpreted to be the result of a convergent plate margin active during Late Devonian and Carboniferous times. The subduction zone within this system dipped westward towards the Australian continent, according to most workers (see Korsch & others, this Bulletin).

Formation of the Bowen Basin in Late Carboniferous or earliest Permian times coincided with a phase of voluminous calc-alkaline volcanism along the trend of the earlier (interpreted) Connors-Auburn volcanic arc (Veevers & others, 1982). This and various other features have been explained by most workers in terms of a retro-arc (foreland) basin model (terminology of Miall, 1983), with formation of the Denison Trough at the western basin margin being ascribed to possible back-arc extension. Hammond (1987), however, viewed this early extensional phase as a consequence of major crustal extension, and proposed an early phase of "passive margin' style tectonics, followed by true foreland basin development from mid-Permian times onward. Hammond (1987) and Mallett & others (1988) recognised a series of northeast-southwest trending structural "corridors" across the Bowen Basin, which they ascribed to transfer fault activity during Early Permian crustal extension. This hypothesis has not yet been rigorously tested, and there is a possibility that these features relate to the much later opening of the Coral Sea.

A few workers ascribe formation and infilling of the Bowen Basin to major strike-slip fault motion. Dextral movement along northwest-trending fault lines is favoured by Evans & Roberts (1980), while Paten & others (1979), Brown & others (1983), and Korsch & others (1989) attribute compressional forces to transpression associated with sinistral shearing along major fault lines. Murray & others (1987) interpreted regional-scale oroclinal bending in the New England Fold Belt of northern NSW as a consequence of major dextral transcurrent faulting in Late Carboniferous or Early Permian times. Further, evidence exists for limited wrench fault activity during Early Permian times in the Denison Trough (Ziolkowski & Taylor, 1985) and in the adjacent Cooper Basin (Nelson, 1985).

Tectonic models proposed for the Bowen Basin have to date ignored perhaps the most important line of evidence available to geologists, - sedimentological characteristics of the basin fill. Modern basin analysis (cf. Miall, 1983) relies upon recognition of distinct styles of basin infilling to a large extent, together with the availability of a closely resolved stratigraphy. The stratigraphy of the Bowen Basin is well-described, although a myriad of local stratigraphic terminology exists (QDM, 1988), and current investigations by the author and colleagues are aimed at improving on earlier sedimentological models generated by BMR workers (e.g. Dickens & Malone, 1973; Jensen, 1975).

In the following passages, the available sedimentological information is combined with stratigraphic and structural data to provide an interpretative history of Bowen Basin evolution. This is followed by a similar treatment of the Surat Basin.

GEOLOGICAL HISTORY

The interpreted development and infilling of the Bowen and Surat Basins are detailed in this section, and summarised in Figure 5.

The Bowen Basin was initiated during Early Permian times with the establishment of a major calc-alkaline volcanic belt on the eastern margin (termed the Camboon volcanic arc). Volcanism seems to have been discontinuous, and concentrated in the northeast (Lizzie Creek Volcanics) and southeast (Camboon Volcanics) margins of the basin. It is possible that a westward-thinning wedge of volcaniclastic rocks extends across the basin, mainly in the subsurface. This would account for the numerous records of Early Permian volcanics over the basin, although further centres of volcanic activity cannot be discounted.

Following the onset of calc-alkaline volcanism (in both subaerial and subaqueous environments), basement-involved extension created a series of north and northwest-trending graben and half-graben. These are best documented along the western margin of the basin, where they collectively form the Denison Trough and also extend southwards across the Roma Shelf (Figs. 2, 3). Such extensional sub-basin fills have also been recognised in the lower part of the Permian section in the Taroom Trough (Elliott & Brown, 1988), although the intervening Comet Platform area remained a non-subsiding structural high at this time. Fault lines in many cases were probably reactivated earlier structures, which penetrated basement. Loading of the crust with arc-derived volcaniclastic material has been suggested as the stimulus for subsidence at this time, although a more fundamental plate tectonic event may have initiated both volcanism and subsidence.

Early sedimentation in the Bowen Basin was restricted to isolated, tectonically active sub-basins. Those close to mountainous terrains (notably the Denison Trough area) received coarse clastic sediment, whereas elsewhere (e.g. the Arbroath Trough) sediment supply was dominantly fine-grained. During deposition of the Reids Dome beds, environments were lacustrine and alluvial, with periodic lake infills by deltaic progradation and subsequent establishment of peat-forming mires (Draper & Beeston, 1985). Substantial coal resources are preserved in the Denison Trough. These events took place at the same time as the decline of a major glaciation across what is now eastern Australia (Crowell & Frakes, 1971). Exotic clasts have been noted from mudrocks in various Early Permian formations, and interpreted as glaciomarine dropstones (Draper, 1983).

Movement along faults bounding graben and half-graben seems to have declined concurrently with waning of volcanism on the eastern margin of the basin (Fig. 5). Subsidence in the Denison Trough changed to a more passive style, interpreted by Ziolkowski & Taylor (1985) to reflect thermal relaxation of the thinned, stretched crust. A marine transgression then inundated the Denison and Taroom Troughs. Facies relationships in the northern Bowen Basin (Martini & Johnson, 1987), and the lack of Early Permian strata in the south, suggest that the transgression came from the east, possibly via the gap between the Lizzie Creek and Camboon volcanic terrains.

A period of mixed carbonate-siliciclastic sedimentation followed on the subsiding eastern volcanic margin (Buffel Formation; Draper, 1988), with, in the west, mainly fine-grained marine clastic deposition (Cattle Creek Formation) and eastward progradation of deltaic wedges across the Denison Trough (Fielding & Lang, 1988). Marine sediment deposition was influenced by both wind-driven waves and tides, the latter implying connection with the ocean to the east of the Australian landmass (Fielding, 1989 a,b). North of the Denison Trough, half-graben development continued with the formation and infilling of the Wolfang and some other small basins (some of which are coal-bearing). In the south, however, no equivalent strata are preserved on the Roma Shelf, implying tectonic stability imparted by the presence of buoyant granite masses in the shallow subsurface.

In seismic reflection records from the Denison Trough, evidence of minor uplift and implied compression is present in the form of local onlap (unconformity) surfaces within the Cattle Creek Formation (Nelson & Bauer, 1981). Evidence of substantially greater compressional stresses is contained within the overlying Aldebaran Sandstone, which has several seismically-defined unconformity surfaces (Nelson & Bauer, 1981; Ziolkowski & Taylor, 1985). The Aldebaran Sandstone is a coarse-grained, quartzose sandstone-dominated interval of broadly deltaic origin. Progradation direction and basinal influences are as for the underlying Cattle Creek Formation. Vertical sequence development is more complex in the Aldebaran Sandstone, however, a consequence of the repeated structural upheavals of the time. Aldebaran Sandstone equivalents are not known from the southeastern side of the basin, although coastal and shallow-marine deposits of the Gebbie Formation and Collinsville Coal Measures occur in the northeastern and northern Bowen Basin, respectively (Martini & Johnson, 1987). This period of time is represented by an unconformity over much of the eastern flank of the Taroom Trough. North of the Denison Trough, half-graben formation and infilling continued in the Clermont area (e.g. Moorlands Basin; Sorby & Scott, 1988). Sedimentation in these basins was evidently of shallow-marine and deltaic aspect, and substantial coal reserves have been proved in some areas.

The mild compressional phase ended towards the close of deposition of the Aldebaran Sandstone. As a consequence, cessation of coarse sediment supply led to a marine transgression over the Denison Trough, while on the eastern flank of the Taroom Trough the carbonate-dominated Oxtrack Formation was formed in shallow-marine and coastal environments. At about this time, the previously non-subsiding Roma Shelf area was finally inundated by the sea (Paten & Groves, 1974) and a widespread marine transgression resulted in marine conditions over much of the basin, including the Springsure Shelf to the west. This may be interpreted either as reflecting the onset of steady, basin-wide subsidence, or as the result of a eustatic sea level change.

In the western half of the basin, the equivalent to the upper part of the Oxtrack Formation is the Freitag Formation, also a quartzose deltaic and shallow-marine deposit which prograded eastward into the Denison Trough and beyond. Above this unit is the fine-grained Ingelara Formation and its overlying Catherine Sandstone, together with their lateral equivalent to the south, the Muggleton Formation. The Catherine Sandstone passes northward into the German Creek Formation, which contains economically important coal deposits.

Sandstones in these five units (Cattle Creek Formation, Aldebaran Sandstone, Freitag and Ingelara Formations, Catherine Sandstone) are quartzose, implying continuation of sediment supply from the stable landmass to the west, a notion supported by regional palaeocurrent distributions. On the eastern flank of the Taroom Trough, the dominantly fine-grained Barfield Formation is equivalent to the Ingelara Formation and Catherine Sandstone. In contrast to these units, however, the Barfield Formation has a volcanic lithic composition, presumably reflecting derivation from the inactive volcanic arc. Depositional environments during this interval were entirely marine.

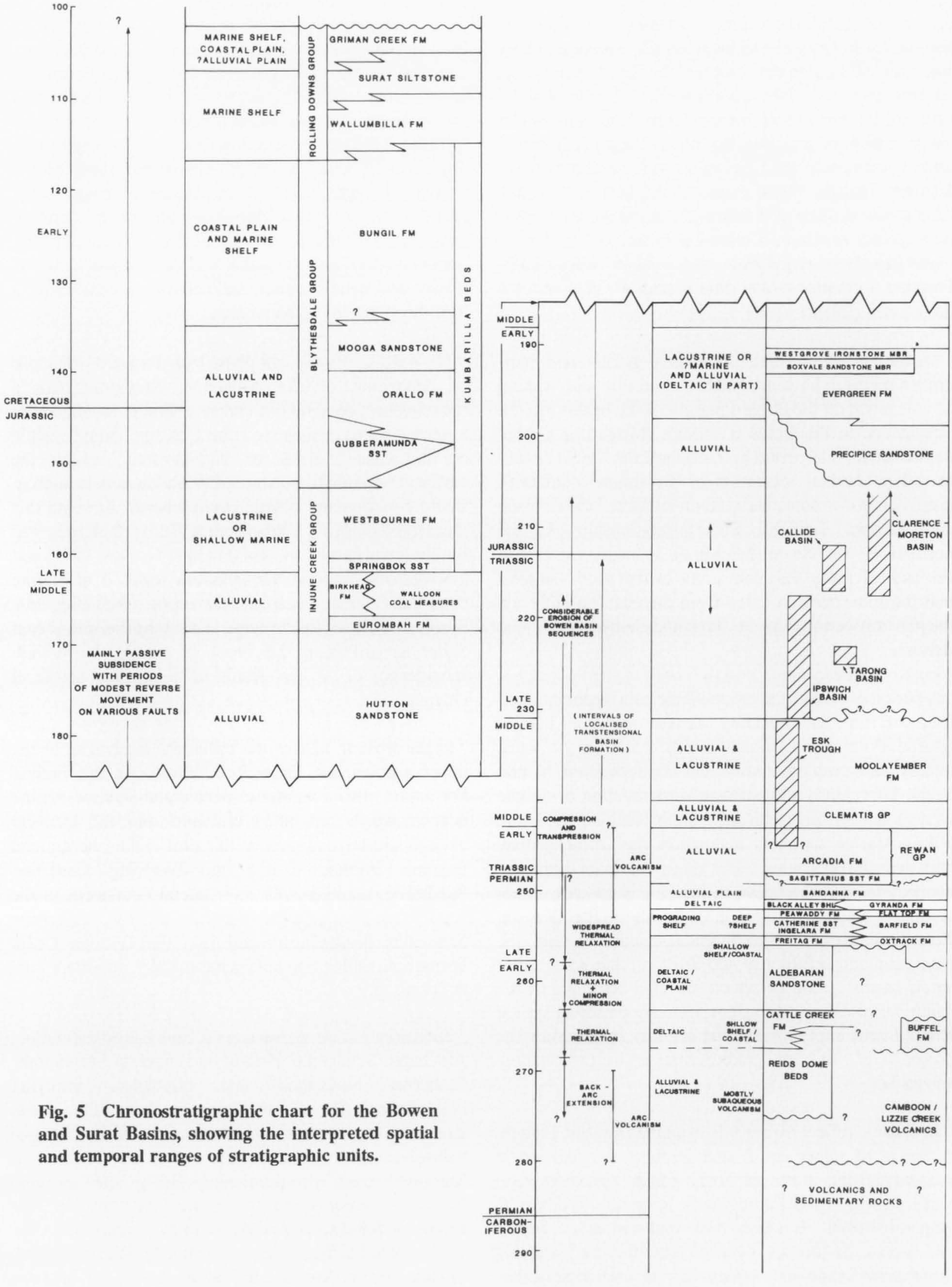

Fig. 5 Chronostratigraphic chart for the Bowen and Surat Basins, showing the interpreted spatial and temporal ranges of stratigraphic units.

The overlying Peawaddy/Flat Top/Tinowon Formation interval is also of shallow-marine to coastal origin, but is more uniformly volcanic lithic in composition. Only on the Roma Shelf are quartzose sandstones recorded, from the Tinowon Formation (Paten & Groves, 1974). The Peawaddy Formation interval has considerable tectonic significance as it marks a basin-wide change to volcanic lithic sediment composition, and precedes infilling of the marine basin. These observations, together with the incoming of abundant primary air-fall tuffs, suggest recommencement of volcanism in the arc system. Initial movement along the Burunga-Leichhardt fault system near Tara is suggested by pronounced thickening of the later Back Creek Group into the "Undulla Embayment" (Fig. 2). Sinistral strike-slip movement is suggested by fold orientations. The Peawaddy Formation and its northern equivalents represent a southward-prograding clastic wedge, which also extended westward across the Springsure Shelf area. This phase of subsidence and sedimentation probably reflects the onset of foreland basin conditions *sensu stricto*.

Following the Peawaddy Formation and its equivalents, the volcanic lithic and in places volcaniclastic Black Alley Shale/ Gyranda Formation interval was deposited in deltaic and coastal plain environments. Minor coals are preserved in the upper part of the succession, and major coal resources are found in the overlying Bandanna Formation, which is interpreted to be of deltaic plain origin. The Bandanna Formation, and its lateral equivalents the Baralaba and Rangal Coal Measures, represent basin-wide establishment of poorly-drained, coal-forming fluvial environments. Each is characterised by a volcanic lithic mineralogy. The introduction of vast amounts of volcanic detritus to the basin, and revitalisation of volcanic activity in the arc, were probably responsible for the final withdrawal of the sea from the Bowen Basin.

The changing tectonic climate during Late Permian times is exemplified in the Gogango Overfold Zone on the central eastern margin of the basin, where compressive deformation led to possibly 50% crustal shortening (Mallett & others, 1988) and uplift of previously deposited sediments, coincident with emplacement of the dominantly fragmental Rockwood Volcanics. These rocks appear localised in the gap between the earlier Camboon and Lizzie Creek volcanic terrains, and may have been responsible for blocking marine access from the east.

The top of the Bandanna Formation and equivalents is conventionally regarded as the Permian-Triassic boundary (Dickens & Malone, 1973), although it is evident that the surface is diachronous, indeed difficult to identify in some areas. The boundary between the coal-bearing Bandanna Formation and overlying Rewan Group is probably conformable in the axial part of the basin, but unconformable at the basin margins. The Rewan Group itself, comprising interbedded, variously coloured clastic sedimentary rocks with a volcanic lithic mineralogy, is interpreted to have formed in partially to well-drained alluvial environments. Recent palynological work (Foster, 1983) suggests that the Permian-Triassic boundary lies within the Rewan Group, and it may correspond to the boundary between the typically drab-coloured Sagittarius Sandstone and overlying reddened Arcadia Formation. Compressive tectonic stresses associated with initiation of thrusting are reflected within the Rewan Group by development of local unconformities (Cosgrove & Mogg, 1985) and the occurrence of alluvial fan conglomerates in the south-east (Cabawin Formation; Exon, 1976). During this period, the eastern basin-bounding faults began to be transformed into high-angle reverse faults by compressive forces, with a component of transcurrent shear.

The Rewan Group is succeeded stratigraphically by the Clematis Group, an alluvial succession of more quartz-rich mineralogy. An increased contribution from the stable craton to the west is implied by this change in mineralogy and by regional palaeocurrent distributions. On the Roma Shelf and farther south, the Clematis Group equivalent in the subsurface, the Showgrounds Sandstone, shows evidence of deposition in standing water, which may have been lacustrine or marine (Butcher, 1984; Schroder, 1988).

The youngest preserved unit in the Bowen Basin is the Moolayember Formation, deposition of which heralded a return to volcanic lithic sediment dominance. The Moolayember Formation was deposited in alluvial environments of deposition, perhaps again with basinal influences to the south (Butcher, 1984; Cosgrove & Mogg, 1985). Coarse conglomerates are preserved along the trend of the Burunga-Leichhardt Fault (Fig. 2). Progressive northward migration of reverse and strike-slip fault movement is indicated by the younging of fault-fringing conglomerates in that direction.

Compressive tectonic deformation reached a climax in Middle to Late Triassic times resulting in regional uplift and erosion of up to 3000 m of strata. Pre-existing normal faults were reversed, to become high-angle reverse faults. During and immediately following this period, right-lateral movement and other adjustments along fault lines created a series of small, intermontane sedimentary basins within the orogenic belt east of the Bowen Basin (Fig. 5), (O'Brien & others, this Bulletin). These basins, which include the Ipswich, Tarong and Callide Basins were associated with high-energy alluvial deposition, synsedimentary tectonism and mostly rhyolitic volcanic activity, and in several cases contain substantial coal resources. The basin fills are not strongly deformed, suggesting formation during the decline of the orogeny.

Renewed subsidence and sediment accumulation characterised Early Jurassic times as the Surat Basin began to form. In contrast to the predominant structural control on subsidence of the earlier Bowen Basin, subsidence in the Surat Basin was areally extensive, largely passive, slow and relatively even (Exon, 1976). A succession of alluvial and lacustrine sequences of Jurassic age accumulated, intermittently under the influence of flexure along pre-existing fault lines (particularly the Hutton-Wallumbilla fault). Mild compressive deformation characterised the Surat basin structuring (Elliott & Brown, 1988).

The Cretaceous sequence in the Surat Basin shows a dominantly marine character in contrast to the underlying Jurassic, which is probably of entirely freshwater origin, and culminates in a widespread marine mudrock interval, the Wallumbilla Formation (Fig. 5). Tectonic events to the east of the Bowen-Surat Basin area led to the intermittent continuation of flexuring along pre-existing fault lines. In Late Cretaceous times, sediment accumulation ceased, and the area was uplifted, giving rise to widespread erosion.

During Palaeocene times, opening of the Coral Sea produced a series of elongate, in many cases graben or half-graben style rift basins over the Bowen-Surat Basin area and to the immediate east. Bounding faults to these basins were of the listric, normal type, and the basins themselves were filled with continental, lacustrine and alluvial sediments with some basaltic volcanics. A later Oligocene-Miocene period of mild compressive structuring has been recognised in both the Bowen/Surat (Brown & others, 1983) and Eromanga (Finlayson & others, 1988) basins. This has been related to the extensive phase of basaltic volcanism which affected eastern Australia at this time. Gentle folding and reactivation of older fault lines characterised this period of mild upheaval. Further basaltic volcanism affected some areas during Pliocene and Pleistocene times. Present-day structural configurations were achieved during these Tertiary events; late Tertiary and Quaternary times have been characterised by crustal stability, and erosion.

DISCUSSION

The balance of evidence strongly supports the notion of a retro-arc foreland basin origin for the Bowen Basin. In particular, the east-west basinal asymmetry (Fig. 3), the abundance of westward-propagating thrust faults and the concentration of compressive deformation at the eastern basin margin, parallelism of the basin with a chain of contemporaneously active volcanic mountains, and sediment infilling patterns are explained satisfactorily only in terms of such a model.

The early extensional phase may be adequately explained in terms of back-arc extension rather than rift-type extension, which is incompatible with simultaneous island arc volcanism at the eastern edge of the basin. Although the mechanism is poorly understood, a similar early extensional phase appears to have occurred in a number of foreland basins (Allen & others, 1986). The evolution of depositional systems from dominantly marine to wholly continental with time, and change in sandstone petrography from craton-derived to orogen-derived through time, are also consistent with a foreland basin origin (Allen & others, 1986; Covey, 1986; Schwab, 1986).

Oblique convergence between the palaeo-Australian and Pacific plates is implied by the northwest trend to many structures, particularly those along the eastern basin margin (Fig. 2). Strike-slip fault motions may well have exercised an active role at various times in the accommodation of stress regimes, but the features described above cannot be explained in terms of a fundamental strike-slip origin.

CONCLUSIONS

This paper provides a review of the stratigraphy, sedimentology and structural features of the Bowen and Surat Basins, and synthesises these data into an interpretation of latest Palaeozoic and Mesozoic geological history for the area. The Bowen Basin was initiated during an Early Permian phase of back-arc extension associated with continental arc volcansim, and subsequently became a "classical" foreland basin during Late Permian to Mid Triassic times, as the palaeo-Australian and -Pacific plates converged. Some of the compressive deformation associated with closure of the basin was accommodated by strike-slip motion along major fault lines. Following a Middle to Late Triassic period of uplift, erosion and local development of intermontane basins through strike-slip fault movement, the Surat basin was initiated in Early Jurassic times. Subsidence was passive, relatively slow and even, and punctuated only intermittently by flexuring along major fault lines. Following cessation of subsidence in mid-Cretaceous times the area was uplifted via mild structuring during Tertiary times to arrive at its present structural configuration.

The interpretations presented herein are based primarily upon review of published literature, and form a basis for re-examination of the geology of the Bowen-Surat area. Research in progress is aimed at critical evaluation of these ideas.

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MESOZOIC BASINS AT THE EASTERN END OF THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT: STRIKE-SLIP FAULTING AND BASIN DEVELOPMENT

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ABSTRACT

Superimposed Late Permian to Jurassic sedimentary basins are set within the New England Orogen at the eastern end of the Eromanga-Brisbane Geoscience Transect. The Mid-Triassic Esk Trough and Late Triassic Ipswich and Tarong Basins overlie unnamed Late Permian to Mid-Triassic basins. The latest Triassic to Jurassic Clarence-Moreton Basin covers large parts of these older basins. Structures identified on seismic reflection profiles show that transtension along dextral strike-slip faults initiated basin formation beneath the Esk Trough as early as Late Permian. In the Early Triassic, strike-slip faulting moved eastward to the site of the present Ipswich Basin and the Logan Sub-basin of the Clarence-Moreton Basin. These more easterly features formed by the interaction of dextral faults so that the basin contains a complex set of basement highs and lows. After transtension ceased, subsidence caused by thermal relaxation led to further deposition in the Esk Trough, Ipswich Basin and finally the Clarence-Moreton Basin. Minor dextral strike-slip continued producing positive flower structures and thrusts in sediments overlying the major faults. Movement ceased some time in the Cretaceous when the locus of movement shifted east to sinistral faults associated with rifting on the eastern Australian margin and the opening of the Tasman Sea.

INTRODUCTION

The eastern end of the Eromanga-Brisbane Geoscience Transect crosses a series of superimposed sedimentary basins overlying the New England Orogen. These are the Clarence-Moreton Basin overlying the Esk Trough and Ipswich Basin, which themselves overlie two unnamed basins (Figs. 1, 2, Korsch & others, 1989; O'Brien & others, in prep.). This set of related basins contains a sedimentary record that extends from the ?Late Permian to the Cretaceous.

It has long been suspected that strike-slip faulting influenced basin development in this area. Leven (1977), Evans & Roberts (1980), and Harrington & Korsch (1985a) suggested that the Esk Trough originated as a pull-apart basin on a dextral strike-slip fault. Popescu (1984) explained the pattern of surface structures in these basins as the result of dextral movement on the Demon Fault and sinistral movement on a fault along the eastern Clarence-Moreton Basin margin. Korsch & others (1989) and O'Brien & others (in prep.) examined the geometry and structure of the Esk, Ipswich and Clarence-Moreton basins, using seismic reflection data, and concluded that the basins developed by dextral transtension commencing as early as Late Permian. This paper summarizes the results of these last two studies.

MAJOR STRUCTURAL ELEMENTS

The sedimentary basins of easternmost Australia are surrounded and underlain by rocks of the New England Orogen which consist of a series of north-northwest trending basement blocks on the northern and northeastern side of the Clarence-Moreton Basin (Fig.1, Murray, this volume). The southeastern part of the Clarence-Moreton Basin rests on the oroclinally folded rocks of the Texas and Coffs Harbour Blocks. The major structural subdivisions of the Mesozoic basins are, from west to east (Fig.1), the Kumbarilla Ridge, the Cecil Plains Sub-basin of the Clarence-Moreton Basin, the Tarong Basin, the Gatton Arch, the Esk Trough, the Laidley Sub-basin of the Clarence-Moreton Basin, the Great Moreton Fault System, the South Moreton Anticline, the Ipswich Basin, and the Logan Sub-basin of the Clarence-Moreton Basin.

The Kumbarilla Ridge is a very broad basement high separating Early Jurassic depocentres in the Cecil Plains Sub-basin and the Surat Basin (Day & others, 1974). The Gatton Arch is a similar basement ridge over which Clarence-Moreton Basin sediments are folded and relatively thin (Gray, 1975).

The Great Moreton Fault System (Cranfield & others, 1976) is a network of braided faults that includes the Eastern Border Fault of the Esk Trough and the West Ipswich Fault forming the western edge of the Ipswich Basin (Cranfield & others, 1976). The West Ipswich Fault also forms the western margin of the South

Moreton Anticline (Fig. 1), and continues south-southwest for at least 100 km (O'Brien & Wells, 1985). The South Moreton Anticline is a broad structural high in basement over which Clarence-Moreton Basin sediments thin and are folded (Figs. 2, 4, 5). In the north, it is usually depicted as a broad anticline with minor faulting on either side (Cranfield & others, 1976, 1981), whereas its southern extension is depicted as a horst, known as the Richmond Range Horst (Ties & others, 1985). Sediments of the Esk Trough are confined to the west of the South Moreton Anticline.

The Mesozoic basins of easternmost Australia can be divided into three provinces (Fig.1): (1) the Cecil Plains Sub-basin in the west, (2) the Esk Trough overlain by the Laidley Sub-basin of the Clarence-Moreton Basin, and (3) the South Moreton Anticline and the Ipswich Basin overlain by the Logan Sub-basin of the Clarence-Moreton Basin in the east. This sub-basin is covered by large areas of Tertiary volcanics along the Queensland - New South Wales border so that outcrop and subsurface data are sparse. Therefore, the northern and southern parts of Logan Sub-basin will be discussed separately.

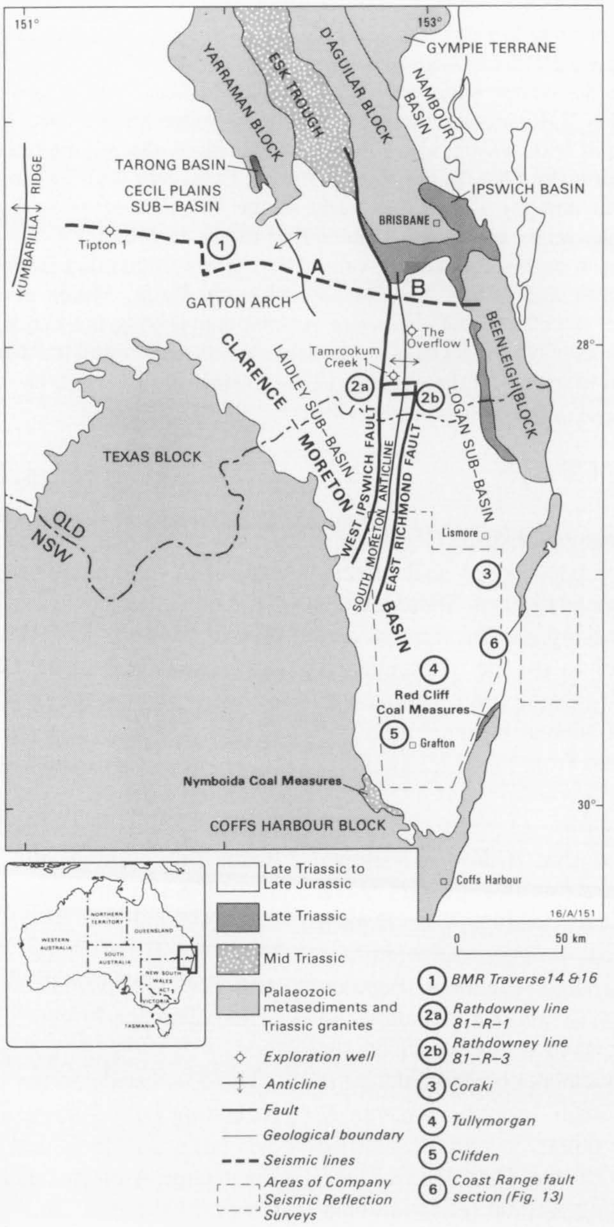

Fig. 1 Simplified geological map of easternmost Australia showing some of the major geological elements of the Clarence-Moreton Basin and location of surrounding geological units. The Red Cliff Coal Measures are included in the Ipswich Basin. BMR seismic lines are shown, as are lines of the Rathdowney seismic survey. Parts of the Logan Sub-basin covered by more extensive company surveys are also shown. Note that the well Tamrookum Creek No.1 is located on seismic line 81-R-1 (3a). Tertiary and Quaternary rocks have been omitted for clarity. A - B is the location of the seismic section shown in Figure 3.

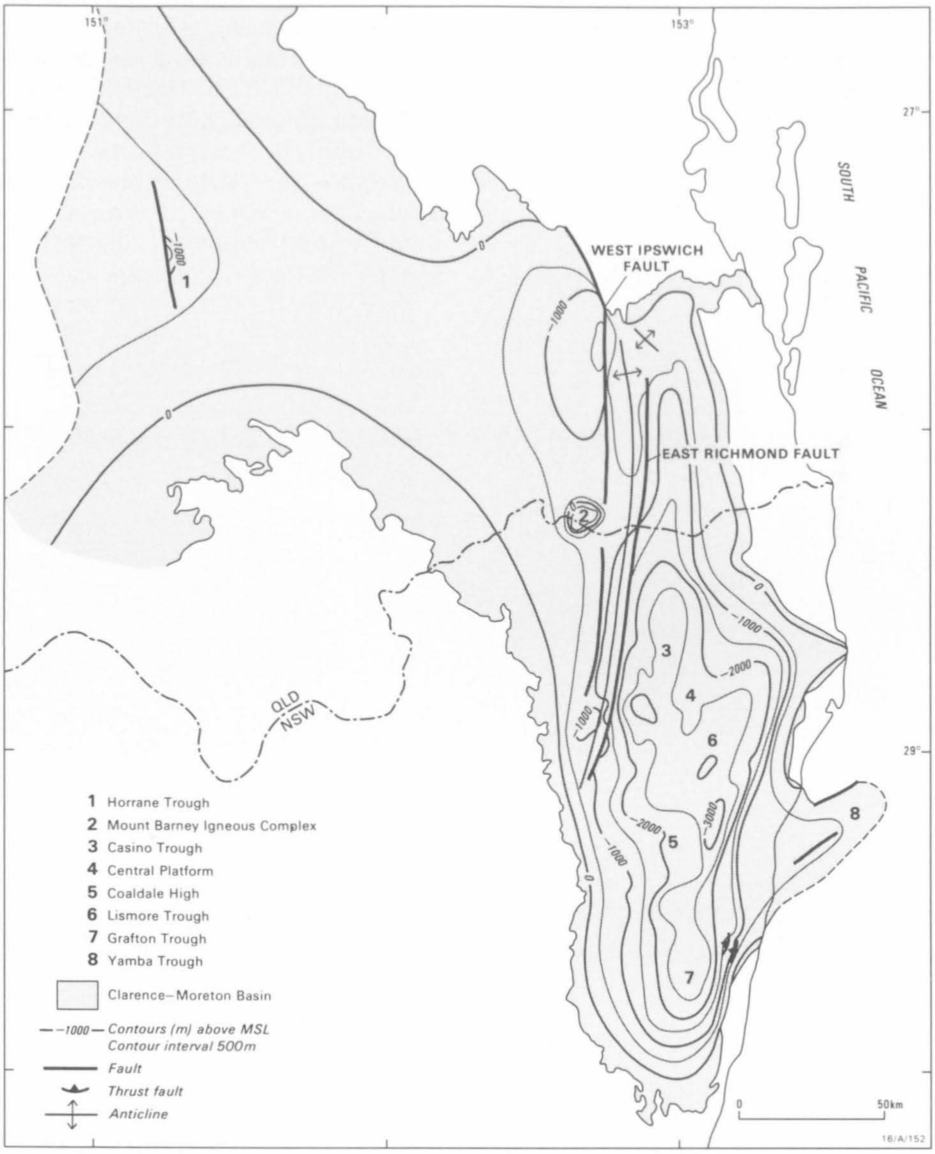

Fig. 2 Generalized structure contour map of the base of the Clarence-Moreton Basin showing features within the major sub-basins. A more detailed (1:500 000 scale) map is available as part of the Clarence-Moreton Basin Map folio (Wells & O'Brien, in prep.).

SEISMIC STRATIGRAPHY AND STRUCTURE

Esk Trough - Laidley Sub-basin

Seismic record sections reveals four major sequences (Fig. 3):

Sequence 1 is seismic basement and is probably Early to Mid- Permian marine sediments and volcanics (Korsch & others, 1989).

Sequence 2 unconformably overlies basement and underlies sediments of the Esk Trough (Fig. 3). Its position between the Mid-Permian basement and Mid-Triassic Esk Trough sediments indicates a Late Permian to Early Triassic age. Reflections in this sequence define an asymmetrical syncline with dips of 7° on the western limb and up to 19° on the eastern limb. The succession laps on to basement on the western side but is truncated in the east by the West Ipswich Fault. In the hinge of the syncline, a maximum thickness of about 3900 m is estimated using Dix corrected interval velocities.

Sequence 3 unconformably overlies sequence 2 (Fig. 3) and consists of the Mid-Triassic Toogoolawah Group, the fill of the Esk Trough. The Toogoolawah Group consists of nonmarine polymict and volcanoclastic conglomerate, sandstone and shale with interbedded andesitic lavas and tuffs (Cranfield & others, 1976). These rocks reach a maximum thickness of about 2350 m in the hinge of the syncline and wedge out gradually on the western side. On the eastern side, the sequence thins rapidly towards the West Ipswich Fault. A thin wedge of these sediments may extend across the fault zone.

Sequence 4 consists of non-marine Clarence-Moreton Basin sediments. They unconformably overlie the Toogoolawah Group and reach a maximum thickness of about 1650 m in the centre of the syncline. To the west they thin gradually on to the Gatton Arch, where the sequence is somewhat condensed. To the east they become abruptly thinner across the West Ipswich Fault and South Moreton Anticline because of onlap and recent erosion (Fig.3). At the surface, subvertical dips in Clarence-Moreton sediments near the West Ipswich Fault method of Gibbs (1983), should be about 11 s TWT (about 33 km depth), assuming a basement velocity of 6 km sec^{-1}. This is about the depth of the reflection Moho beneath the Esk Trough. To extend this deep, the fault must cut through the brittle-ductile transition. It is more realistic to expect the detachment to flatten within the ductile zone. The steep dip of the fault, and lack of evidence for a detachment, suggest that the rift did not form by pure extension in the plane of the seismic section.

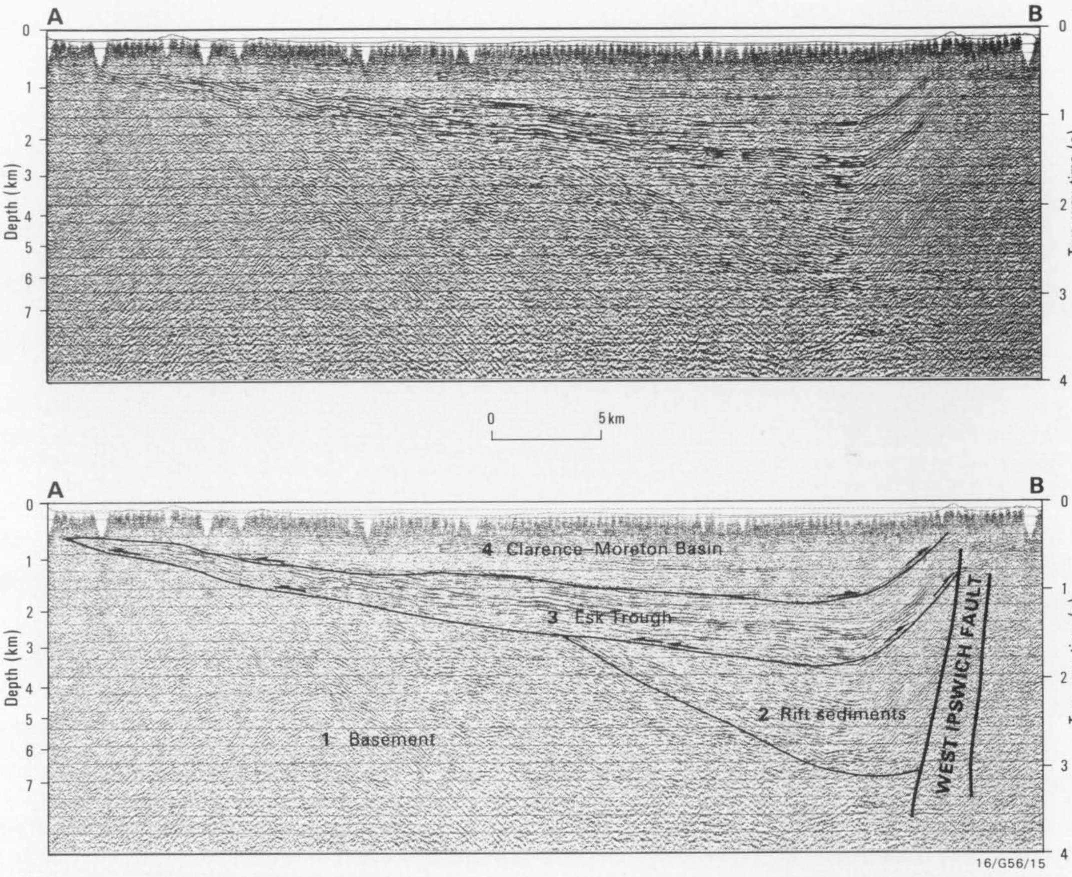

Fig. 3 Unmigrated seismic reflection profile across the Esk Trough. Profile shown is the AB portion of BMR seismic traverse 16 (Fig. 1), a six-fold common depth point, 20 s record length, dynamite source, split-spread with 2 km far offset profile. The seismic sequences are: (1) basement (pre-rift Cressbrook Creek Group); (2) rift sediments and volcanics; (3) Esk Trough sequence (Toogoolawah Group); and (4) Clarence-Moreton Basin sequence (Bundamba Group and Walloon Coal Measures). Arrows indicate direction of onlap. The arbitrary seismic datum is approximately 250 m above sea level; note that it is above ground elevation which is shown by an irregular line at the top of the section. The depth scale is from seismic datum.

indicate late fault reactivation and tectonic inversion.

The structures and sediments of the Esk Trough have the pronounced asymmetry typical of half-grabens and most rift basins (Bally, 1982). This implies that it developed by extension controlled by a west-dipping normal fault (West Ipswich Fault of the Great Moreton fault system). However, a detachment fault has not been detected extending to depth from the West Ipswich Fault. The depth to any detachment, calculated using the

To accommodate the geometry in the seismic section (Fig. 3), a horizontal (strike-slip) component of movement is required parallel to the trough margins. However, the seismic data and fault patterns do not indicate that the basin is a pull-apart basin formed between two parallel strike-slip faults or within a major bend in a single strike-slip fault (Crowell, 1974). Extension was therefore oblique to the trough margin (West Ipswich Fault), and the basin is transtensional in character (Harland, 1971). This suggests that the faults of the Great Moreton system are surface traces of a major strike-slip fault system at depth. Low-angle reverse faults in the system are probably part of positive flower structures on the crest of the strike-slip system, as are associated *en echelon* folds (Fig. 2).

The geometry of the southern end of the Esk Trough and its fill in the region of the transect can be used to determine the sense of movement on the West Ipswich Fault. Company seismic data, well logs, and BMR seismic data show that the sequence thickens and dips to the south, and then terminates abruptly. Hence, the extensional bounding fault must be located south of BMR Line 16 and probably dips to the north. Movement on the West Ipswich Fault was therefore dextral during formation of the Esk Trough. This interpretation of movement sense is consistent with the northwest-southeast trends of *en echelon* fold axes on the crest of the south Moreton Anticline (Korsch & others, 1989).

South Moreton Anticline - Northern Logan Sub-Basin

BMR seismic Line 16 crossing the Ipswich Basin and South Moreton Anticline displays one seismic sequence only, most of which represents Late Triassic Ipswich Coal Measures and the underlying felsic to basaltic Chillingham Volcanics (see fig. 2 in Korsch & others, 1986); the Clarence-Moreton sequence is relatively thin in this area. The Ipswich Coal Measures - Chillingham Volcanics package appears to be 2800 m thick; however, on displays of the deep crustal data (Korsch & others, 1986) events at 1.8 s suggest that there could be up to 4000 m of sediments and volcanics. The Ipswich Basin seismic sequence thins to the west onto the South Moreton Anticline and is truncated by the West Ipswich

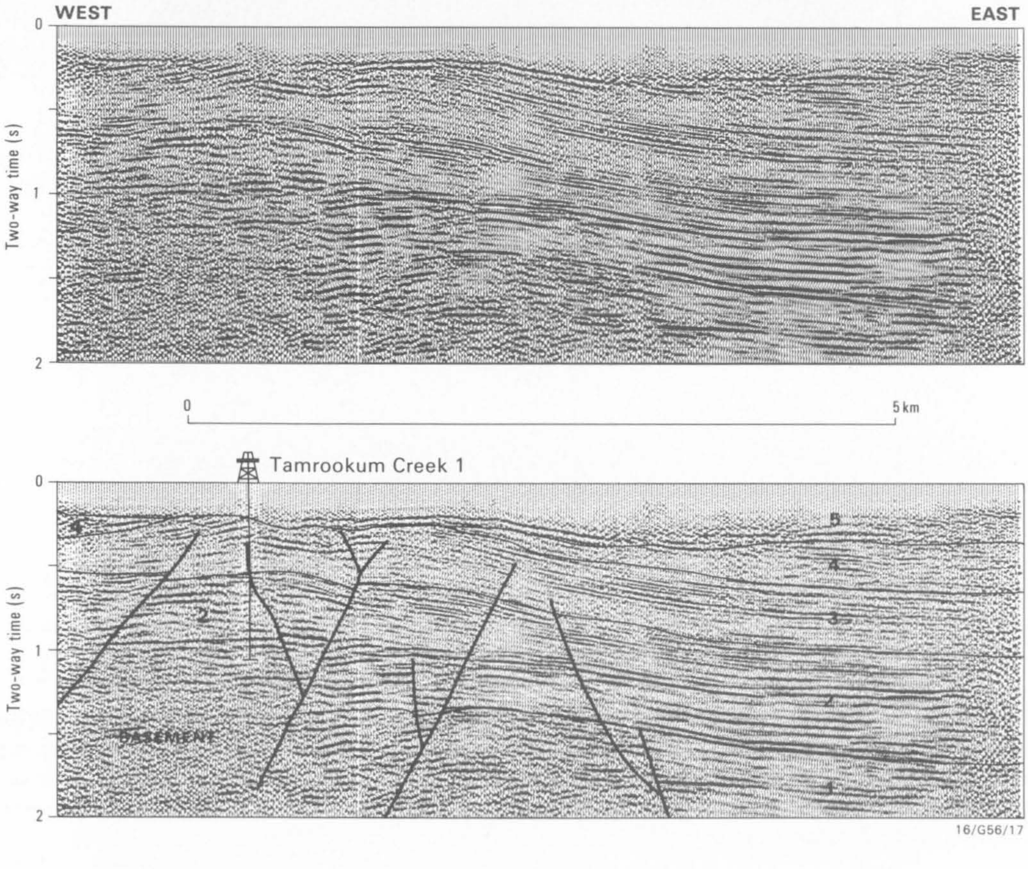

Fig. 4 Interpretation of unmigrated Line 81-R-1 of the Rathdowney Seismic Survey (Milner & Milner 1981). The seismic sequences are: (1) a sequence of sediments and volcanics that is truncated by the fault on the eastern edge of the South Moreton Anticline; (2) the Chillingham Volcanics and some sediments - this sequence thins onto basement in the South Moreton Anticline; (3) the Ipswich Coal Measures; (4) the Ipswich Coal Measures and/or Raceview Formation of the Clarence-Moreton Basin; and (5) the Ripley Road Sandstone and overlying units of the Clarence-Moreton Basin. Sequences (1) and (2) are interpreted as "rift" fill sediments and volcanics. Also shown is position of Tamrookum Creek No. 1 well which intersected 187 m (to 0.269 s TWT) of Ripley Road Sandstone (sequence 5), 588 m (to 0.591 s TWT) of Ipswich Coal Measures (sequence 3) and 1159 m (total depth 1934 m, 1.18 s TWT) of volcanics considered to be Chillingham Volcanics and possibly part of an older volcanic sequence (Project Oil Exploration Limited 1982).

Fault. Ipswich Coal Measures are absent west of the West Ipswich Fault. The sequence also gradually laps on to the Beenleigh Terrane to the east.

The Rathdowney Seismic Survey (Milner & Milner, 1981) provides more detail on relationships on, and east of, the South Moreton Anticline (Fig.1). Line 81-R-1 (Fig.4) displays five seismic sequences, three of which can be tied directly to the Tamrookum Creek No.1 well. Sequence 5 consists of the Ripley Road Sandstone of the Clarence-Moreton Basin. It rests unconformably on the Ipswich Coal Measures (Sequence 3) on the crest of the anticline at Tamrookum Creek No.1 well, but to the east unconformably overlies another sequence (Sequence 4) which is above the identified Ipswich Coal Measures sequence. Sequence 4 may be either the upper part of the Ipswich Coal Measures or the lowest units of the Clarence-Moreton Basin. The Ipswich Coal Measures conformably overlie Chillingham Volcanics (Sequence 2). On Line 81-R-3 (Fig. 5), the Ipswich Coal Measures onlap basement. Part of the Chillingham Volcanics, with a seismic character similar to the Ipswich Coal Measures, gradually onlaps basement, but the bulk of the volcanics form a wedge beneath the Ipswich Coal Measures that thins rapidly on to basement and is faulted (Figs. 4, 5).

Beneath the Chillingham Volcanics (Sequence 2) on the eastern half of Line 81-R-1 (Fig. 4), another seismic sequence (Sequence 1) terminates against basement beneath the eastern edge of the South Moreton Anticline. Sequence 1 may consist of more Chillingham Volcanics, or may include older sediments and volcanics similar to the Toogoolawah Group in the Esk Trough. Sequence 1 is interpreted as the lower part of a rift sequence that has the Chillingham Volcanics (Sequence 2) as its upper part.

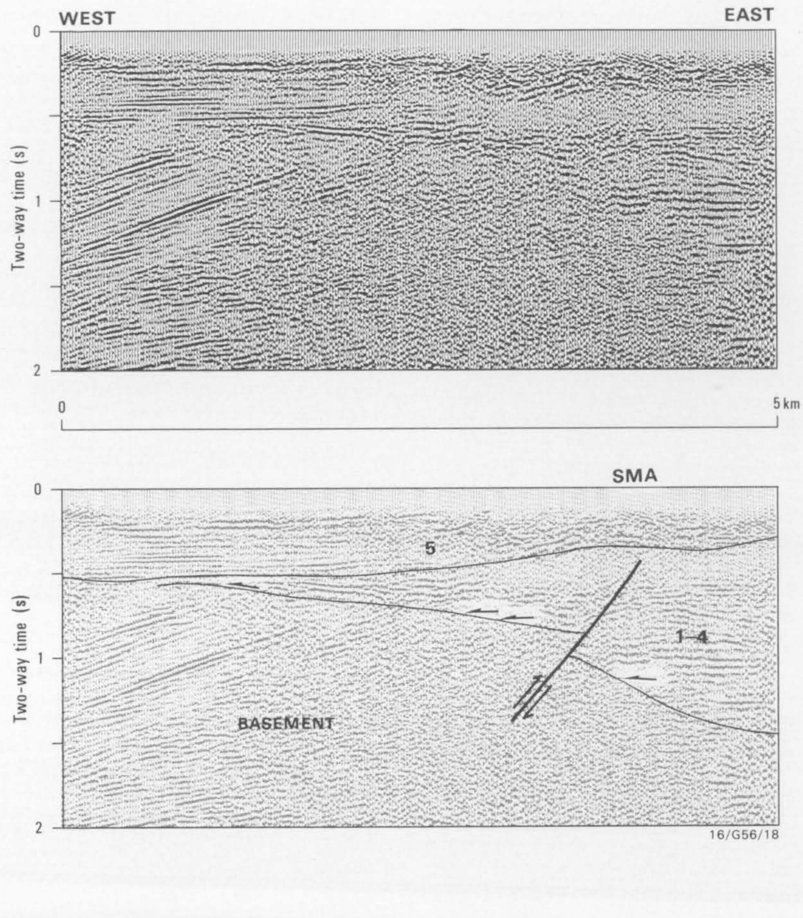

Fig. 5 Rathdowney Seismic Survey Line 81-R-3 showing thinning by onlap (direction of arrows) of sequences 1 to 4 onto basement on the eastern side of the South Moreton Anticline. Sequence numbering is the same as Figure 3 although deformation prevents precise separation of sequences 1 to 4 in this area. SMA refers to the location of the surface expression of the anticline (positive flower structure) in the Clarence-Moreton sequence above the eastern margin of the basement high.

The considerable volume of volcanics east of the South Moreton Anticline implies thinning of the crust and an extensional mechanism for basin formation. Both seismic sections (Figs. 4, 5) indicate onlap of sequences 1 to 4 on to the eastern margin of the South Moreton Anticline, with the total sediment package thickening to the east. These inferences imply that the major bounding fault must be located on the eastern edge of the South

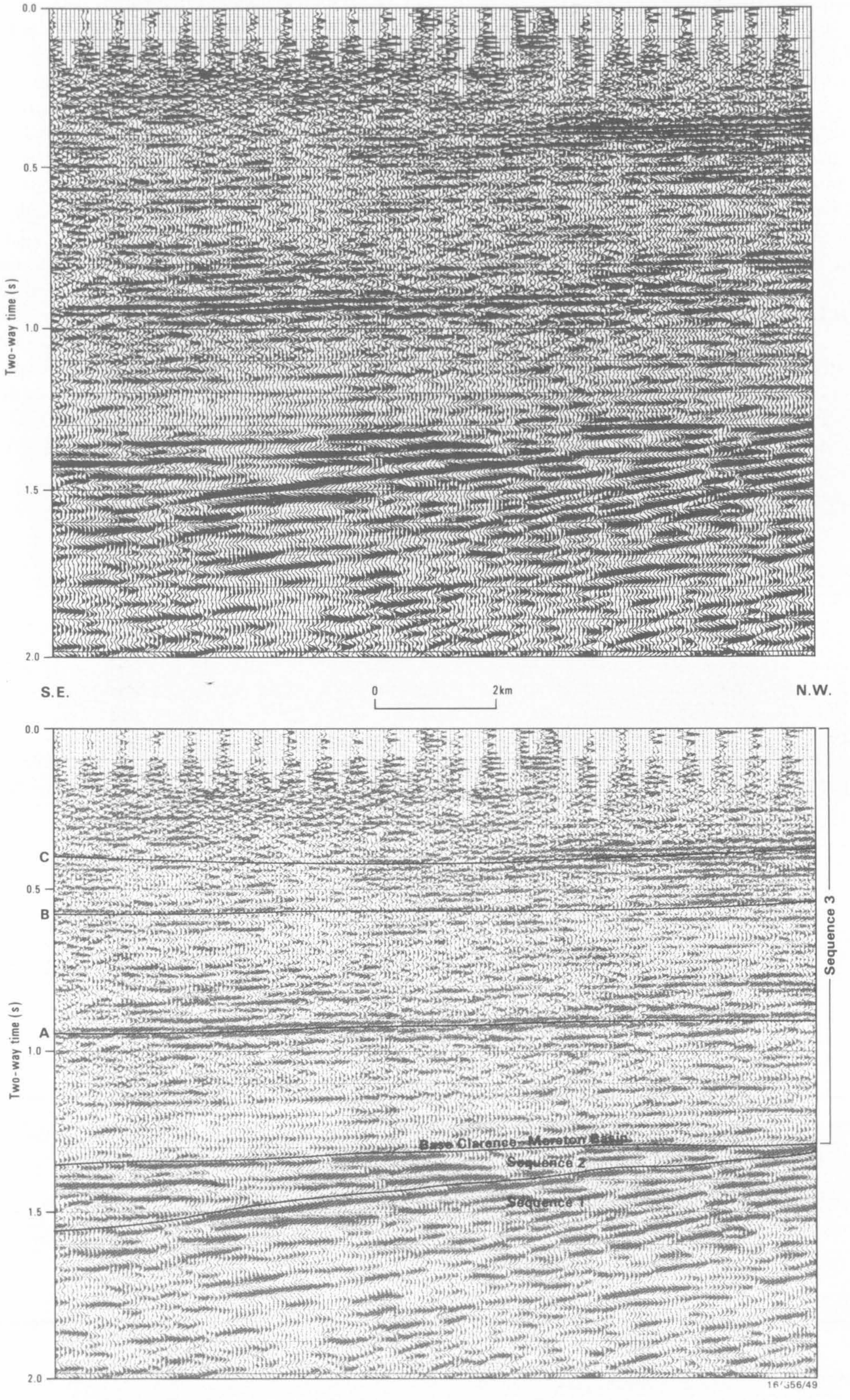

Fig. 6 Seismic stratigraphy of the Logan Sub-basin. Sequences are: (1) interbedded sandstone, conglomerate, shale, coal and thick felsic volcanics. It is probably equivalent to Sequence (1) or (2) on the Rathdowney seismic sections (Figs. 4 & 5); (2) probable Ipswich Coal Measures; and (3) Clarence-Moreton Basin sediments. Horizons within the Clarence-Moreton Basin sediments are: (A) near top Gatton Sandstone, (B) near top Koukandowie formation, and (C) near top Walloon Coal Measures.

Moreton Anticline. The bounding fault could be that at the western limit of sequence 1 (Fig. 4).

The Rathdowney seismic lines provide additional evidence for the interpretation of the South Moreton Anticline as a positive flower structure. Line 81-R-1 (Fig. 4) displays a series of steep, branching faults with small normal and reverse displacements, whereas the parallel Line 81-R-3 (Fig. 5) displays a reverse fault dipping west. Such rapid, along-strike changes in structural style are typical of strike-slip fault systems (Harding, 1985; Gibbs, 1987).

Southern Logan Sub-basin

Three seismic sequences can be identified beneath the southern Logan Sub-basin (Fig 6). The upper one is the Clarence-Moreton Basin fill which has several mappable horizons in it. Beneath the Clarence-Moreton Basin sediments are two other seismic sequences. One thin, well-defined sequence (Sequence 2, Fig. 6) laps out on basement highs and is overlain by the Clarence-Moreton Basin with a subtle unconformity. This sequence overlies a deeper one (Sequence 1, Fig. 6) with pronounce truncation of reflectors. The deepest sequence is structurally complex with discontinuous reflectors. The lower sequence was probably penetrated by one petroleum well which encountered interbedded sandstone, conglomerate, shale, coal and thick felsic volcanics. It is probably equivalent to Sequence 1 or 2 in the northern Logan Sub-basin and thus partly equivalent to the Chillingham Volcanics that are thought to be of mid Triassic age (Korsch & others, 1989). Sequence 2 (Fig. 6) is probably equivalent to Ipswich Coal Measures of Late Triassic age. The structure contour map of the base of the Clarence-Moreton Basin (Fig. 2) shows the main features of the Logan Sub-basin. Unlike the simple structure of the Laidley Sub-basin and Esk Trough, the Logan Sub-basin contains a series of troughs and ridges named by O'Brien & others (in prep.) (Fig. 2).

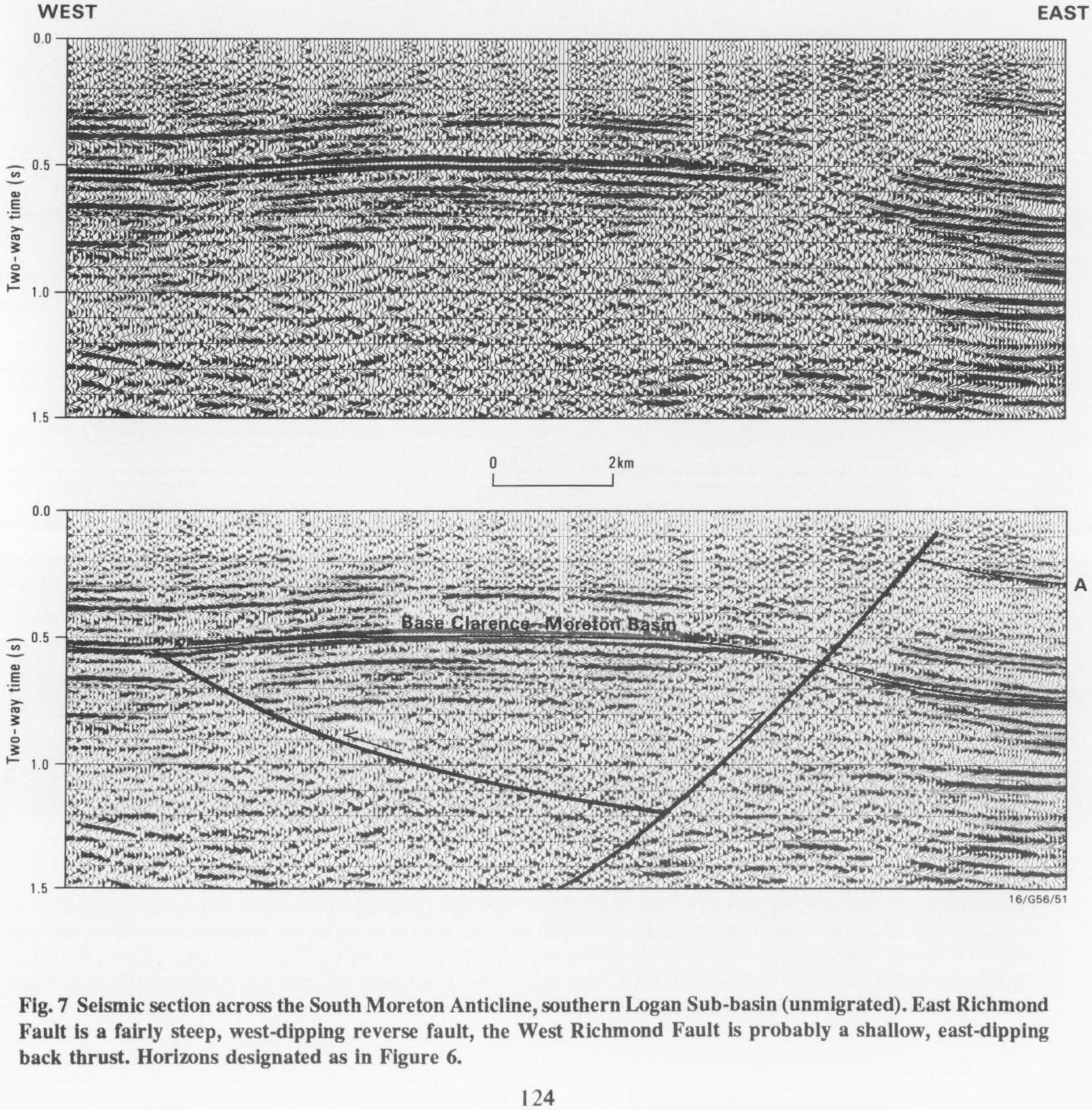

Fig. 7 Seismic section across the South Moreton Anticline, southern Logan Sub-basin (unmigrated). East Richmond Fault is a fairly steep, west-dipping reverse fault, the West Richmond Fault is probably a shallow, east-dipping back thrust. Horizons designated as in Figure 6.

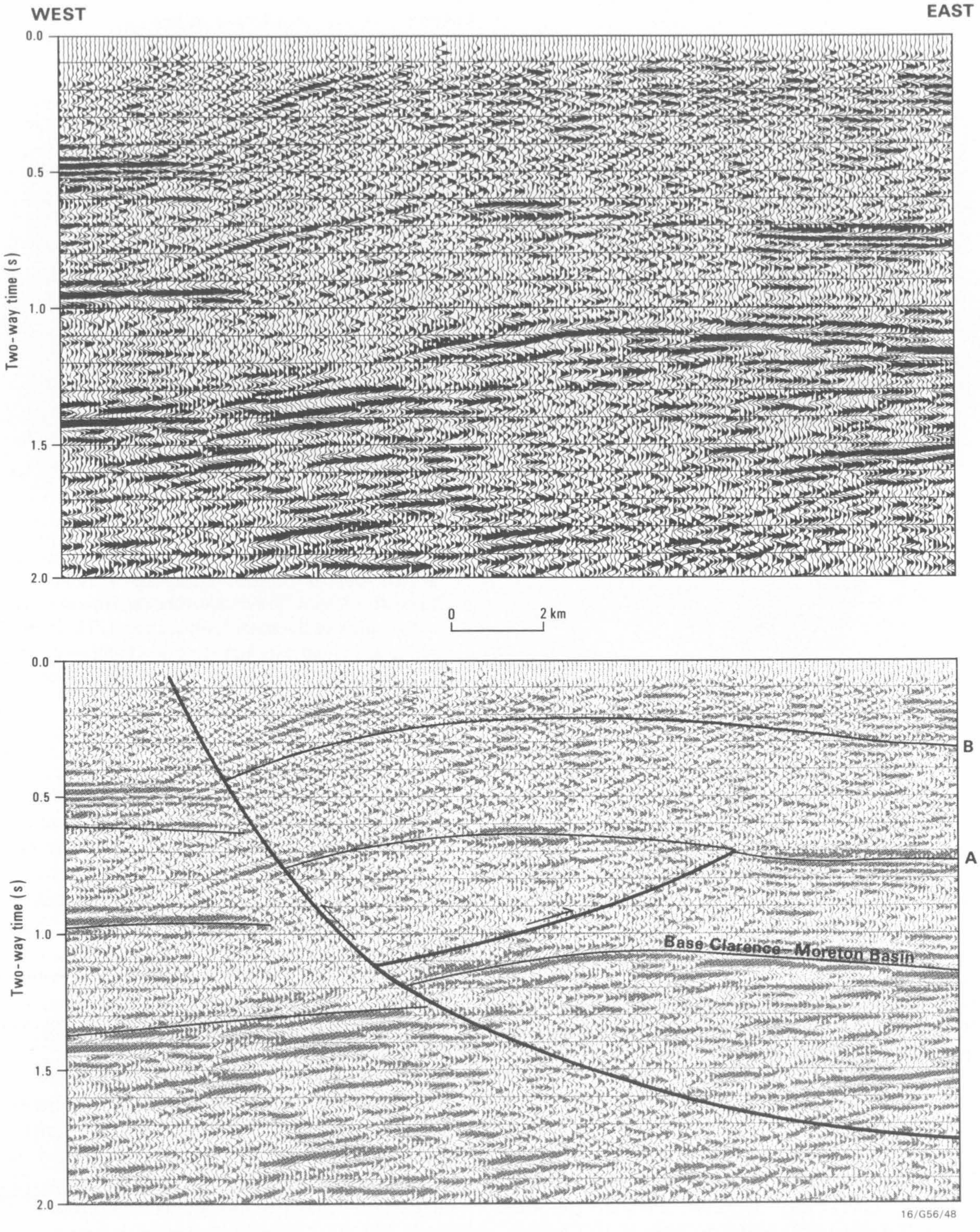

Fig. 8 East-dipping thrust with west-dipping backthrust near Coraki (unmigrated section). Hanging wall anticlines affect most horizons.

The southern extension of the South Moreton Anticline features a west-dipping reverse fault (East Richmond Fault; Fig. 7) that changes to a monocline in shallow horizons along strike. Hanging-wall anticlines are present on the crest of the structure. Backthrusts linked to the East Richmond Fault are visible in places (Fig. 7).

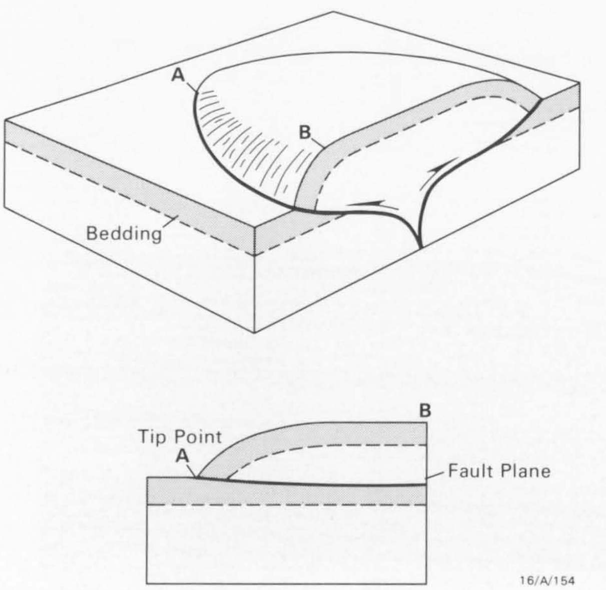

Fig. 9 Interpretation of opposing thrust faults in the Coraki area as a flower structure. The main faults steepen at depth and join a vertical strike-slip fault. They end in tip points, so that in sections along the fault plane strike, the fault appears as a roughly horizontal line ending at a monoclinal

Surface anticlines occur along a south-southwest line crossing the southern Logan Sub-basin (O'Brien & others, in prep.). Seismic sections across these anticlines reveal them as hanging wall anticlines on reverse faults (Fig. 8). These faults end at tip points in the Clarence-Moreton sequence and have arcuate trends. They are commonly arranged in groups with east-dipping faults opposed by west-dipping faults a few kilometres to the east (O'Brien & others, in prep.)

This arrangement of opposing thrusts strongly suggests flower structures over a deep strike-slip fault (Fig. 9; Harding, 1985). The most likely geometry is that the thrusts steepen down into the main fault below 2 s (Fig. 9, Gibbs, 1987). Poor reflector continuity prevents tracing of the faults far below the base of the Clarence-Moreton Basin. A line drawn between the sets of flower structures also corresponds to several near-vertical faults cutting Clarence-Moreton sediments (O'Brien & others, in prep.). Their orientation and the north-south orientation of the hangingwall anticlines suggests that the flower structures are developed on restraining bends along a dextral strike-slip fault (Fig. 10; Crowell, 1974; Christie-Blick & Biddle, 1985). O'Brien & others (in prep.) named it the Coraki Fault (Fig. 12).

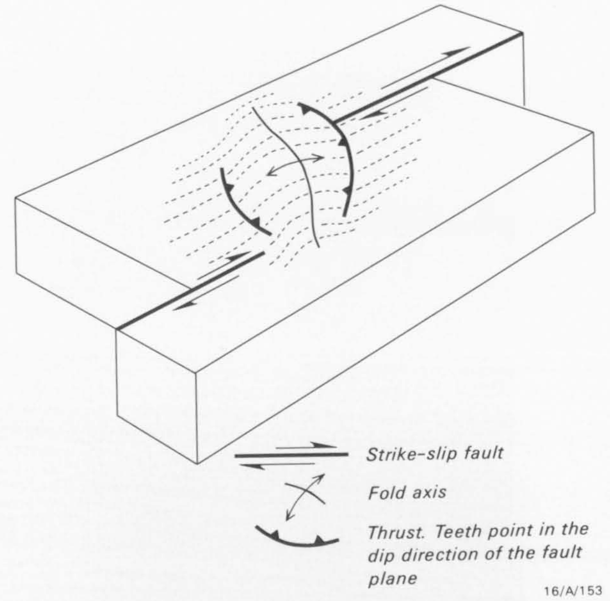

Fig. 10 Folding and faulting on a restraining bend or side-step of dextral strike-slip fault (after Crowell, 1974). A flower structure develops oblique to the strike of the main fault segments. The angle between the fold axis and the main faults depends on the amount of strain and the degree of overlap between the fault segments.

Seismic lines along the eastern basin margin reveal a series of west dipping thrusts with hanging-wall anticlines in places (e.g. Fig. 11). None of the lines examined show opposing thrusts similar to those along the Coraki Fault but this might be because few of the seismic lines extend far enough west. Alternatively, the stress field may have been such that the fault was transpressional along most of its length (Fig. 12). We therefore tentatively interpret these features as part of another dextral strike-slip fault zone running about 2 to 6 km west of the present eastern basin margin. O'Brien & others (in prep.) refer to this as the Coast Range Fault after McElroy (1962).

An offshore seismic surveys (Alin, 1970) provides evidence supporting a transpressional interpretation for the eastern basin margin. The survey detected east-northeast trending structures, including a graben filled with probable Triassic to Jurassic sediments (the Yamba Trough, Fig. 2). These faults can be interpreted as a series of splay faults off the Coast Range Fault with the graben formed because of dextral movement on the main fault (Fig. 12; Harding & others, 1985; Christie-Blick & Biddle, 1985).

Cecil Plains Sub-basin

The Cecil Plains Sub-basin contains two sequences (Fig. 13). The lower (Sequence 1, Fig. 13) is thickest in a fault-bounded half-graben, named the Horrane Trough (about 2500 m). The Horrane Trough is about 20 km across and extends at least 35 km parallel to its boundary fault which strikes slightly west of north (Fig. 2). This boundary fault on the western side of the trough dips steeply east (Fig. 13). Cecil Plains-1 well penetrated the top 380 m of the trough fill and encountered reddish pebble conglomerate. Wells drilled outside the Horrane Trough penetrate thin reddish shales and intermediate volcanics. Sequence 2 consists of relatively thin Clarence-Moreton sediments. Clarence-Moreton sediments immediately overlying Sequence 1 contain Late Triassic palynomorphs. The Horrane Trough half-graben is similar to that of the Tarong beds (Flood & Garces, 1986) farther north, which are Late Triassic in age, equivalent to the Ipswich Coal Measures (Carnian; Day & others, 1974). This suggests a probable Late Triassic age for Sequence 1.

The Horrane Trough underlies the thickest Clarence-Moreton section in the Cecil Plains Sub-basin, which is a broad, relatively undeformed depression from the

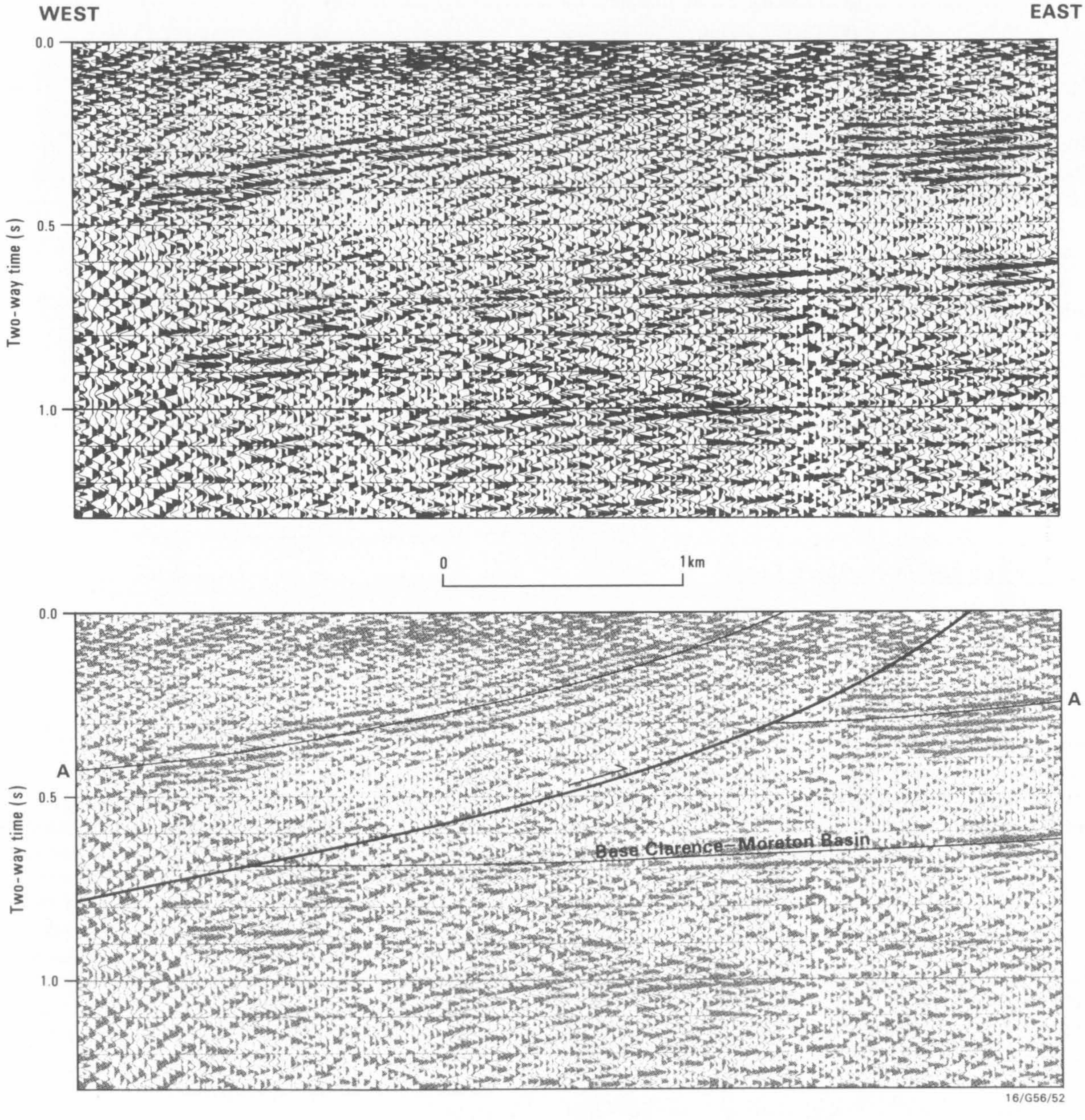

Fig. 11 West-dipping thrust from near the eastern basin margin. Such faults result from dextral transpression along a strike-slip zone near the basin margin (Coast Range Fault).

Gatton Arch in the east to the Surat Basin in the west (Figs. 1 & 2). A few gentle folds mapped in outcrop on the northeastern edge of the Cecil Plains Sub-basin (Cranfield & others, 1976) are the only signs of deformation in the Clarence-Moreton section.

SUBSIDENCE HISTORY

Poor age control on previously unknown sediments beneath the Esk Trough (Fig. 3, Sequence 2), on the Ipswich Coal Measures, and on the sediments of the Esk Trough and Clarence-Moreton Basin, only enabled the construction of approximate subsidence curves (Korsch & other, 1989). Curve A (Fig. 14) uses data from the BMR seismic line across the Laidley Sub-basin to estimate sequence boundary depths. The shape of the curve is typical for a sedimentary basin initiated by extension followed by a protracted period of subsidence caused by thermal cooling of the lithosphere (e.g. McKenzie, 1978; Dewey, 1982; Chadwick, 1985). The curve indicates that the most rapid subsidence took place during initial sedimentation and that the subsidence rate decreased with time, the subsidence rate of the Clarence-Moreton Basin being less than that of the Esk Trough.

A raw subsidence curve (Fig. 14, curve B) for the Logan Sub-basin, constructed using depths obtained from seismic and well data, has a pattern similar to that west of the South Moreton Anticline, again suggesting that deposition was controlled mainly by thermal subsidence after extension. This pattern is repeated in the Cecil Plains Sub-basin (Curve C), except that extension started later and the total amount of subsidence was much less.

BASIN EVOLUTION

Laidley Sub-basin - Northern Logan Sub-basin

The Esk Trough, Ipswich Basin and Clarence-Moreton Basin represent a genetically related, linked system of basins. These basins were controlled by strike-slip faulting and show subsidence curves indicative of a period of crustal extension followed by thermal relaxation. Fault geometries and the mechanics of basin formation can be invoked to explain details of basin geometry and make predictions about deep basin structures.

In the Laidley Sub-basin, the Gatton Arch remained a relatively high area through the transtensional and thermal relaxation phases. Its position and history suggest that it is the peripheral bulge caused by isostatic adjustment of the lithosphere in response to loading by the basin fill (Beaumont, 1978). Clarence-Moreton sediments overlap Esk Trough sediments on the Gatton Arch because, with time, peripheral bulges decrease in amplitude, increase in wavelength and move away from the depocentre (Beaumont, 1978). Esk Trough and Clarence-Moreton sediments also thin across the South Moreton Anticline. This probably resulted from continued faulting along the anticline rather than it acting as a peripheral bulge, because, in asymmetrical extension, the peripheral bulge forms on the upper plate (hanging-wall block) (Houseman, 1987). In this case, the upper plate is the western side of the West Ipswich Fault.

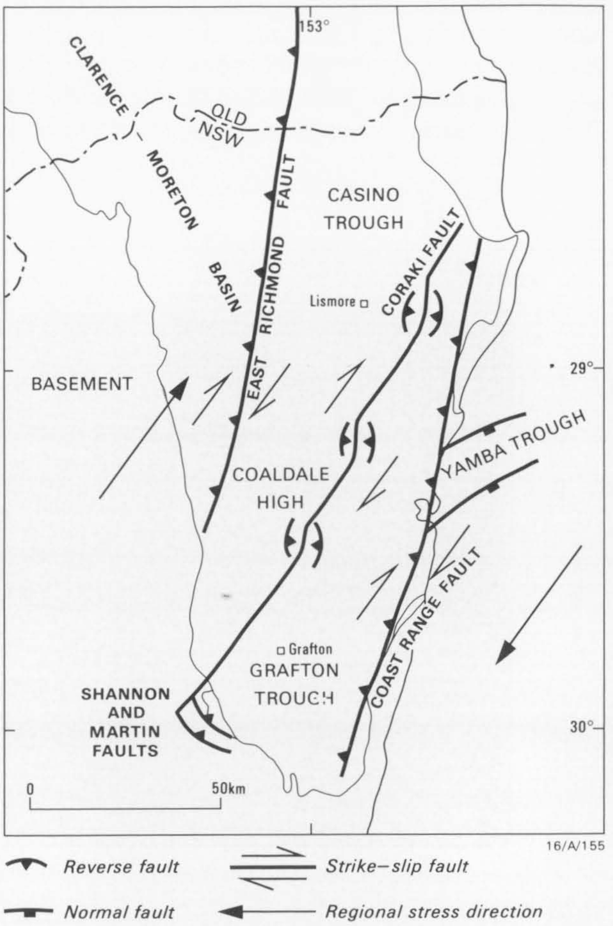

Fig. 12 Tectonic model for the southern Logan Sub-basin. West-dipping thrusts on the East Richmond and Coast Range Faults indicate transpressional movement; the discrete flower structures on the Coraki Fault suggest dextral movement with thrusting and folding on restraining bends or side-steps. The regional stress direction is that necessary to produce the sense of movement seen on the major faults. The deeper parts of the Logan Sub-basin occupy zones of diverging faults, whereas the Coaldale High occupies the zone where the East Richmond and Coraki Faults converge. In this interpretation, the Yamba Trough is a pull-apart basin formed between splay faults of the Coast Range Fault and the Martin and Shannon Faults are thrust faults splaying from the Coraki Fault.

The unconformity between the Esk Trough sediments and the Clarence-Moreton succession represents a break of about 12 Ma during which time sediments and volcanics of the Ipswich Basin were deposited farther to the east. The unconformity represents a period of time when movement on the West Ipswich Fault was transpressional and the Esk Trough was elevated.

For the Ipswich Basin, the Rathdowney seismic survey (Milner & Milner, 1981) suggests that the fault on the eastern margin of the South Moreton Anticline is the bounding fault to the rift sequence (fig. 8 of Korsch & others, 1989). If this is the case, the rift sequence probably thins gradually towards the east, away from the fault and laps out rapidly to the west on to the footwall block, the South Moreton Anticline. In this area, the Beenleigh Block represents the elevated peripheral bulge on the upper plate. The Ipswich Coal Measures wedge out on to the South Moreton Anticline but the Clarence-Moreton succession continues across it (Figs. 4 & 5). By the time deposition of the Clarence-Moreton sequence commenced, thermal relaxation was the dominant driving force for subsidence allowing sediments to spread over both rift sequences and basement highs.

Southern Logan Sub-basin

The structure of the southern Logan Sub-basin is controlled by several major strike-slip fault systems. These faults were probably most active prior to deposition of the Ipswich Basin and Clarence-Moreton successions, forming transtensional basins, as in the north, but have reactivated since to produce positive flower structures in Clarence-Moreton sediments. The orientation of these faults can explain the distribution of structural highs and lows in the Logan Sub-basin. Convergence of strike-slip faults with the same sense of movement produces uplift of the wedge of rock between them (Fig. 15a; Crowell, 1974), whereas diverging faults produce an area of subsidence (Fig. 15b; Crowell, 1974). In the southern Logan Sub-basin, the Coaldale High (Fig. 2) occupies the area where the South Moreton

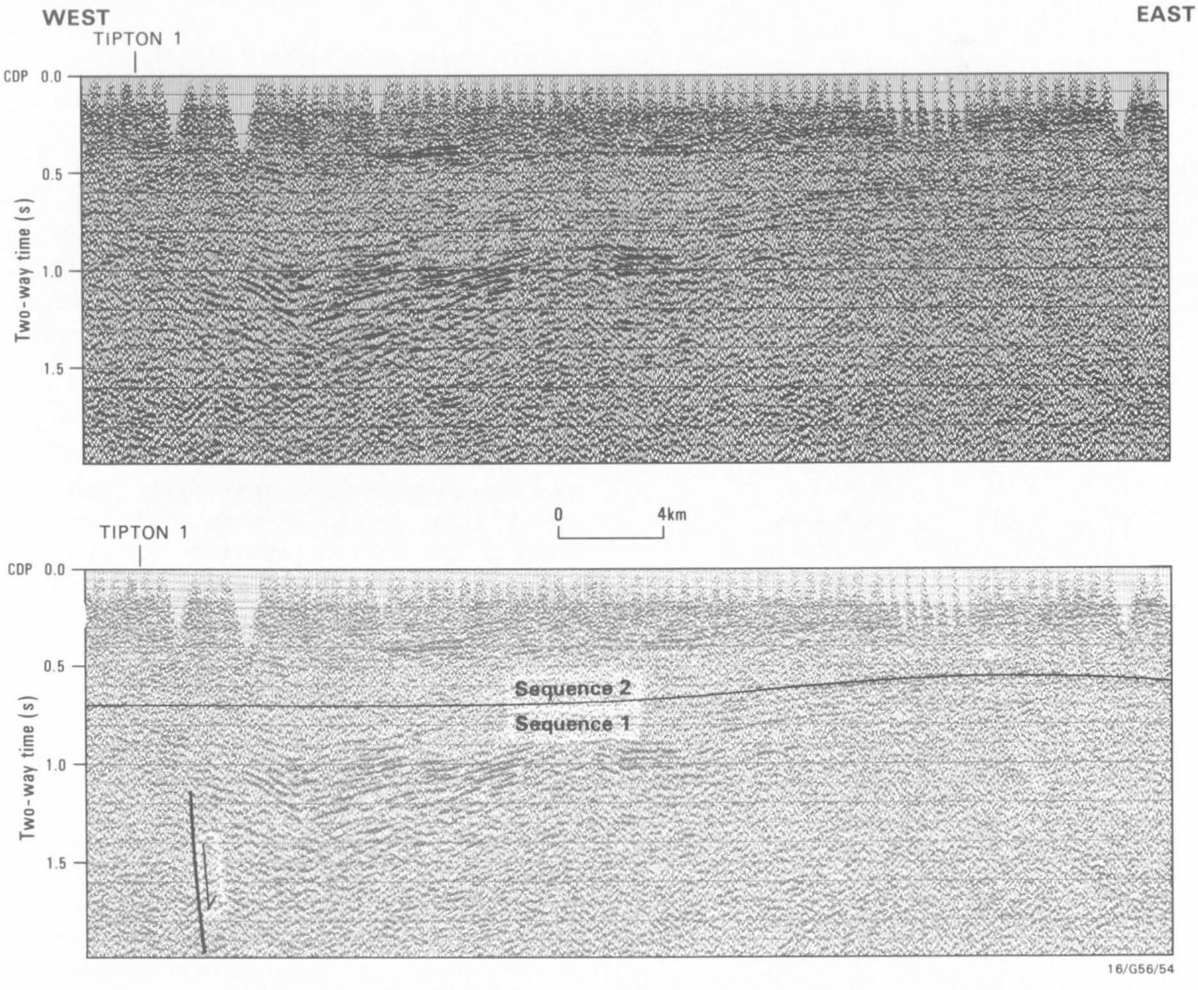

Fig. 13 East-west seismic line across the Horrane Trough (Segment of BMR Line 14, unmigrated). Seismic sequences are: (1) conglomerate, sandstone, mudstone and coal, probably of Late Triassic age; and (2) Clarence-Moreton Basin sediments.

Anticline and the Coraki Fault converge and the basin floor slopes northeast from there to the Casino and Lismore Lows. The Grafton Low is located where the Coraki and Coast Range Faults diverge.

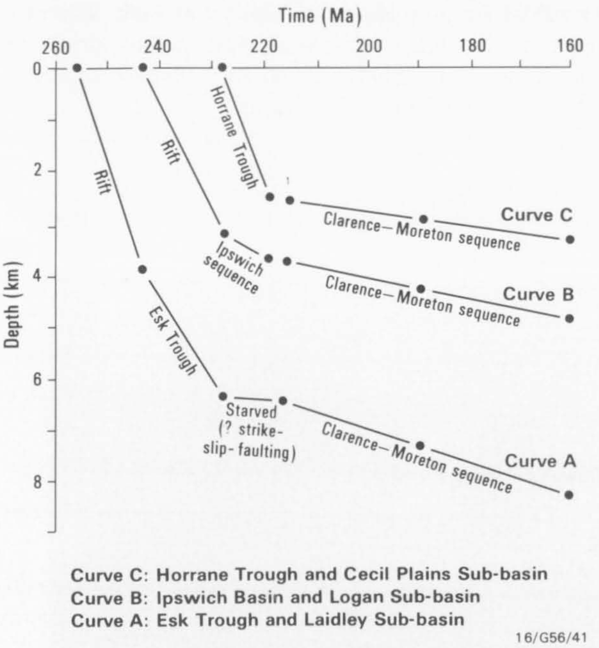

Fig. 14 Raw subsidence curves showing cumulative thickness of sediment versus time for the fill of the Esk Trough and Laidley sub-basin (Curve A), and the Ipswich Basin and Logan sub-basin (Curve B). Sediment thicknesses for Curve A were derived from two-way time-depth conversions of data from BMR seismic traverse 16. Curve B thickness data derives from a composite of seismic data from the Rathdowney survey and BMR traverse 16, and well data from the northern Logan Sub-basin. Thickness data from company seismic and well data from the southern Logan Sub-basin produced a virtually identical curve.

In the northern Clarence-Moreton Basin, the maximum subsidence in both the Ipswich Basin and the overlying Clarence-Moreton Basin coincides with areas of older transtensional rifting (Korsch & others, 1989). This model can be used to predict the geometry of some of the deep, extensional-phase basins underlying the southern Clarence-Moreton Basin. The dextral strike-slip nature of the main basin-forming faults suggests that the deep structure consists of half-grabens trending southwest-northeast (Harding & others, 1985). The north-easterly trend of Mid-Triassic volcanics beneath the southern Logan Sub-basin inferred from aeromagnetic data by Wellman (his fig. 4, this bulletin) is consistent with this model.

Cecil Plains Sub-basin

The Horrane Trough appears to be bounded by a near-vertical fault, although the data are less clear than in sub-basins to the east. This fault geometry and the presence of a pull-apart basin of similar age, the Tarong Basin, along the northeast margin of the Cecil Plains Sub-basin (Flood & Garces, 1986), suggest that the Horrane Trough also formed along a strike-slip fault. This fault was probably active during the Late Triassic, contemporaneous with deposition of the Ipswich Coal Measures and formation of the Tarong Basin (Flood & Garces, 1986). Subsidence by thermal relaxation then followed through the latest Triassic and Jurassic.

IMPLICATIONS OF BASIN EVOLUTION AND STRUCTURAL MODEL FOR REGIONAL TECTONICS

Strike-slip faults have played a significant role in the development of basement structures in the New England Orogen. They controlled the transport and accretion of displaced terranes and Late Carboniferous-Early Permian oroclinal bending of the mid-Palaeozoic accretionary wedge sequence (Korsch & Harrington, 1987; Korsch, this Bulletin). Late Carboniferous to Early Permian displacements of the order of hundreds of kilometres have been suggested for the major faults (Harrington & Korsch, 1985b; Murray & others, 1987; Korsch, this Bulletin) but proven displacements are an order of magnitude lower (Korsch & others, 1978; McPhie & Fergusson, 1983). Basin structures indicate that this tectonic style continued into the Mesozoic, controlling basin formation and deformation. The geometry of the basins suggests that they and the basement orocline are both products of strike-slip faulting, but that the orocline itself did not exert much influence on basin geometry.

The transtensional phase beneath the Ipswich Basin is probably younger than that beneath the Esk Trough, reflecting a general easterly younging of large sedimentary basins in eastern Australia (Harrington & Korsch, 1985a). This may be explained by two possible mechanisms. During extension, lateral heat flow causes cooling of the lithosphere, particularly in narrow rifts like that beneath the Esk Trough (Houseman & England, 1986). This cooling strengthens the lithosphere until it becomes stronger than unstretched lithosphere. At this point, extension ceases and relocates to unstretched lithosphere (Dewey, 1982). The eastward stepping of basins may also be explained by the tendency of strike-slip faults that contain bends to straighten themselves with time. This produces lateral shifts of active, parallel faults through time. Either mechanism, or a combination of both, could have caused the eastward stepping in the study area.

Although the main area of extension stepped east, some extension occurred west of the Esk Trough, where

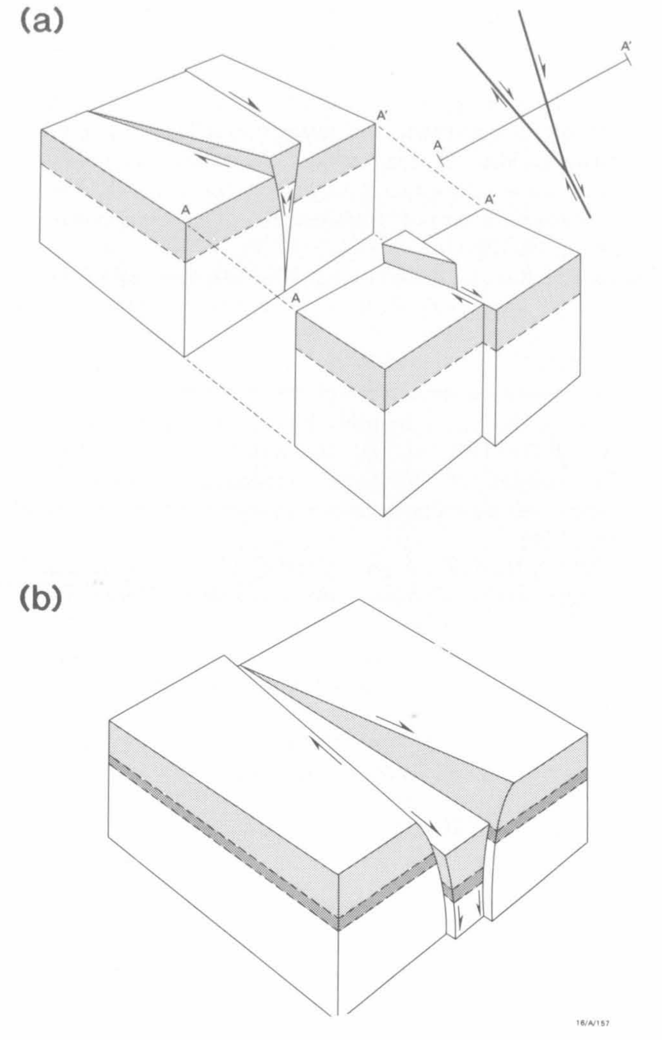

Fig. 15 (a) Uplift of a block between converging strike-slip faults (after Crowell, 1974). This type of uplift produced the Coaldale High between the East Richmond and the Coraki faults. (b) Subsidence of a block between diverging strike slip faults. The Grafton Trough is interpreted as the result of divergence of the Coraki and Coast Range Faults.

the relatively small (75 x 10 km) Tarong Basin and Horrane Trough developed at the same time as the Ipswich Basin to the east. The Tarong Basin is dominated by alluvial fan sedimentation and is a pull-apart basin related to dextral, braided, strike-slip faulting (Flood & Garces, 1986).

The styles of deformation along the South Moreton Anticline, and the Coraki and Coast Range Faults are best explained by almost pure dextral strike-slip along the Coraki Fault (Fig. 12). Such a stress direction would produce transpression, and hence thrusting, on the East Richmond and Coast Range Faults, whereas flower structures formed on eastward side-steps/restraining bends on the Coraki Fault (Fig. 12). Compressional structures, such as anticlines (Cranfield & others, 1981; Korsch & others, 1986), and thrust faults (Korsch & others, 1989) on the South Moreton Anticline in the north, probably involved late-stage transpressional movement on the West Ipswich Fault, thus indicating reactivation and tectonic inversion after deposition of the Clarence-Moreton sequence.

The involvement of the Late Jurassic Grafton Formation in thrusting and folding along the Coraki Fault indicates that movement continued past this time. Shaw (1978) demonstrated sinistral movement along this part of the eastern Australian margin during rifting. Break-up first occurred between 72 and 76 my BP (Shaw, 1978) so the last dextral movements on the faults in the Logan Sub-basin probably took place during the early to mid-Cretaceous. The implications of this timing for hydrocarbon prospectivity are discussed elsewhere by Powell & others (in prep.) and Gleadow & O'Brien (in prep.).

CONCLUSIONS

Dextral strike-slip faults produced crustal transtension that led to the formation of the basins crossed in the eastern sector of the Eromanga-Brisbane Geoscience Transect and continued to deform the basin fill until the Cretaceous. The largest movements took place along the West Ipswich Fault and associated faults, beneath the South Moreton Anticline. Significant movement also took place along the Coraki and Coast Range Faults, forming Triassic basins and a complex pattern of highs and lows beneath the Logan Sub-basin. Smaller-scale strike-slip movement in the west formed the Horrane Trough, initiating development of the Cecil Plains Sub-basin. Later movement along major faults produced positive flower structures and thrusts in Clarence-Moreton sediments.

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PERMIAN TO CRETACEOUS SUBSIDENCE HISTORY ALONG THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT

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ABSTRACT

The Permian to Cretaceous subsidence histories of 44 wells along the Eromanga-Brisbane Geoscience Transect have been calculated using standard backstripping techniques. Some consistent features are revealed and these are discussed in terms of relevant basin formation mechanisms.

The Permian and Triassic subsidence history is consistent with the Bowen Basin evolving as a foreland basin, and the Cooper and Galilee Basins subsiding on a more stable part of the craton. The subsidence data alone cannot resolve the role of extension in the evolution of these Permian basins, but it is concluded that surface loading and in-plane stresses probably acted to varying degrees at the same time.

A widespread period of erosion occurred in the Late Triassic but, by the Jurassic, sediments were being deposited over most of southern Queensland. It is suggested that this areally extensive subsidence was related to subduction-induced convection below the continental platform. This model is not wholly consistent with the evolution of the Eromanga Basin at the western end of the transect; here an additional mechanism needs to be invoked. A simple thermal model is favoured, and the rapid increase in subsidence during the Early Cretaceous is attributed to an excess sediment load at the surface, rather than a deeper lithospheric process.

These interpretations imply that a variety of different tectonic subsidence mechanisms controlled the evolution of the Permian to Cretaceous cratonic sedimentary basins in eastern Australia and the convergent plate margin off the eastern coast of Queensland had a significant influence on subsidence and sedimentation in this region.

INTRODUCTION

The subsidence history of a sedimentary basin reflects mainly vertical tectonic movement. The presence of unconformities within a sedimentary basin or between successive sedimentary sequences can be interpreted in terms of secondary influences acting together with a primary subsidence mechanism. Alternatively, unconformities may record major events resulting in significant changes to the subsidence mechanism and overall tectonic regime. There are three physical factors which can contribute to the tectonic evolution of a sedimentary basin : thermal, stress-based, and gravitational influences. Thermal effects are manifested as heating or cooling with concomitant thermal expansion (uplift) or contraction (subsidence), stress effects as in-plane forces (extension, compression and shear) and gravity effects as the response to surface or internal loading of the lithosphere. In general, all three mechanisms will operate and it is the relative importance of each through time that determines the evolutionary character of a basin.

Geohistory analysis, or backstripping, provides a quantitative method of examining the tectonic component of the subsidence history of a sedimentary basin. Most simply, the progressive loading effect of sediment is removed backwards through time and additional corrections may be incorporated to account for variations in the water depth and sea-level. The remaining part of the subsidence is termed the tectonic contribution and is defined as that part of the observed subsidence attributable to the primary physical process(es) driving basin formation. The backstripping technique has been widely used to provide supporting evidence for particular basin-forming mechanisms and the processes involved during basin evolution (e.g. Sclater & Christie, 1980; Bond & Kominz, 1984; Barton & Wood, 1984; Shaw & others, 1990). Additionally, where assumptions have been made regarding the form of tectonic subsidence, backstripping has been used to examine the magnitude of sea-level variations (Watts & Steckler, 1979).

In this paper, the backstripping method has been applied to 44 wells lying along or close to the Eromanga-Brisbane Geoscience Transect. The results of the backstripping calculations illustrate the variations in uplift and subsidence along the transect from the eastern margin to the intracratonic basins in the west. The late Palaeozoic to Cretaceous subsidence history is discussed qualitatively with a review of basin formation models relevant to the tectonic evolution of eastern Australia.

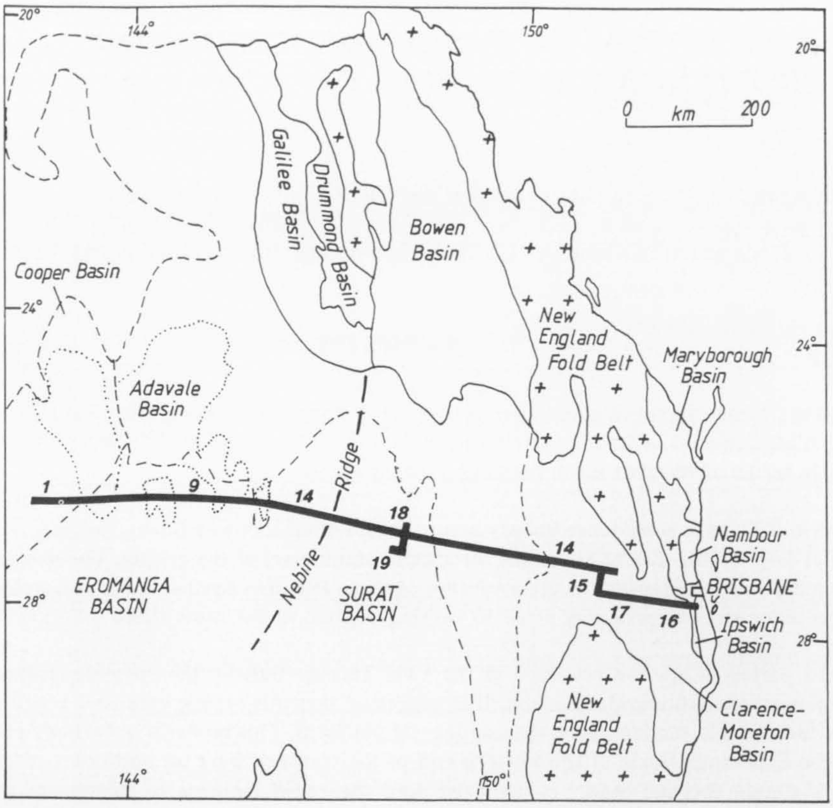

Fig. 1 Sedimentary basins in southern Queensland. The dashed boundaries represent concealed Permo-Triassic basins and the dotted boundaries represent concealed Devonian basins. The principal BMR seismic lines are also shown.

GEOLOGICAL HISTORY

Only a brief summary of the geological evolution of the region from the late Palaeozoic onwards is given here. More detail is given in other papers in this Report. Figure 1 illustrates the major tectonic features discussed in this paper, and Figure 2 shows the general stratigraphy across the region.

In the Middle Carboniferous, the Adavale and Drummond Basins were subjected to E-W compressional deformation, the accompanying uplift and erosion resulting in the distribution of troughs and intervening basement blocks existing today. This deformation was probably directly related to a convergent plate margin to the east. This margin subsequently developed a significant transform component, as indicated by the Late Carboniferous-Early Permian development of large oroclinal structures in the New England Fold Belt (e.g. Murray & others, 1987). During the Permian and Triassic, the New England Fold Belt was a tectonically active region, undergoing uplift and substantial volcanism. To the west, subsidence commenced during the Early Permian, with marine sedimentation in the Bowen Basin, and non-marine sedimentation in the Cooper and Galilee Basins. Korsch & others (1988) have suggested that subsidence in the Bowen basins was controlled by contemporary transtensional faulting. Alternatively, Veevers (1984) suggested that the Bowen Basin represents a foreland basin, with subsidence being related to easterly directed thrusting during the Permian and Triassic. Murray (this Bulletin) discusses the evidence for these models in more detail, but, although the physical mechanisms differ considerably, it is significant that both of these concepts invoke a strong influence from the tectonically-active eastern margin during this time. Kuang (1985) suggested that the Early Permian topography in the Cooper Basin was rather irregular, with major escarpments and horst/half graben structures formed by the preceding compressional orogenic activity. During the Early Permian, subsidence appears to have been largely fault controlled. It is possible that early subsidence in the Cooper Basin was the result of a reduction in the magnitude of these compressional forces. Following a period of non-deposition/erosion in the Middle Permian, fluvial and lacustrine sedimentation also occurred in the Cooper and Galilee Basins. The depositional environments and rates of subsidence and sedimentation in these two basins were more subdued than in the Bowen Basin, and these differences are reflected in the relative development of coal measures within the basin sequences (Hunt, 1988).

The Late Triassic marks a period of widespread uplift and erosion across much of eastern Australia, commonly referred to as the Hunter-Bowen Orogeny. The degree of deformation increases towards the east, again suggesting a dominantly compressional regime at the eastern margin. By the latest Triassic to Early Jurassic, fluvial sediments were deposited over the much of the region. The onset of sedimentation was earliest in the east (Clarence-Moreton and Maryborough Basins, and possibly parts of the Surat Basin) and commenced later in the Eromanga and Surat Basins. The depositional areas expanded with time, so that by the end of the Jurassic most of southern Queensland was covered by a blanket of sediment. Minor marine influences are apparent in the Clarence-Moreton Basin during the Early Jurassic (Day & others 1983), but sedimentation appears to have declined by the Middle to Late Jurassic.

The lack of post-Jurassic sediment in the eastern region (except for the Maryborough Basin) may be the result of erosion. The observation that the Cretaceous marine transgression into the Surat and Eromanga Basins initially appears to have come from the east (Exon & Senior, 1976) suggest that subsidence and deposition may have continued through into the Cretaceous in this region. During the Cretaceous, the area of shallow-marine sedimentation covered the Surat and Eromanga Basins and extended north to the Carpentaria Basin. The depositional environment fluctuated between shallow-marine, paralic and fluvial during the Early Cretaceous, and a period of fluvial sedimentation during the Cenomanian marked the effective cessation of deposition in these basins. It is thought that sedimentation ceased in the Surat Basin earlier than in the Eromanga Basin and sedimentary facies suggest that the sea moved out through what is now the Gulf of Carpentaria (Exon & Senior, 1976; Burger, 1986). Once again erosion may be responsible for the absence of younger Cretaceous sediments in the Surat Basin.

SUBSIDENCE ANALYSIS: METHOD AND ASSUMPTIONS

The references cited in the introduction summarise the backstripping method, and the details will not be given here. For a given thickness of sediment, H_s, the water-loaded tectonic subsidence, H_b, under the assumption of Airy isostasy, is given as

$$H_b = \{[(\rho_m - \rho_s)H_s - \rho_m H_{sl}]/(\rho_m - \rho_w)\} + H_w \qquad (1)$$

where ρ_s is the mean sediment density, ρ_m and ρ_w are the densities of the mantle and sea-water respectively, H_w is the water depth during deposition (palaeobathymetry), and H_{sl} is the difference in sea-level, positive for a sea-level rise relative to the present day. The results are presented without palaeobathymetry or sea-level corrections, although the magnitude of these corrections will be discussed later.

The assumption of Airy, or local, isostasy is made primarily for simplicity although, considering the large

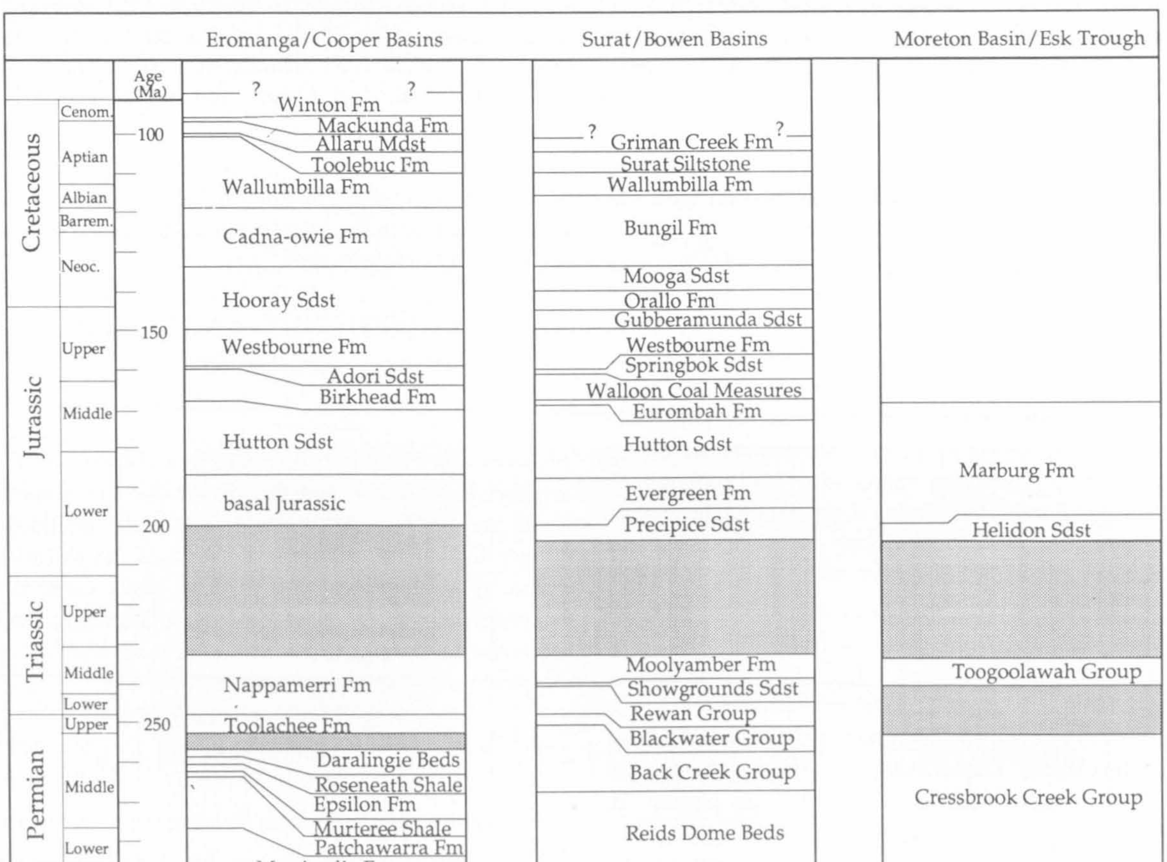

Fig. 2 General Permian to Cretaceous stratigraphy of the sedimentary basins in southern Queensland.

extent of the depositional area over the relevant time period, this assumption is not unreasonable. Following the methodology and formulations outlined in Gallagher & Lambeck (1990), the backstripping calculations were made under the assumption that porosity is reduced either solely by compaction, in which case the thickness of the unit decreases as it is buried, or solely by externally sourced cement, in which case the thickness of the unit remains constant. In both cases, the mean density of a given sedimentary unit increases as it is progressively buried. Porosity-depth relationships for sandstone, siltstone and shale were parameterised in the form

$$\phi(z) = \phi_o \exp(-cz) \qquad (2)$$

where $\phi(z)$ is the porosity at depth z, ϕ_o is the porosity at the surface (z=0) and c is a constant. To assess the errors that may arise because of an inappropriate porosity-depth relationship, the calculations were also carried out with different values of ϕ_o and c chosen as bounding values on the observed porosities. These were then used to generate vertical error bars on the subsidence curves. The values of ϕ_o and c for each lithology were based on those given in Gallagher & Lambeck (1990) and are given in Table 1.

The locations of the wells considered in this study are shown in Figure 3. The formation tops were taken from the Queensland Department of Mines Queensland Energy Resources Data Base (QERDB) to provide an internally consistent stratigraphy. The absolute ages of formations were assigned using the QERDB stratigraphic nomenclature and the timescale of Harland & others (1982). It should be noted that although the formation tops represent lithostratigraphic, rather than chronostratigraphic, boundaries it was considered that uncertainties in the absolute ages were greater than any potential errors arising from diachronous formation boundaries. Gallagher (1988) has discussed the uncertainties associated with the assignment of absolute ages using different timescales. The Harland & others (1982) scale tends to give ages some 5-10 Ma older during the Jurassic and Triassic, compared with the timescales of Odin (1982) and Snelling (1985). Error bars on the absolute ages, however, were not included to improve the clarity of the diagrams. As this study focuses on the relative forms of the subsidence curves along the transect, the interpretations are not sensitive to the errors in absolute ages.

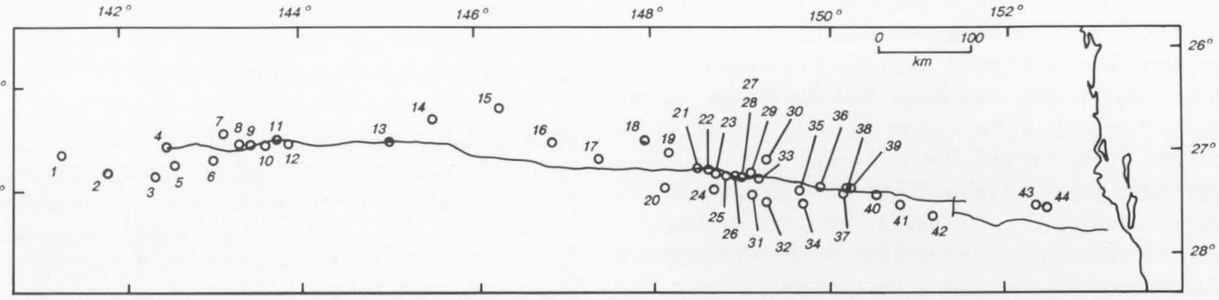

Fig. 3 Locations of wells used for subsidence analysis. 1-Cook North 1; 2-Barrolka East 1; 3-Wareena 1; 4-Mt Howitt 2; 5-Boldrewood 1; 6-Berellem 1; 7-Mt Bellalie 1; 8-Black Stump 1; 9-Kenmore 1; 10-Eromanga 1; 11-GSQ Eromanga 1; 12-Mongarlo 1;13-Quilpie 1; 14-Quilberry 1; 15-Charleville 1; 16-Lowood 1; 17-Scalby 1; 18-Strathmore 1; 19-Glenroy 1; 20-Hoolah 1; 21-Tory Boy 2; 22-Avondale South 1; 23-Narelle 1; 24-Basketyard Creek 1; 25-Kincora 1; 26-Taralga 1; 27-Newington 1; 28 Newington 2; 29-Bainbilla 1; 30-Appletree 1; 31-Grantham 1; 32-Red Cap 1; 33-Myall Creek 1; 34-Paloma 1; 35-Coalbah 1; 36-Meandarra 1; 37-Arlington 1; 38-Lorraine 1; 39-Leichhardt 1; 40-Alick Creek 1; 41-Kumbarilla 1; 42-Cecil Plains West 1; 43-Baylam 1; 44-Lockrose 1.

SUBSIDENCE ANALYSIS: RESULTS

The calculated Permian to Cretaceous subsidence histories for 29 of the 44 wells are shown along the transect in Figure 4 and the results for the remaining wells are given in Appendix 1. The wells included on Figure 4 give a good representation of the subsidence trends along the transect, with those wells listed in the appendix having similar subsidence histories to nearby wells shown on Figure 4.

PERMIAN-TRIASSIC SUBSIDENCE

Permian-Triassic sediments occur at the western end of line 1 (northern Cooper Basin), the eastern half of line 14 (Taroom Trough of the Bowen Basin), and in the eastern region of the transect (e.g. Moreton Basin, Esk Trough, and the Yarraman Block which lies just west of the Esk Trough, see Day & others; 1983). The general trend of the subsidence histories is one of greater irregularity and more rapid subsidence towards the east, although erosion has complicated the Permian-Triassic sedimentary record in many of the wells. However, this observation, together with the easterly increase in the volcanogenic component of the sediments and fluctuations between marine and non-marine depositional environments, suggests the tectonic activity at the eastern margin of the continent had a significant influence on subsidence and sedimentation. Although variable, with intermittent periods of erosion, the overall form of the subsidence in the Bowen Basin (e.g. Apple Tree 1, Bainbilla 1, Red Cap 1) shows rapid Early Permian subsidence, possibly slowing during the remainder of the Permian and then increasing until the Middle Triassic, although this trend could be partly an artefact of the absolute ages assigned to formation tops.

The Permian-Triassic subsidence at the western end of the transect in the Cooper Basin is not particularly well established owing to a lack of stratigraphic data. In comparison with the eastern end of the transect it appears that subsidence was slower, and this is reflected in the sediment thicknesses in the two areas - the Bowen Basin having up to 3-5 times the thickness of equivalent Cooper Basin sediments (Day & others, 1983; Veevers, 1984). Battersby (1976) has suggested that the increase in sedimentation in the northern Cooper Basin during the Triassic (e.g. Barrolka East 1 and Mount Howitt 2) may reflect a change in the location of sediment source region, away from the area to the south of the Cooper Basin. However, the increase in subsidence rate may not be as severe as illustrated because the Middle Triassic age assigned to the top of the Nappamerri Formation may be too old. Youngs & Boothby (1985) have suggested that, in the northern Cooper Basin, the formation may in fact continue into the Late Triassic.

LATEST TRIASSIC-CRETACEOUS SUBSIDENCE

After a period of widespread erosion of the eastern Australian basins in the Late Triassic, subsidence and non-marine sedimentation was reinitiated by the Jurassic. Sedimentation commenced by the Late Triassic in the eastern region of the transect (Moreton, Nambour, and Maryborough Basins) and subsequently spread eastwards to the Surat and Eromanga Basins, the depocentres initially being located over the thicker Permian-Triassic sequences (Day et al., 1983). In the Jurassic, the depositional areas in the latter two basins rapidly expanded, while the sedimentation rate was reduced in the eastern basins and may even have ceased by the Middle to Late Jurassic. Alternatively, the sequence may have been subsequently truncated by erosion. In either case, the observations suggest uplift of the eastern region.

Examination of the subsidence curves along the transect reveals that the subsidence is more or less linear through the Jurassic, although the rate of subsidence may decrease slightly in the Late Jurassic-Early Cretaceous. Once again it is felt that the absolute age control does not allow unequivocal conclusions to be drawn regarding the significance of small kinks in the curves. There does, however, appear to be an overall trend of more rapid subsidence by approximately a factor of 2 from west to east during the Jurassic (e.g. compare Coalbah 1 or Taralga 1 with Mount Bellalie 1 or Barrolka East 1). An interesting aspect of this observation is that there does not seem to be any significant change in subsidence rate eastwards to the Nebine Ridge, implying that this region was subsiding at the same rate as the eastern Eromanga Basin. Additionally, a west-to-east progression is seen in the age of the surface strata. In the Eromanga Basin, the youngest preserved Cretaceous sediments are about 90 Ma compared with 100-110 Ma in the Surat Basin, and the stratigraphic record in some of the eastern basins ceases in the Early-Middle Jurassic. This trend of progressively younger outcrop towards the west may reflect a period of non-deposition in the eastern region,

TABLE 1

Best fit, upper and lower bound values derived for the parameters ϕ_0 and c in the exponential porosity-depth function (equation 2), and grain density for sandstone, siltstones and shale.

	Best fit		Upper		Lower		Grain Density
	ϕ_o	c	ϕ_o	c	ϕ_o	c	ρ_g
Lithology	%	km^{-1}	%	km^{-1}	%	km^{-1}	$kg.m^{-3}$
Sandstone	43.0	0.718	48.0	0.567	38.0	0.987	2670
Siltstone	45.7	1.158	51.0	0.965	41.0	1.310	2680
Shale	50.4	1.616	55.0	1.300	45.0	1.760	2680

although recent fission track analysis in the northern Bowen Basin has led to the conclusion that 1-2 km of erosion occurred about 100-120 Ma, stripping off any Jurassic and Cretaceous cover (Marshallsea, 1988). It is quite possible that contemporaneous erosion, albeit possibly reduced in magnitude, may also have occurred farther to the south in Queensland and reduced the thickness of the Cretaceous section.

Another significant feature in the subsidence curves is the apparent increase in subsidence rate during the Cretaceous, most noticeably in the Eromanga Basin. It was briefly stated earlier that the Cretaceous marks a period of marine transgression in the Surat and Eromanga Basin, and it may be argued that the increase in the rate of tectonic subsidence is merely a result of the sea-level correction in the backstripping equation (eqn. 1) has been neglected. The Cretaceous sea-level high, relative to present-day mean sea-level, has been estimated to lie between 150 and 350 m, with a preferred value of about 200-250 m and the timing of this maximum is somewhere in the middle of the Cretaceous (Watts & Steckler, 1979; Kominz, 1984; Haq & others, 1987). If a sea-level peak of 200-250 m is adopted, then the maximum correction to the subsidence curves would be about 290-360 m, and the increase in subsidence would effectively disappear. However, this neglects the palaeobathymetry correction, which acts in the opposite sense to the sea-level term and is equal to the depth of water under which the sediments were deposited. This is a difficult parameter to estimate with any confidence, and Gallagher & Lambeck (1990) assumed a maximum water depth of 50 m during the deposition of the marine Cretaceous sediments in the Eromanga Basin; so the correction is then not too substantial. However, while this period of shallow-marine sedimentation was occurring in the Eromanga Basin, a non-marine to paralic environment existed in the Surat Basin to the east. This may have marked the final stage of Cretaceous sedimentation in the Surat Basin, while deposition continued in the Eromanga Basin, with over 1200 m of fluvial sediment being deposited in less than 10 Ma. During this time, global sea-level was at, or approaching, its maximum value, although this interpretation is not in accord with that made from the stratigraphy in eastern Australia, which suggests a progressively more non-marine environment.

It is not obvious what influence sea-level variations will have on sedimentation within a continental interior. This will clearly depend on a variety of factors, such as (a) the relative elevation of the depositional area, (b) the access the encroaching sea may have to the interior platform, and (c) any changes in the erosional base-level and sediment source area. Given these caveats, as well as the fact that the sedimentary and subsidence record can be demonstrated to be strongly influenced by tectonic factors, it is considered that the apparent increase in subsidence rate in the Cretaceous is not merely a consequence of neglecting the sea-level term in the backstripping procedure.

DISCUSSION

Some of the basin-forming mechanisms and tectonic controls, which may be appropriate to the evolution of the sedimentary basins along the Eromanga-Brisbane Geoscience Transect, can now be considered. The form of subsidence in the Permian-Triassic Bowen Basin (e.g. Bainbilla 1, Grantham 1, Myall Creek 1, Red Cap 1) is consistent with predictions made from foreland basin models (Beaumont, 1981) and also with models which involve compressional forces acting on the lithosphere (Lambeck, 1983; DeRito & others, 1983). Both of these mechanisms can produce rapid, rather irregular subsidence patterns as the rate of subsidence is controlled by variable externally applied forces. Veevers (1984) favours a foreland basin environment for the Bowen Basin, drawing an analogy with present-day Papuan Basin environment. In contrast, Korsch & others (1988) have used structural interpretations from seismic data to suggest that predominantly extensional processes controlled subsidence in the Bowen Basin, and also the Clarence-Moreton Basin (Baylam 1, Lockrose 1). The presence of widespread volcanism may also be cited as additional evidence of extension (see Murray, this Bulletin). Taken alone, the subsidence data presented here cannot discriminate between these various mechanisms. Indeed it is probable that more than one mechanism influenced the evolution of this and the other Permian-Triassic basins in eastern Australia, with surface loading and in-plane stresses acting at the same time to varying degrees to produce the complicated structural and stratigraphic relationships seen today.

Subsidence to the west in the Cooper and Galilee Basins was slower and more subdued than in the Bowen Basin, although still somewhat irregular (Gallagher, 1988). The mechanisms responsible for the subsidence in these two basins are again not apparent, although contemporaneous faulting seems to have accommodated at least some of the Permian subsidence. Overall, the slower rate of subsidence and the nature of the sediments in the Cooper and Galilee Basins reflect their positions on a relatively stable cratonic environment, away from the active eastern margin. However, the correlation of some significant unconformities and depositional styles across the Cooper, Galilee, and Bowen Basins implies that, at times, these basins were indeed influenced by the same regional scale-tectonic

Fig. 4 (Fold-out map opposite) Results of subsidence analysis for selected wells along the Eromanga-Brisbane Geoscience Transect. Appendix 1 gives the results for the wells not shown in this figure.

processes, mainly operating at the eastern margin of the continent.

Subsidence during the Late Triassic and Jurassic in the Clarence-Moreton Basin (Baylam 1, Lockrose 1) has been interpreted by Korsch & others (1988) in terms of a post-rift thermal relaxation phase, and they extend this interpretation to the Surat Basin. Simple extensional models (McKenzie, 1978) predict an exponentially decreasing subsidence rate after a rapid initial fault-controlled phase but, while it may be argued that the data in the Clarence-Moreton Basin is of this form, the Surat Basin subsidence tends to be more linear in time. The lack of data from the Clarence-Moreton Basin and the obvious erosional episodes precludes a definitive statement regarding the validity of an extensional model for this basin. The data from the eastern Eromanga Basin also show a linear form although the rate of subsidence is about 50% that observed in the Surat Basin. Additionally, Gallagher (1988) has shown that the rate of subsidence in the western Eromanga tends to be greater than in the east of the basin, but still only about 75% of the Surat Basin rate. The data from the Surat and Eromanga Basins are inconsistent with the typical exponential form expected during post-rift subsidence, but the observations do not necessarily rule out a thermal influence. Similar linear subsidence has been observed in other intracratonic or platform basins, e.g. the Williston Basin and the Hudson Bay Basin in North America, and the Canning Basin in Western Australia. A variety of mechanisms have been proposed to explain these subsidence patterns, as discussed briefly below.

Fowler & Nisbet (1985) interpret the subsidence of the Williston Basin in terms of deep crustal metamorphism, involving a phase change from gabbro to eclogite. They state that this interpretation is supported by results from a seismic refraction survey which indicated the presence, at the base of the crust, of a layer with a seismic velocity similar to eclogite. No such layer has been recognised under the basins along the transect and such interpretations tend to be equivocal. Also very large-scale intrusive and/or metamorphic processes would be required to generate the observed subsidence across the Surat and Eromanga Basins, an area of approximately 1.5×10^6 km^2.

Middleton (1989) developed a model which relates the Early Palaeozoic evolution of the Canning Basin to mantle convection by examining the effect of downwelling plumes on surface elevation. To be consistent with his mathematical formulation, Middleton discusses the model in terms of exponential subsidence. However, over the time interval he considers (~100 Ma), the calculated subsidence in the Canning Basin and the model prediction could also be approximated by an effectively linear function, especially after allowing for uncertainties in the calculated subsidence. This model relies on a removal of heat from the lithosphere, but does not require an anomalous heat input as in, for example, extension models, and therefore provides an explanation for the circular shape of some intracratonic basins with little evidence of normal faulting or widespread thermal activity. However, without appealing to a series of downwelling plumes, it seems unlikely that this model alone is an appropriate mechanism for the extensive subsidence in eastern Australia.

In considering the Hudson Bay Basin, Quinlan (1987) concluded that seaward tilting of the onshore platform region may be caused by sub-lithospheric convection cells associated with subduction (Mitrovica & others, 1989). The tilting may occur over distances of more than 1000 km from the trench, and such a mechanism may be a plausible explanation for the widespread subsidence in eastern Australia during the Jurassic. Additional factors that favour this mechanism are the clear influence of the eastern margin on subsidence during this time (e.g. as a sediment source) and the observation that subsidence rates tend to increase towards the east. A rearrangement of the margin by the Cretaceous, such as a change of convergence rates or direction, would have resulted in uplift in the eastern basins, resulting in erosion and a reduction in the thickness of Mesozoic sediments. Although this platform subsidence mechanism is an attractive first order explanation for the subsidence of eastern Australia, the observation that the major depocentre in the Eromanga Basin lies in the western part of the basin is not consistent with model predictions, suggesting that an additional, or different, mechanism was operative in this region. Also, the increase in subsidence rate during the Cretaceous is difficult to explain with any of the models discussed above, implying a significant contribution from a secondary influence at this time. Given that the platform subsidence model can explain the first-order features of the subsidence along the transect, some additional mechanisms are now considered that may have influenced the evolution of the Eromanga Basin and caused the rapid increase in subsidence.

Middleton (1978, 1980) proposed a deep crustal metamorphism model for the Eromanga Basin which, though adequately predicting the observed two-stage subsidence history up to the middle Cretaceous, requires a rapid doubling in thickness of the thermal lithosphere at the beginning of the Cretaceous. It is difficult to envisage a process that would account for this thickening, although Middleton suggested that progressive cratonisation of eastern Australia may be an explanation. However, the evidence presented in the papers in this Bulletin imply that this occurred during the Palaeozoic.

Houseman & England (1986) outline a model related to upwelling mantle plume and the associated dynamic uplift. Their mechanism relies on the tendency for the lithosphere to extend when its potential energy is increased. Initially the predicted subsidence is similar to the pure shear extension model of McKenzie (1978), but an additional subsidence component will occur as the mantle plume is removed. The form of this component is difficult to specify realistically, and Houseman & England (1986) calculated subsidence histories for a variety of plume removal rates. Clearly, such a mechanism could explain the rapid subsidence in the Cretaceous by a similarly rapid removal of the plume support. Objections to this mechanism include the general lack of obvious extensional features within the Eromanga Basin, and the fact that the continent would have had to have been static with respect to the mantle plume for the duration of the subsidence history of the basins.

Lambeck (1983) and DeRito & others (1983) have discussed another mechanism that predicts an increase in subsidence rate, whereby the flexural interaction of a pre-existing lithosphere deflection and an applied in-plane compressional force results in an increase in the curvature of the deflection, i.e. enhanced subsidence. There are also problems with applying this mechanism to the Mesozoic basins in eastern Australia in that the length scale of these basins is considerably greater than the commonly cited flexural wavelengths of 50-150 km, and also unreasonably large stresses (>10 kbars) would be required merely to explain the magnitude of the increase in subsidence.

A striking feature of the Mesozoic sediments is their cyclic character, with each cycle starting with generally quartzose sand, progressively showing more labile detritus and finer-grained material. This observation has been related to global sea-level variations and changes in the base level of erosion and energy of deposition (Exon & Burger, 1981; Burger, 1986). The global nature of sea-level fluctuations is somewhat controversial, especially on timescales of 5-10 Ma. Gallagher & Lambeck (1990) have suggested that the sediments in the Eromanga Basin may be reflecting a combination of first order sea-level variations and higher-frequency fluctuations in the intensity of tectonic activity on the eastern margin of the continent and the subsequent availability of volcanogenic detritus. In this model, the pseudo-linear subsidence during the Jurassic is considered to be consistent with a simple thermal contraction mechanism (e.g. Nunn & Sleep, 1984; Ahern & Mrkvicka, 1984), although the cause of the initial thermal perturbation is unclear. Gallagher (1988) considered the possibility that the same overall subsidence mechanism may be appropriate for both the Eromanga and the underlying Cooper Basins and suggested that the presence of granite intrusions of Late Carboniferous age directly under the southern Cooper Basin implies an elevated thermal regime in the lithosphere prior to subsidence.

The irregular character of the Cooper Basin (and also the Galilee Basin) sequences contrasts with the relatively uniform and widespread subsidence in the Eromanga Basin sequence and unconformities of varying temporal and geographical significance complicate the interpretation of the subsidence data. These differences may be the result of the greater importance of applied stresses and fault-controlled subsidence during the Permian and Triassic and the subsequent subsidence reflects the increasing contribution of the thermal-contraction driving load acting over a broader area. Unfortunately, the timing of the initiation of thermal subsidence cannot be adequately resolved with the data discussed by Gallagher (1988), and it is also possible that the primary mechanism responsible for the formation of the Eromanga Basin was unrelated to the underlying Permian/Triassic basins and was not initiated until the Late Triassic.

Irrespective of this problem, the late stage increase in subsidence represents a significant departure from the form of subsidence predicted by thermally driven mechanisms. Gallagher & Lambeck (1990) attribute the departure from the thermal model to a rapid influx of sediment as a result of enhanced tectonic activity to the east. The sediment acted as a surface load rather than a passive infill, making a significant additional contribution to the primary thermal subsidence. The volcanogenic nature of much of the Cretaceous sediment and the facies variations between marine and non-marine at a time when global sea-level was at, or rising to, its peak during the Cretaceous are consistent with the simple overfill model which predicts that the depositional base-level rises close to or above the rising contemporary sea-level.

CONCLUDING REMARKS

The subsidence histories of the Late Palaeozoic-Mesozoic basins along the Eromanga-Brisbane Geoscience Transect have been calculated from well data using standard backstripping techniques. Periods of erosion during this time have complicated the sedimentary record, but some consistent features can been seen in the results.

The tectonic activity at the eastern margin of the continent, together with a variety of interacting tectonic subsidence mechanisms (surface loading, faulting, thermal effects), appears to have controlled the evolution of the basins on the platform from the Permian to the Cretaceous. Variations in the state of stress at the margin probably influenced the stress state in the continental interior. The resultant uplift and volcanism provided an active and renewable sediment source. On a larger scale,

Fig. 4 (foldout page opposite) Results of subsidence analysis for selected wells along the Eromanga-Brisbane Geoscience Transect. Appendix 1 gives the results for the wells not shown in this figure.

146°00'
150°00'
154°00'
25°30'
28°30'
COOK NORTH 1
MOUNT HOWITT 2
MOUNT BELLALIE 1
KENMORE 1
MONGARLO 1
QUILBERRY 1
CHARLEVILLE 1
SCALBY 1
GLENROY 1
AVONDALE SOUTH 1
TARALGA 1
MYALL CREEK 1
ARLINGTON 1
ALICK CREEK 1
KUMBARILLA 1
BAYLAM 1
BARROLKA EAST 1
BOLDREWOOD 1
BERELLEM 1
EROMANGA 1
QUILPIE 1
LOWOOD 1
HOOLAH 1
NARELLE 1
GRANTHAM 1
COALBAH 1
MEANDARRA 1
CECIL PLAINS WEST 1
LOCKROSE 1
BLACK STUMP 1
GSQ EROMANGA 1
WAREENA 1
STRATHMORE 1
APPLE TREE 1
NEWINGTON 1,2
TORY BOY 1
KINCORA 1
BAINBILLA 1
BASKETYARD CREEK 1
RED CAP 1
PALOMA 1
LEICHHARDT 1
LORRAINE 1
BMR SEISMIC PROFILE
Adavale
Eromanga
Quilpie
Charleville
Morven
Mitchell
Roma
Miles
Dalby
Moonie
Esk
Toowoomba
BRISBANE
GOLD COAST
Thargomindah
Eulo
QLD
NSW
0
100 km
Subsidence (m)
Age (Ma)

sub-lithospheric convection related to subduction is considered to have been a controlling influence on the subsidence history of the platform in the Jurassic and Cretaceous.

The Eromanga Basin requires an additional or alternative tectonic mechanism as the geometry of the basin is inconsistent with the platform subsidence model. One interpretation of the subsidence history in this basin relies on a thermally based mechanism, with the final rapid subsidence phase in the Cretaceous being attributed to an excess surface load as a result of a large influx of volcanogenically derived detritus from the east at a time when sea-level was rising. This again points to the significant influence of plate margin processes on intracratonic subsidence and sedimentation.

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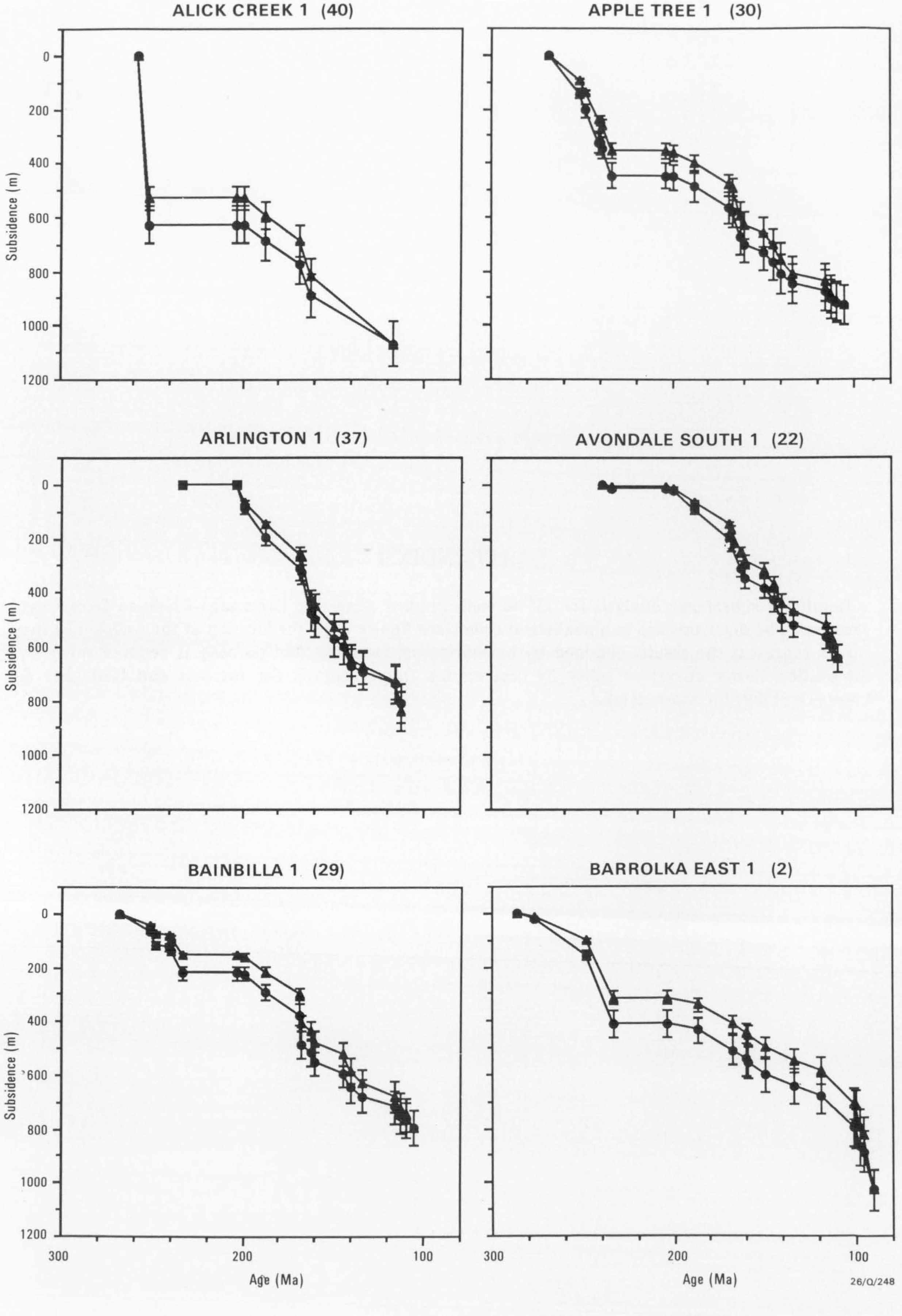
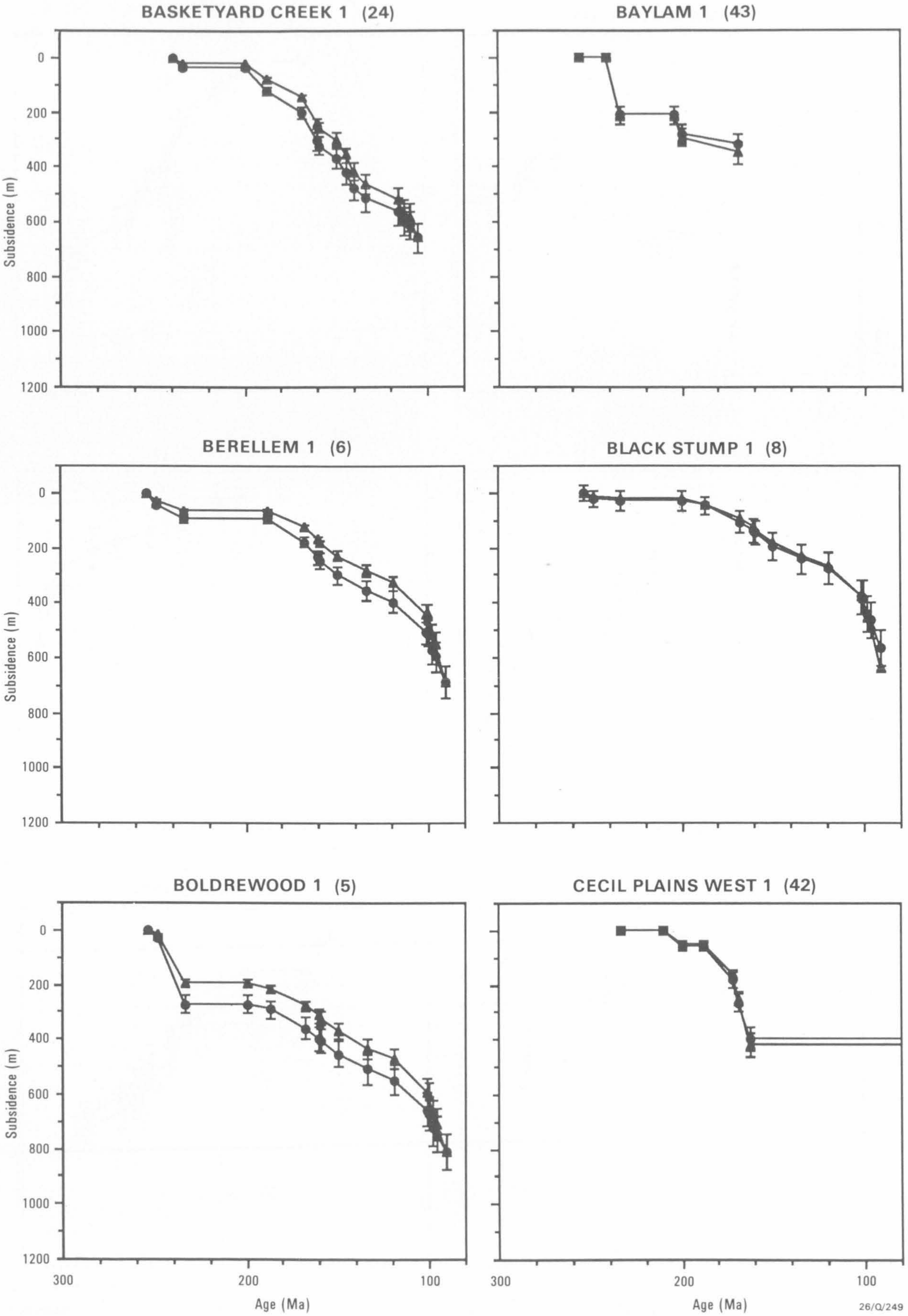
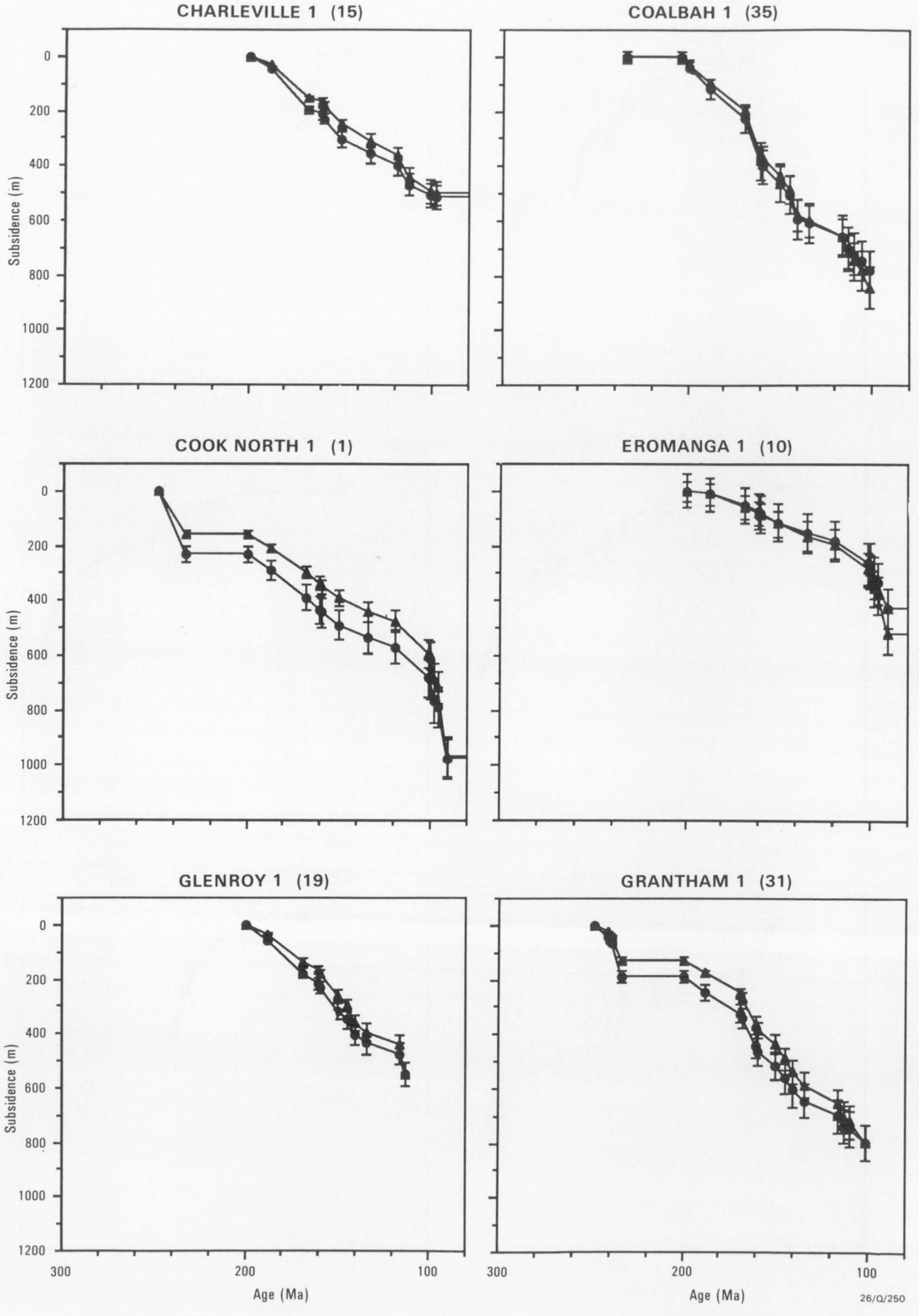
APPENDIX 1

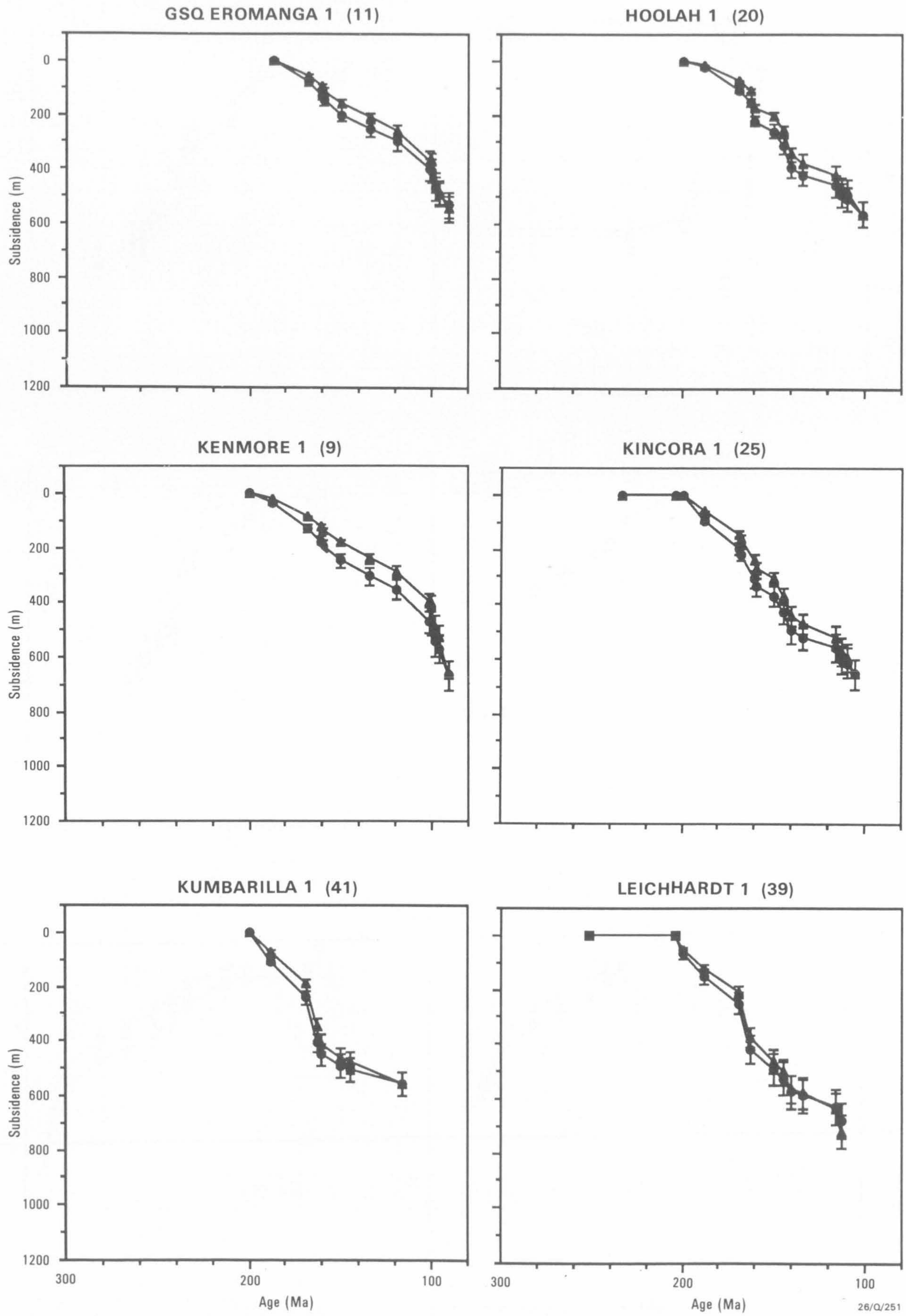
Results of subsidence analysis for all 44 wells studied along the Eromanga-Brisbane Geoscience Transect. The diagrams are in alphabetical order (see Figure 3 for the location of the wells). The two curves represent the results obtained by backstripping assuming that porosity is reduced solely by compaction (lower curve) or solely by cementation (upper curve). See the text and Gallagher & Lambeck (1989) for more detail.

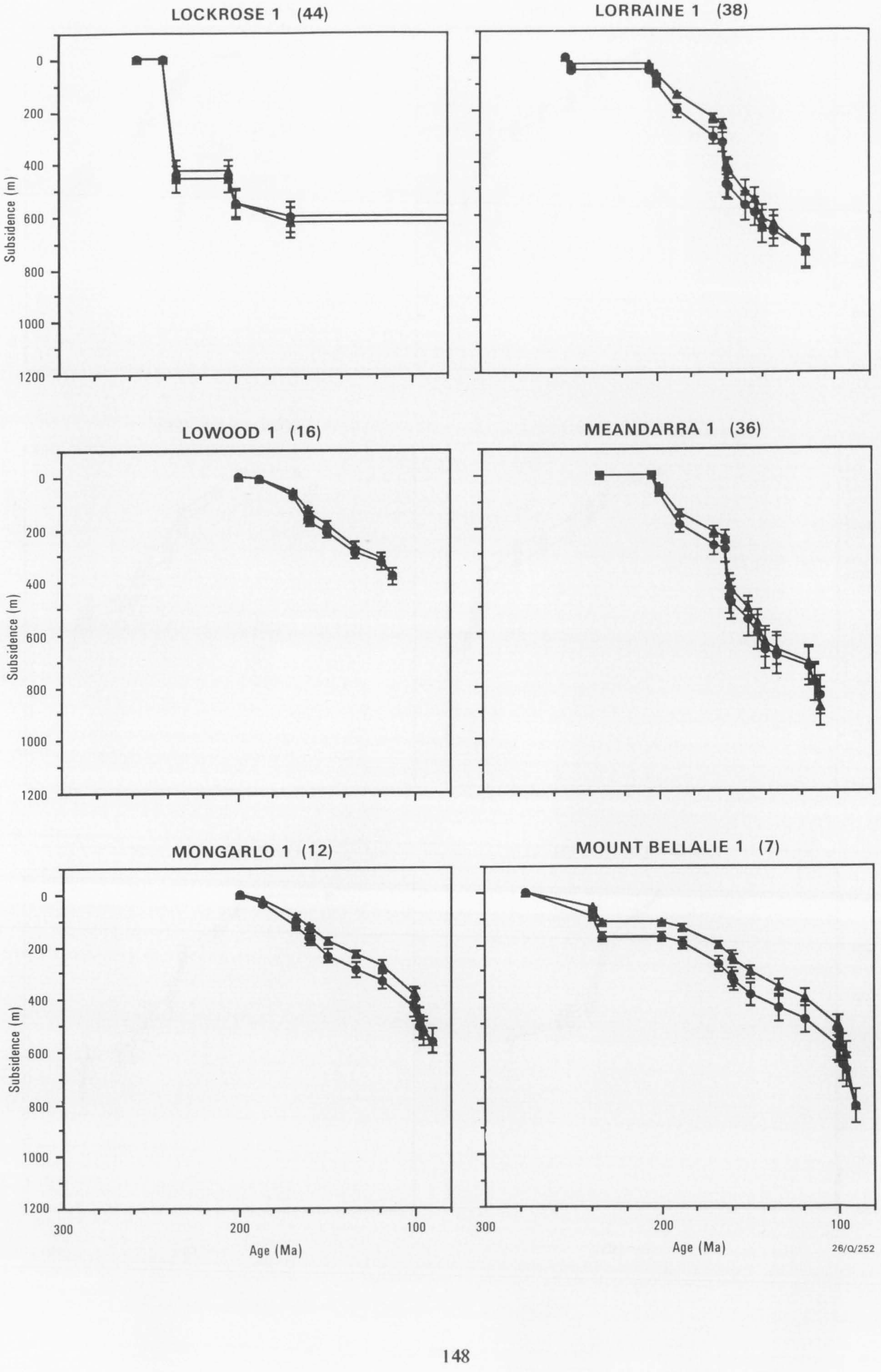

ALICK CREEK 1 (40)
APPLE TREE 1 (30)
ARLINGTON 1 (37)
AVONDALE SOUTH 1 (22)
BAINBILLA 1 (29)
BARROLKA EAST 1 (2)
Subsidence (m)
0
200
400
600
800
1000
1200
300
200
100
Age (Ma)
26/Q/248

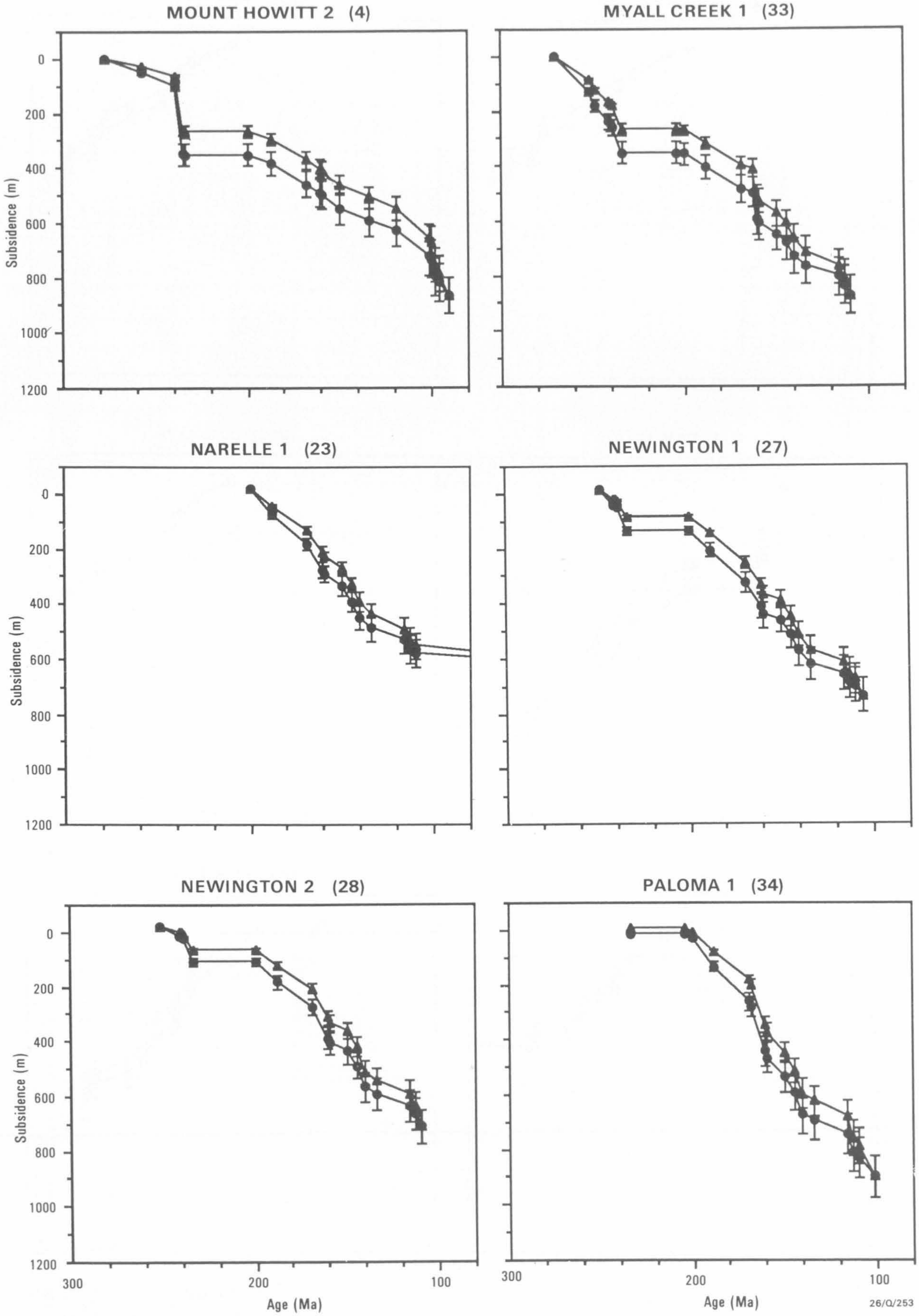


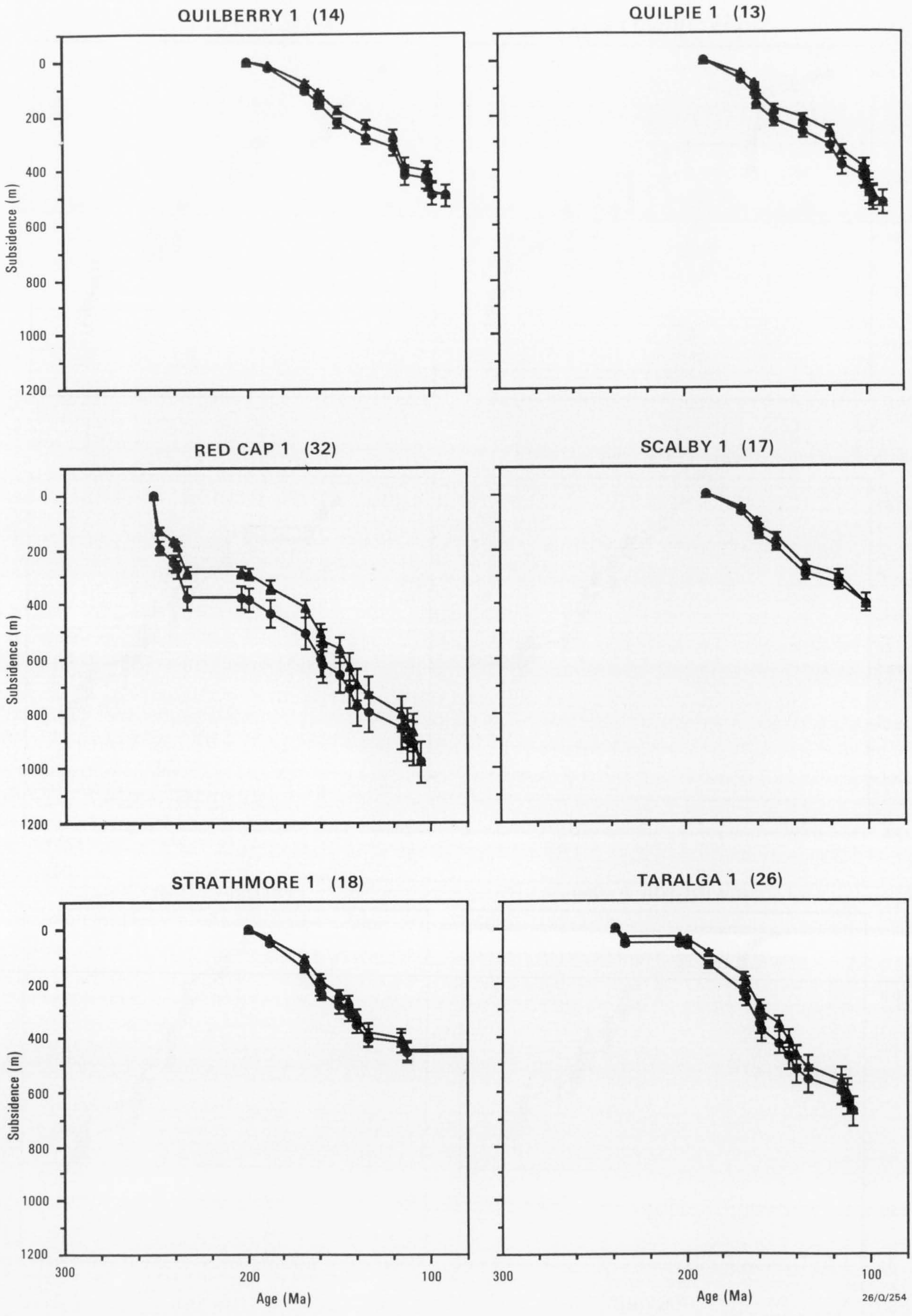
BASKETYARD CREEK 1 (24)
BAYLAM 1 (43)
BERELLEM 1 (6)
BLACK STUMP 1 (8)
BOLDREWOOD 1 (5)
CECIL PLAINS WEST 1 (42)
Subsidence (m)
0
200
400
600
800
1000
1200
300
200
100
Age (Ma)
26/Q/249

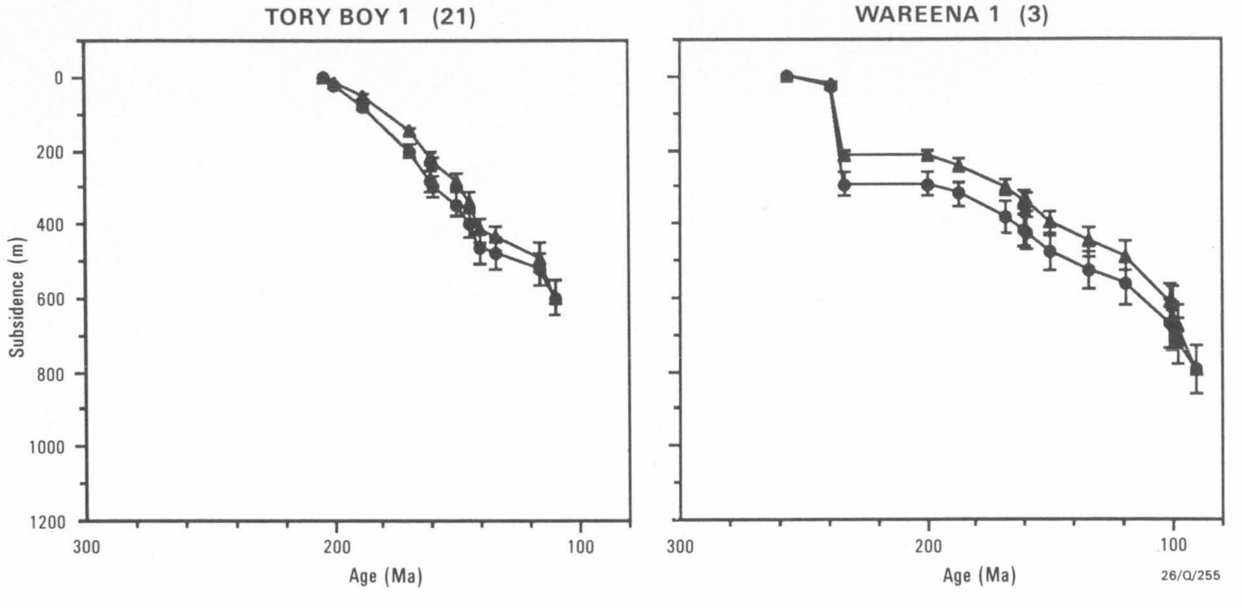

CHARLEVILLE 1 (15)
COALBAH 1 (35)
COOK NORTH 1 (1)
EROMANGA 1 (10)
GLENROY 1 (19)
GRANTHAM 1 (31)
Subsidence (m)
0
200
400
600
800
1000
1200
300
200
100
Age (Ma)
26/Q/250


GSQ EROMANGA 1 (11)
HOOLAH 1 (20)
KENMORE 1 (9)
KINCORA 1 (25)
KUMBARILLA 1 (41)
LEICHHARDT 1 (39)
Subsidence (m)
0
200
400
600
800
1000
1200
Age (Ma)
300
200
100
26/Q/251


LOCKROSE 1 (44)
LORRAINE 1 (38)
LOWOOD 1 (16)
MEANDARRA 1 (36)
MONGARLO 1 (12)
MOUNT BELLALIE 1 (7)
Subsidence (m)
0
200
400
600
800
1000
1200
Age (Ma)
300
200
100
26/Q/252


MOUNT HOWITT 2 (4)
MYALL CREEK 1 (33)
NARELLE 1 (23)
NEWINGTON 1 (27)
NEWINGTON 2 (28)
PALOMA 1 (34)
Subsidence (m)
0
200
400
600
800
1000
1200
300
200
100
Age (Ma)
26/Q/253


QUILBERRY 1 (14)
QUILPIE 1 (13)
RED CAP 1 (32)
SCALBY 1 (17)
STRATHMORE 1 (18)
TARALGA 1 (26)
Subsidence (m)
0
200
400
600
800
1000
1200
300
200
100
Age (Ma)
26/Q/254


TORY BOY 1 (21)
WAREENA 1 (3)
Subsidence (m)
0
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A CRUSTAL IMAGE UNDER THE BASINS OF SOUTHERN QUEENSLAND ALONG THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT

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ABSTRACT

Deep reflection profiling data and interpretations along the Eromanga-Brisbane Geoscience Transect constrain models for the tectonic evolution of Phanerozoic Australia, and hence constrain likely models for basin evolution in southern Queensland. Basement under the central Eromanga Basin and its Devonian and Permian infra-basins is distinctly divided into upper crustal Thomson Fold Belt rocks and a lower crust of more mafic composition and/or higher metamorphic grade. Lateral differential movement of several 10s of kilometres is interpreted between these two crustal units. This movement took place along west-dipping ramps at mid-crustal depths during an Early-Middle Carboniferous orogenic period, called here the Quilpie Orogeny. The ramps probably formed during early Palaeozoic orogeny. The reflective lower crust could have been formed during two extensional/transtensional basin-forming episodes with accompanying mafic intrusion: the first during the early Palaeozoic and the second during the Devonian. Other contributing mechanisms for producing deep reflections are also likely, e.g. ductile shearing.

The Nebine Ridge, forming the southeast margin of the Thomson Fold Belt, has a different seismic fabric from that under the central Eromanga Basin farther west, suggesting a thickened and uplifted crustal feature resulting from the Carboniferous Quilpie Orogeny. A major west-to-southwest dipping geosuture (the Foyleview Geosuture) is interpreted from deep seismic data as the boundary between the Thomson Fold Belt and the crust under the Taroom Trough of the Bowen Basin to the east. The seismic fabric of the crust under this trough is markedly different from that under the Devonian basins farther west. It has few distinct features, suggesting a different evolutionary mechanism, e.g. trans-tensional model(?), foreland basin model(?). The north-south trending Meandarra Gravity Ridge along the centre of the trough is evident as topography on basement, but has little seismic expression deeper within the crust. The interpreted mass excess at mid/lower crustal levels is thought to result from mafic intrusions.

The north-south trending Burunga-Mooki Geosuture is interpreted as a high-angle/vertical fault separating the Taroom depocentre of the Bowen Basin from the uplifted New England Fold Belt to the east. In the New England Fold Belt, seismic data support the identification of a possible regional mid-crustal detachment, above which Late Carboniferous - Early Permian oroclinal bending occurred. Other decoupling surfaces are also thought possible in the upper crust. Within the orocline, a trans-tensional basin-forming mechanism appears appropriate for the Triassic basins within the fold belt. At the eastern end of the transect the west-dipping Greenbank Deep Layered Sequence, interpreted as an imbricate thrust stack/accretionary wedge in the upper-middle crust is imaged above the Brisbane Mid-crustal Detachment.

The Moho under the Thomson Fold Belt at the western end of the transect has numerous cross-cutting reflections at 36 to 42 km depth, interpreted to have formed during the Early-Middle Carboniferous Quilpie Orogeny. East of the Nebine Ridge, the Moho is well-defined by a "package" of sub-horizontal reflectors at about 12-13 s two-way time. The Moho is deepest under the Nebine Ridge at 44 km depth. East of the Foyleview Geosuture, the Moho probably re-established its present 36 km depth after the major Palaeozoic events in the region.

INTRODUCTION

During the period 1980-86, the Bureau of Mineral Resources, Geology and Geophysics (BMR) conducted seismic reflection profiling in southern Queensland along twenty-one lines totalling 2365 km in length. This included a continuous 1100 km east-west profile which extended from the central Eromanga Basin to the coast near Brisbane. This profile forms the basis for the Eromanga-Brisbane Geoscience Transect. Recording of 20 s two-way-time (TWT) records enabled structures to be imaged to depths of 60 km (assuming an average P-wave velocity of 6 km/s). The locations of these BMR seismic profiles in southern Queensland are shown in Figure 1.

Operational information for the surveys in southern Queensland has been reported by Wake-Dyster & Pinchin (1981), Sexton & Taylor (1983), and Wake-Dyster & Johnstone (1985). The recording details can be obtained from these reports; only a summary is given in this paper, together with some information on the seismic processing. There have also been a number of interpretations of the deep seismic data by Moss & Wake-Dyster (1983), Mathur (1983a,b,c, 1984), Wake-Dyster & others (1983), Finlayson & Mathur (1984), Johnstone & others (1985), Wake-Dyster & others (1985), Moss & Mathur (1986), Korsch & others (1986, 1988), Leven & Finlayson (1987), Wake-Dyster & others (1987), Finlayson & Leven (1987), Finlayson & others (1989a,b, 1990), and Leven & others (1990).

In this paper, we describe the major features of the deep reflection data which have geological significance in terms of basin development across Phanerozoic Australia. Further comments on the integration of the geological/tectonic information with the deep seismic data are contained in the precis and analogues paper at the end of this Bulletin (Finlayson, this Bulletin). The transect can be divided into four sectors for the purpose of discussing interpretations. They are, (1) the central Eromanga Basin and its infra-basins, (2) the Nebine Ridge and its overlying basins, (3) the Taroom Trough of the Bowen Basin and overlying Surat Basin, and (4) the New England Fold Belt and overlying basins. Data from these four sectors will be discussed and compared in this paper.

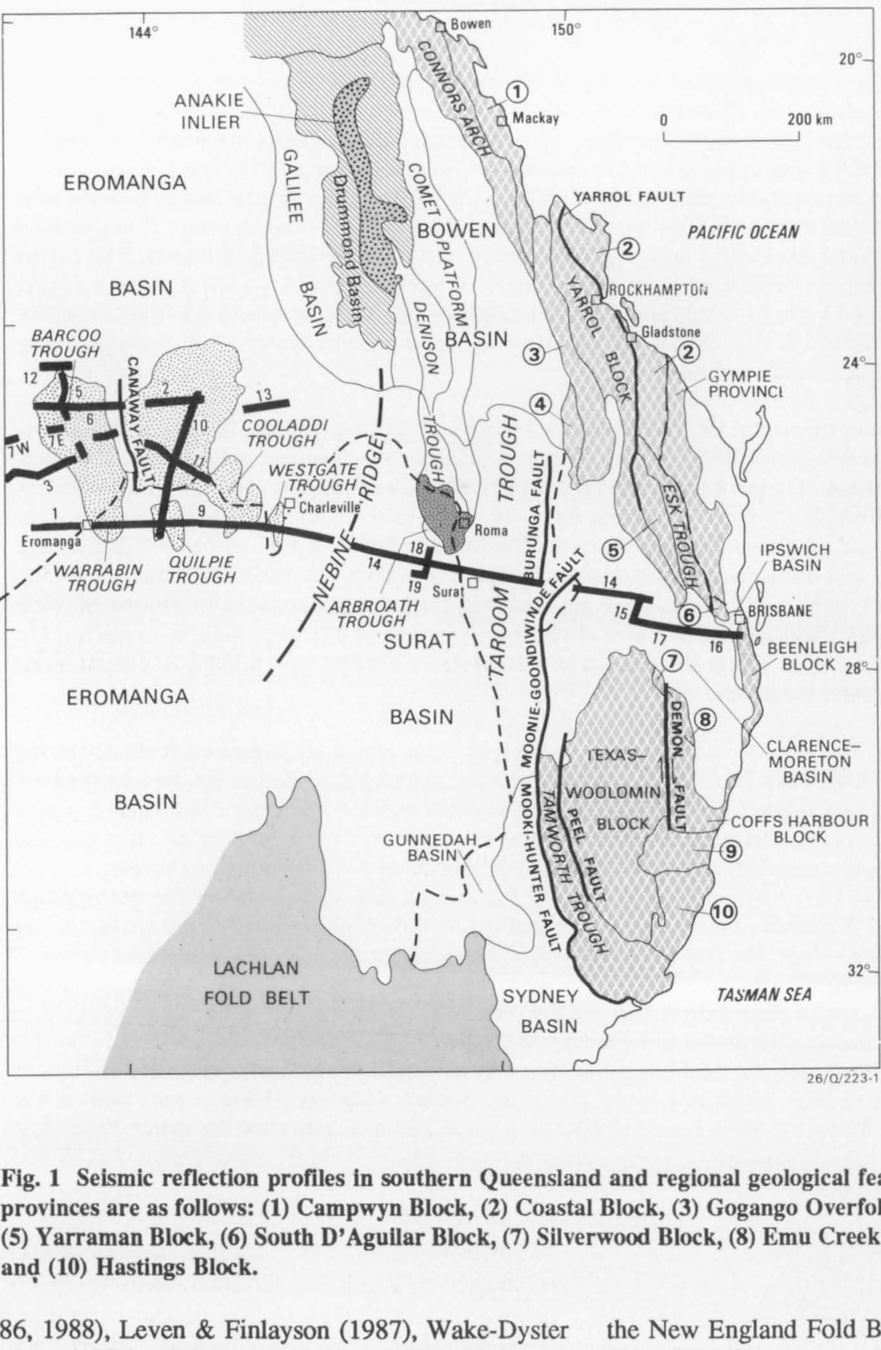


Fig. 1 Seismic reflection profiles in southern Queensland and regional geological features. Numbered geological provinces are as follows: (1) Campwyn Block, (2) Coastal Block, (3) Gogango Overfolded Zone, (4) Auburn Arch, (5) Yarraman Block, (6) South D'Aguilar Block, (7) Silverwood Block, (8) Emu Creek Block, (9) Numbucca Block, and (10) Hastings Block.

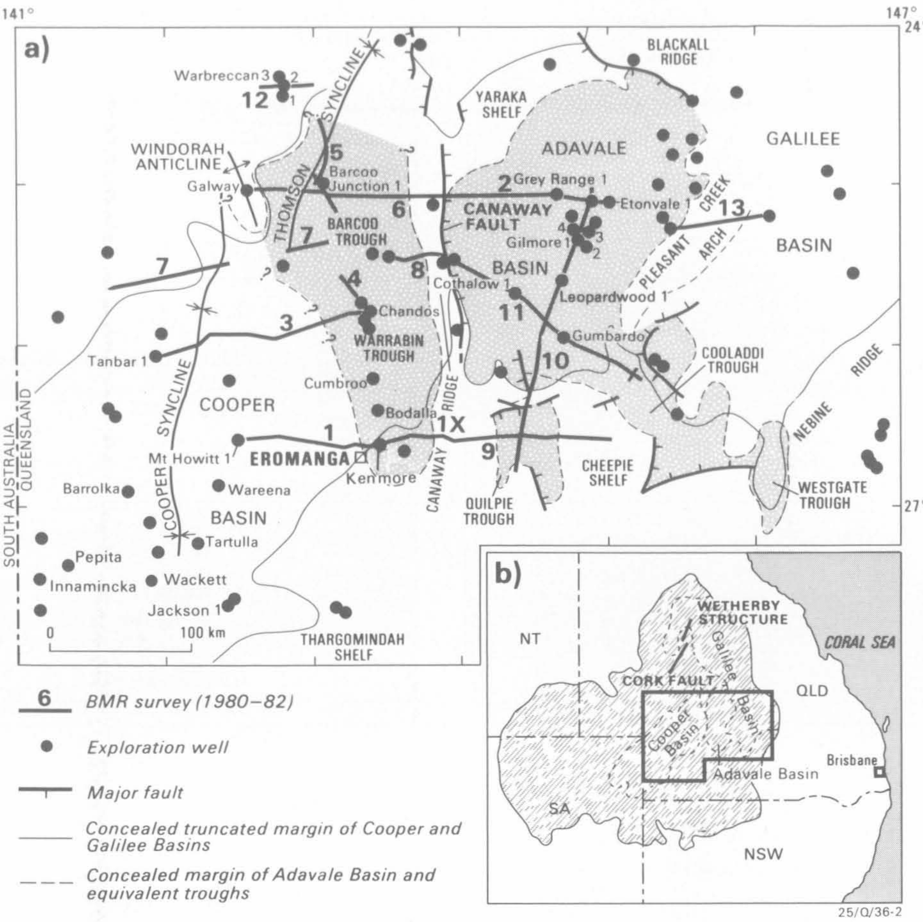


Fig. 2 General geology of the central Eromanga Basin region and the location of BMR regional deep seismic reflection profiling lines 1 to 14.

SEISMIC RECORDING AND PROCESSING

The deep seismic data in southern Queensland comprise profiling Lines 1-19 shot by the Bureau of Mineral Resources, Geology and Geophysics (BMR) during 1980 to 1986 (Fig. 1). In addition, two short separate lines were recorded at the eastern end of the transect on the Beenleigh Block, referred to as Beenleigh Lines 1 and 2. Explosive energy sources were used, commonly 8-12 kg in drillholes at 40 m depth; the geophone group interval was 83.3 m during 1980-84 and 60.0 m during 1986; generally shotpoint locations enabled 6-fold common-mid-point coverage of the subsurface; the sampling rate was 2 ms on a 48 channel Texas Instruments DFS-4 recording system during 1980-84, and on a 96 channel Sercel 368 system during 1986. Data processing was conducted by contractor during 1980-83 (GSI Australia and Digicon Australia) and by BMR processors during 1984-86 using Disco software; deep reflection imaging was improved after final stack using the Digistack enhancement routine. Along the seismic transect there are also crustal velocity profiles interpreted from BMR wide-angle reflection and refraction data (Finlayson & others, this Bulletin). In addition there are refraction data from the region near reflection Line 18 (Fig. 1), and also refraction data from the Taroom Trough about 250 km north of the transect.

Deep seismic reflection sections to 20 s two-way-time (TWT) are difficult to reduce and reproduce as illustrations. Hence, all significant reflection events from the profiles were digitized to produce line diagrams. These were then migrated using a velocity of 6 km/s to give some indication of the true positions of dipping events, assuming that the reflections were from within the plane of the seismic section. Both the line diagrams and migrated line diagrams are presented later in this paper, together with the interpretation of major features within the crust to depths of about 40 km. Also included are a few seismic sections from some areas to illustrate the quality of data. Only in the region of near-surface Tertiary volcanics in the eastern part of the transect were seismic reflection data of lesser quality because of poor signal penetration.

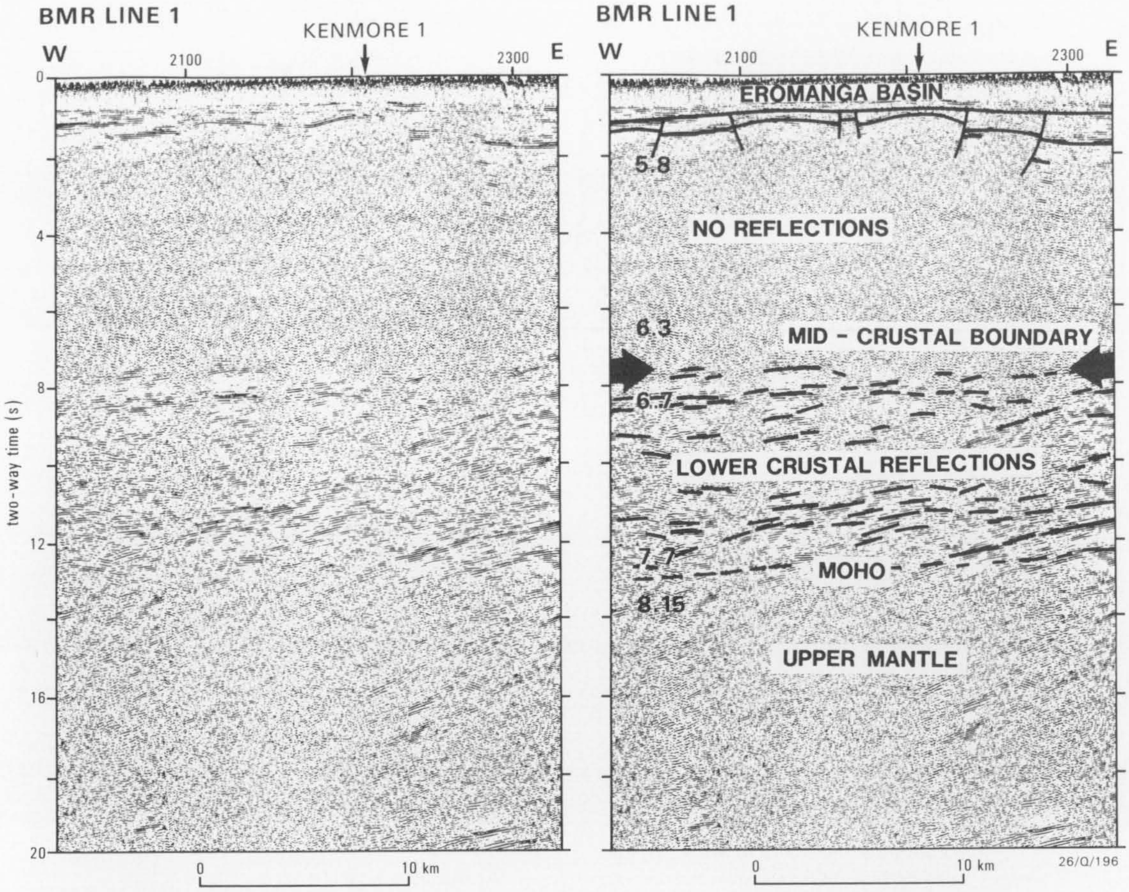


Fig. 3 Example of deep reflections from the Warrabin Trough in the central Eromanga Basin region, illustrating the major subdivisions of the continental crust based on the pattern of seismic reflections (seismic "fabric").

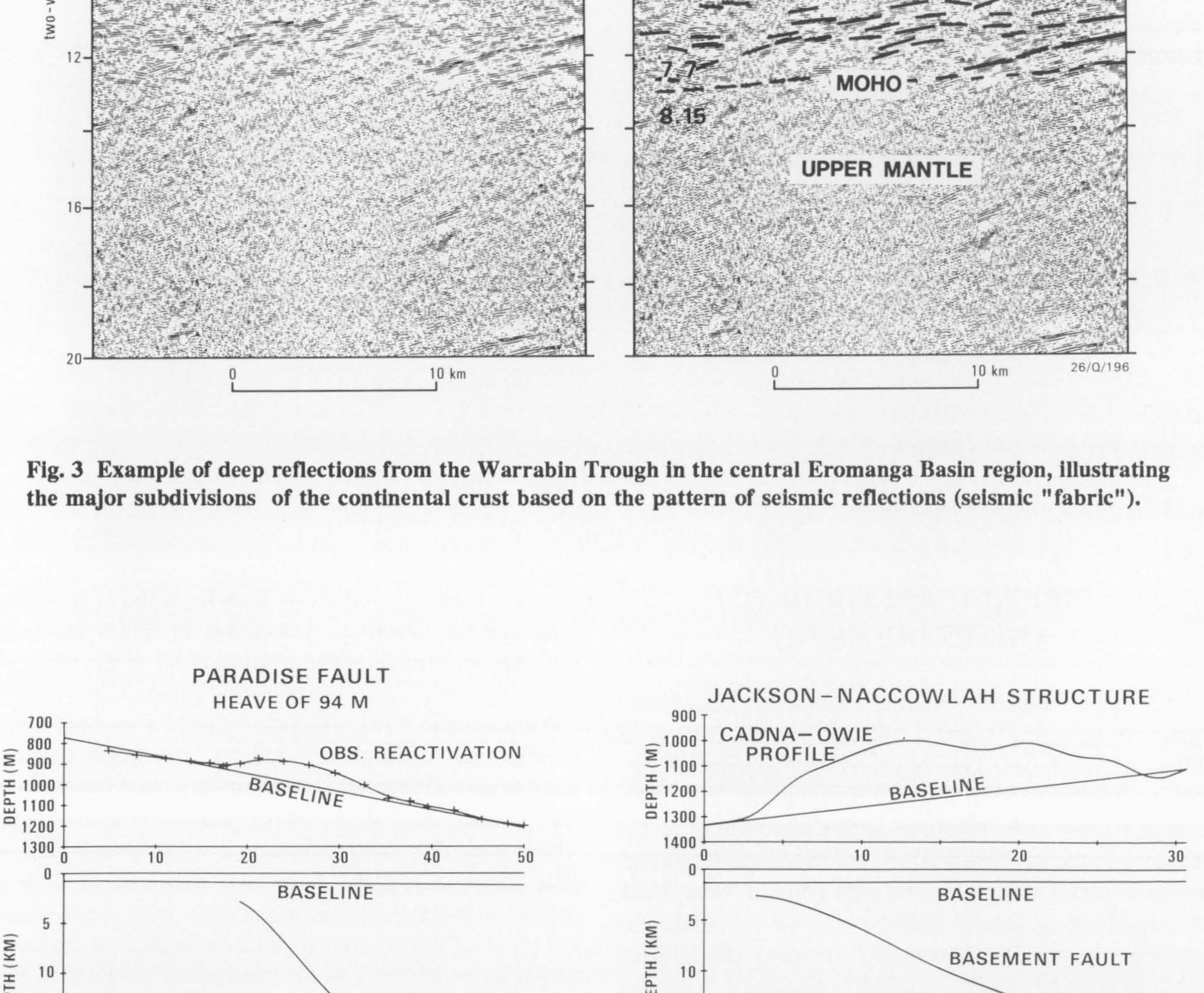


Fig. 4 Basement fault geometry calculated from the deformation of the Cadna-Owie reflection event during Tertiary reactivation of basement structures across the Paradise Fault (southern Adavale Basin) and the Jackson-Naccowlah Structure (eastern margin of the Cooper Basin) (Leven, 1986).

SEISMIC REFLECTION PROFILES

Diagrams of the BMR seismic lines 1, 9, 14, 17 and 16, at a scale of 1:1 000 000 are shown as a continuous profile in Map 4 of this Bulletin. These seismic lines comprise the main part of the Eromanga-Brisbane Geoscience Transect, and total 1100 km in length. The line diagrams of seismic data in Map 4 have been curved using an equivalent Earth's radius of 6371 km to emphasize the dimensions of the geological features being examined in southern Queensland. The lower part of the line diagram contains the migrated line data. The elevation datum level for the profile at this scale is effectively the ground elevation, which for most of the transect is between 150 and 300 m above sea level. Only near Toowoomba, at the eastern end of the transect, do elevations rise to 500-600 m across the Great Divide before dropping to 100 m and less near the coast.

The other BMR regional seismic lines are illustrated as line diagrams in this paper at 1:1 000 000 scale. The lower part of each figure has the migrated line diagrams. Intersections of the profiles are shown, illustrating places where there is some control on the 3-dimensional dip of deep horizons.

INTERPRETATION OF DEEP REFLECTIONS

Reflection seismic methods commonly applied in sedimentary basins depend on the acoustic impedances between sub-horizontally layered sequences returning seismic energy to the surface. The lithology of the sedimentary units can be interpreted using tie lines intersecting exploration wells. Lithological changes within the sedimentary sequence, such as facies changes, porosity, fluid content, etc., are considered to be the major cause of the reflected energy.

Within basement, the reflected energy still arises from acoustic impedance contrasts, but the geological reasons for these impedance contrasts are not as well understood. Within the upper crust, energy may be reflected from fault structures, plutons, metamorphic layering, intrusions, fluids, and lower crustal layered sequences thrust to higher levels in the crust (Klemperer, 1984; Allmendinger & others, 1987; Clowes & others, 1987; Green & others, 1988).

At deeper levels in the crust, the increasing temperature and pressure affect the physical properties of rocks, such as rheology. In the lower crust, therefore,

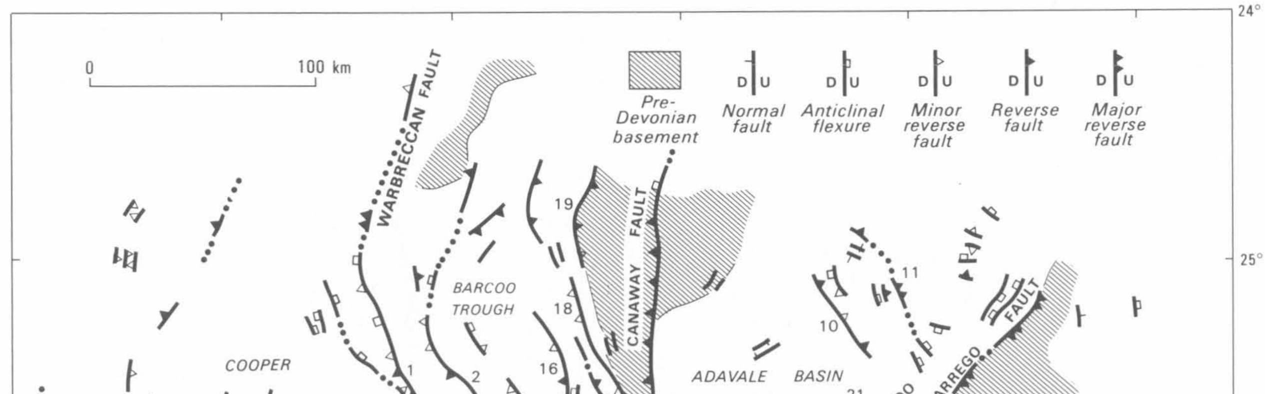


Fig. 5 Faults in the central Eromanga Basin region mapped at the pre-Permian unconformity (Finlayson & others, 1988). Numbers indicate the following faults: 1 - Windorah-Ingella; 2 - Moothandella-Chandos; 3 - Monkey-Coolah; 4 - Kenmore; 5 - Tallyabra; 6 - Bodalla; 7 - Cumbroo; 8 - Terrachie; 9 - Whynot; 10 - Grey Range; 11 - Etonvale; 12 - Paradise; 13 - Corona; 14 - Como; 15 - Como Splay; 16 - Thunda; 17 - Bulgroo; 18 - Cheviot; 19 - Jedburgh; 20 - Pingine; 21 - Leopardwood; 22 - Mount Howitt; 23 - Tintaburra.

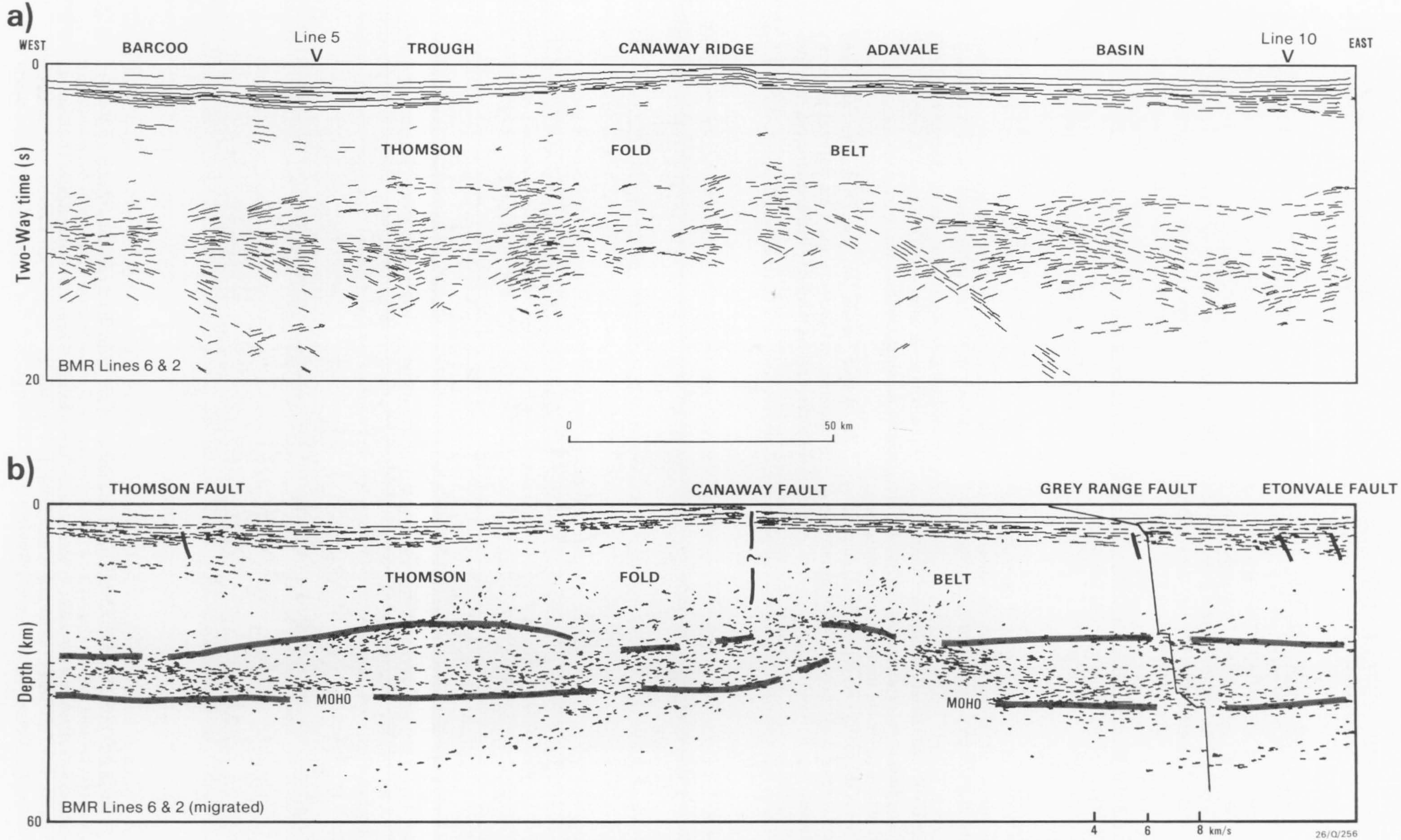


Fig. 6 Deep reflection events along BMR Lines 6 and 2 across the Barcoo Trough and northern Adavale Basin; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

ductile strain and intrusive layering resulting from the injection of sub-horizontal sheets of mantle-derived magma must be considered as the likely causes of many lower-crustal reflection characteristics (McCarthy & Thompson, 1988; Klemperer, 1987). Warner (1990) argued that the large impedance contrasts measured in the lower crust in some areas can only be derived from contrasts between lower crustal metamorphic rocks and mantle-derived mafic igneous rocks, not from ductile shearing or fluid effects. He suggests that underplating of the crust, in a process associated with lithospheric extension and/or elevated mantle temperatures, is the most plausible explanation for many lower crustal reflections. Fountain (1987), Fountain & others (1987), Smithson (1987), and Smithson & others (1987) emphasize the multiple nature of processes affecting the lower crust, especially during episodes of high heat flow and extension. In particular, the transfer of mafic material from the mantle into high-grade metamorphic rocks of the lower crust, ductile shearing, and the underplating of the lower crust are regarded as likely processes.

However, there are also lower crustal reflections in some localities, which are interpreted as fault structures penetrating deep into the lower crust. The Wind River Thrust, western U.S.A. (Smithson & others, 1979; Lynn & others, 1983) can be traced from the surface to at least 32 km depth, and in a spectacular case from northwest Scotland, reflections interpreted as being associated with a through-going crustal fault/shear zone systems are seen at depths ranging from the surface to 80 km depth in the mantle (the Outer Isles Fault and the Flannan reflections) (McGeary & Warner, 1985; Flack & Warner, 1990). Dipping reflections in the lower crust can also be interpreted as the boundaries between major terranes, e.g. the Iapetus suture in northeast England (Klemperer & Matthews, 1987). In Australia there are

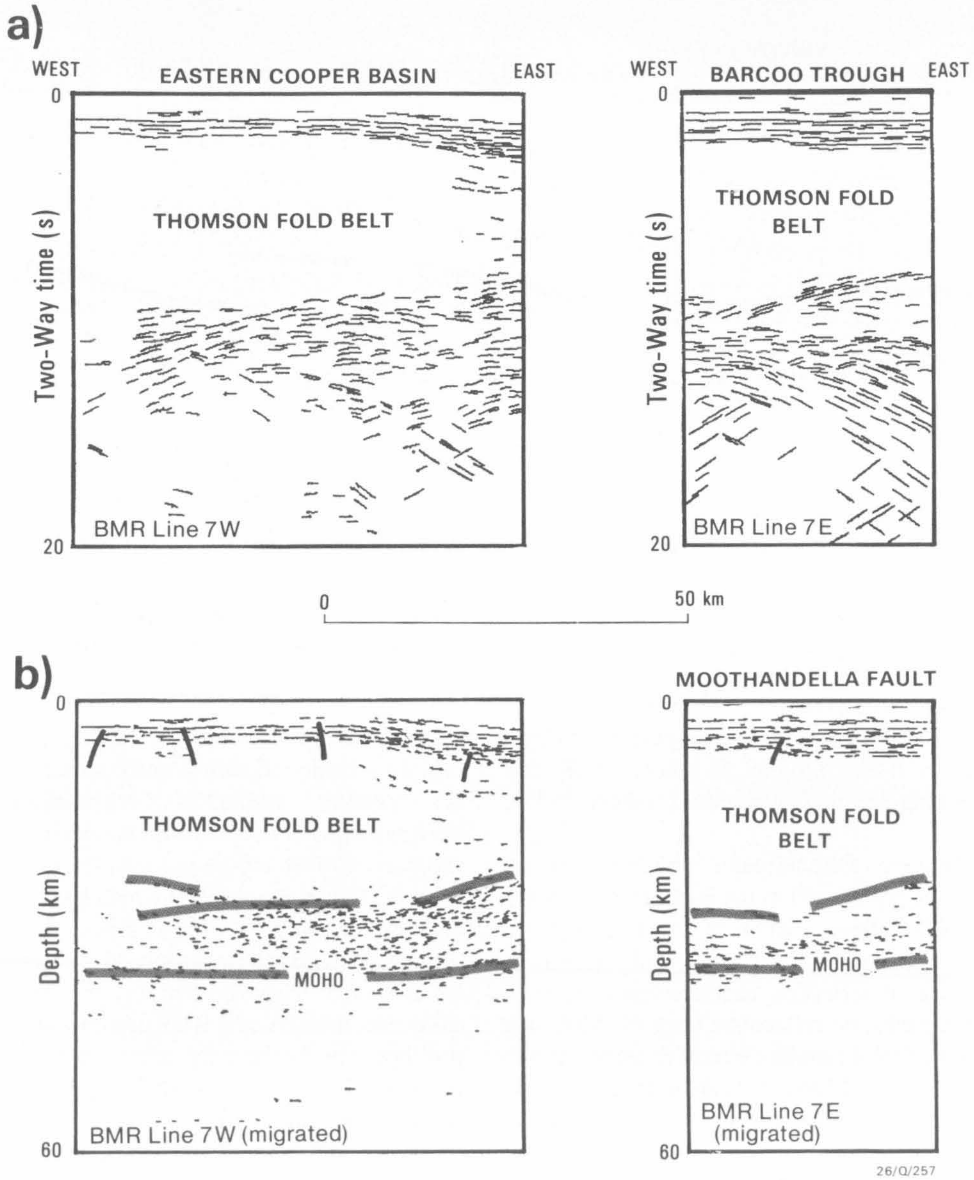


Fig. 7 Deep reflection events along BMR Lines 7(E and W) across eastern Cooper Basin and western Barcoo Trough; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

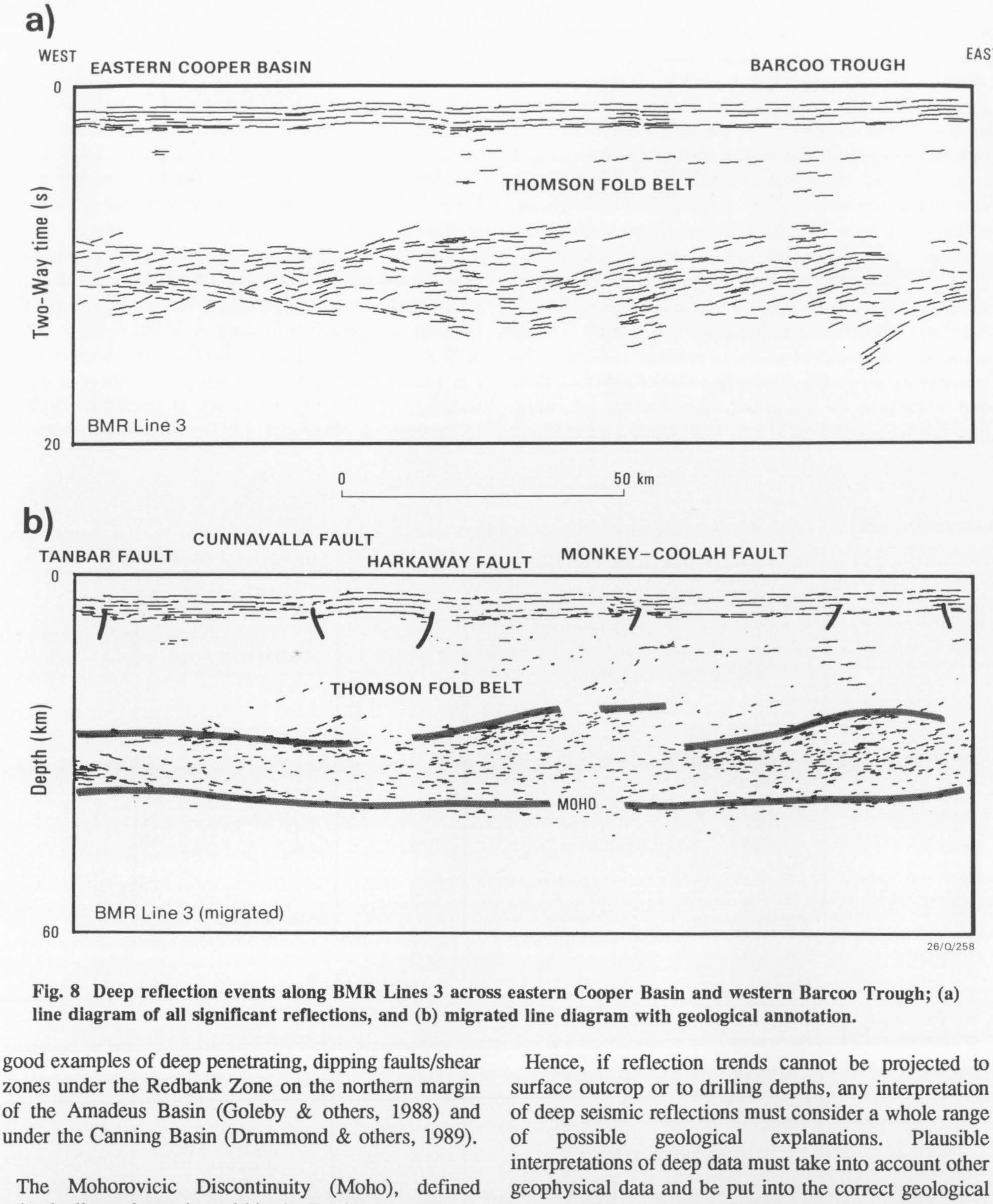


Fig. 8 Deep reflection events along BMR Lines 3 across eastern Cooper Basin and western Barcoo Trough; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

good examples of deep penetrating, dipping faults/shear zones under the Redbank Zone on the northern margin of the Amadeus Basin (Goleby & others, 1988) and under the Canning Basin (Drummond & others, 1989).

The Mohorovicic Discontinuity (Moho), defined seismically as the region within the Earth where P-wave velocities stabilize above a value of about 7.8 km/s (a value in the range 7.6 to 8.5 km/s is usually accepted), also has a range of reflection characteristics. In some Precambrian terranes, no reflections from the Moho are observed, e.g. central Australia (Wright & others, 1987). In younger areas, the Moho is evident as the deepest limit of lower crustal reflections above a non-reflective upper mantle, or as a discrete narrow band of reflections and may be interpreted as a mobile feature post-dating major orogenic events (Brown & others, 1987; Mooney & Brocher, 1987; Oliver, 1988).

Hence, if reflection trends cannot be projected to surface outcrop or to drilling depths, any interpretation of deep seismic reflections must consider a whole range of possible geological explanations. Plausible interpretations of deep data must take into account other geophysical data and be put into the correct geological context. This is the strategy which has been adopted for the interpretation of deep seismic data along the Eromanga-Brisbane Geoscience Transect. In this paper, we take into consideration (a) the features seen on extensive, industry and BMR shallow-seismic reflection profiling, (b) the velocity information from wide-angle reflection and refraction data, (c) the regional gravity and magnetic information, and (d) geological constraints on the possible models for the deep seismic reflection data.

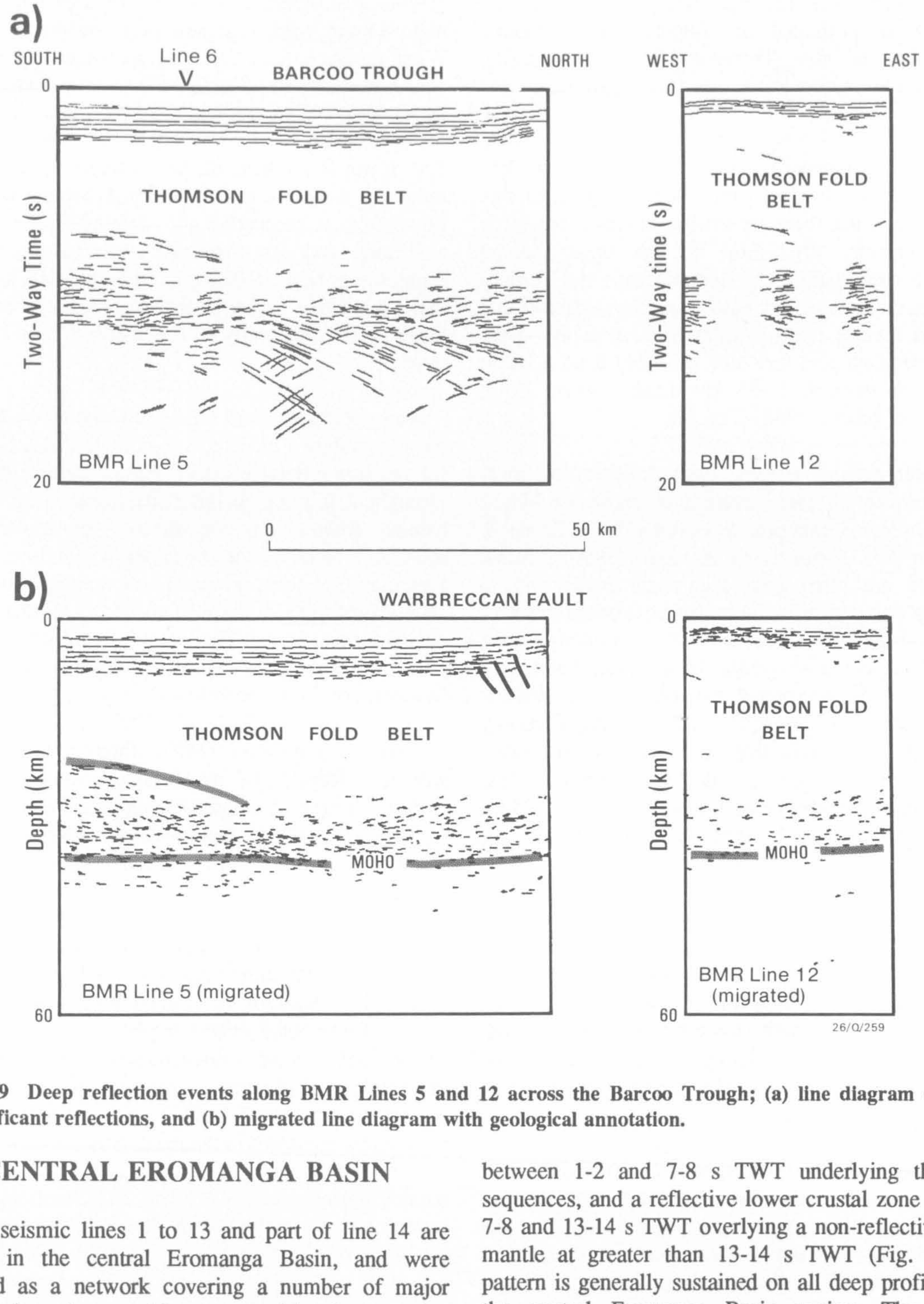


Fig. 9 Deep reflection events along BMR Lines 5 and 12 across the Barcoo Trough; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

CENTRAL EROMANGA BASIN

BMR seismic lines 1 to 13 and part of line 14 are located in the central Eromanga Basin, and were designed as a network covering a number of major geological provinces and features considered prospective for oil and gas (Moss & Wake-Dyster, 1983; Finlayson & others, 1987). Underlying the Eromanga Basin sequence are the Cooper/Galilee and Adavale Basin sequences (Evans & others, this Bulletin). The regional BMR seismic network enabled many features in the crust to be examined in detail for the first time. The BMR seismic reflection profiling network in the central Eromanga Basin is shown in Figure 2, together with the general geology of the central Eromanga Basin region.

Mathur (1983a) described the prominent features of the deep reflections in terms of a non-reflective basement between 1-2 and 7-8 s TWT underlying the basin sequences, and a reflective lower crustal zone between 7-8 and 13-14 s TWT overlying a non-reflective upper mantle at greater than 13-14 s TWT (Fig. 3). This pattern is generally sustained on all deep profiles from the central Eromanga Basin region. The marked difference between the seismic characteristics (seismic "fabric") in the upper and lower parts of the crust indicates that the processes applying in these two regions were significantly different at some stages in their history.

The non-reflective upper crustal basement to 7-8 s TWT (21-24 km depth) is identified with the Thomson Fold Belt rocks as seen in drillcores. These steeply dipping, low-grade, metasediments are intruded by Siluro-Devonian granites, and are considered to form basement under the whole of the Thomson province delineated by Wellman (this Bulletin) from gravity and magnetic data. The non-reflective seismic character ("fabric") is attributed to folding and extensive deformation of the Thomson Fold Belt during middle-to-late Ordovician orogeny (Murray, this Bulletin).

Despite the non-reflective nature of the Thomson Fold Belt rocks, detailed velocity interpretations (Collins, this Bulletin) show that there are significant lateral as well as vertical velocity differences in the upper crustal basement. Also, analysis of Carboniferous and Tertiary faulting in the sedimentary basin sequences indicates that numerous reverse faults must be present in the upper crustal basement and that they probably have a listric form which soles at 15-18 km depth (Leven, 1986; Finlayson & others, 1988) (Fig. 4).

Basement geology derived from drillcore and most faults interpreted within basement are shown on Map 1 of this Bulletin. Finlayson & Leven (1987), Leven & Finlayson (1987), and Leven & others (1990) indicate that there are also some high-angle faults with a strike-slip component in the region controlling much of the basin structure, and that these faults probably extend at least into the lower crust. In addition, Pinchin & Anfiloff (1986) interpreted the gravity and shallow seismic data over the north-south trending Canaway Fault (Fig. 2) as indicating an elogate upper crustal granite body 1-2 km in thickness intruding the metasediments on the western side of the fault. Thus, although the upper crustal basement is non-reflective, it is certainly not homogeneous or unfaulted.

The structural patterns in the central Eromanga Basin region can be divided into those (a) west and (b) east of the Canaway Ridge (Fig. 5) (Finlayson & others, 1988). West of the ridge, a series of northwest-trending, Tertiary reactivation fault structures are evident across the Barcoo and Warrabin Troughs; line diagrams of the deep seismic profiles west of the ridge are shown in Figures 6 to 10.

Finlayson & others (1989b) have drawn attention to some of features of the lower crustal reflections west of the Canaway Ridge in the Devonian Barcoo Trough area. This area did not undergo major Carboniferous deformation as did the Devonian basins/troughs farther south and east, and therefore may be more representative of pre-Carboniferous crust. They suggested that the lower crustal reflections can be divided into those that define the upper and lower limits of a lower crustal "lenticle" (Figs. 6, 7), the interior of the lenticle sometimes being relatively non-reflective.

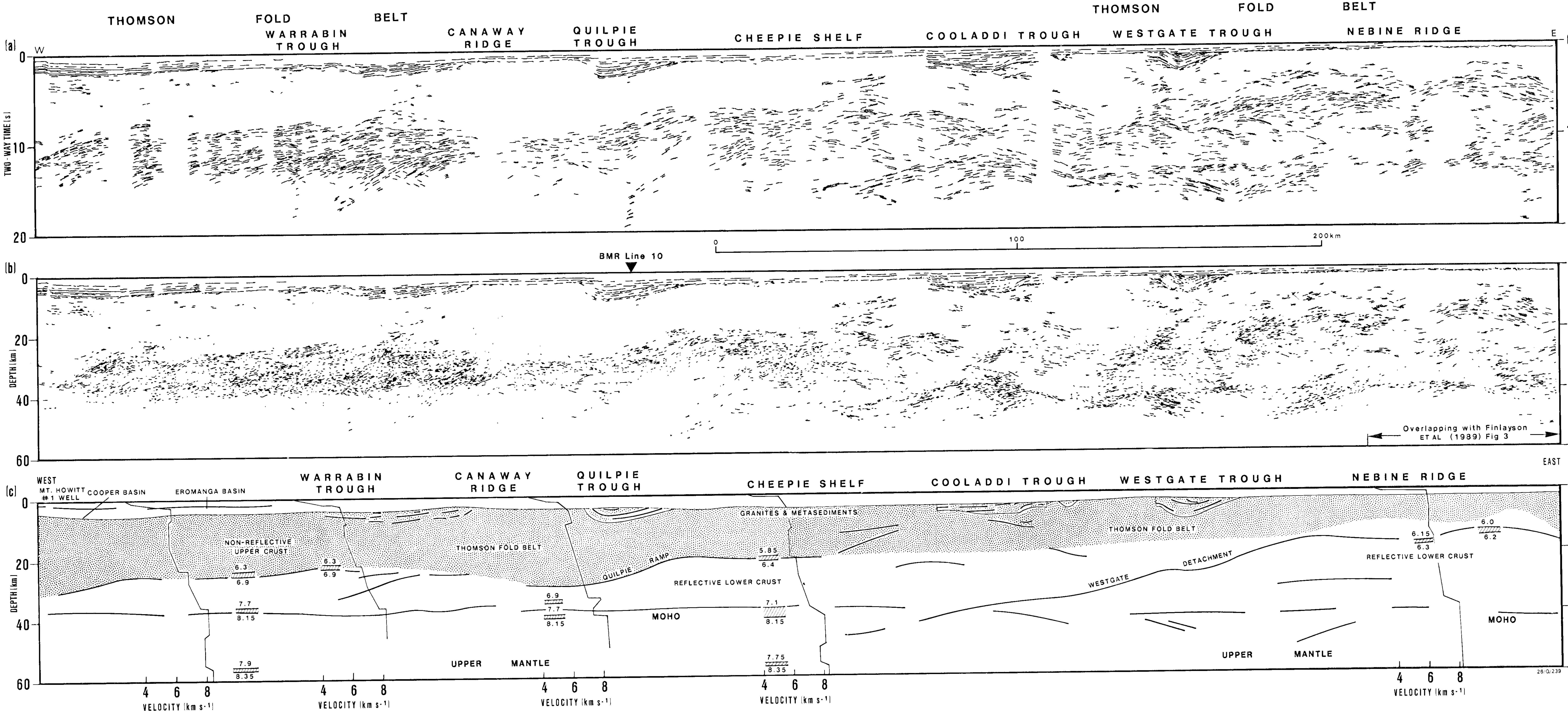
The upper boundary of the lenticles, at about 7 s TWT at their shallowest, tend to dip towards the basement highs to the west, east and north of the trough (the Windorah Anticline, Canaway Ridge and Warbreccan Dome, respectively). The upper boundary corresponds to an increase in P-wave velocity from about 6.3 km/s in the upper crust to about 6.7-6.9 km/s in the lower crust. The lower boundary of the lenticle is formed by reflectors near the poorly defined Moho, which are interpreted as truncating the dipping upper-boundary reflectors, and are therefore likely to be younger. Finlayson & others (1989b) interpret the data to indicate that a lenticle of higher density material has been added to the lower crust early in the history of the Devonian basin.

Deep reflections from the region east of the Canaway Ridge are shown in Figures 6, and 10 to 13. Generally, (a) the non-reflective upper crustal basement persists throughout this region but thins towards the east (the Nebine Ridge), (b) the density of lower crustal reflections tends to increase under the southern Adavale Basin, and (c) there are persistent dipping reflectors at Moho depths (13-14 s TWT) and deeper which obscure a well-defined Moho. The tectonic significance of these observations, together with some definite lower crustal features, are discussed below.

Leven & Finlayson (1987) interpreted the major structural features of the southern Adavale Basin - Quilpie Trough - Cooladdi Trough region as the result of Devonian - Early Carboniferous crustal-shortening events involving the lower crust. The first caused major faulting from north-south compression, e.g. the Como Fault on the northern margin of the Quilpie Trough (Fig. 14a), and the second resulted in major folding from east-west crustal shortening, e.g. the Westgate Trough (Fig. 14b). The style of lower crustal involvement in these deformational events implies that the upper and lower crust were mechanically decoupled. The deformation in the upper crust occurred on a linked system of orthogonal faults, which were significantly affected by processes in the lower crust, reverse faults in one episode being rectivated with a shear component in another episode, and vice versa. The Grenfield Fault is interpreted to result from lower crustal thrusting identified on the deep seimic data (Fig. 11).

The Early-Middle Carboniferous orogeny in the central part of the Thomson Fold Belt was the last major event to affect the region. It is of major significance and greatly affected the structures in the region and formed (among others) the Quilpie Trough. Because the events in the Qulpie Trough typify those for the region, we here refer to the event as the Quilpie Orogeny. It was

Fig. 10 (fold-out page opposite) Deep reflection events along BMR Lines 1, 1X and 9 across the eastern Cooper Basin, Warrabin Trough, Canaway Ridge, Quilpie Trough, Cheepie Shelf, and Cooladdi and Westgate Troughs; (a) line diagram of all significant reflections, (b) migrated line diagram, and (c) geological interpretation.



THOMSON FOLD BELT
WARRABIN TROUGH
CANAWAY RIDGE
QUILPIE TROUGH
CHEEPIE SHELF
COOLADDI TROUGH
THOMSON FOLD BELT
WESTGATE TROUGH
NEBINE RIDGE
(a)
W
E
TWO-WAY TIME (s)
0
10
20
0
100
200km
BMR Line 10
(b)
DEPTH (km)
0
20
40
60
Overlapping with Finlayson ET AL (1989) Fig 3
EAST
(c)
WEST
MT. HOWITT #1 WELL
COOPER BASIN
EROMANGA BASIN
WARRABIN TROUGH
CANAWAY RIDGE
QUILPIE TROUGH
CHEEPIE SHELF
COOLADDI TROUGH
WESTGATE TROUGH
NEBINE RIDGE
GRANITES & METASEDIMENTS
NON-REFLECTIVE UPPER CRUST
THOMSON FOLD BELT
THOMSON FOLD BELT
QUILPIE RAMP
REFLECTIVE LOWER CRUST
REFLECTIVE LOWER CRUST
WESTGATE DETACHMENT
MOHO
MOHO
UPPER MANTLE
UPPER MANTLE
6.3
6.9
6.3
6.9
7.7
8.15
7.9
8.35
6.9
7.7
8.15
5.85
6.4
7.1
8.15
7.75
8.35
6.15
6.3
6.0
6.2
4 6 8
VELOCITY (km s-1)
26/Q/239

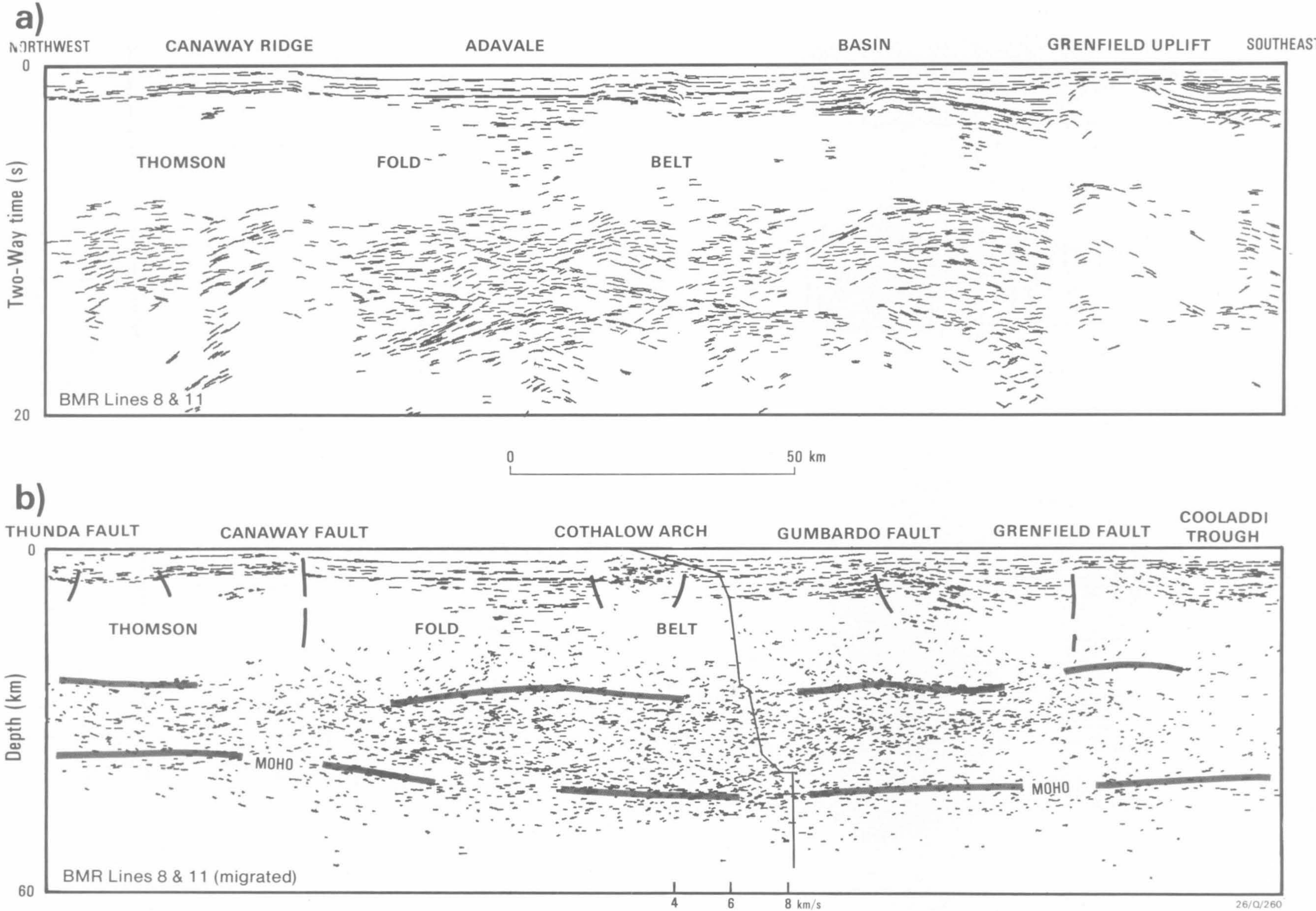


Fig. 11 Deep reflection events along BMR Lines 8 and 11 across the Adavale Basin and Cooladdi Trough; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

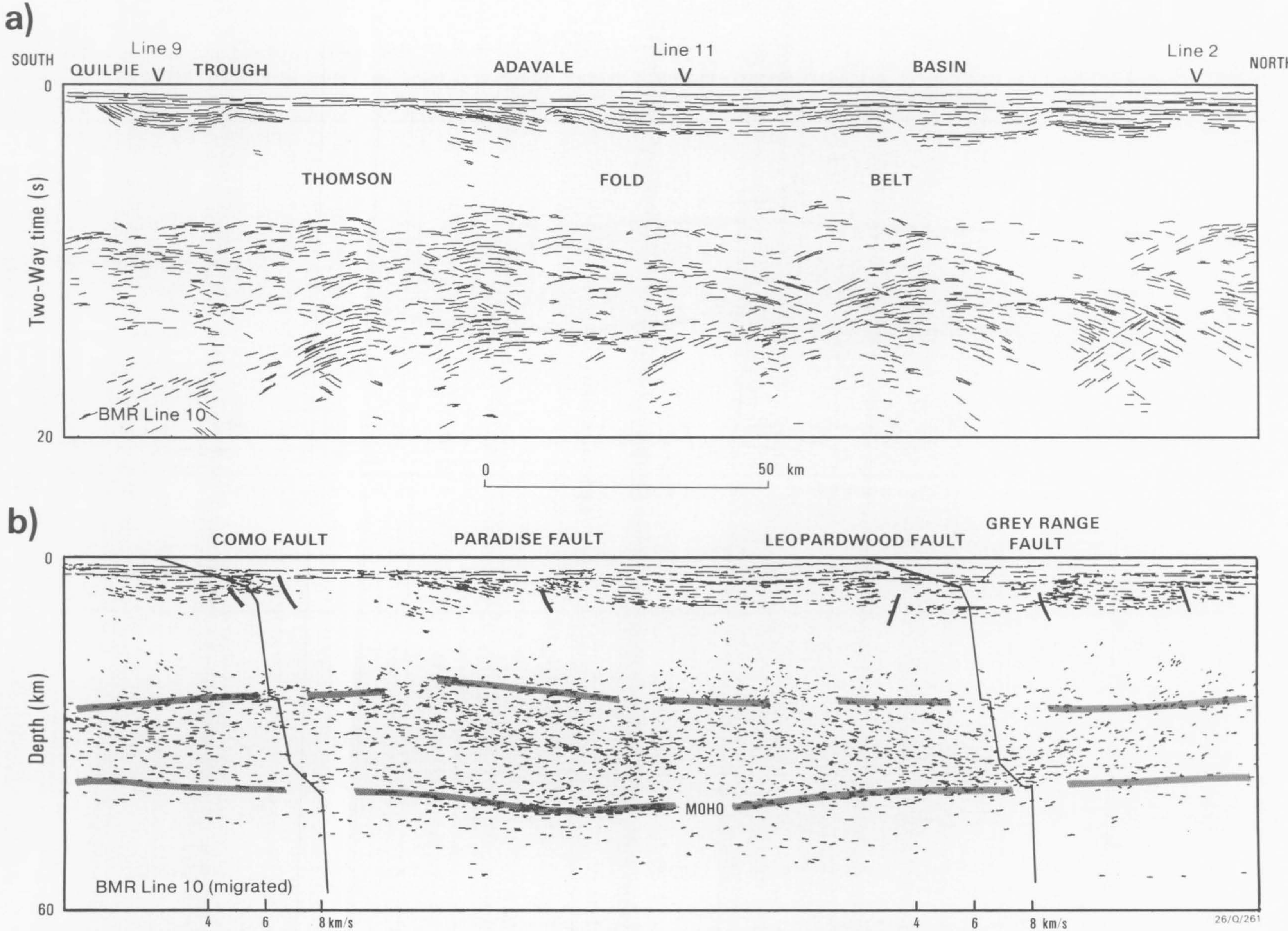


Fig. 12 Deep reflection events along BMR Line 10 across the Quilpie Trough and Adavale Basin; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

contemporaneous with the Alice Springs Orogeny in central Australia and the Kanimblan Orogeny in southeastern Australia, but, since these geological provinces are structurally distinct from the Thomson Fold Belt, we prefer a different name. Li & others (1989) have drawn attention to the rapid poleward movement of eastern Gondwanaland at that time, and it seems probable that the orogenic events in different parts of Australia are all related to this movement.

In further analysis of seismic data from the troughs containing Devonian sequences, Leven & others (1990) sought to quantify the differential movements between the upper and lower crust during Early Carboniferous times. They interpreted the deformation of the Devonian sequences of the Quilpie Trough as resulting from the movement of the upper crust over a lower crustal ramp identified in the deep seismic data (Fig. 10). The Quilpie Trough is interpreted as a ramp syncline resulting from 35 km of westward movement of the upper crust (Thomson Fold Belt and overlying Devonian sequences) over a ramp surface at mid-crustal depths dipping westward at 14°. Figure 15a illustrates the mechanism of deformation. Farther east the Devonian Westgate Trough is interpreted as being formed/deformed by a similar mechanism, only in this case there has been a two-stage process, a ramp basin being formed after an initial ramp syncline (Fig. 15b). This interpretation suggests strongly that there are possibly major west-dipping ramp detachments in the crust, which may have been reactivated during the Carboniferous Quilpie Orogeny.

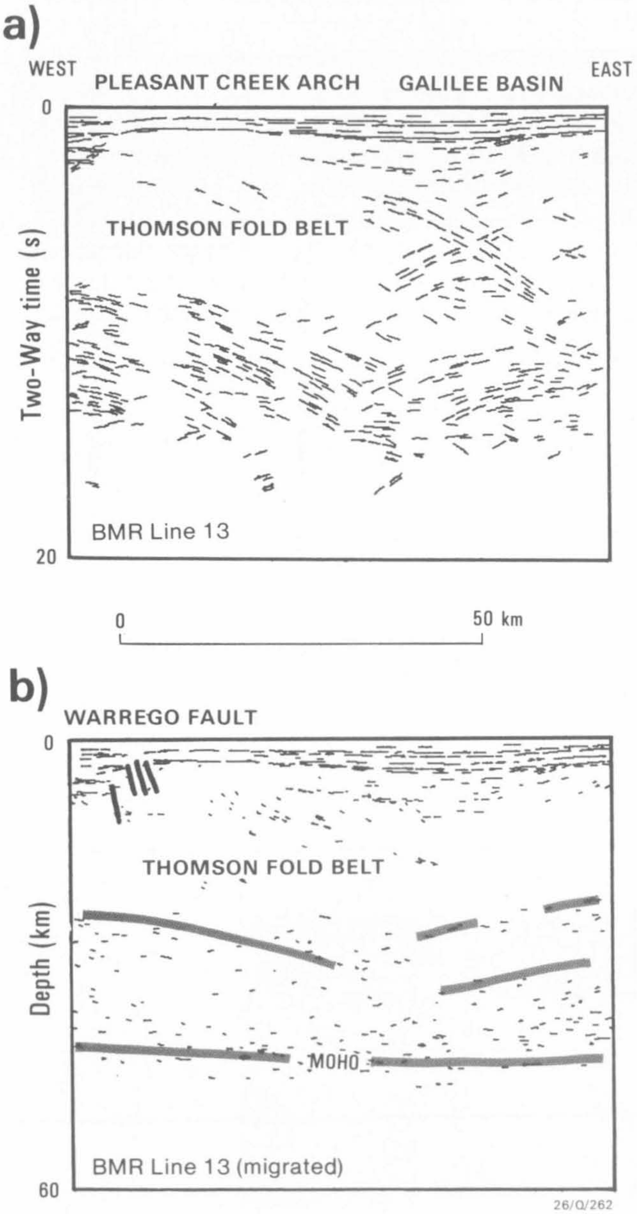


Fig. 13 Deep reflection events along BMR Line 13 across the Pleasant Creek Arch and part of the Galilee Basin; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

Many lower crustal reflections have significant dip, some continuing below 13-14 s TWT, the two-way time to the Moho identified from seismic refraction data (Finlayson & others, this Bulletin). On migration, many of these dipping events move higher in the section but do not form clear structures within the vertical plane through the seismic line. Consequently, we interpret many of the dipping events as resulting from structures out of the plane of the seismic section, but mostly still within the lower crust. We also interpret the latest time that these lower crustal reflectors could have formed/reformed as being during the Early-Middle Carboniferous Quilpie Orogeny, the time of major structuring of the Devonian sequences described above, and also the time of major continent-wide compressional/transpressional events (Murray, this Bulletin; Evans & others, this Bulletin).

The early history of the lower crustal reflections must necessarily remain conjectural but here we consider some of the likely possibilities. As discussed earlier, it seems likely that the lower crustal reflections are the product of ductile shearing and mafic intrusion into the lower crust from the mantle. There is unlikely to be agreement on the relative importance of each mechanism, but the higher velocity observed from refraction data (Finlayson & others, this Bulletin) must suggest that the lower crust is significantly more mafic than the upper crust. Ductile shearing on the detachment is an essential element of the Quilpie Trough model described above (Leven & others, 1990).

Most models for the formation of extensive basin systems involve crustal extension. In young basins there are often lower crustal reflections, suggesting that such events are associated with a predominantly extensional process. In the central Eromanga Basin region we must consider, firstly, the early Palaeozoic (presumed extensional) events terminated by the middle-to-late Ordovician Thomson Orogeny (Murray, this Bulletin) and, secondly, the Devonian extension which formed the Adavale Basin sequences and terminated by the Early-

Middle Carboniferous Quilpie Orogeny. The lower crustal reflections might well be a product of polyphase layering from both these extensional episodes. We must separate, if we can, the events which led to lower crustal reflections and those which caused its present geometry.

Finlayson & others (1989b) indicate that the character of lower crustal reflections appears to change under the basement highs between the Devonian basins/troughs, where the reflections are weaker, altered, and fewer in number. If we take the locus of extention to be under the remnant Devonian sequences where there are greater densities of lower crustal reflectors, then we infer that the majority of deep reflections are of Devonian origin with a background of less dense early Palaerozoic events. Other senarios are, however, also possible.

NEBINE RIDGE

The Nebine Ridge lies at the southeastern margin of the Thomson Fold Belt and is traversed by BMR Line 14 (Fig. 1). Murray (this Bulletin), Wellman (this Bulletin), and Finlayson & others (1990) have summarised the geological and geophysical features of the ridge. Where the Eromanga-Brisbane Geoscience Transect crosses the ridge, basement is concealed by Eromanga/Surat Basin sequences. There are several exploration seismic traverses across the ridge which assist the interpretation of basement features (Angove, 1986), as well as the BMR regional line.

Seven different reflection domains are recognized on BMR line 14 across the Nebine Ridge (Fig. 16). The uppermost layered domain is the Surat Basin sedimentary sequence. The basin is about 700 m thick at the crest of the Nebine Ridge on the seismic traverse. The underlying basement is non-reflective and considered to comprise Thomson Fold Belt rocks, whose thickness is much less than that farther west under the central Eromanga Basin (Fig. 10). The minimum thickness to the west is about 15 km, whereas the minimum thickness over the Nebine Ridge is about 6 km. The Thomson Fold Belt rocks over the ridge are thought to be uplifted (ramped upwards?) and eroded.

The mid-crustal domain underlying the crest of the Nebine Ridge has a different seismic fabric from that under the central Eromanga Basin (Finlayson & Collins, 1987). Refraction data indicate no prominent velocity increase at mid-crustal depths, only an increase from 6.0-6.15 km/s in the Thomson Fold Belt rocks to 6.2-6.3 km/s at 14-18 km depth (Fig. 16) near the top of upwardly convex reflectors. The mid-crustal Nebine Ridge structure, therefore, differs in reflection character

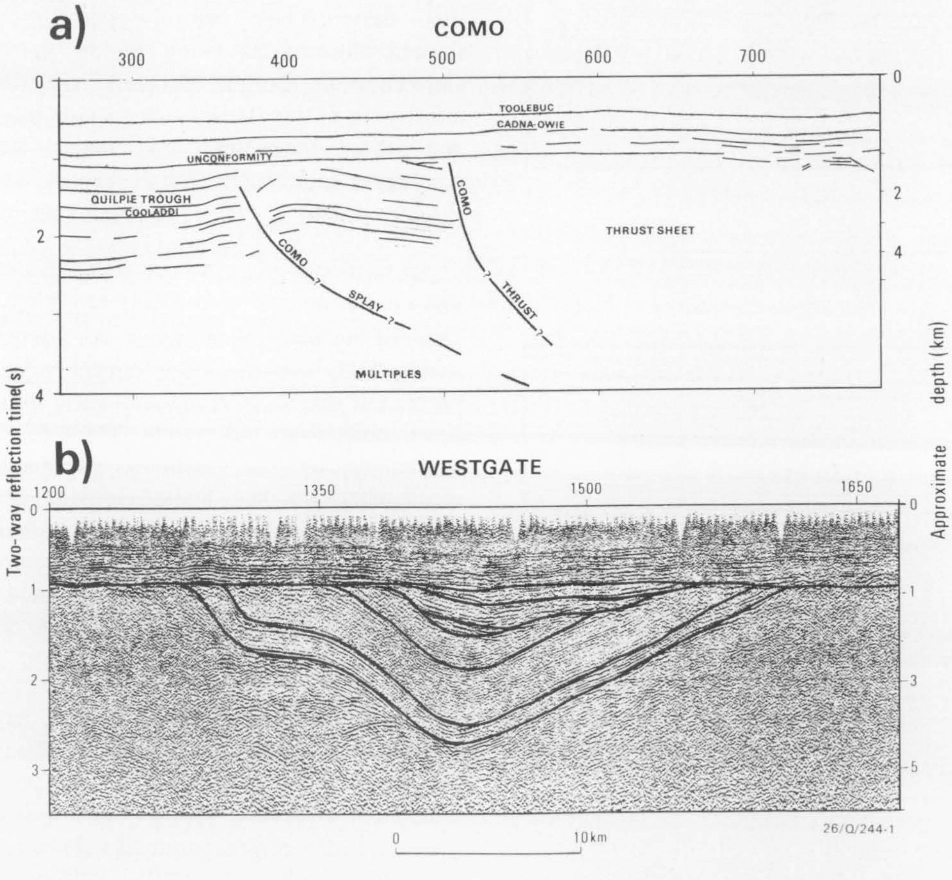


Fig. 14 Examples of major structural features in the southern Adavale Basin region: (a) faulting on the Como Fault (BMR Line 10), and (b) folding across the Westgate Trough (BMR Line 14).

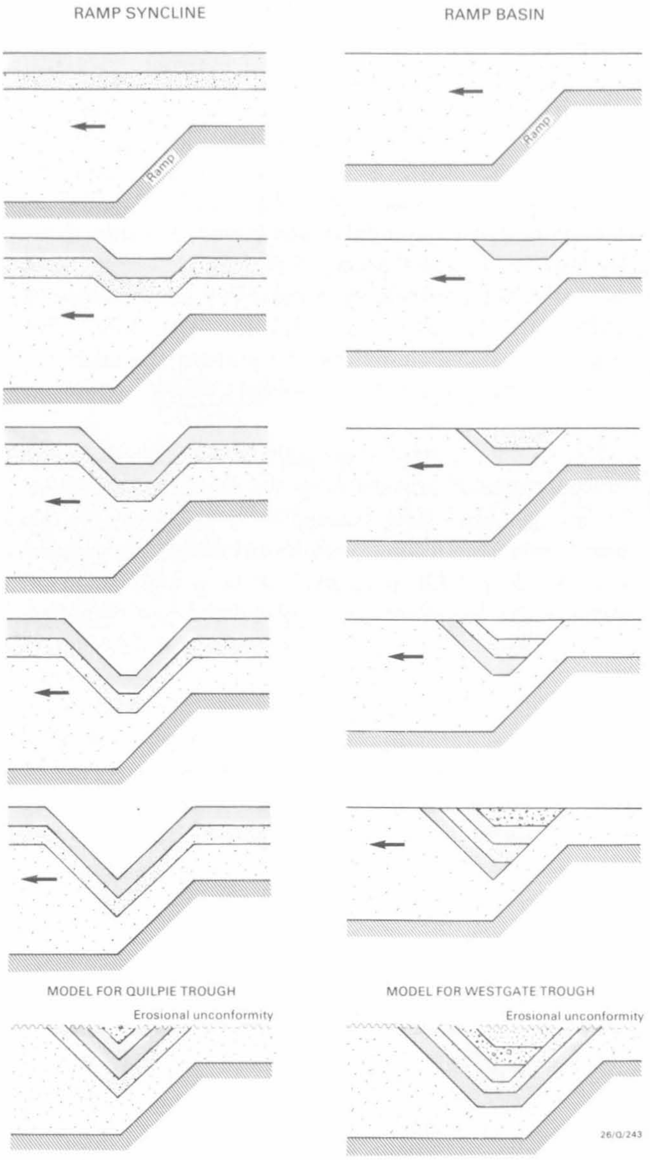


Fig. 15 Deformation mechanism for Devonian sequences by movement over mid-crustal ramp structures to form; (a) the Quilpie Trough (ramp syncline), and (b) the Westgate Trough (ramp syncline followed by a ramp basin).

and seismic velocity structure from that under the central Eromanga Basin. However, the velocity and reflectivity changes in the upper/middle crust may give clues to possible decoupling planes for any crustal deformation.

Drillcore data on the eastern side of the Nebine Ridge identifies the non-reflective upper crustal domain with the Timbury Hills Formation (folded Devonian metasediments) intruded by Carboniferous Roma granites (Murray, this Bulletin). This domain is up to 15 km thick on the western side of the Arbroath Trough. No boundary is recognised between the Timbury Hills Formation and the Thomson Fold Belt basement rocks. What is recognised from seismic, drillcore, and potential field data is the granitic terrane sub-cropping at the base of the Surat Basin sequence over the Nebine Ridge and we suggest that some of the upwardly convex reflectors at 4-6 s TWT, which coincide with increased velocity, can be interpreted as a more mafic melt fraction associated with Carboniferous plutonism.

The most significant feature at mid-crustal depths under the central and eastern side of the Nebine Ridge is a prominent series of events dipping westward (on the east-west traverse) from about 4 s TWT under the Arbroath Trough to about 10 s TWT near the apex of the Nebine Ridge (Fig. 16). This series of events clearly separates different seismic fabrics. Farther east we call it the Foyleview Structure on BMR Lines 14 and 18 (Figs. 16, 17). Under the Nebine Ridge we refer to it as a mid-crustal detachment. In total, the reflection events underlie at least 180 km of the transect. It is clearly imaged on BMR Crossline 18 (Fig. 18), with a southerly dip component as the brightest reflection events on the southern Queensland seismic lines. The velocity increase across the boundary is determined to be about 0.3 km/s from coincident refraction data. About 50 km west of the Foyleview Structure, the detachment is clearly imaged on expanding spread seismic recordings (Finlayson & others, this Bulletin).

We interpret this detachment feature to be the suture between the Thomson Fold Belt/Nebine Ridge and the crust under the Taroom Trough of the Bowen Basin; we will refer to it as the Foyleview Geosuture. The definition of a geosuture is taken from the Glossary of Geology (American Geological Institute, 1987) - "a boundary zone between contrasting tectonic units of the crust; in many places a fault which probably extends through the entire thickness of the crust." The Foyleview Geosuture coincides with the major gravity boundary identified to the west of Roma (Wellman, this Bulletin). We interpret it to be a shallow-dipping structure with an apparent dip of 5-10°, separating major crustal provinces. We speculate that the geosuture formed during the Early-Middle Carboniferous Quilpie Orogeny, accompanying the deformation and faulting in the Devonian basins farther west.

The Foyleview Geosuture does not, however, have the appearance of a simple through-going crustal fault structure as may be interpreted for, say, the Outer Isles Fault to the northwest of Scotland (McGeary & Warner, 1985), where a package of semi-continuous dipping reflectors is identified. Rather, the Foyleview Geosuture is characterised by a progressive deepening of prominent upwardly convex reflectors (diffractions?) separating domains with different seismic fabrics. Further three-dimensional work in the area would help to define the true dip and strike of the geosuture. Current data indicate that there could be a southerly component of dip on BMR Line 18 as well as the westerly dip on BMR Lines 14 and 19.

Beneath the Foyleview Geosuture is a relatively non-reflective lower crustal domain above a series of sub-horizontal events at 13-14 s TWT identified as a Moho transition zone. This latter zone is in marked contrast to the Moho events seen farther west under the central Eromanga Basin (Fig. 10). The Moho to the west has a complex structure often with cross-cutting events. East of the apex of the Nebine Ridge there is a well-defined series of reflections events above a (generally) non-reflecting upper mantle (Fig. 16).

TAROOM TROUGH OF THE BOWEN BASIN

To the east of the Foyleview Geosuture lies a simpler pattern of crustal reflectors under the Taroom Trough (Fig. 19). In the upper crust, the Jurassic-Cretaceous sequences of the Surat Basin are clearly identified across the trough, and the underlying Permo-Triassic Bowen Basin sequences are well-defined down to at least 3 s TWT. The oldest identified basin sequences are the Early Permian Back Creek Group. Some events underlying the primary reflections are multiples but, generally, the reflections from within basement under the Taroom Trough and its western margins are weaker in amplitude and continuity, and more sub-horizontal than the regions farther west. The line diagram in Figure 16, because of equal line weight everywhere, can give a false impression of strong reflectors within basement. There is, however, a well-defined package of events defining the Moho Transition Zone at 12-13 s TWT.

Within basement, no indication of a single horizon or other structure above the Moho exists, which could form a simple detachment in any purely extensional model of basin formation. However, gravity data indicate a prominent, north-south trending, positive anomaly, the Meandarra Gravity Ridge, extending the length of the trough (Wellman, this Bulletin). This gravity ridge has a value of about +20 $\mu m/s^2$ above the regional field near the seismic traverse. Qureshi (1984, 1989) interpreted the continuation of this anomaly in the Sydney Basin in terms of mafic volcanics at a middle/high level in the crust. If the same interpretation is accepted for the Meandarra Gravity Ridge, then the volcanics do not produce strong reflections.

On the eastern margin of the Taroom Trough, the Permo-Triassic sequences have been uplifted across both the Burunga and Moonie Faults with an intervening embayment. Industry seismic data show that there are Carboniferous sequences (Kuttung Formation) east of the Moonie Fault with many features of overthrusting. This eastern boundary of the Taroom Trough is formed by a major north-south trending fault system, the Burunga and Moonie-Goondiwindi Faults. These faults extend south into the Mooki-Hunter Faults (Murray, this Bulletin). The BMR seismic line crosses this fault system where the northeast-trending Moonie Fault diverges from the north-trending Burunga Fault. The deep seismic data show this structural complexity with non-reflective zones under both major faults (Fig. 19; Fig. 16, U1 and U2). This whole region is referred to here as the Burunga-Mooki Geosuture. The deep seismic fabric changes markedly across the geosuture with the reflections to the east in the New England Fold Belt displaying greater continuity and complexity than those to the west. Just to the east of the Burunga Fault there are mid-crustal reflections at 8 s TWT similar to those under the central New England Fold Belt, and interpreted as indicating that the uplifted crustal block may be related to the New England Orogen.

The Burunga-Mooki Geosuture is the second major crustal boundary traversed by the BMR seismic line. Unlike the Foyleview Geosuture to the west, all the indications from the Burunga-Mooki Geosuture suggest that the deep fault structures are at a high angle or vertical. As has been found elsewhere, e.g. the San Andreas Fault, the images of near-vertical faults/fault zones on seismic reflection profiles are poor. However, the reflection fabric across such zones is commonly different, and this is the case across the Burunga-Mooki Geosuture between Taroom Trough of the Bowen Basin and the New England Fold Belt. Within the upper crust, however, the structural features within the Permo-Triassic and Carboniferous sequences suggest thrust structures that may be detached at relatively shallow depths (9-12 km).

NEW ENGLAND FOLD BELT

The seismic data across the central part of the New England Fold Belt can be related to three upper crustal domains associated with accretionary wedge development, oroclinal bending, and Triassic basin formation (Korsch & others, this Bulletin). These domains are (1) the region near the Texas Orocline in the west, (2) the Triassic trans-tensional basins in the central part of the orogen, and (3) the Beenleigh Block and Greenbank Deep Layered Sequence in the east.

The crustal fabric about 30 km north of the gravity/magnetic expression of the Texas Orocline (Korsch & others, this Bulletin) is illustrated in Figure 20. Within basement, a complex series of reflections throughout the crust above a Moho Transition Zone at 11-12 s TWT is observed. In the upper crust, reflections tend to be convex upwards and are interpreted to be related to complexly deformed and faulted upper crustal

Fig. 16 (fold-out page opposite) Deep reflection events along BMR Lines 14, 17, and 16 across the Nebine Ridge, Taroom Trough and New England Fold Belt; (a) line diagram of all significant reflections, (b) migrated line diagram, and (c) geological interpretation.

THOMSON FOLD BELT
BOWEN BASIN
NEW ENGLAND FOLD BELT
a)
NEBINE RIDGE
FOYLEVIEW STRUCTURE
ARBROATH TROUGH
TAROOM TROUGH
BURUNGA FAULT
MOONIE FAULT
TEXAS OROCLINE (30KM SOUTH)
MAIN RANGE VOLCANICS
ESK TROUGH
IPSWICH BASIN
BEENLEIGH BLOCK
TWO WAY TIME (s)
0 100 200 km
Vertical / Horizontal scales = 1 approximately
b)
BMR Line 18
BMR Line 15
DEPTH (km)
OVERLAPPING WITH FIG 10
c)
NEBINE RIDGE
SURAT BASIN (JURASSIC - CRETACEOUS)
FOYLEVIEW GEOSUTURE
BOWEN BASIN / TAROOM TROUGH (PERMOTRIASSIC)
BURUNGA-MOOKI GEOSUTURE
CECIL PLAINS SYNCLINE (TRIASSIC)
TERTIARY VOLCANICS
CLARENCE - MORETON BASIN (JURASSIC - CRETACEOUS)
ESK TROUGH (TRIASSIC)
IPSWICH BASIN (TRIASSIC)
BEENLEIGH BLOCK
THOMSON FOLD BELT
TIMBURY HILLS BEDS
FOYLEVIEW STR
DETACHMENT
MULTIPLES
UNCONFORMITY
MAFIC VOLCANICS ?
MAFIC INTRUSION
POOR DATA
NEARA VOLCANICS
GREENBANK DEEP LAYERED SEQUENCE
BRISBANE MID - CRUSTAL DETACHMENT
MID - CRUSTAL DETACHMENT
TEXAS MID - CRUSTAL DETACHMENT
LOWER CRUST
MOHO TRANSITION ZONE
MOHO
UPPER MANTLE
VELOCITY (km s⁻¹)
28/Q/225

tend to be convex upwards and are interpreted to be related to complexly deformed and faulted upper crustal rocks of a Palaeozoic forearc sequence (Finlayson & others, 1990). At 7-8 s TWT, a series of strong reflections divides the upper and lower parts of the crust. Tectonic modelling requires that oroclinal bending occurs on an extensive regional detachment surface and the mid-crustal reflections is tentatively interpreted as being associated with that surface, here called the Texas

another (shallower) decoupling surface cutting down to the Texas Mid-crustal detachment under the Cecil Plains Syncline. This decoupling surface could be associated with the overthrust features seen in the Carboniferous sequences (Kuttung Formation) near the Moonie Fault.

In the central part of the New England Fold Belt, the Main Range (Tertiary) Volcanics inhibit imaging of the deep crust. However, a marked difference in crustal

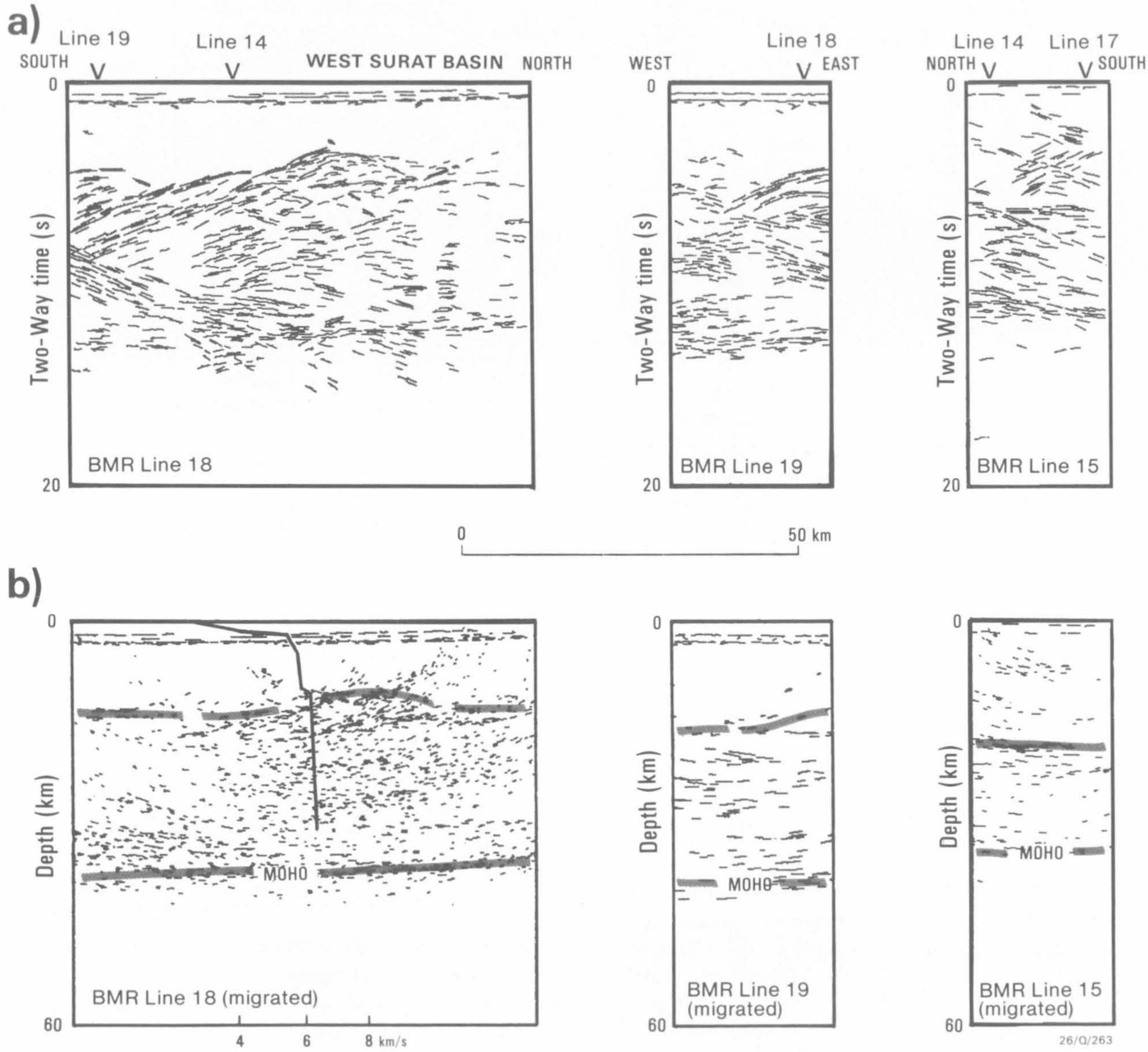


Fig. 17 Deep reflection events along BMR Lines 18, 19 and 15, in the Mitchell Shelf and Kumbarilla Ridge regions; (a) line diagram of all significant reflections, and (b) migrated line diagram with geological annotation.

Mid-crustal Detachment. It is clearly defined on both east-west BMR Lines 14 & 17 and connecting north-south BMR Line 15. From seismic refraction data a mid-crustal velocity increase is interpreted, and the transition to the Moho is at 36 km depth (Fig. 16).

Within the upper crust east of the Moonie Fault (Fig. 16), reflections at 3-4 s TWT may be interpreted as

fabric east and west of the volcanics highlights a gross crustal difference. Wellman (this Bulletin) demonstrates that the northward continuation of the Demon Fault, which has 23 km of dextral displacement, also displaces gravity trends near the seismic line. Murray & others (1987) surmised that this location is the position of a hypothetical Palaeozoic strike-slip fault, the Gogango - Baryugil Fault. The seismic data strongly suggest

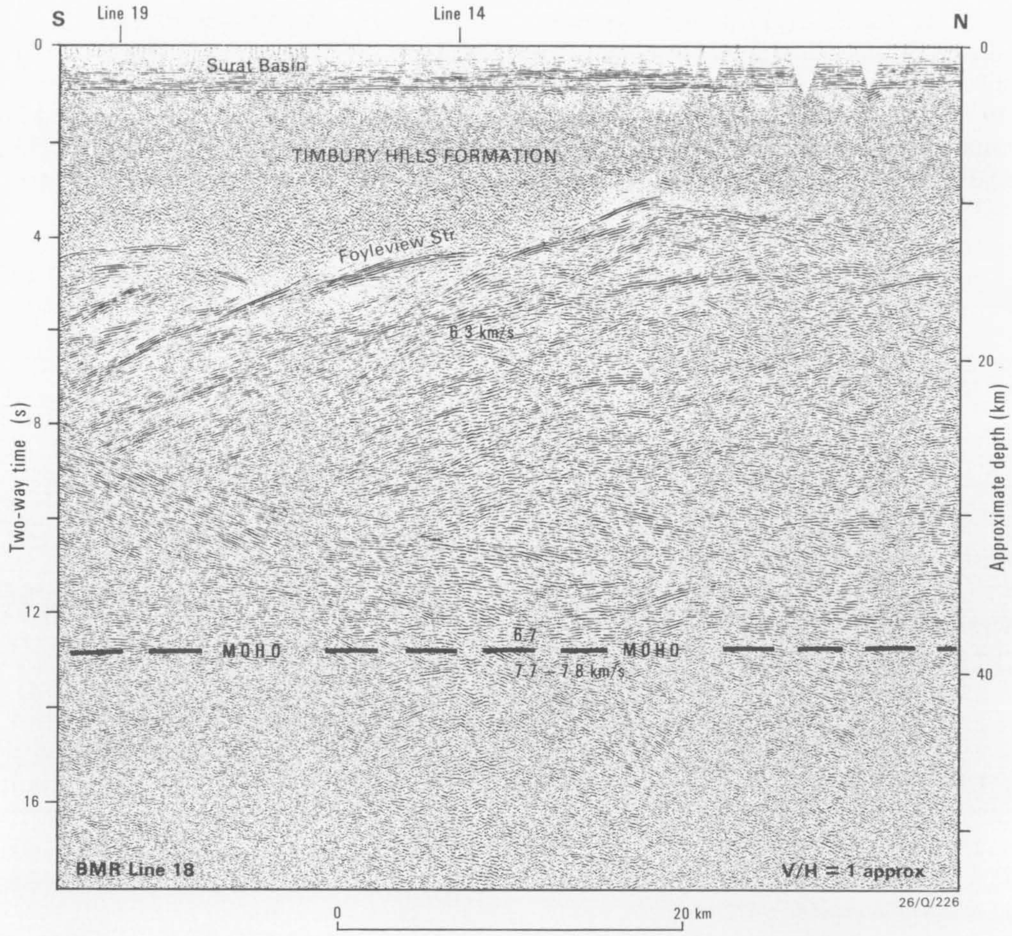


Fig. 18 Seismic reflection data along BMR Line 18 with major horizon at about 4 s TWT, the Foyleview Structure.

significantly different crustal domains to the east and west of the volcanics.

Between the Main Range Volcanics and the Beenleigh Block (Fig. 16) lie a series of elongate Triassic basins, interpreted by Korsch & others (1988) to have formed by a transtensional mechanism. Basement under the Esk Trough, imaged by the deep seismic data, contains eastward dipping reflection events in the upper crust interpreted to be associated with a volcanic rift sequence (Fig. 21). The eastern margin of the Esk Trough is a high-angle fault, the West Ipswich Fault. The South Moreton Anticline separates the Esk Trough from the Ipswich Basin. This basin also has early transtensional features, but has the main bounding fault on its western side, as seen on exploration seismic data farther south (O'Brein & others, this Bulletin).

The most significant feature of the basement under the Ipswich Basin is a series of west-dipping reflections between 2 and 8 s TWT. This feature, here called the Greenbank Deep Layered Sequence, extends to mid-crustal levels (8-9 s TWT) where there are sub-horizontal reflections. We interpret these latter reflections to be the detachment (here called the Brisbane Mid-crustal Detachment) above which oroclinal bending occurred. Korsch & others (1986) interpreted the Greenbank Deep Layered Sequence as an imbricate thrust stack of sedimentary rocks 20 km thick. A remnant accretionary wedge model such as that seen under Vancouver Island (Clowes & others, 1987) is thought possible. At the eastern end of the seismic transect is the Beenleigh Block, a non-reflective wedge with a lower boundary dipping shallowly to the east. This block is either an exotic terrane (Korsch & others, 1986) or a displaced and rotated fragment of the accretionary wedge (Murray & Lohe, 1987).

Right across the New England Fold Belt, the Moho varies only slightly from 12 s TWT. This agrees with the seismic refraction depth of 36 km (Finlayson & others, this Bulletin). Variations in the crustal fabric in the eastern and western parts of the oroclinal system tend to be most obvious above the Texas and Brisbane Mid-crustal Detachments. The lower crustal domain is a significant feature under the detachments, being at least 12 km thick. Its origin is speculative. In the east, it possibly truncates dipping features of the Greenbank Deep Layered Sequence, inferring that it postdates the formation of upper crustal sequences.

MOHO TRANSITION ZONE

The depth to the Mohorovicic discontinuity is determined in several places by seismic refraction measurements, indicating that the top of the non-reflective zone at 12-14 s TWT can be equated with the top of the upper mantle (Finlayson & others, this Bulletin). Although some reflections have greater two-way times there is no strong evidence that they are primary events in the plane of the section from within the upper mantle. Above this non-reflective region, right along the transect from the Nebine Ridge to the coast, is a package of discontinuous, sub-horizontal events about 3 km thick (1 s TWT), interpreted to be the Moho Transition Zone. This zone is a prominent feature of the seismic data.

The thickness of crustal basement rocks along the transect varies from 40 km under the Nebine Ridge to 28 km under the Taroom Trough. The deepening of the Moho west of the Taroom Trough appears to mirror, in part, the deepening of the Foyleview Geosuture. This suggests that the whole of the lower crust responded to the process resulting in the geosuture. Under the Burungi-Mooki Geosuture there is no evidence of a displacement of the Moho. If tectonic evolution in the region involved either substantial crustal thinning during basin formation or any form of major crustal thickening and loading in the process of oroclinal bending/foreland thrusting, then these effects have largely disappeared during subsequent mobilization of the Moho.

CONCLUSIONS

Across Phanerozoic eastern Australia, four major crustal sub-divisions are identified from seismic profiling data in southern Queensland. Their crustal architecture can be recognised from seismic fabric characteristics and velocity information. They occur under (west to east) the central Eromanga Basin, the Nebine Ridge, Taroom Trough, and New England Fold Belt, with each crustal sub-division having its own internal architecture. Two major bounding structures (geosutures) are also recognised.

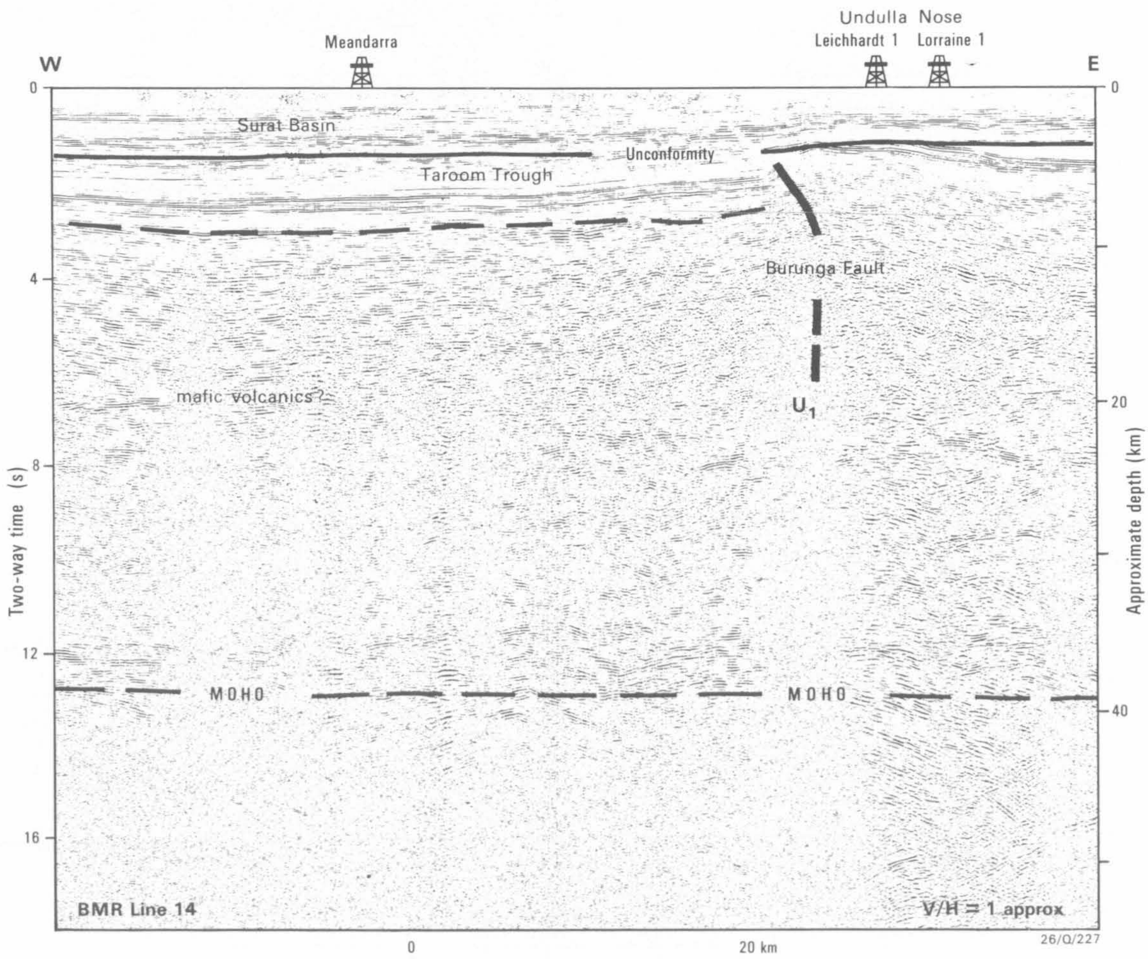


Fig. 19 Seismic reflection data along BMR Line 14 across central Taroom Trough and its eastern faulted margin (U1), the Burunga-Mooki Geosuture.

The significant features are as follows:

(1) Under the central Eromanga Basin there is a distictly layered crustal architecture. A Palaeozoic-Mesozoic basin sequence overlies a non-reflective upper crustal basement, the Thomson Fold Belt. The lower crust exhibits distinctive reflectivity from a region of higher velocity deeper than 20 km under the principal Devonian depocentres. Under basement highs, such as the Canaway Ridge, the lower crustal reflectors are less significant, suggesting at least two stages of crustal extension caused the reflectivity, one prior to the Thomson Orogeny (middle to late Ordovician), the other during the formation of the Devonian Adavale Basin. Reactivation of Thomson Fold Belt structures as reverse faults and folds during the Early-Middle Carboniferous Quilpie Orogeny and during Tertiary compressional events has been the major factor controlling deformation of the Eromanga Basin and its infra-basins.

(2) Major Early-Middle Carboniferous folding of the Devonian sequences during the Quilpie Orogeny indicates that the upper crust probably moved with respect to the lower crust by at least 35 km along detachment surfaces at mid-crustal levels in some places. These movements may have been along major west-dipping mid-crustal ramp structures established during the Thomson Orogeny.

(3) The Nebine Ridge forms the southeastern part of the Thomson Fold Belt, but has a crustal architecture distinctly different from that under the central part of the fold belt, tentatively supporting an evolutionary model involving fragmentation and eastward stretching and thinning of a Precambrian Australian craton. The ridge crust has possibly been considerably altered by Carboniferous plutonism and compressional events.

(4) Between the Nebine Ridge and the crust under the Taroom Trough of the Bowen Basin, a major low angle (5-10°) geosuture (the Foyleview Geosuture) is interpreted, extending from upper to lower crustal levels along a series of prominent horizons. It was probably active during the major Early-Middle Carboniferous Quilpie Orogeny, which severely affected Devonian basins west of the Nebine Ridge. The geosuture forms the southeastern boundary of the Thomson Fold Belt.

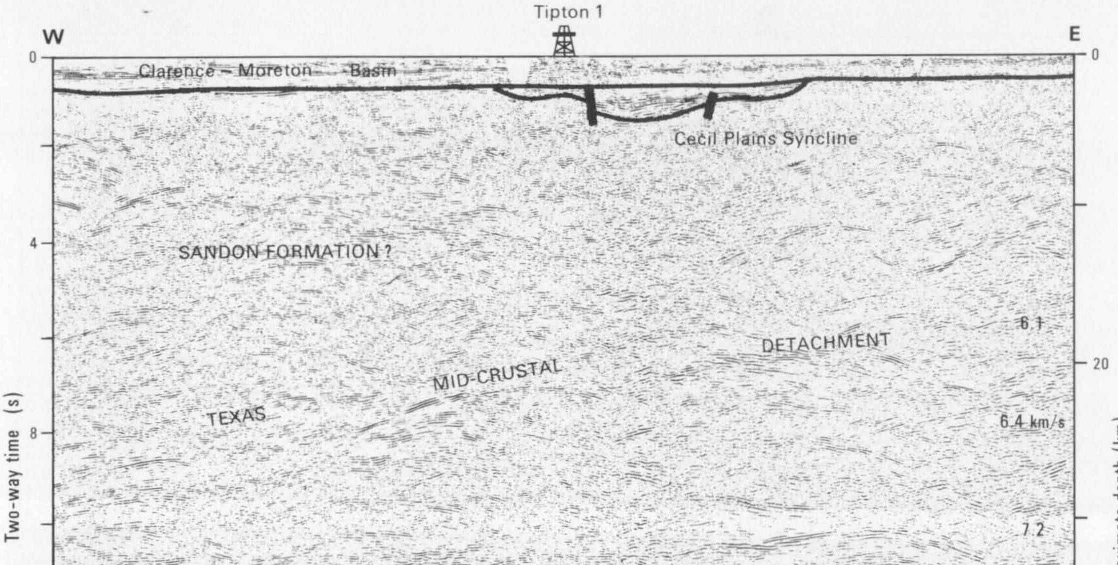


Fig. 20 Seismic reflection data along BMR Line 14 across the western part of the New England Fold Belt about 30 km north of the gravity and aeromagnetic expression of the Texas Orocline.

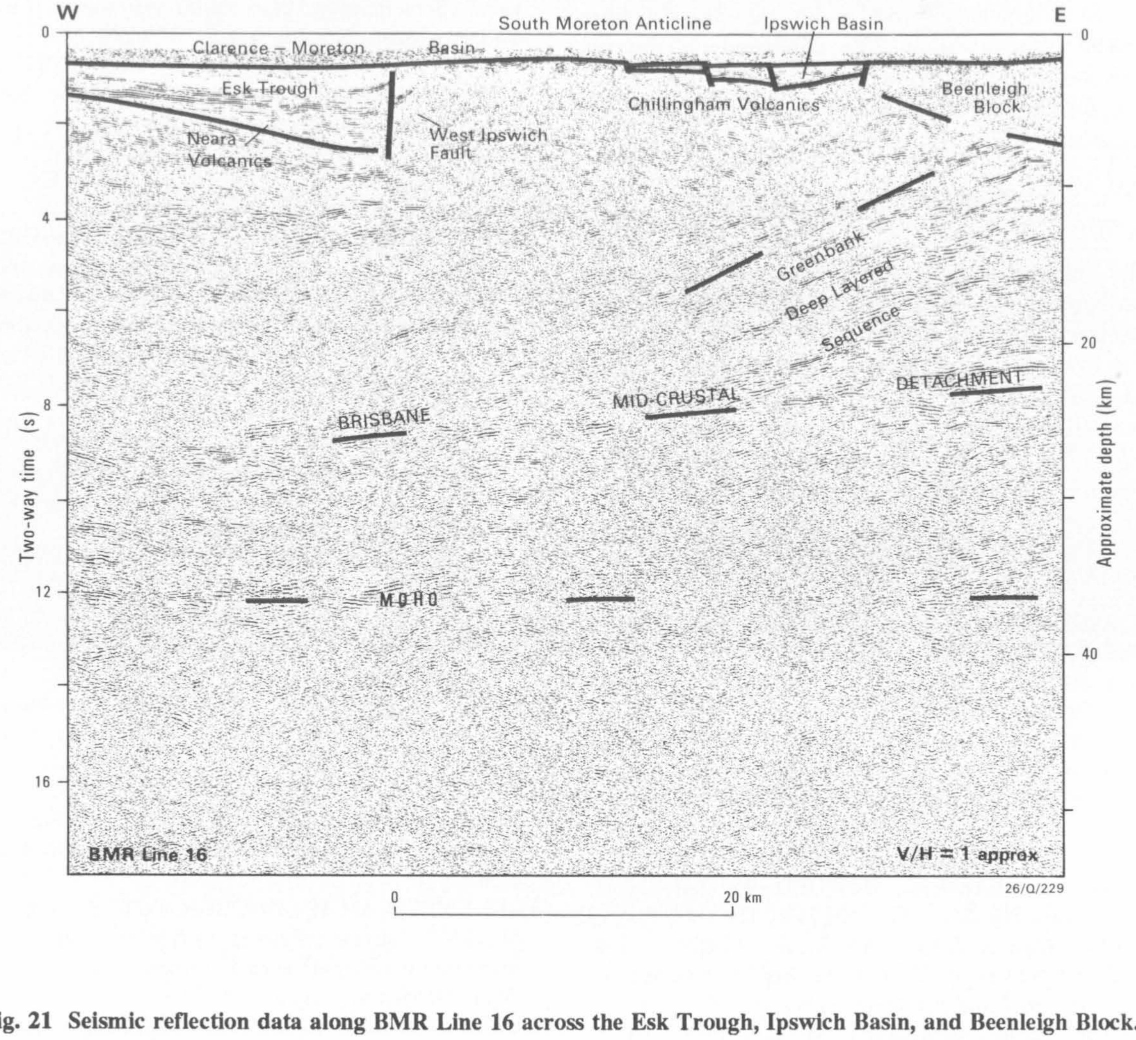


Fig. 21 Seismic reflection data along BMR Line 16 across the Esk Trough, Ipswich Basin, and Beenleigh Block.

(5) The crustal architecture under the Taroom Trough has little obvious structure above an outstanding Moho transition zone, contrasting it with early to middle Palaeozoic basins to the west. The prominent Meandarra Gravity Ridge striking north-south along the trough axis suggests the possibility of high-density mafic volcanics within the crust.

(6) Between the Taroom Trough and the New England Fold Belt, a high-angle geosuture is interpreted, probably extending deep within the crust (the Burunga-Mooki Geosuture). Reactivation during late Palaeozoic - early Mesozoic times produced a series of reverse faults (overthrusting? positive flower structures?) along the uplifted eastern margin of the trough. The geosuture separates regions with distinctly different seismic fabrics.

(7) Within the New England Fold Belt, crustal domains can be recognized associated with late Palaeozoic oroclinal bending and early Mesozoic transtensional basins. Major oroclinal bending of the upper crust in the west of the fold belt is interpreted above a mid-crustal detachment (the Texas Detachment). Within the upper crust there are reflections which may indicate another detachment associated with thrust faulting. In the central part of the fold belt, the seismic data support a transtensional mechanism for the formation of Triassic basins with steeply dipping bounding faults. Under the east of the fold belt, an imbricate thrust stack/accretionary wedge (the Greenbank Deep Layered Sequence) is clearly imaged above a mid-crustal horizon (the Brisbane Mid-crustal Detachment). Upper crustal deformation appears to have occurred above this detachment.

(8) From the Nebine Ridge to the coast, the Moho is clearly defined below a 3 km thick Moho transition zone. East of the Thomson Fold Belt (the Foyleview Geosuture), the Moho level is identified as a gently undulating feature at 36-38 km depth, probably established at this level after the major late Palaeozoic tectonic events which formed the crust under the Taroom Trough and New England Fold Belt. This contrasts with the middle Palaeozoic, more rugged, lower crustal/Moho morphology at 36-42 km depth preserved farther west under the Thomson Fold Belt.

(9) The thickest crust is under the Nebine Ridge (about 44 km) and seems to be associated with lower crustal wedging from the east. There is an apparent thinning of the non-sedimentary crust under the deepest basin (Taroom Trough) of about 30%, compared with the crust to the west and east. No crustal thickening is evident in the region of oroclinal bending within the New England Fold Belt.

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VELOCITY VARIATIONS WITHIN THE UPPER CRUSTAL BASEMENT OF THE CENTRAL EROMANGA BASIN

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ABSTRACT

Seismic refraction studies along a 262.5 km traverse in the Central Eromanga Basin have determined the detailed velocity structure of the basin, its infra-basins, and the upper crustal basement. Data were coincident with six-fold CMP reflection profiling and regional deep refraction profiling across the eastern Cooper Basin, Warrabin Trough, Canaway Ridge, Quilpie Trough and Cheepie Shelf. Iterative forward modelling was used to interpret the data, using ray-tracing and synthetic seismograms to match the observed refracted and reflected arrivals, including multiples.

The Eromanga Basin forms a continuous cover across the entire profile and varies in thickness from about 2.4 km in the west to about 1.1 km in the east. The near-surface velocity varies between 2.0 and 2.2 km/s and, at the base of the sequence, the velocity varies from about 3.8 km/s to 4.2 km/s. Generally, the velocities decrease from west to east. Below the Eromanga Basin, the Permo-Triassic Cooper Basin rocks are 549 m thick at Mt Howitt No. 1 well, and gradually wedge out at the western margin of the Warrabin Trough. The Devonian Warrabin and Quilpie Troughs are separated by a basement ridge, the Canaway Ridge. The Warrabin Trough sequence, bounded by high-angle faults, varies in thickness from 1.1 to 3.0 km; the Quilpie Trough, synclinal in form, varies in thickness from 0 to 3.8 km. The velocities in the troughs vary from about 4.0 km/s near the unconformity with the overlying Eromanga Basin, to between 5.0 and 5.5 km/s at the base of the Devonian sequence. The bottom of the troughs is not clearly defined, but reflections interpreted as coming from volcanics in the Gumbardo Formation are assumed to be from near basement.

Velocities within the top of basement Thomson Fold Belt rocks vary from 4.8 km/s under the Canaway Ridge and Cheepie Shelf, to about 5.7 km/s beneath the troughs. The velocity increases with depth to 6.0 km/s at a depth of 4.0 km under the Cheepie Shelf, and at a depth of over 8 km elsewhere. Steep velocity gradients near the top of the basement are probably caused by weathering, when the basement surface was exposed before subsequent deposition. The velocity models were tested against observed gravity data and it was found that variations in observed gravity can be satisfactorily attributed to structures within the top 10 km of the crust, with a thinner crust in the west accounting for a small regional gravity gradient.

INTRODUCTION

In 1980 and 1981, detailed refraction studies were undertaken by the Bureau of Mineral Resources (BMR) in the Jurassic-Cretaceous central Eromanga Basin to obtain the velocity structure of the basin, the underlying Devonian basins/troughs, and in particular, the early Palaeozoic basement rocks of the Thomson Fold Belt. These studies were made along an east-west traverse, coincident with six-fold CMP profiling along BMR Lines 1, 1X, and 9 (Wake-Dyster & Pinchin, 1981; Sexton & Taylor, 1983) and deep refraction profiling (Finlayson & others, 1984).

Near-vertical incidence reflection profiles contain almost no reflections from within the basement at two-way times (TWT) less than 8 s (Mathur, 1983), and rarely can faults be followed from the Eromanga and Adavale Basins into basement. However, this basement was deformed during middle to late Ordovician orogeny and Silurian plutonism (Murray & Kirkegaard, 1977) and the detailed refraction method was expected to show some of these features within basement from velocity changes. The velocities derived from the detailed refraction data presented here were also used to improve the velocity models of the deeper crust and upper mantle from regional refraction data (Finlayson & others, this Bulletin).

The total length of the traverse was 262.5 km and extended from Mount Howitt No.1 exploration well in the west to the Geological Survey of Queensland (GSQ) stratigraphic well Quilpie No.1 in the east (Figure 1). The western part of the traverse from Mount Howitt No.1 well to Eromanga township was discussed by Lock & Collins (1983), and the eastern part from Eromanga No.1 to Quiplie No.1 wells was discussed by Collins & Lock (1983). The interpretation of the data across the Warrabin Trough between Eromanga town and Eromanga No.1 well is presented here and integrated with the previous interpretations.

DATA RECORDING AND PROCESSING

Details of the recording scheme used to collect the data are given in Lock (1983). The maximum distance of recorders from the shots was 75 km. Shots were fired in a pattern of drill-holes, 40-50 m deep with 100-150 kg of explosive in each hole. A total charge weight of 200 kg was used in shots recorded between 0 and 37.5 km offset from the shot, while shots recorded between 37.5 and 75 km were 400 kg in weight.

Twenty-one analogue recorders (Finlayson & Collins, 1980) were used to record these shots at a station spacing of 1.875 km. The analogue records were later digitized and compiled into record sections. The data recorded across the Warrabin Trough are shown in Figures 2, 3 and 4. The traces were digitally filtered with a bandpass of 3-12 Hz and plotted on a time scale reduced by a velocity of 6.0 km/s. The record sections have been trace-normalized, so that the first arrival phase has constant amplitude to enhance the first-arrival travel-time branch (shown as P in the figures). Travel-times of the first arrivals used in the interpretation were picked from the original unprocessed records.

INTERPRETATION TECHNIQUES

Approximate models of depths and velocities were derived initially assuming planar dipping layers between shot points. Detailed velocity structure within the Eromanga Basin sedimentary sequence could not be resolved because the station spacing was too large. However, two prominent reflectors identified on reflection profiling records from within the Eromanga sequence are the Toolebuc Formation and Cadna-Owie Formation (Table 1) (Moss and Wake-Dyster, 1983; Wake-Dyster & others, 1983; Finlayson & others, 1988). Depths and velocities to these reflectors were tied at the three wells along the line, Mount Howitt No.1, GSQ Eromanga No.1 and GSQ Quilpie No.1. Velocity information for the wells was available from well shoots (Velocity Data Pty. Ltd., 1982; Hegarty, 1983). Assuming the velocity above and below these reflectors does not vary significantly along the traverse, the reflectors (depositional time horizons) can be regarded as iso-velocity lines on the reflection profiling section. Agreement of the velocities at the wells from one end of the traverse to the other confirms the validity of this assumption.

Using the near-vertical reflection records in this way, a detailed velocity/depth profile can be obtained for the Eromanga Basin sediments. Velocity information from wells was not available below the Eromanga Basin

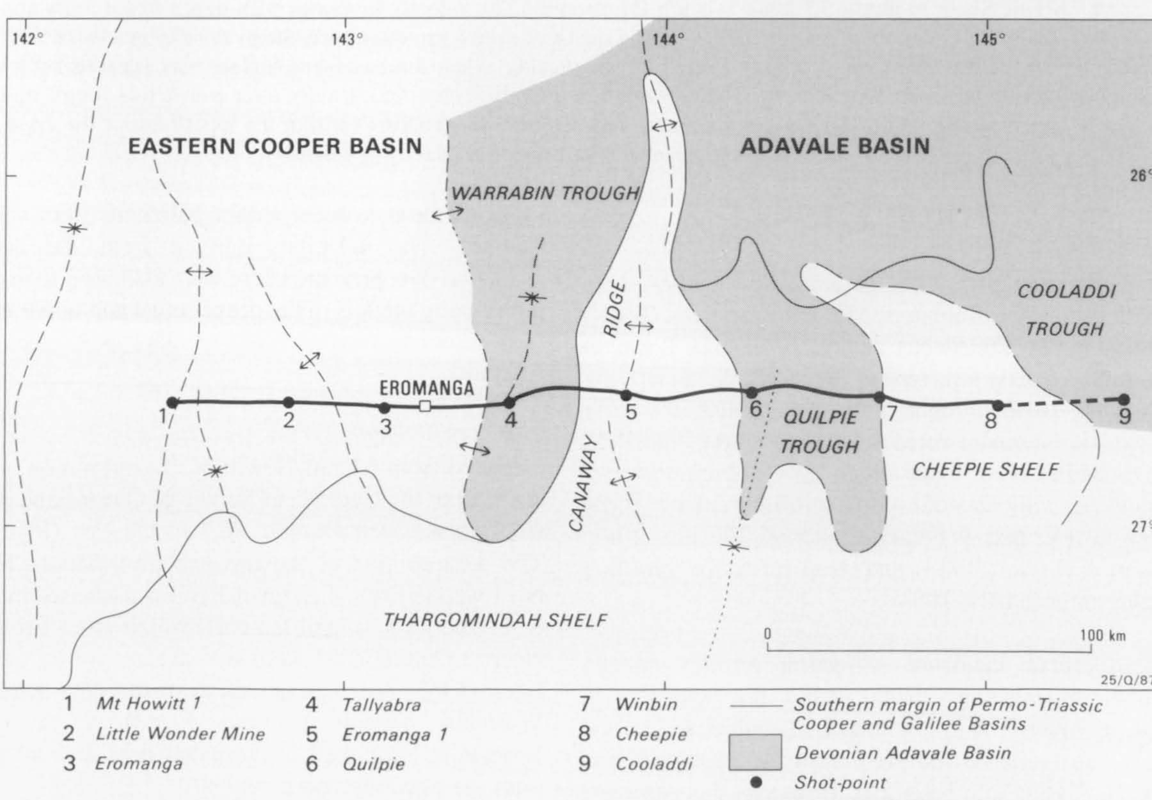

Fig. 1 Location of recording traverses and sub-basins of the Eromanga Basin.

TABLE 1

Simplified stratigraphy of the Central Eromanga Basin (based on Moss and Wake-Dyster, 1983) showing prominent reflectors used for control of detailed interpretation of structure.

Age	Basins, stratigraphic units	Seismic reflector
	Eromanga Basin	
Late		
Cretaceous	Winton Fm	
	Mackunda Fm	
Early	Allaru Mudstone	
Cretaceous	Toolebuc Fm	*
	Wallumbilla Fm	
	Cadna-Owie Fm	*
	Hooray Sst	
Late		
Jurassic	Westbourne Fm	
	Adori Sst	
Middle	Birkhead Fm	
Jurassic		
	Hutton Sst	
Early	Evergreen Fm	
Jurassic	Precipice Sst	
	Cooper Basin	
Late		
Triassic	Nappamerri Fm	
	Gidgealpa Group	
Late	Merrimelia Fm	
Carboniferous		
	Adavale Basin and Associated Troughs	
Late	Buckabie Fm	*
Devonian	Etonvale Fm	*
	Cooladdi Dolomite/Bury Limestone	*
	Lissoy Sst	
	Log Creek Fm	
Middle		
Devonian		
	Eastwood Beds	
Early		
Devonian	Gumbardo Fm	*
	Basement	
Silurian	Granitoids, metamporphics,	
Ordovician	volcanics	

sedimentary sequence. Prominent reflectors within the Devonian sedimentary rocks of the Warrabin Trough were used in the same way as in the Eromanga Basin sequences to control the detailed velocity/depth structure. The reflectors used were the Etonvale Formation, Cooladdie Dolomite or equivalent Bury Limestone, and the Gumbardo Volcanics at the base of the Devonian sequence (Table 1) (Finlayson & others, 1988).

An iterative forward modelling approach was used to interpret the data, using the simple layered models as the starting model. Travel-times through the model from each shot point were calculated by ray-tracing using the method of Collins (1980). The model was then modified and re-tested until close agreement with the observed arrivals was achieved. Ray-tracing was also used to calculate the two-way travel-times of near-vertical reflections from the prominent reflecting horizons in the model. These were then compared to the two-way times recorded on the coincident BMR reflection profiles. This provided an added constraint on the model, as the travel-paths are different to the travel-paths of the refracted arrivals. Again, the model was modified where necessary.

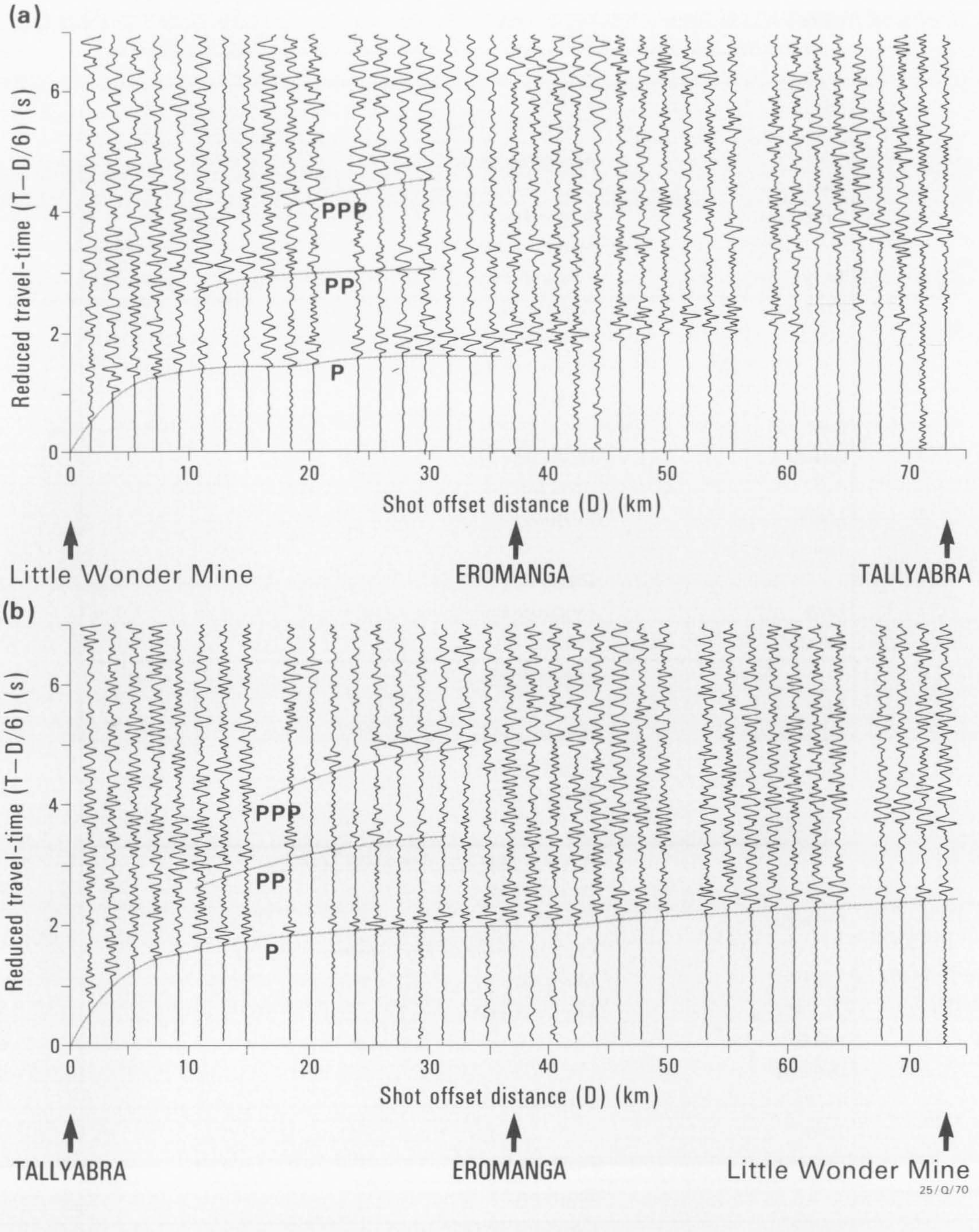

Fig. 2 Record section from (a) shots at Little Wonder Mine recorded towards Tallyabra, and (b) shots at Tallyabra recorded towards Little Wonder Mine.

Prominent multiple refracted arrivals can be seen in the record sections (shown as PP, PPP, etc. in Figures 2 and 3). These are refracted arrivals which, on returning to the surface, are reflected at the air/rock interface and "bounce" farther along the traverse (Lock and Collins, 1983). Their amplitudes vary with location, and up to five sets of multiples are observed on some records.

Multiples of this type have been discussed by Meissner (1965) and McMechan & Mooney (1980). Their generation and propagation depends on a high velocity gradient at the base of the sediments and/or top of the basement, and a low velocity at the surface. The first requirement is best fulfilled where the Eromanga Basin sediments are underlain by infra-basins, e.g. either the Cooper or Adavale Basins. The second requirement is fulfilled throughout the Eromanga Basin, where surface velocities are about 2.0 km/s. The geometry of the sediment/basement interface also affects the generation and propagation of the multiples.

Travel-times of multiples generated by ray-tracing through the interpreted model can be compared with the observed travel-times of the multiples. Since the travel-paths of the multiples through the model are different to both the primary refraction arrivals and the vertical reflections, they provide an additional tight constraint on the interpretation, any errors being "amplified" by the multiplicity of the travel-path.

Synthetic seismograms were computed for some of the shot points, so that amplitudes of the various seismic phases could be compared with the recorded data. Two types of synthetic modelling were used, one based on the reflectivity method, and the other on asymptotic ray theory. The reflectivity method computes the amplitudes of most phases more closely than asymptotic ray theory, but can only be applied to horizontal plane-layered models. Asymptotic ray theory was used where there was significant lateral variation in the model. The reflectivity program applied here was developed by Ha (1984) based on earlier developments by Fuchs (1968). Figure 4a shows the record section between Eromanga No. 1 well and Tallyabra, while Figure 4b illustrates the synthetic seismogram generated from a plane-layered model averaged from the velocity-depth structure. The prominent multiples which are observed have been closely reproduced in the synthetic section. To model

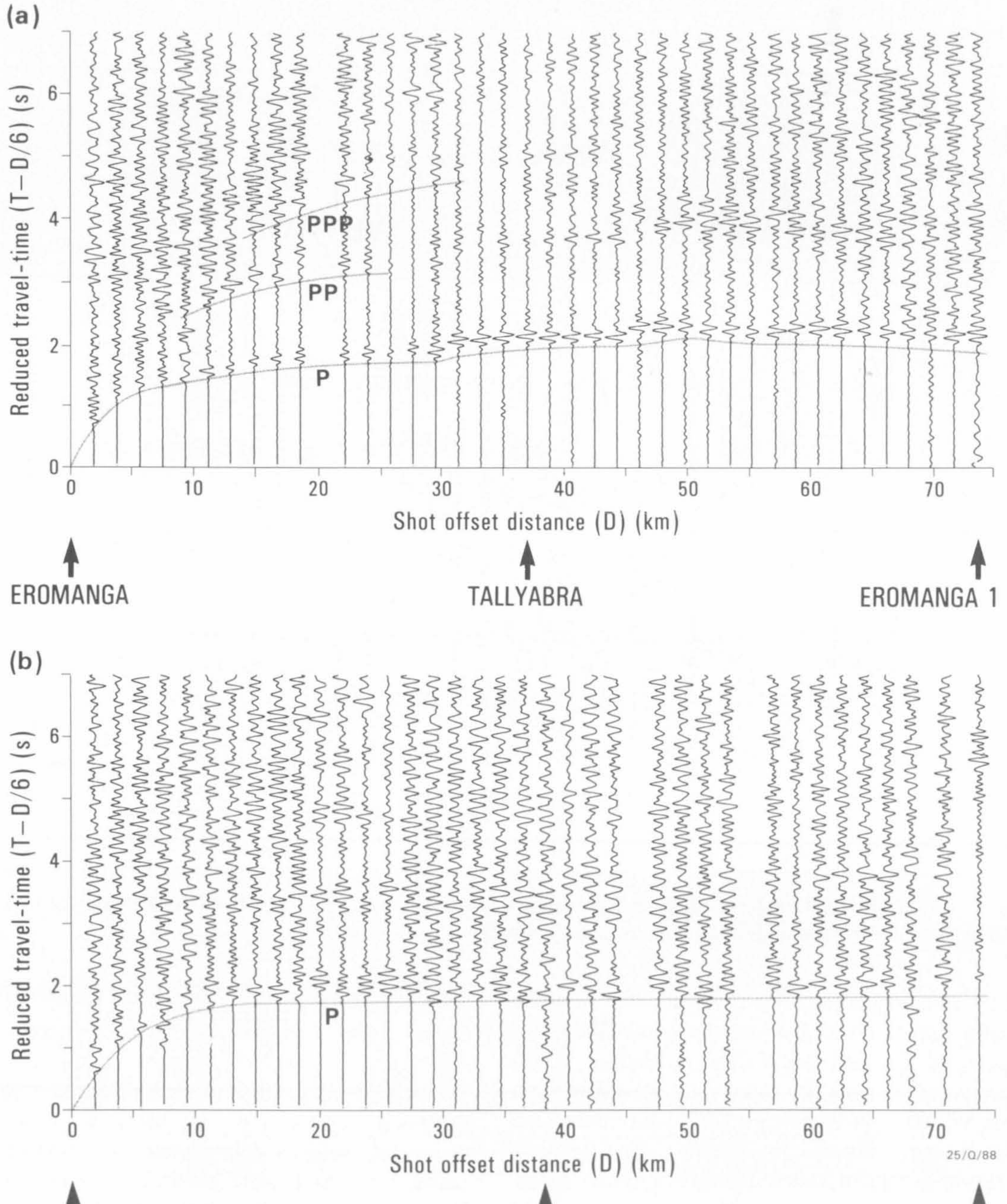

Fig. 3 Record section from (a) shots at Eromanga township recorded towards GSQ Eromanga No.1 well, and (b) shots at Tallyabra recorded towards Quilpie.

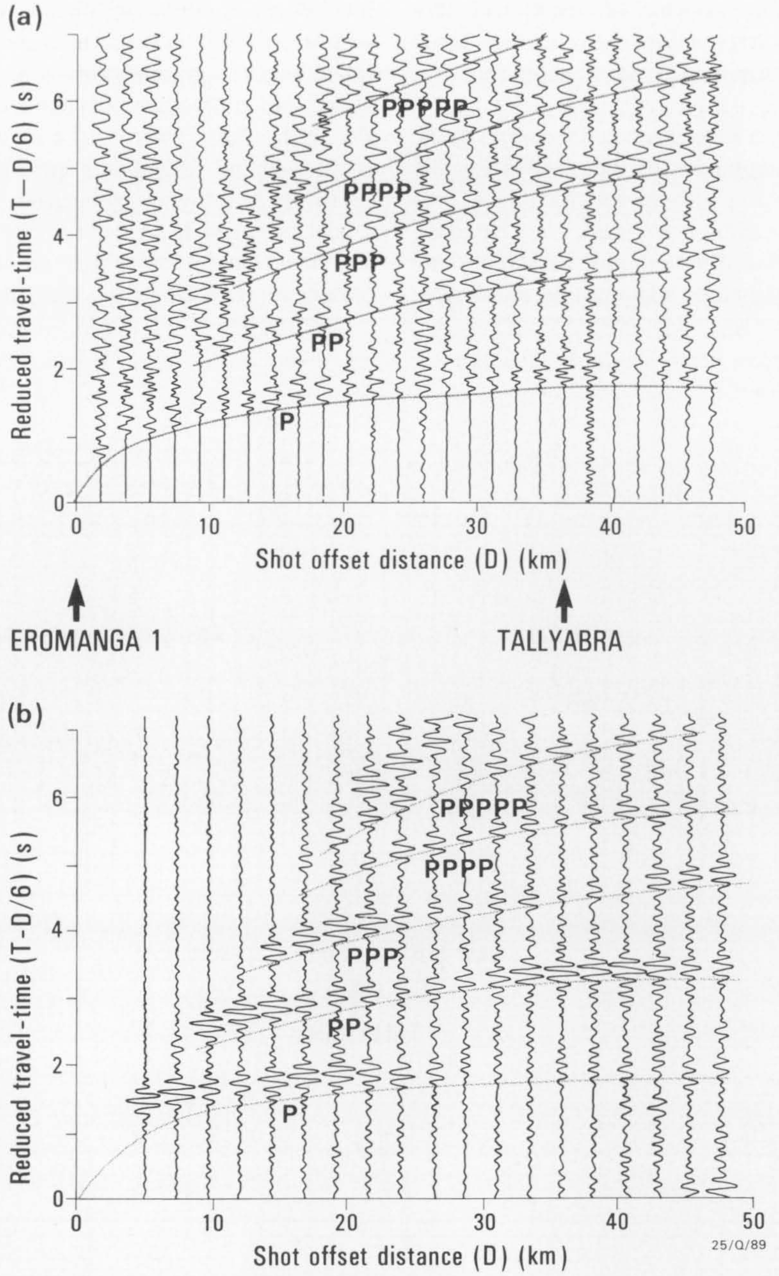

Fig. 4 (a) Record section from shots at Eromanga 1 recorded towards Tallyabra. (b) Reflectivity synthetic seismograms, GSQ Eromanga No.1 well towards Tallyabra.

laterally inhomogeneous sections of the traverse containing dipping layers and lateral velocity changes, a ray-theoretical program (McMechan & Mooney, 1980) was employed. Figure 5a shows the record section between Winbin and Eromanga No.1 well, across the Quilpie Trough. Figure 5b contains the synthetic seismograms generated from a slightly simplified version of the interpreted velocity-depth structure; the first multiple arrivals (PP) can be clearly seen in the synthetic section.

SUMMARY OF VELOCITY-DEPTH STRUCTURES

Velocity/depth profiles at each of the shot points are illustrated in Figure 6 and listed in Table 2. The interpreted velocity-depth structure between Mount Howitt No. 1 well and Quilpie No.1 well is shown in Figure 7. The structure between Mount Howitt and Eromanga township, and between Eromanga No.1 and Quilpie No.1 wells has been discussed by Lock & Collins (1983) and Collins & Lock (1983) respectively. These results are summarised below with a discussion of the Warrabin Trough and its adjacent basement structure.

The Eromanga Basin forms a continuous cover along the entire traverse, varying from about 2.4 km thick in the west to 1.1 km in the east. The velocity at the surface varies between 2.0 and 2.2 km/s and increases to between 2.25 km/s in the east and 3.2 km/s in the west at the top of the Toolebuc Formation (Fig. 6; Table 2). Below this, the velocity at the top of the Cadna-Owie Formation varies between 2.5 km/s in the east to 3.6 km/s in the west. The velocities at the base of the Eromanga basin sequence vary from 3.8 km/s in the east to about 4.3 km/s in the west. Higher velocities at the western end are associated with the the underlying Permo-Triassic Cooper Basin sediments, which cannot be resolved from the refraction data alone. They are 549 m thick in Mount Howitt No.1 well, and gradually wedge out at the western margin of the underlying Warrabin Trough.

The Canaway Ridge, a basement ridge, separates the Warrabin Trough from the Quilpie Trough (Pinchin & Anfiloff, 1986). The Devonian Warrabin Trough sequence varies in thickness from 1.1 km at the eastern boundary to about 3.0 km at its deepest part near the centre, and 1.5 km at the western boundary. Both the eastern and western boundaries are defined by steep faults. The velocity near the top of the sediments within the Trough is about 4.5 km/s, and increases to 4.8 km/s at the Buckabie Formation. Below the Buckabie Formation, the velocity increases to 5.0 km/s at the top of the Cooladi Dolomite, and 5.5-5.6 km/s at the base of the Devonian sediments. The bottom of the Trough is not clearly defined, but reflections interpreted as coming from the Gumbardo Volcanics are assumed to be from near basement (Wake-Dyster & others, 1983; Leven & others, 1990).

The Quilpie Trough is synclinal in form and varies in thickness from 0 to 3.8 km. The velocities in the Quilpie Trough increase from about 4.0 km/s near the unconformity with the overlying Eromanga Basin, to 4.3 km/s at the top of the Etonvale Formation. The velocity at the top of the Cooladdi Dolomite, or equivalent Bury Limestone, is 4.8 km/s, and is between 5.0 and 5.5 km/s at the base of the Devonian sediments. As with the Warrabin Trough, the bottom is not clearly defined, and the basement is assumed to be near the level from which reflections interpreted as Gumbardo volcanics are recorded.

Velocities within the top of the basement vary from 4.8 km/s in the Cheepie Shelf, to about 5.6-5.7 km/s beneath the troughs. The velocity increases with depth to 6.0 km/s at a depth of 4.0 km under the Cheepie Shelf, but not until over 8.0 km elsewhere. The higher velocities nearer the surface below the Cheepie Shelf coincide with a shallowing of mid-crustal reflections,

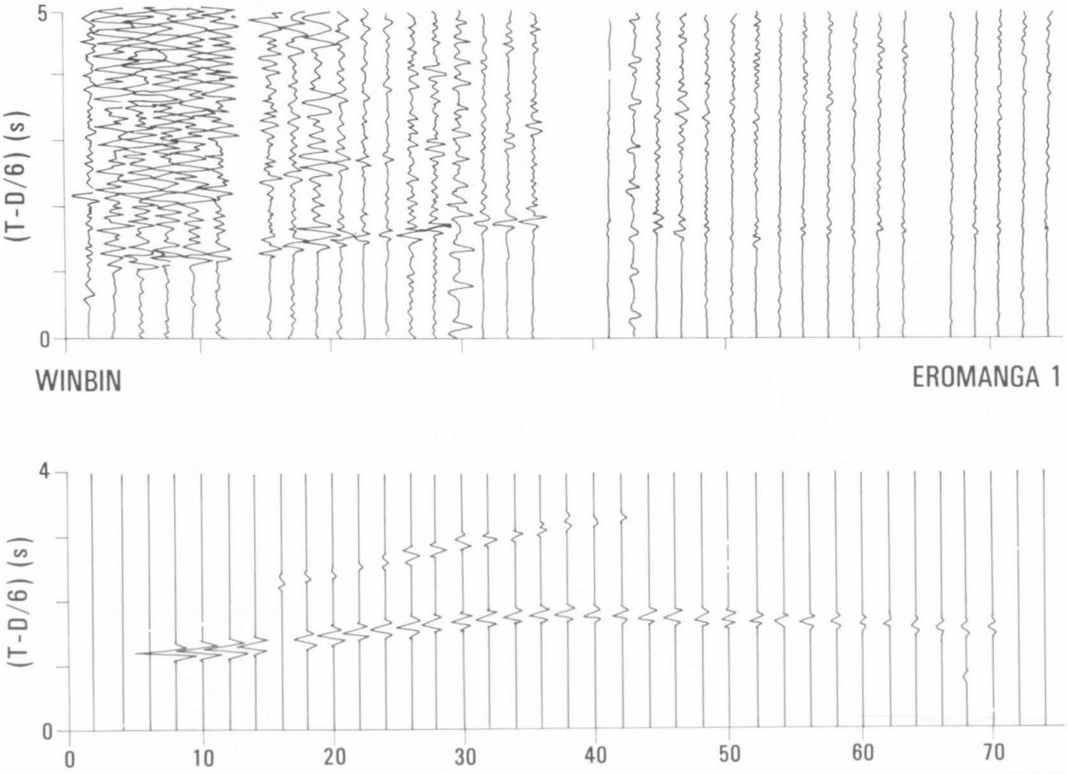

Fig. 5 (a) Record section from shots at Winbin towards GSQ Eromanga No.1 well, (b) Ray theoretical synthetic seismograms, Winbin towards GSQ Eromanga No.1 well.

TABLE 2

Velocity/depth data at the shot sites along the traverse between Mount Howitt No. 1 and Quilpie No. 1 wells.

Top of Formation	Mt. Howitt No. 1		Little Wonder Mine		Eromanga Town		Tallyabra	
	km	km/s	km	km/s	km	km/s	km	km/s
	0.00	2.30	0.00	2.20	0.00	2.20	0.00	2.10
Toolebuc	0.80	3.20	1.20	3.00	1.10	3.00	0.90	2.60
Cadna-Owie	1.20	3.60	1.50	3.60	1.30	3.40	1.10	3.10
	--	--	--	--	--	--	2.00	4.30
Top Dev. Unc.	--	--	--	--	--	--	2.10	4.60
Buckabie	--	--	--	--	--	--	2.70	4.80
Cooladdi	--	--	--	--	--	--	3.20	5.00
Gumbardo	--	--	--	--	--	--	4.60	5.50
Basement	2.40	5.00	2.50	5.00	2.40	5.00	4.61	5.50
	3.50	5.60	3.50	5.60	3.50	5.50	6.10	5.70
	5.00	5.90	4.50	5.70	5.00	5.80	7.80	5.90
	8.00	6.00	8.00	5.80	8.00	5.90	10.0	6.00
Top of Formation	**Eromanga No. 1**		**Quilpie Town**		**Winbin**		**Quilpie No. 1**	
	km	**km/s**	**km**	**km/s**	**km**	**km/s**	**km**	**km/s**
	0.00	2.00	0.00	2.00	0.00	2.00	0.00	2.00
Toolebuc	0.45	2.25	0.55	2.25	0.42	2.25	0.32	2.25
Cadna-Owie	0.70	2.50	0.75	2.50	0.65	2.50	0.50	2.50
	1.25	3.80	1.30	3.80	1.20	3.80	1.10	3.80
Top Dev. Unc.	--	--	1.31	4.05	--	--	--	--
Cooladdi	--	--	2.45	4.55	--	--	--	--
Gumbardo	--	--	3.60	4.90	--	--	--	--
Basement	1.26	4.50	4.40	5.50	1.21	5.00	1.11	4.80
	1.50	5.40	--	--	--	--	--	--
	2.25	5.50	--	--	1.95	5.50	2.30	5.00
	7.20	6.00	7.20	6.00	4.00	6.00	4.00	6.00
	9.00	6.10	9.00	6.10	9.00	6.10	9.00	6.10

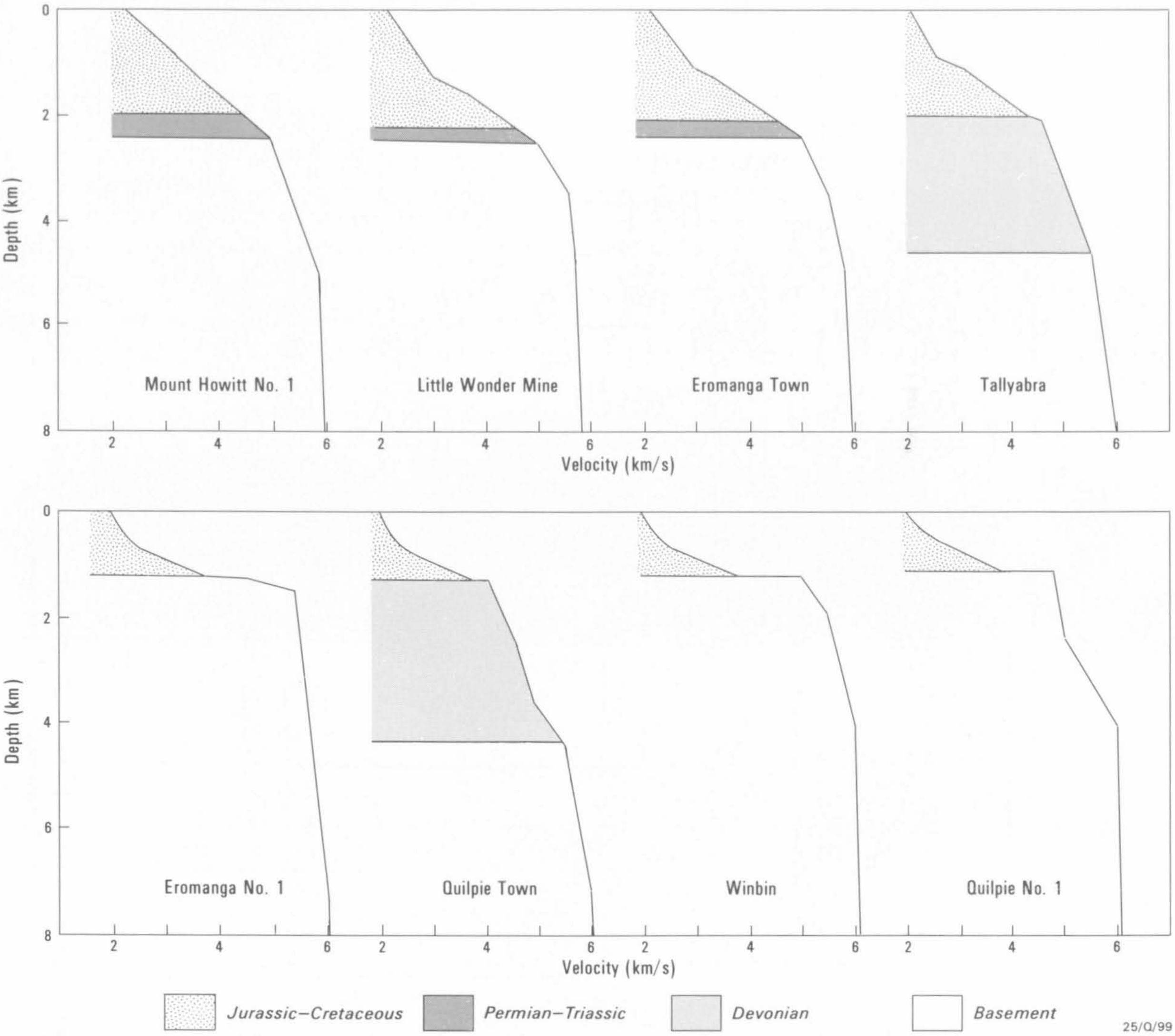

Fig. 6 Velocity/depth profiles along the traverse between Mount Howitt No.1 and GSQ Quilpie No.1 wells.

from about 8 s TWT to about 6 s TWT. Strong velocity gradients near the top of the basement may be due to weathering, when the basement surface was exposed prior to deposition of the Eromanga Basin. This gradient, along with the low surface velocity, is the likely cause of the very prominent multiple refractions between the surface and the basement.

The velocities in the Eromanga Basin sediments generally increase towards the west, as do the velocities in the Devonian troughs. The interpretation of the detailed velocity structure in the Eromanga Basin and the Devonian trough sequences is constrained by the geometry of the prominent reflectors, whilst basement structure cannot be resolved with the same detail because of the lack of reflectors. The velocity gradient zones defined by the broken lines in Figure 7 give the best fit to the observed travel times of both primary and multiple refracted arrivals through the basement. Lateral variations of velocity within basement appear to be related to the faulting within the overlying sediments (Fig. 7) and, as noted earlier, the depth at which the velocity reaches 6.0 km/s and over appears related to the depth of the mid-crustal reflective layer. Local areas of lower velocity (4.0-4.5 km/s and 4.5-5.4 km/s) occur at the top of the basement in the Cheepie Shelf and Canaway Ridge (Fig. 7); these coincide with weak reflections and may be remnants of Devonian sediments.

GRAVITY MODELLING

The velocity model was tested by computing its gravitational effect and comparing this with the observed gravity. To model the gravitational effects, the P-wave velocities were converted to densities using the empirical relation given by Dooley (1976) for continental sedimentary rocks. As the velocity within each layer is gradational, the average velocity within the layer was used to derive its mean density. The modelling was done using a computer program adapted from Milsom &

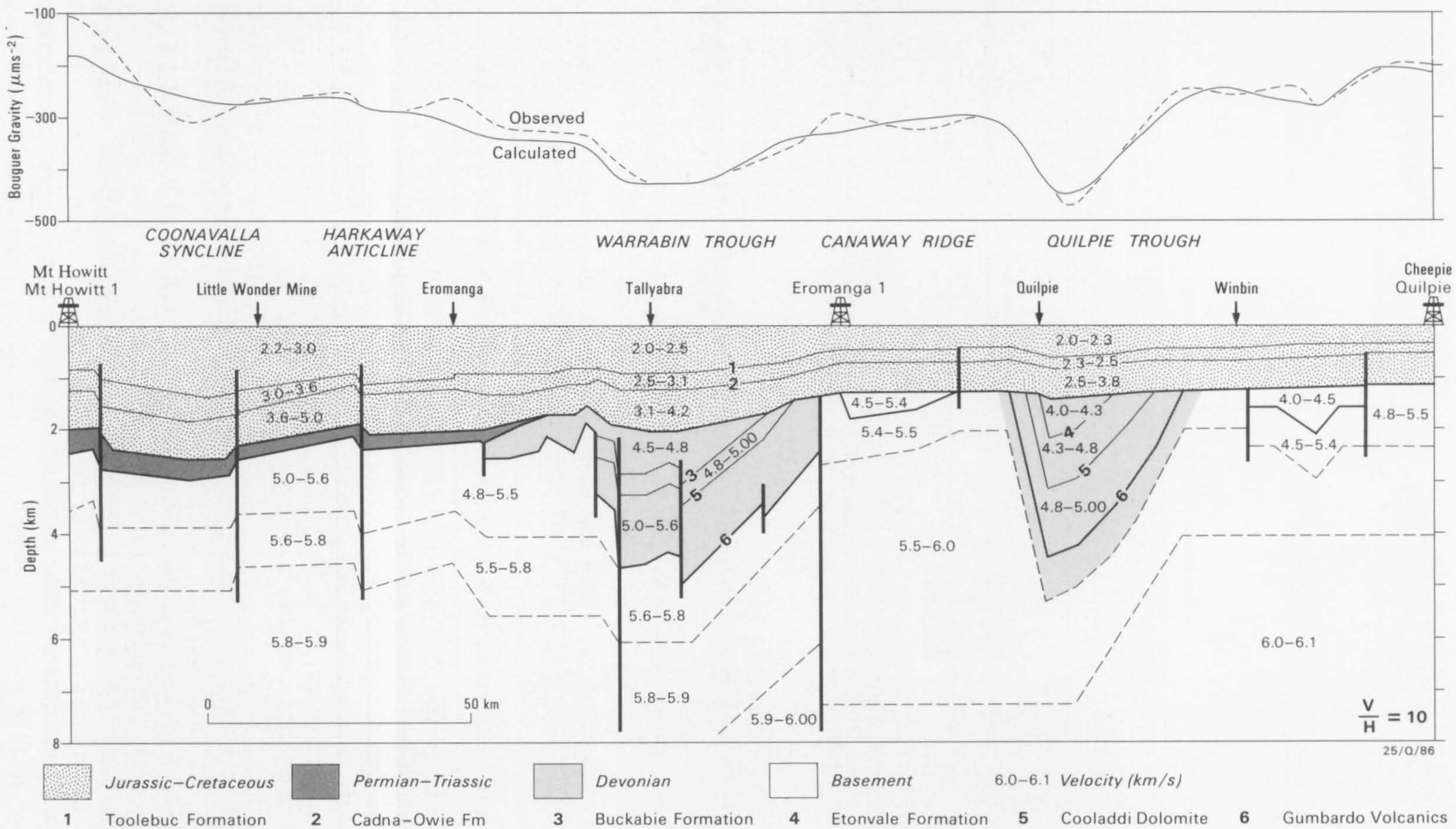

Fig. 7 Velocity structure between Mount Howitt No.1 and GSQ Quilpie No.1 wells. Observed and calculated Bouguer gravity shown above the model. Numbered horizons indicate the tops of the formations listed in the legend.

Worthington (1977) (B.J. Drummond, personal communication, 1982). This computes the gravitational attraction of two-dimensional bodies of limited strike length. The agreement with the detailed Bouguer gravity recorded along the traverse (Anfiloff, 1984) is considered good, within 50 $\mu m.s^{-2}$ of the observed gravity data (Fig. 7).

The average crustal velocity model between Mount Howitt No.1 and Cheepie No.1 wells for depths below 10 km (Finlayson & others, 1984) was included in the gravity model to acount for a regional effect observed after modelling the top 10 km. The regional gradient changes slope about half way along the traverse, and it was found that this could be satisfactorily modelled by shallowing the crust-mantle boundary from 41 km in the east to 39 km in the west, over a distance of 25 km near the western margin of the Warrabin Trough. The refraction interpretation shows the crust-mantle boundary at about 36.4 km in the west, which is shallower than inferred from the regional gravity data. However, the agreement is considered good, bearing in mind the assumption of two-dimensionality for the model and the uncertainties in the velocity/density conversion relationships at depth.

ACKNOWLEDGEMENTS

We wish to acknowledge our many co-workers in the Eromanga Basin project whose contributions have had a bearing on the results presented here. In particular we thank Doug Finlayson, who critically read a draft of this paper and made many useful comments, and Chris Rochford, Jim Whatman, John Williams and the BMR seismic crew. Barry Drummond wrote the gravity modelling program.

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SEISMIC VELOCITY MODELS OF THE CRUST AND UPPER MANTLE UNDER THE BASINS OF SOUTHERN QUEENSLAND

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ABSTRACT

P-wave velocity models of the crust and upper mantle under the basins of southern Queensland, interpreted from seismic refraction/wide angle reflection profiles, indicate lateral inhomogeneities at many levels within the lithosphere. Velocity/depth models for basement below the central Eromanga Basin, the Nebine Ridge, the Taroom Tough of the Bowen Basin, and the New England Fold Belt have significantly different velocity characteristics which must be taken into account in the construction of any tectonic models for the region. For instance, the depth at which velocities over 6.3 km/s occur is quite variable and suggests that processes such as ramping/thrusting of lower crustal rocks to higher levels and mafic volcanism must have been significant throughout the region.

The interpretation of the seismic data is enhanced when velocity information is combined with reflection profiling. In many places high velocity gradients coincide with prominent reflection features, strongly suggesting that they represent "seismic terrane" boundaries which abut rocks of different compositions/metamorphic grades and may form sites for decoupling of crustal units during episodes of crustal deformation. P-wave velocities for the shallow sedimentary rocks have been derived for some localities from expanding reflection spreads and "piggy back" refraction experiments that use routine reflection profiles shots as energy sources. Expanding spreads have also provided localized velocity estimates deep within basement.

In the central Eromanga Basin the major crustal velocity changes occur at mid-crustal depths (21-25 km) where P-wave velocities increase from 5.6-6.3 km/s in the upper crust to 6.4-6.9 km/s in the lower crust. The zone of increased velocity corresponds to the zone of increased reflectivity evident on near-vertical reflection profiling records. The Moho depth increases from 36-37 km in the west to 41 km under the Cheepie Shelf. The velocity in the uppermost mantle is 8.15 km/s, but increases to 8.35 km/s at 56-57 km depth.

Under the Nebine Ridge and Mitchell Shelf, a velocity gradient occurs at shallower depth (14-15 km) than farther west, with P-wave velocities changing from 6.02 km/s in the uppermost crustal basement to 6.2-6.3 km/s; there is a velocity gradient to 6.7 km/s in the lower crust above the Moho at 44 km depth. The velocity increase in the upper crust correlates with strong reflections on profiling records. Under the Taroom Trough, crustal velocities are less well determined. North of the transect area, a velocity of 6.39-6.40 km/s has been interpreted within basement at 7-30 km depth and the Moho at greater than 36 km depth, with an upper mantle velocity of 8.10 km/s. Expanding spread data from the transect indicate an increase in reflectivity at 17 km depth with a velocity of 6.7 km/s in the lower crust. Within the New England Fold Belt, there are wide-angle reflections interpreted to indicate a velocity increase at mid-crustal depths (19 km); the Moho is interpreted at 36 km depth and the upper mantle velocity is a low 7.7 km/s.

INTRODUCTION

Along the Eromanga-Brisbane Geoscience Transect, seismic wide-angle reflection and refraction data from BMR surveys provide velocity information throughout the crust and upper mantle. From such data, where there are long transmission paths for seismic "phases", the velocity structure of the crust can be determined, and, from it, the significant vertical and horizontal compositional/metamorphic changes can be judged. Generally, near-vertical seismic reflection profiling data only provide useful velocity information in the sub-horizontal sedimentary rocks within the top 5 km of the crust. Deeper velocities within basement are generally best determined from the wide-angle reflection and refraction data. However, expanding reflection spreads, which provide reflection data from the deep crust at moderate angles of incidence, can also provide velocity information through the entire crust, provided there are persistent strong reflections at depth. In some cases, reflection horizons coincide with significant velocity changes, thus providing important constraints on the interpretation of both datasets. Often only the P-wave velocities are determined from wide-angle reflection and refraction data. But from such data, the nature of velocity gradients within the crust can characterize the crust in a particular tectonic setting. This can be used to discriminate between areas which otherwise might be regarded as having the same crustal signature. If both P-wave and S-wave velocity can be measured, additional information can be determined on the physical properties of the deep basement rocks, further constraining the geological interpretation.

Finlayson (1983), Finlayson & others (1984), Finlayson & Collins (1986) have described the crustal and upper mantle velocity structure under the central Eromanga Basin at the western end of the transect. Finlayson & Collins (1987), Wright & Finlayson (in press) describe the velocity structure under the Nebine Ridge; and Finlayson & others (1988) describe crustal and upper mantle velocities under the New England Fold Belt. In addition there are BMR refraction data from the region near BMR reflection line 18 (Finlayson & others, this Bulletin), and also refraction data from the Taroom Trough about 250 km north of the transect discussed here (Collins, 1978, 1980; Leven, 1980). Wright & Barker (1989) and Wright & others (1990a,b) have used expanding reflection spreads to measure seismic velocities across the Nebine Ridge and Taroom Trough.

This paper reviews the velocity information for the crust and upper mantle along the geoscience transect and puts it in the context of possible geological interpretations. Data are discussed from four regions; the central Eromanga Basin, the Nebine Ridge and Mitchell Shelf, the Taroom Trough of the Bowen Basin, and the New England Fold Belt.

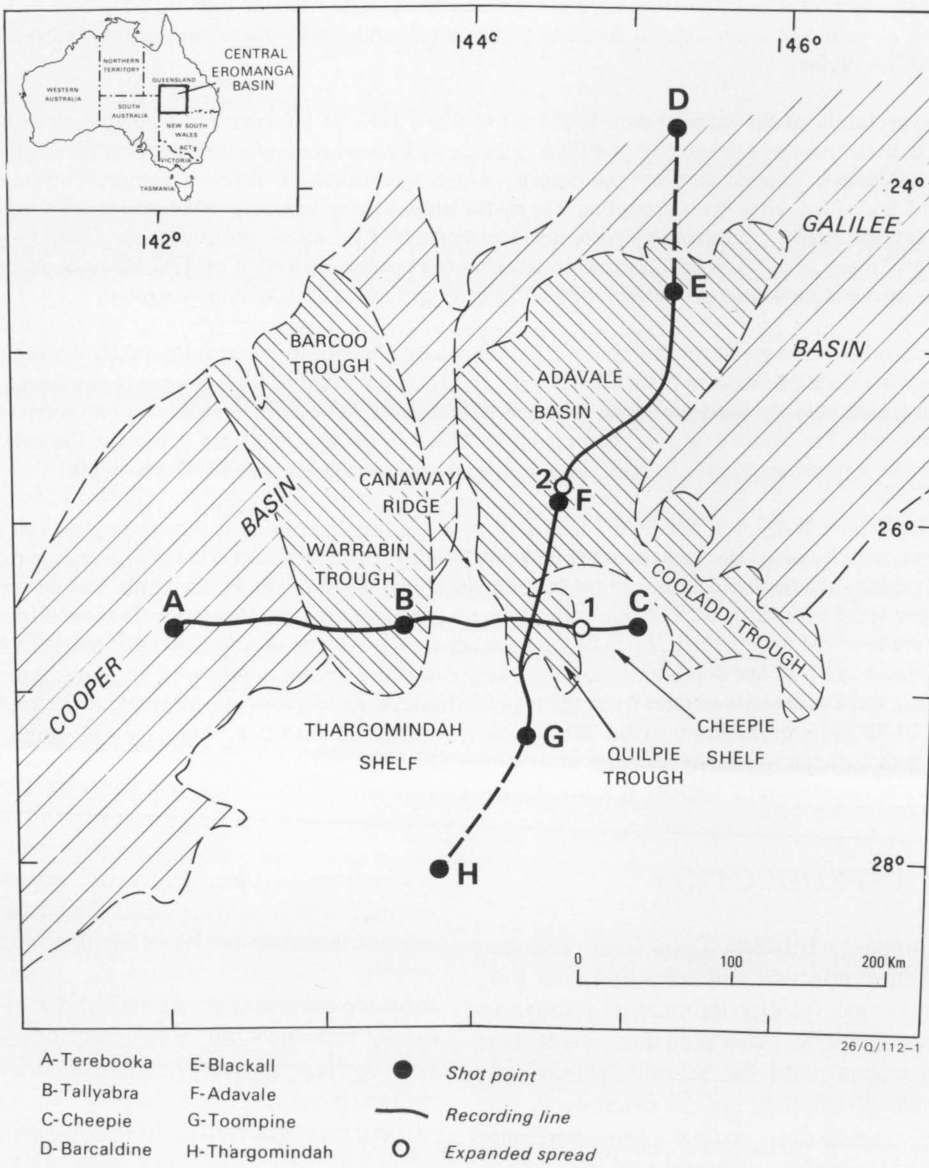

Fig. 1 Location of wide-angle reflection and transmission seismic profiles and expanding spread profiles in the central Eromanga Basin region.

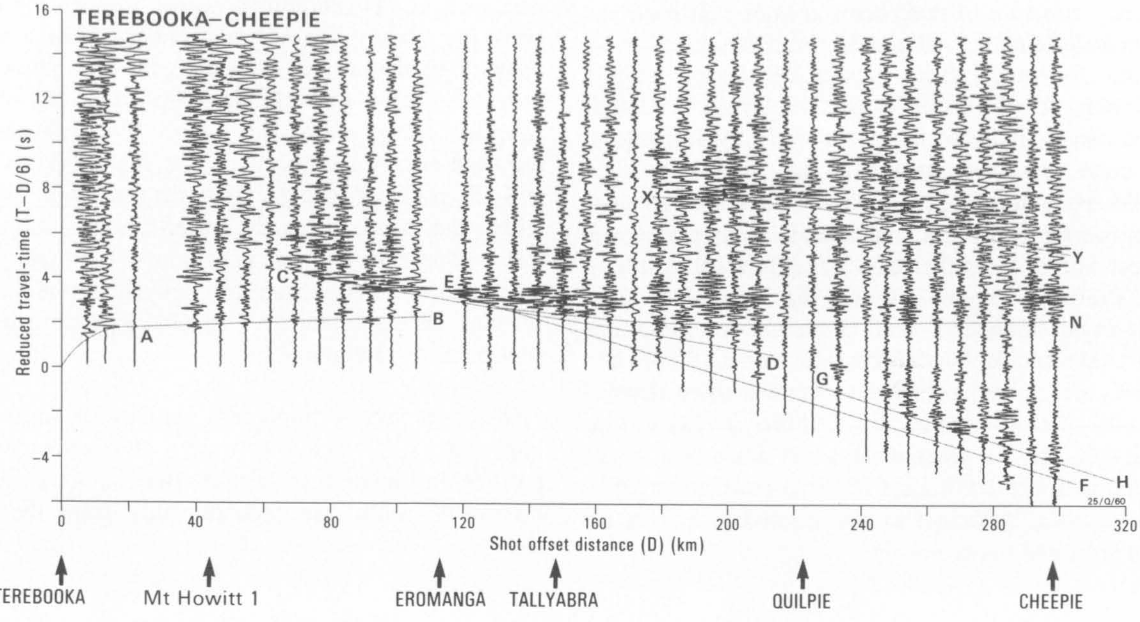

Fig. 2 Example of wide-angle reflection/refraction seismic data from the central Eromanga Basin region illustrating the seismic phases used to derive velocity/depth models. Both the travel-times and amplitudes of the various phases were used to derive preferred velocity/depth models. A-B = Pg phases; N-C-D = PcP and Pc phases; D-E-F = PmP and Pn phases; G-H = sub-Moho reflected phases; X-Y = upper/middle crustal reflected refractions.

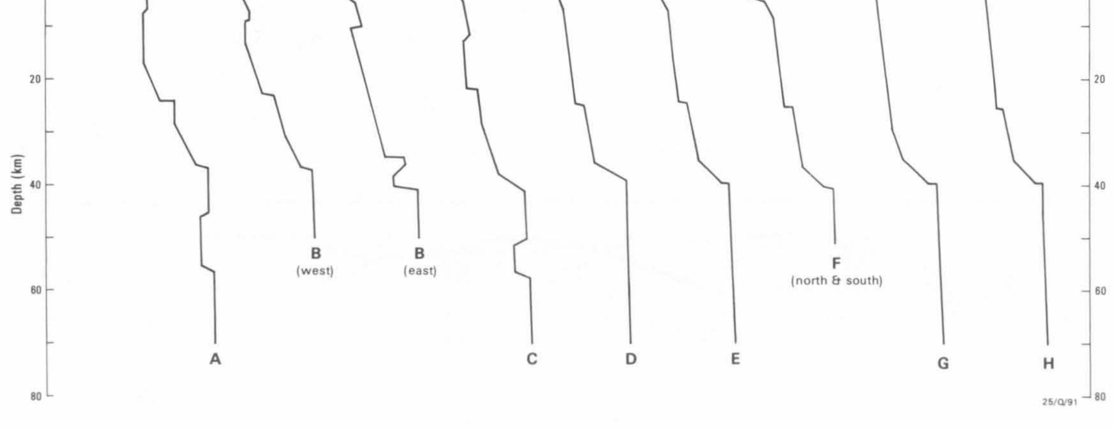

Fig. 3 Velocity/depth models of the crust and upper mantle from the central Eromanga Basin region. The letters against each model refers to the shot point locations indicated in Figure 1.

CENTRAL EROMANGA BASIN

Collins & Lock (this Bulletin) have outlined the velocity structure of the Thomson Fold Belt basement rocks underlying the sedimentary basin sequences in the central Eromanga Basin region. Finlayson (1983), Finlayson & others (1984), and Finlayson & Collins (1986) have described the overall velocity structure of the crust and upper mantle in that region and their results are summarised here. These results were obtained from two major profiles coincident in many places with reflection profiles in the same area (Finlayson & others, this Bulletin). Figure 1 shows the location of these wide-angle reflection and refraction profiles: one east-west coincident with BMR reflection profiling Lines 1, 1X, and 9, and coincident with the more detailed velocity data described by Collins & Lock (this Bulletin), and the other north-south coincident in the central part with BMR Line 10. Both profiles were over 300 km long, sufficient to obtain good data from the deep crust and upper mantle.

Figure 2 shows data from the east-west profile, illustrating the seismic phases which can be identified and interpreted to obtain velocity information. These data include, (1) transmitted energy through the upper crust (Pg phase), (2) reflected energy from a strong velocity gradient at mid-crustal depths (PcP phase), (3) refracted energy through the lower crust (Pc phase), reflected energy from the Moho (PmP phase), (4) refracted energy through the upper mantle (Pn phase), and (5) reflected energy from deeper (over 50 km) within the mantle. Both the travel-time and amplitude information from all these phases have been interpreted to provide velocity-depth profiles along the various sections of the traverses (Fig. 3), depending on the shot-point distribution.

There are some important similarities and some important differences between the velocity-depth profiles. In Figure 3, profiles A, B(west), C, D, E, and F(north & south) are predominantly from the main

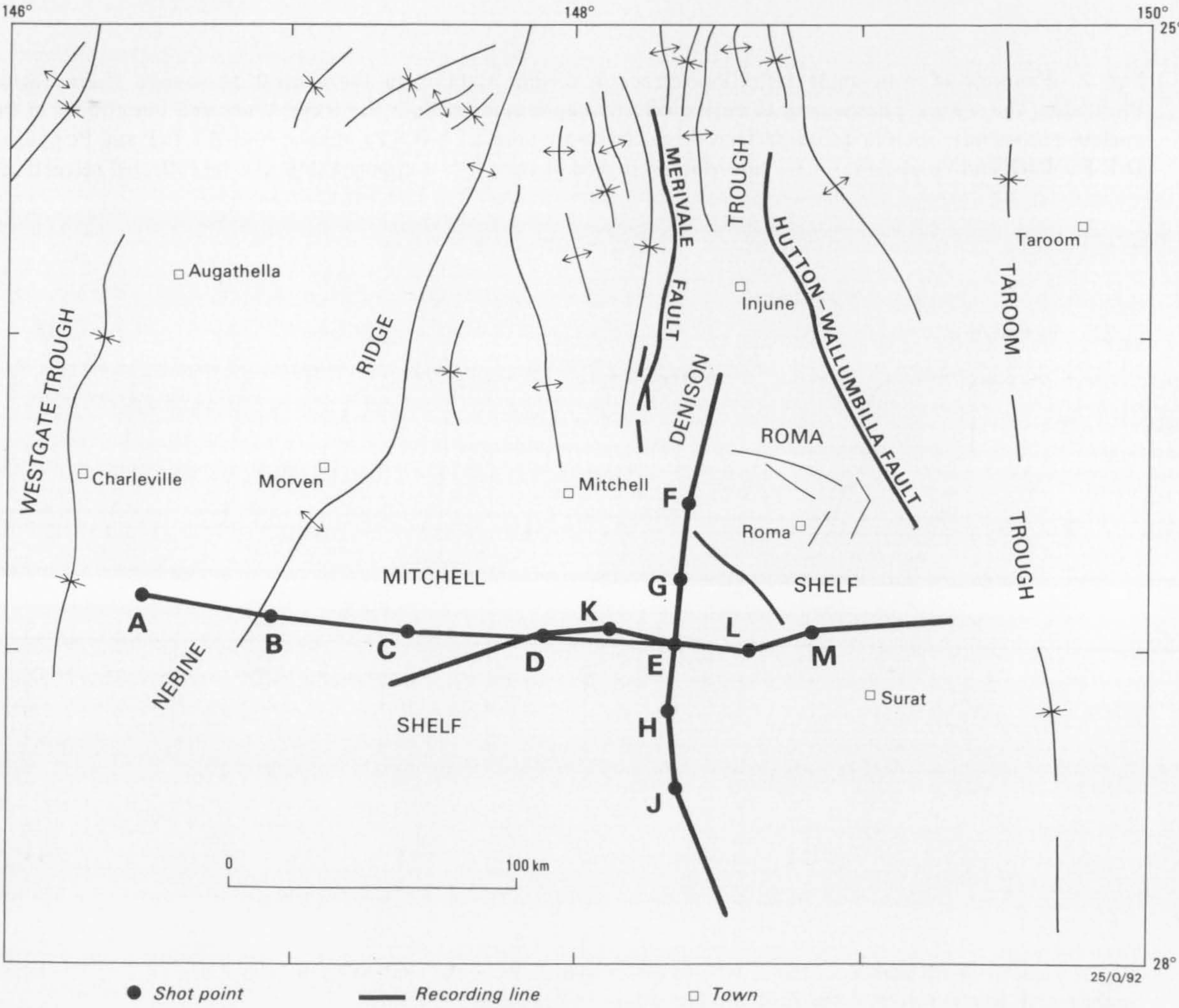

Fig. 4 Location of wide-angle reflection/refraction seismic profiles and expanding spread profiles across the Nebine Ridge, Mitchell Shelf and western Surat Basin.

depocentres of the Devonian Adavale Basin and its associated troughs (Evans & others, this Bulletin). These profiles indicate that the upper crust has a velocity in the range 5.6 to 6.3 km/s, sometimes with a velocity decrease at 7-10 km depth. These are, therefore, the velocities in the Thomson Fold Belt rocks. The important feature of these profiles is that, at mid-crustal depths (21-25 km) there is a prominent velocity increase (to 6.4-6.9 km/s) marking the top of the lower crust. This velocity increase coincides with the top of the lower crustal reflectivity zone discussed by Mathur (1983), Finlayson & others (this Bulletin). Velocity-depth profiles B(east) and G, (Fig. 3) apply mostly under the Canaway Ridge and do not have a velocity increase at mid-crustal depths; instead they have a steady velocity gradient throughout most of the crust with some strong velocity increases in the lower crust and at the Moho. The Canaway Ridge is, therefore, interpreted as having a significantly different basement velocity structure from that under the surrounding Devonian basins and troughs.

In the lower crust, the velocities are mostly in the range 6.4 to 7.1-7.5 km/s above a velocity gradient zone at the Moho. These velocities are interpreted to be indicative of more mafic material and/or higher metamorphic grades in the lower crust (Christensen & Fountain, 1975; Smithson & others, 1981; O'Reilly and Griffin, this Bulletin). Finlayson & others (this Bulletin) show the velocity-depth profiles superimposed on line diagrams of crustal reflections and there is a close correspondence between the higher velocities in the lower crust and the reflective zone at depths greater than 20 km. This correspondence has been found elsewhere (Mooney & Brocher, 1987) and associated with extensional tectonic environments where velocities over 7 km/s are determined for the lower crust. These high crustal velocities can be attributed to intruded, laminated mafic and ultramafic material in a lower crust in association with high-grade metamorphic rocks, the mafic/felsic layering being seen as the cause of the deep reflections. Fountain (1987) indicates that during extensional events, magmatic underplating and dike/sill emplacement are important processes in the lower crust resulting in a complex layered geology accentuated by ductile stretching. Smithson (1987) has emphasized the superposition of many different processes in the formation of the present-day seismic characteristics of the crust.

The P-wave velocity of the upper mantle in the central Eromanga Basin region is determined as 8.15 km/s on both the east-west and north-south traverses. This velocity is reached at the base of a Moho transition zone which is 2-4 km thick. The depth of the Moho is at 36-37 km under the western end of the east-west traverse, increasing to 41 km under the Cheepie Shelf. The Moho depth under most of the central Adavale Basin is 39-40 km, shallowing to 38-39 km depth at the northern end of the basin. Under the Canaway Ridge, the Moho is still interpreted at 39-40 km depth, but there is a high-velocity feature at about 34 km which indicates a structural complication not resolved by the available data. There is no evidence for major short-wavelength changes in Moho depth. The deep reflection profiles from the region (Finlayson & others, this Bulletin) do not show a clear Moho feature. Rather, there is a marked decrease in reflections at two-way travel-times (TWT) greater than 12 s, often with cross-cutting events at 12-14 s TWT. Despite this unclear picture, the Moho depth

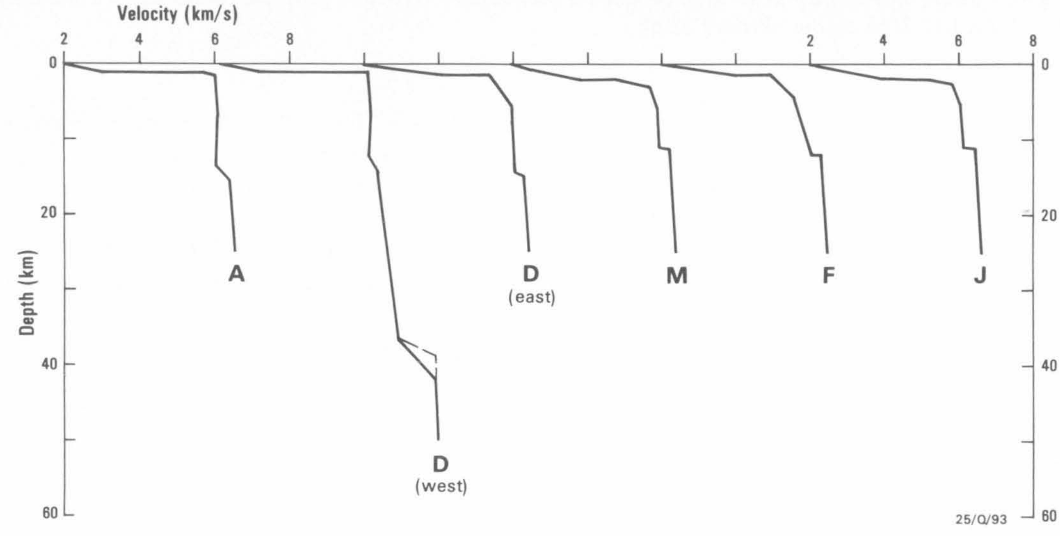

Fig. 5 Velocity/depth models for the crust from the Nebine Ridge, Mitchell Shelf, and western Surat Basin regions. The letters against each model refer to the shot point locations indicated in Figure 4.

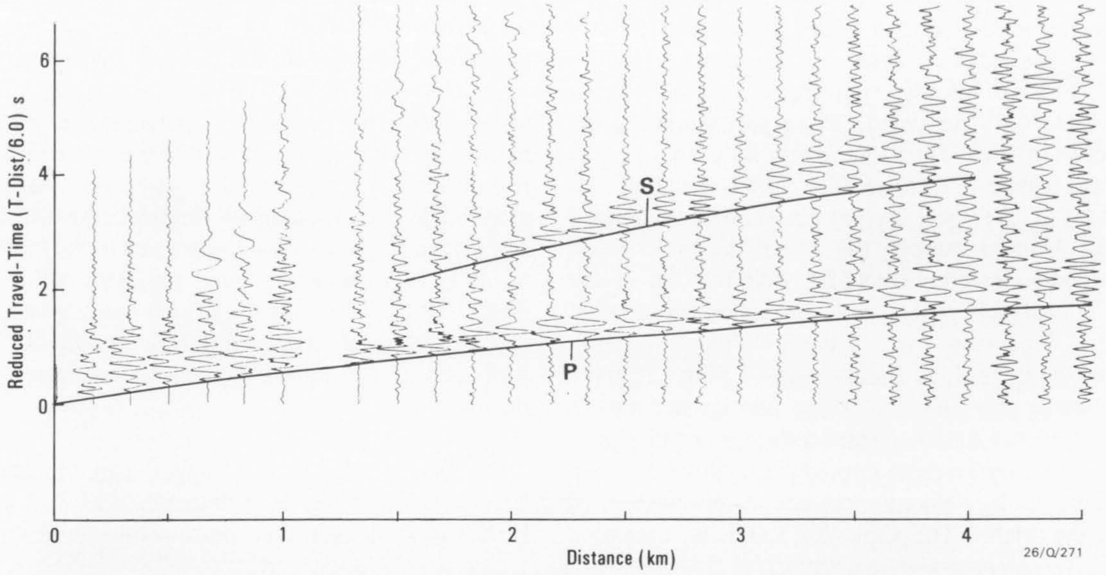

Fig. 6 Seismograms recorded by an array of geophones used in deep reflection profiling connected to a stand-alone recorder at location 3166 on the Mitchell Shelf.

determined from the wide-angle reflection and refraction data does correspond closely with that interpreted from near-vertical reflection data, bearing in mind that the wide-angle data give a depth value which must be averaged over distances of about 70-80 km (Finlayson & others, this Bulletin, Figures 10, 11 & 12).

Deeper in the upper mantle, seismic data indicate a further increase in P-wave velocity to 8.35 km/s at depths of 56-57 km. The data do not enable the detail of this velocity change to be determined. It is interpreted below a low velocity zone 7-9 km thick under the east-west traverse, but associated with a simple velocity gradient under the north-south traverse. Some velocity anisotropy is possible at these depths but is not conclusively demonstrated.

NEBINE RIDGE AND MITCHELL SHELF

Across the Nebine Ridge a number of datasets can be interpreted to give velocity information within basement rocks. During 1984-86, two wide-angle relection and refraction seismic surveys were undertaken and also an expanding spread profile was recorded (Finlayson &

TABLE 1

R.M.S. velocities for basement reflections from expanding spread recorded across the Taroom Trough.

Expanding spread location	Vertical two-way time, (s)	R.m.s. velocity (km/s)	Interval velocity (km/s)	Approximate depth range of interval velocity, (km)
Taroom Trough	4.22	4.41 ± 0.06		
(crustal thick-	7.10	4.86 ± 0.24+	5.44	9.3 - 17.2
ness ~ 39 km)	13.41	5.77 ± 0.16+	6.66	17.2 - 38.7*

* interpreted as base of crust. + estimates by linear regression.

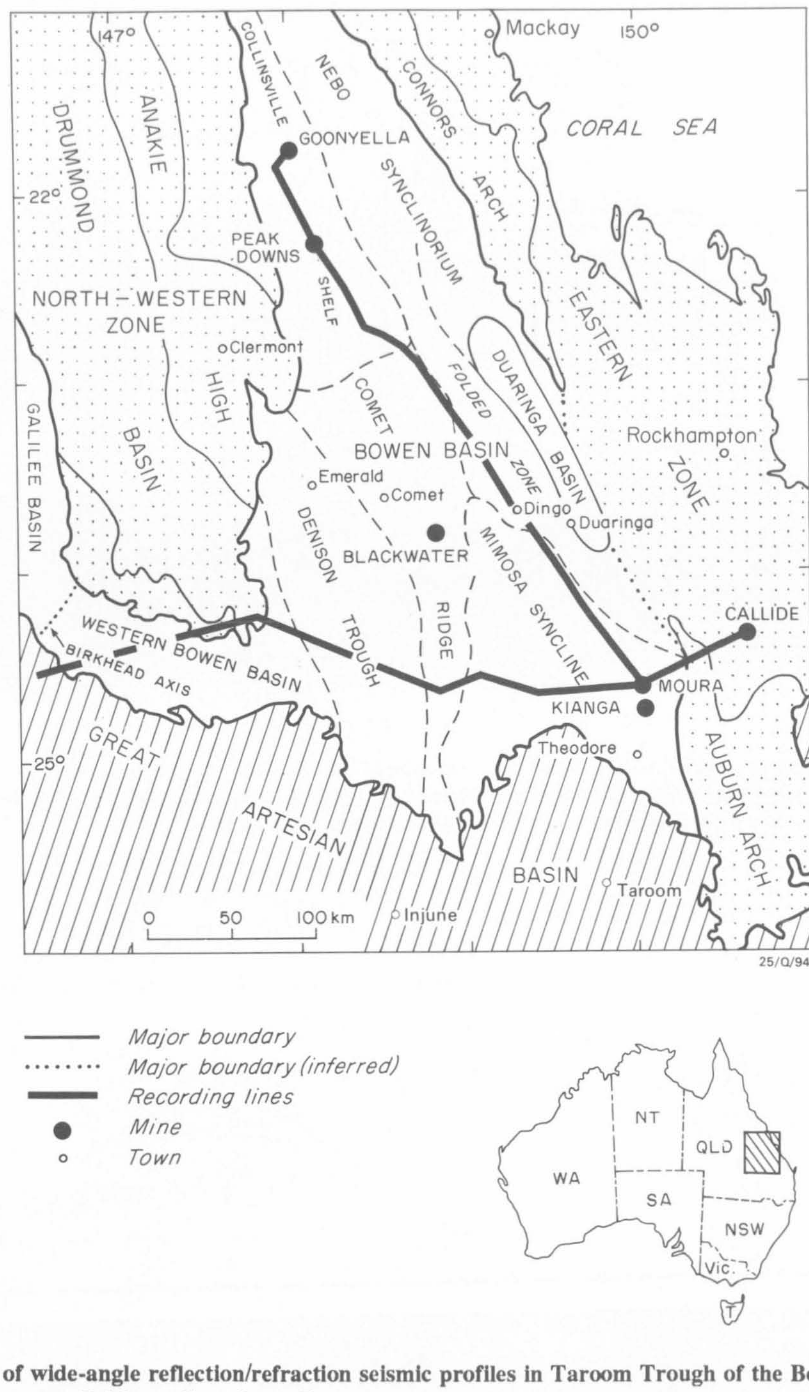

Fig. 7 Location of wide-angle reflection/refraction seismic profiles in Taroom Trough of the Bowen Basin to the north of the Eromanga-Brisbane Geoscience Transect.

Collins, 1987; Wake-Dyster & others, 1987). Figure 4 shows the location of the various profiles across the apex of the Nebine Ridge and over the Mitchell Shelf area of the western Surat Basin. Figure 5 contains the velocity/depth plots for the crust under the Nebine Ridge. Only one of the wide-angle reflection and tranmission profiles was long enough to record events which could give an indication of Moho depths and the velocity within the upper mantle. Even this line was marginal for the interpretation of an accurate Moho depth. The maximum offset recorded by Finlayson & Collins (1987) to interpret Moho depth was only 196 km and their interpreted depth of 38 km is considered to be too shallow. The two-way time of near-vertical reflections considered to be from the bottom of the Moho transition zone is 14 s on average (maximum 14.5 s) giving an estimated average depth of 42 km with a maximum depth of 44 km.

The important feature interpreted by Finlayson & Collins (1987) from the Nebine Ridge data was that there was no marked increase in velocity at mid-crustal depths (20-25 km) as there was under the central Eromanga Basin. Prominent PcP reflections from

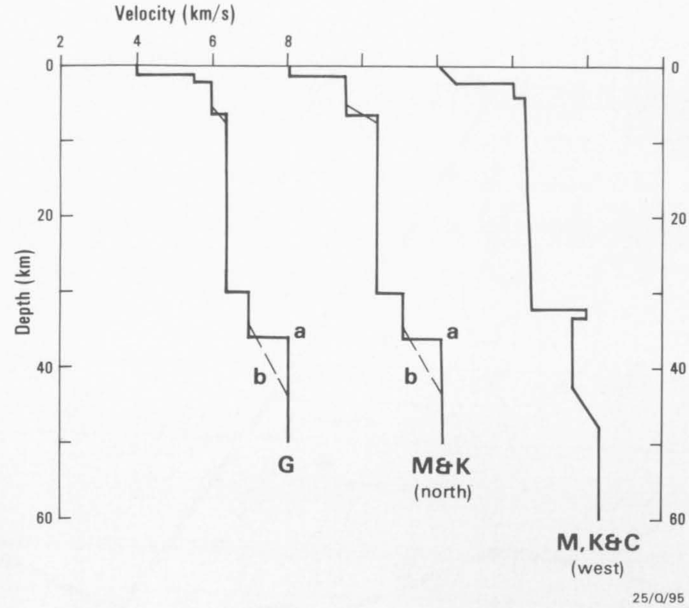

Fig. 8 Velocity/depth models of the crust and upper mantle in the Taroom Trough of the Bowen Basin to the north of the Eromanga-Brisbane Geoscience Transect. The letters against each model refer to the shot point locations indicated in Figure 7: G = Goonyella; M = Moura; K = Kianga; C = Callide.

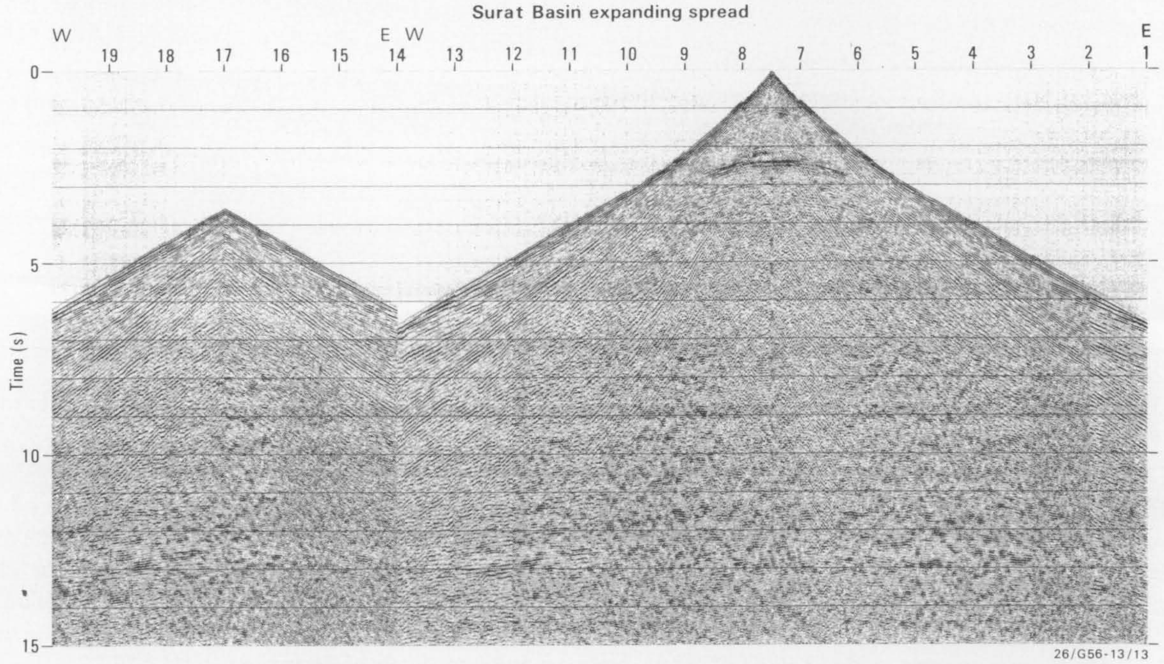

Fig. 9 Seismograms from and expanding spread reflection profile recorded across the eastern part of the Taroom Trough showing an increase in the strength of reflections at two-way times greater than 7 s, corresponding to mid-crustal depths. The field parameters for this expanding spread work have been given by Wright & others, (1989b).

mid-crustal depths were not recorded. Instead, the interpretation included a steady velocity gradient at mid-crustal depths, below a small velocity increase at 14-15 km depth from P-wave velocities of 6.02 km/s in the upper crustal basement to about 6.2-6.3 km/s at mid-crustal levels. The lower crustal velocities (6.2-6.7 km/s) are lower than the P-wave velocities in the lower crust farther west, emphasizing that there is a significant difference in the velocity profile of the Nebine Ridge from that under the central Eromanga Basin at the western end of the transect. Finlayson & Collins (1987) indicated that the Nebine Ridge velocities were more like those of the Canaway Ridge, lending weight to the idea that the two were possibly rifted fragments of early Palaeozoic continental margin.

Under the apex of the Nebine Ridge, the velocity increase at 14-15 km depth corresponds in some places with reflections seen on continuous profiling records (Finlayson & others, this Bulletin). Geologically, this may correspond to a boundary between the Roma granites and Timbury Hills metasediments in the upper crustal basement, to a more mafic sequence or higher-grade metamorphic rocks in the middle crust. Such boundaries may indicate the location of any detachment surfaces between upper and middle crustal rocks during episodes of crustal deformation.

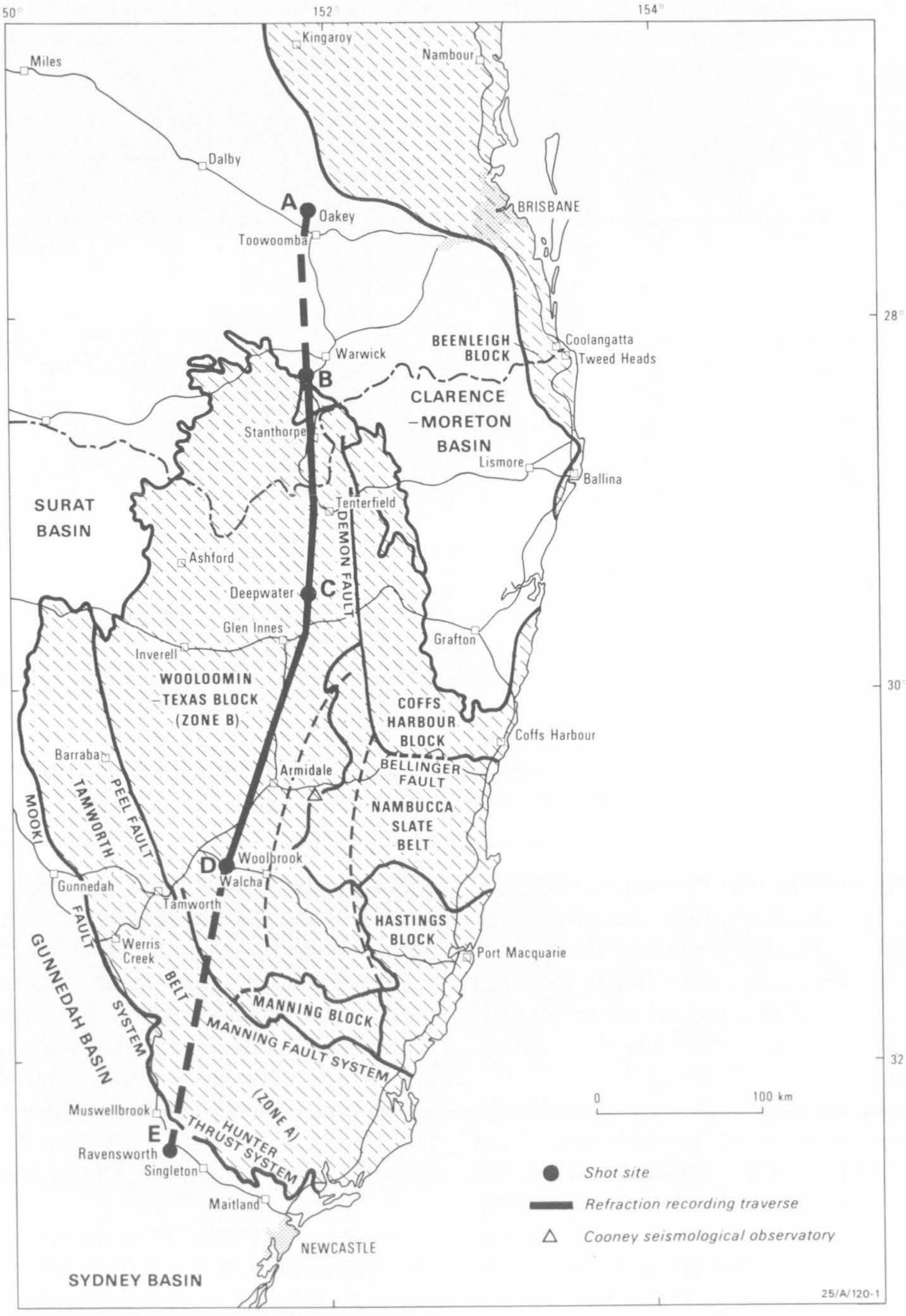

Fig. 10 Location of wide-angle reflection/refraction seismic profiles across the New England Fold Belt.

The expanding spread reflection profile recorded across the Nebine Ridge (maximum shot-receiver offset of 25 km) provides more detailed information on seismic velocities in a localised area than is obtainable from the refraction/wide-angle reflection profile of Finlayson & Collins (1987). From a combined analysis of the reflections from the boundary between sedimentary rocks and basement, and the refracted arrivals from the expanding spread that have propagated through basement rocks, Wright & Barker (1989) inferred the presence of an increase in shallow basement velocities from 5.5 km/s in the west to 5.9 km/s in the east over a distance of about 20 km centered on the crest of the ridge. Furthermore, the S waves recorded on the expanding spread indicate a surprisingly low P wave to S wave velocity ratio (V_P/V_S) of between 1.5 and 1.6, implying an average Poisson's ratio of 0.15 for uppermost basement rocks. The most reasonable explanation of the the S wave velocities is moderate, these reflections indicate an average V_P/V_S ratio of 1.67, which compares favourably with the value of 1.69 for the upper crust below the Nebine Ridge derived from the refraction data of Finlayson and Collins (1987).

Across the Mitchell Shelf a number of short wide-angle reflection and refraction lines are interpreted to give velocity/depth profiles for the upper crust. The important feature identified from these profiles is that there is a velocity increase associated with the strong reflectors termed the Foyleview Structure by Finlayson & others (1990). The location of the seismic lines is shown in Figure 4: they were specifically targeted at the prominent reflectors of the Foyleview Structure. The velocity/depth profiles (Fig. 5) show that the P-wave velocity increases from 5.8-6.1 km/s in the upper crustal basement to 6.3-6.4 km/s deeper than 10-11 km,

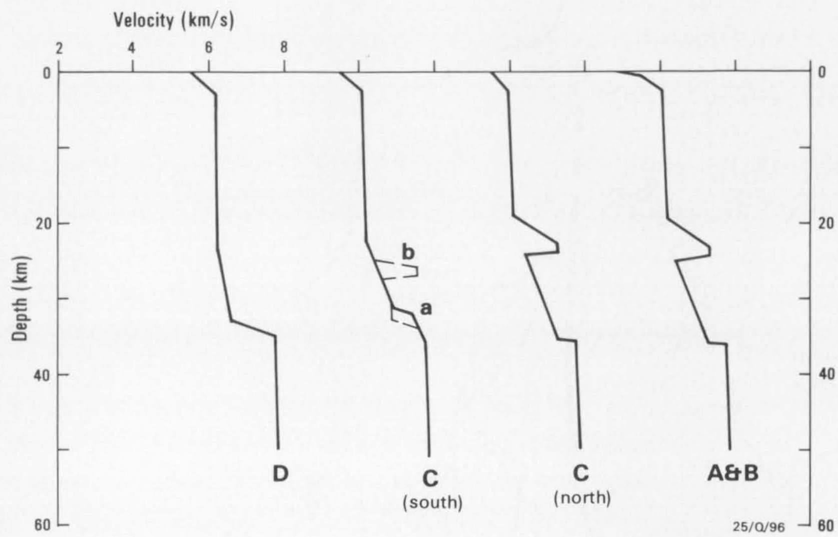

Fig. 11 Preliminary velocity/depth models of the crust and upper mantle across the New England Fold Belt. The letters against each model refer to the shot point locations indicated in Figure 8.

low value of Poisson's ratio is the presence of rocks with high quartz contents (Kern, 1979, 1982) and a high degree of fracturing possibly related to alteration of feldspars (Hall & Al-Haddad, 1979). This is consistent with the presence of gneisses and schists in borehole samples from the Nebine Ridge area (C. Murray, personal communication).

Reflections from the deep crust were generally not sufficiently strong or persistent to enable r.m.s. (root mean square) velocities to be measured. However, a series of prominent localised reflections (r.m.s. velocity terms in range 4.3-5.1 km/s), observed at shot-receiver offsets between 9.5 and 19.5 km and at times between 8.7 and 9.3 s, are interpreted as P-to-SV or SV-to-P reflections from a boundary that gives rise to a 'bright spot' in the near-vertical incidence P reflection data at a time of 6.3 s (depth 18 km). Although the control over corresponding closely with the onset of prominent reflections at 3-4 s two-way reflection time. Thus, here again, a velocity is identified which can probably be equated with a more mafic sequence or higher metamorphic grade rocks in the middle crust. The reflectors, associated with this velocity increase, dip towards the west-to-southwest, and one geological interpretation is that an upthrust middle-crustal sequence possibly exists, the upper crust being detached from the mid-crustal rocks along the velocity boundary.

To obtain more detailed information on near-surface seismic velocities, stand-alone recorders were deployed to record both the Nebine Ridge expanding spread shots and the normal shots used in the near-vertical incidence profiling across the Nebine Ridge and on the Mitchell Shelf (Bracewell & Finlayson, 1985). Figure 6 shows the seismograms recorded by a geophone array attached to a portable recorder for a series of 12 kg shots at offset distances between 0 and 4.5 km on the Mitchell Shelf. The P and S wave trains are generally clear, and yield average P and S wave velocities of 2.24 and 1.21 km/s, respectively ($V_P/V_S = 1.85$) in the distance range 1.0-4.0 km/s; these results apply to the Mesozoic sedimentary rocks at depths less than 1 km.

TAROOM TROUGH OF THE BOWEN BASIN

At the latitude of the Eromanga-Brisbane Geoscience transect there are no wide-angle reflection and refraction seismic data that enable a direct interpretation of velocities in the crust and upper mantle in the Taroom Trough. However, expanding spread data have been recorded on BMR reflection Line 14, and farther north, in the outcrop area of the Bowen Basin, there are data which give some indication of the velocity structure of the crust and upper mantle (Collins, 1978, 1980; Leven, 1980). These latter data were collected using coal mine blasts as seismic sources and are, therefore, tied to the Taroom Trough between latitudes 21° and 25° South, in the region between the Comet Ridge and the New England Fold Belt (Fig. 7).

Collins (1978, 1980) interpreted crustal velocity models shown in Figure 8 along reversed seismic lines. The P-wave velocities at depths less than 7 km (within the basin sequences) are in the range 4.0 to 5.54 km/s, but within basement, in the depth range 7-30 km, the velocity is 6.39-6.40 km/s. In the lower crust, the data can be modelled to interpret a velocity of 7.07 km/s in the depth range 30-36 km, or as a velocity gradient of 7.08-8.22 km/s in the depth range 34-44 km below a step increase in velocity to 7.07 km/s at 30 km depth. Either way, the data indicate a high-velocity sequence in the lower crust above a Moho at 36 km depth (possibly deeper if a velocity gradient is thought appropriate).

Leven (1980), interpreting single-ended seismic data westward across-strike from the coal mines at Moura and Callide (Fig. 7), determined a basement velocity of 6.3-6.49 km/s in the depth range 4-32 km (Fig. 8), in quite good agreement with Collins above. Leven, however, interpreted the velocity increase at 32 km as being the Moho, with a velocity varying in the range 7.6-8.3 km/s at depths of 33-48 km. It is possible, however, that the recording geometry did not permit the detailed structure of the lower crust to be determined. Collins' Moho depth of greater than 36 km is considered the better-determined value. Collins (1978) interpreted the upper mantle velocity as 8.10 km/s from reversed seismic data.

One important feature of the velocity profiles under the Taroom Trough of the Bowen Basin is the lack of

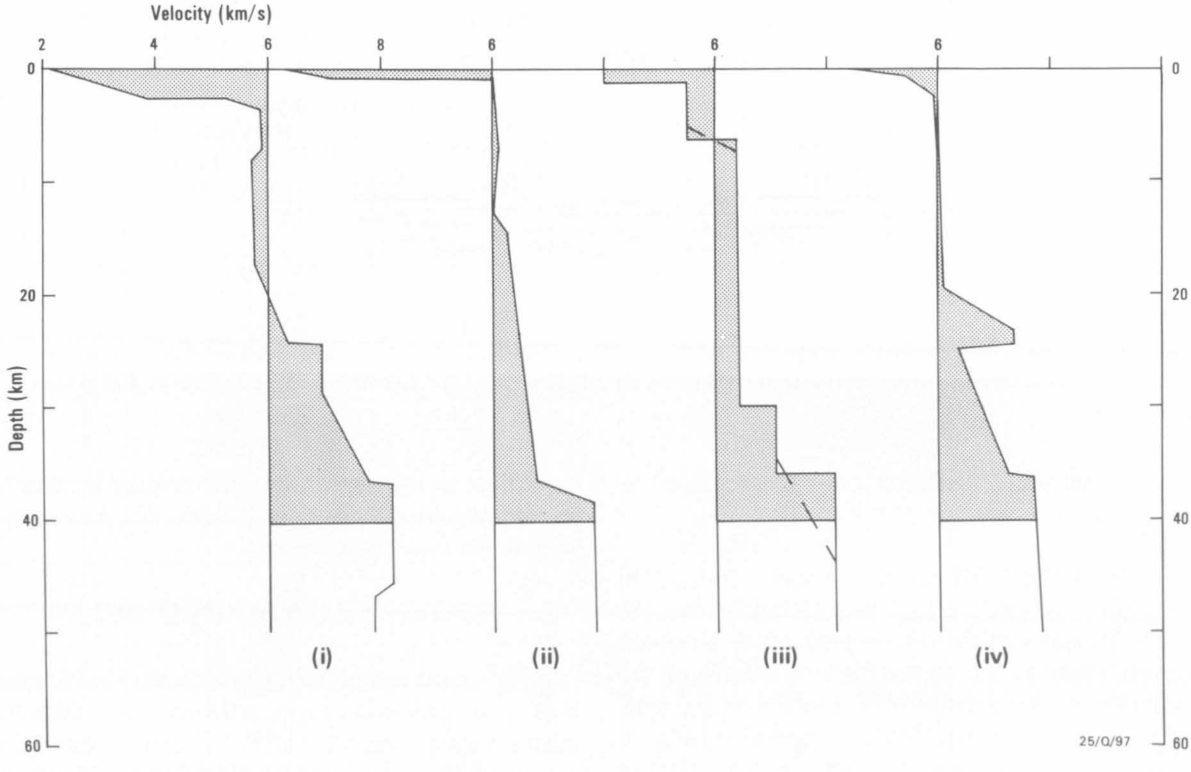

Fig. 12 Comparison of four velocity/depth models along the Eromanga-Brisbane Geoscience Transect (the Taroom Trough model is from over 200 km north of the transect). (i) central Eromanga Basin; (ii) Nebine Ridge; (iii) Taroom Trough of the Bowen Basin; (iv) New England Fold Belt.

velocity boundaries within basement down to depths of 30-32 km. This contrasts the crust under the trough with that under the central Eromanga Basin and the Nebine Ridge. This quite uniform velocity of about 6.4 km/s is in accord with the reflection profiling data across the trough on BMR Line 14 (Finlayson & others, this Bulletin), which is characterised by a lack of prominent events or horizons below the Permo-Triassic sedimentary sequences. Another significant feature is the lack of a thick upper crustal basement with a velocity of about 5.9-6.2 km/s as exists to the west within the Thomson Fold Belt rocks.

At depths below 30-32 km the velocity increases to over 7 km/s, strongly suggesting that there is an increase in mafic content in the lower crust as the Moho is approached.

An expanding reflection spread (maximum shot-receiver offsets 25 km) was recorded over the Taroom Trough with the eastern limit of the shots and receivers close to the faulted eastern margin of the trough (Fig. 9). Velocity analysis of reflection segments (Wright & others, 1990a) yields an interval velocity of 5.4 km/s for the uppermost, relatively non-reflective basement (Table 1). This estimate is a little low for uppermost basement, and is not well-constrained owing to a poor estimate of r.m.s. velocity for the entire sedimentary section (from only four reflection segments). Although there is not a distinct, non-reflective upper crust (Finlayson & others, 1990), a marked increase in reflectivity is evident in the expanding spread seismograms at depths close to 17 km (Fig. 9). The interval velocity for this more reflective region, which extends to depths of nearly 39 km, is 6.7 km/s. The underlying non-reflective zone is interpreted as upper mantle (Wright & others, 1990a,b). The interval velocity for the entire basement region, which provides a reasonable approximation to the average velocity, is 6.3 km/s. This is a relatively low average, in qualitative agreement with both the refraction velocity model across the Nebine Ridge to the west (see previous section and Finlayson & Collins, 1987) and the preliminary velocity model for the New England Fold Belt (see next section), but differs in the suggestion of a distinct increase in seismic velocities at mid-crustal depths that is associated with the increase in reflectivity.

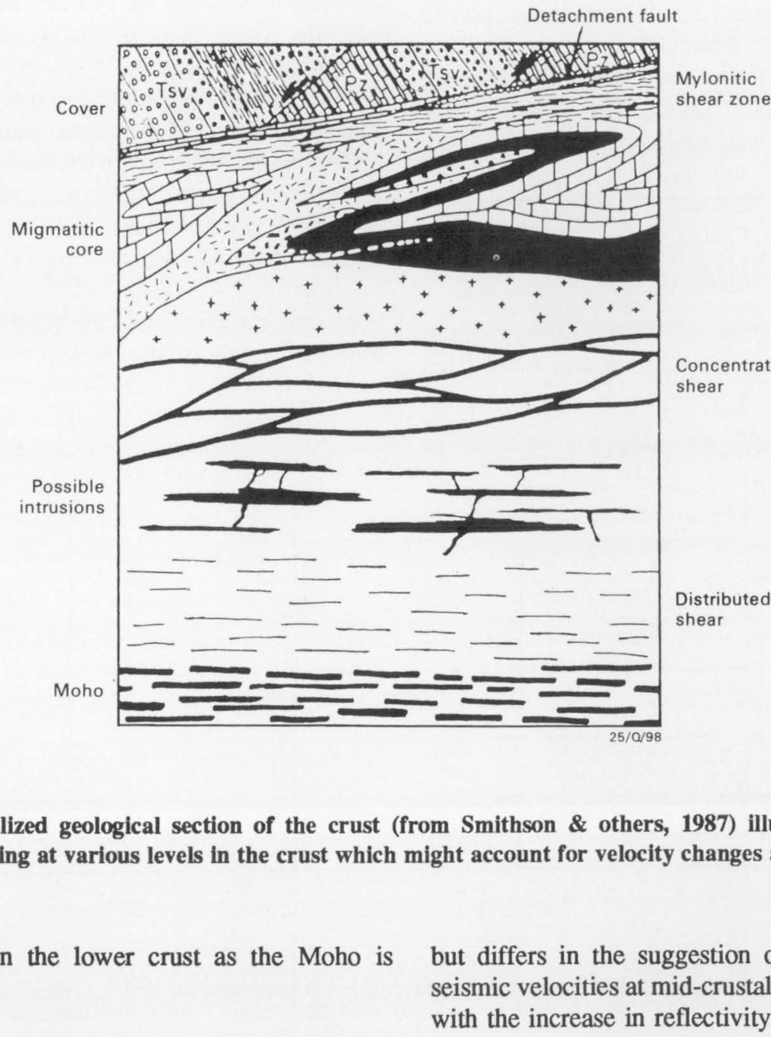

Fig. 13 Generalized geological section of the crust (from Smithson & others, 1987) illustrating the possible processes occurring at various levels in the crust which might account for velocity changes and reflection events.

NEW ENGLAND FOLD BELT

At the eastern end of the transect, in the New England Fold Belt, one wide-angle reflection and refraction seismic survey gives some idea of crustal velocities. The data were obtained across the New England Block, the southern part of the fold belt system (Fig. 10). The northern end of this seismic line intersects the east-west reflection profile.

Figure 11 shows some of the preliminary velocity/depth models, interpreted largely from travel-time and qualitative amplitude information. Further work is required to finalise the models but some features stand out. In the southern part of the New England Block, the crustal velocity model is comparatively simple, with a velocity of 5.5 km/s at the surface increasing to 6.03 km/s at 2.5 km depth; the majority of the crustal section, 2.5-30.5 km, has a velocity of 6.03-6.45 km/s. Below this, between 30.5 km and the Moho at 35.5 km depth, there is a transition to an upper mantle velocity of 7.7 km/s. The crustal velocities are significantly less than the basement velocities interpreted farther west under the Nebine Ridge and Taroom Trough at the same depths.

In the northern part of the New England Block, the preliminary models give some idea of the crustal velocities under the reflection profiling data. The main difference from the southern part of the block is in the middle crust, where a high velocity (7.3 km/s) nose is tentatively interpreted at 19-24 km. The Moho is interpreted at 36 km depth. The high velocity feature at mid-crustal depths corresponds to prominent reflection events at 7-8 s two-way time on continuous profiling records, interpreted as a possible detachment surface, above which oroclinal bending occurred (Finlayson & others, 1990).

DISCUSSION

The velocity/depth profiles obtained along the Eromanga-Brisbane Geoscience Transect demonstrate that no single velocity model can be considered appropriate for the whole region. The crust is inhomogeneous, and the data for the upper mantle indicate that there are inhomogeneities below the Moho as well. The velocity data are insufficient to examine any particular part of the transect in great detail, but the velocity values and the nature of the velocity gradients within the crust and at the Moho must be considered different in the four areas studied: the central Eromanga Basin, the Nebine Ridge, the Taroom Trough of the Bowen Basin, and the New England Fold Belt. Figure 12 shows a comparison of a single, representative, velocity/depth profile from each of these four areas of the transect.

The interpretation of the crustal velocity data is best done in conjunction with the continuous reflection profiling data. Within the crust the sharp velocity increases must be regarded as discontinuities between rocks of different composition (more mafic?) or different (higher?) metamorphic grade. Either way, the locations of these changes must also be considered as sites for changes in the physical properties of rocks and potentially as sites of shearing, ductile flow, or fluid channelling, and must also be considered as sites for detachment of crustal units during episodes of crustal deformation. Hence, it is to be expected that they are also potential sites for near-vertical reflections on continuous profiling records. Figure 13 (Smithson & others, 1987) illustrates, in a general way, the possible processes resulting in significant changes in P-wave velocity and the generation of sites for reflections.

If one considers only the simple increases in velocity from values less than 6.3 km/s to those over 6.3 km/s, it is evident that such increases tend to occur at quite different depth ranges along the transect. The location of such a change can be considered as a "seismic terrane" boundary. In the central Eromanga Basin such a change occurs at 21-25 km depth and is associated with the top of a lower crustal reflective zone. Under the apex of the Nebine Ridge, velocities of over 6.3 km/s are only reached at about 18-20 km depth in a gradient zone and may not be a geological boundary. But under the eastern Nebine Ridge, such a change occurs at 10-12 km depth and is associated with dipping reflectors, the Foyleview Structure. Under the Bowen Basin, such a velocity change occurs immediately under the Permian basin sequences at 6-7 km depth; and under the northern New England Block, the change does not occur until a high velocity nose is seen at about 20-22 km. This infers that significant lateral changes in the bulk velocity of the upper crust exist, and that the tectonic models for the region must account for them. Ramping/thrusting of higher density rock from deeper levels in the crust is one mechanism that can lead to the shallowing of rocks with velocities over 6.3 km/s; the presence mafic volcanic rocks is another that must be considered.

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GEOPHYSICAL AND PETROLOGIC PROPERTIES OF THE CRUST/MANTLE BOUNDARY REGION, EASTERN AUSTRALIA: RELEVANCE TO THE EROMANGA-BRISBANE TRANSECT

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ABSTRACT

Petrologic (geologic) information derived from deep-seated rock fragments (xenoliths) in basaltic rocks from eastern Australia has been used to construct a stratigraphic section of the lower crust and upper mantle beneath the basaltic provinces of eastern Australia. This information is combined with the seismic, gravity and magnetic data available for the Eromanga Transect, as well as MAGSAT and heat flow data, to interpret the nature and evolution of the lower crust and crust-mantle boundary across the Eromanga Transect profile. It is suggested that the change in style, orientation and depth of reflectors beneath the western part of the Roma Shelf marks the boundary between older, more silicic deep-seated sequences to the west and a more mafic, younger component to the east. This younger component probably represents deep-seated intrusive events accompanying the Tertiary (and younger) basaltic volcanic activity.

INTRODUCTION

Studies of the lower crust, uppermost mantle, crust-mantle boundary (CMB) and Mohorovicic (Moho) discontinuity require the integration of all available petrologic and geophysical data to produce geologically realistic models. Petrologic constraints are now available in considerable detail for the deep-seated lithologies characterizing many regions in eastern Australia and can be incorporated into seismic interpretations. The specific regions where such constraints are documented or known in detail are shown in Figure 1. They coincide with the band of basaltic provinces approximately paralleling the eastern continental margin and generally coinciding with the uplifted region of the Great Dividing Range.

The direct petrological data are derived from xenoliths (fragments) of the normally inaccessible deep-seated rock types, carried to the surface in basaltic magmas from depths up to about 70 km depth. At present, eastern Australia is the most suitable region for interpreting the nature of the deep and middle continental lithosphere, because the xenoliths are available over a region about 4000 km long. Eastern China may prove to be of similar value in providing a statistically meaningful sampling of deep-seated lithologies, but at this stage few data are available. Data of varying detail are also available for continental lithospheric columns from areally restricted localities including, e.g. Moses Rock, western USA (McGetchin & Silver, 1972) and Spitsbergen (Amundsen & others, 1987), which are valuable for comparison. In addition to the geographically extensive sampling in eastern Australia, there are also many individual xenolith localities with a mix of rock types that allow well-constrained construction of upper mantle-lower crust stratigraphies.

The most outstanding examples of these are at Bullenmerri/Gnotuk in Victoria (Griffin & others, 1984), Anakie (Wass & Hollis, 1983), north Queensland (Rudnick & Taylor, 1987; 1990), and east-central Queensland (Griffin & others, 1987). Corroborating but less comprehensive information comes from over 100 other localities.

There is no known xenolith locality directly situated on the Eromanga-Brisbane Geoscience Transect which might yield a complete data set for the upper mantle/lower crust region. However, as discussed below, some judicious southerly extrapolation can be made using the well-characterized crust-mantle stratigraphy for east-central Queensland, some 200 km north of the Transect. In addition, the general concepts reviewed below on the lithology and physical properties of the eastern Australian region should be applicable in a general way to features observable on the seismic sections of the transect.

The aeromagnetic and gravity data enhance the interpretation of the seismic data from the Eromanga-Brisbane Geoscience Transect (Wellman, this Bulletin), especially at upper to mid-crustal levels. When considering deeper features in the region of the CMB (about 9-14 seconds two-way travel-time), it is also critical to take into account the geothermal profile of the lithospheric section and long-wavelength crustal-source magnetic anomalies observed at satellite altitudes above Australia (MAGSAT data). The three in-situ petrophysical properties of intrinsic seismic velocity, thermal state (temperature), and magnetic intensity are interdependent for similar rock types, but may vary significantly with rock type.

LOWER CRUSTAL AND UPPER MANTLE ROCK TYPES IN EASTERN AUSTRALIA

Realistic interpretation of seismic data critically depends on knowing, firstly, the mineral assemblages and specific petrophysical characteristics of the rock types through which the seismic waves have propagated, and, secondly, their ambient pressure (P) and temperature (T). It is therefore useful to define the dominant high-pressure rock types for eastern Australia and to indicate their spatial (vertical mainly) relationships. For discussion of cited V_P values, refer to O'Reilly & Griffin (1985), Griffin & others (1987), and O'Reilly & others (1990).

Four main rock types are represented in the xenolith suites:

(1) Mantle wall-rock. The main mantle rock type beneath eastern Australia from the CMB down to about 55-60 km is spinel lherzolite. This comprises olivine (40-80%), orthopyroxene (10-40%), clinopyroxene (Cr-diopside, not jadeite; 5-40%), and spinel (0-5%), ± amphibole, ± mica, ± apatite, ± CO_2-rich cavities). V_P range (900°C, 10kb) is 7.5 to 8.0 km/sec.

At depths greater than about 55-60 km, garnet lherzolite constitutes the major mantle wall rock. Garnet lherzolite represents the higher pressure equivalent of the spinel lherzolite assemblage. The equilibrium describing the major phase change is:

clinopyroxene(1) + spinel = clinopyroxene(2) + garnet ± orthopyroxene

This does not represent a simple transformation of spinel to garnet; the reaction involves clinopyroxene, so that the proportion of garnet is higher than that of spinel for a given bulk composition. The modal range for garnet is 5-20%, and the V_P range (1100°C) is about 8.0 to 8.3 km/sec.

(2) Lower crustal rocks. Dominantly mafic granulites containing plagioclase (10-80%), clinopyroxene (10-90%), ± orthopyroxene (0-30%) ± garnet (0-30%) ± wide range of minor minerals (< 5%). Their V_P range (700°C, 8kb) is 6.7 to 7.3km/sec. P-T calculations, and the relative abundance of these rock types, indicate that the lower crust is dominantly mafic, although rare quartzfeldspathic granulites have been reported (Rudnick & Taylor, 1990).

(3) Pyroxenites. These occur as sills and lenses clustered around the CMB (Fig. 4), but also scattered to deeper levels. They represent cumulates and frozen melts and contain clinopyroxene (50-100%) ± olivine (0-20%) ± amphibole ± mica ± orthopyroxene ± CO_2-filled cavities. Their V_P range (900°C, 10kb) is 7.4 to 7.7 km/sec.

(4) Garnet pyroxenites. These occur as lenses within the mantle and represent the metamorphosed equivalents of the pyroxenites described above. They contain clinopyroxene (20-80%), garnet (5-30%) ± orthopyroxene ± spinel ± CO_2-filled cavities. Their V_P range (900°C, 10kb) is 7.6 to 8.0 km/sec.

THE CRUST-MANTLE BOUNDARY IN EASTERN AUSTRALIA

Types of Evidence from Xenoliths

Data from different xenolith rock types enable the construction of a lithologic stratigraphy at depth using calculated pressure(P) and temperature(T) (or just T) for individual xenoliths. Such geothermobarometry calculations locate the depths of occurrence of given rock types, and the temperature-pressure (or depth conversion) plot defines the geothermal gradient existing at the time of entrainment of the rock fragments (i.e. the age of the host carrier basalt). Thus, a rock type sequence and a thermal profile result.

The geothermal profile for regions below the basaltic provinces of eastern Australia (shown in Figure 2) was constructed empirically by plotting pressure(P) and temperature(T) values calculated for individual xenoliths. The original geotherm was constructed using xenoliths from one locality (Bullenmerri/Gnotuk) in western Victoria. The homogeneity of the minerals and the xenoliths used to define this geotherm (Griffin & others, 1984, 1987) strongly suggest that the P-T estimates reflect ambient conditions in the mantle at the time the xenoliths were plucked by the rising basaltic magmas (O'Reilly & Griffin, 1985). Despite a greater degree of lithological and microstructural heterogeneity, all subsequently acquired data from other eastern Australian localities, including eastern Queensland, are compatible with this locus (O'Reilly, 1989). This geotherm reflects a high geothermal gradient down to about 80 km and shows a strong curvature at about 25-35 km. The shape and path of this geotherm require heat input by underplating (basaltic intrusion) around the 25-35 kmdepth range (Cull & others, 1989). Although this geotherm records the P/T regime at the time of emplacement of the host magmas, if the basaltic hosts are very young, it can also be taken to represent the present-day heat flow, e.g. in western Victoria and east-central Queensland.

This geotherm contrasts with the geotherm for the eastern margin of the Australian craton (EMAC) derived from xenolith data, using the same methodology. The EMAC geotherm is 150°C to 200°C cooler than that for eastern Australia at any given P. A full discussion of this cratonic thermal profile is beyond the focus of this paper, but it does emphasise the general thermal

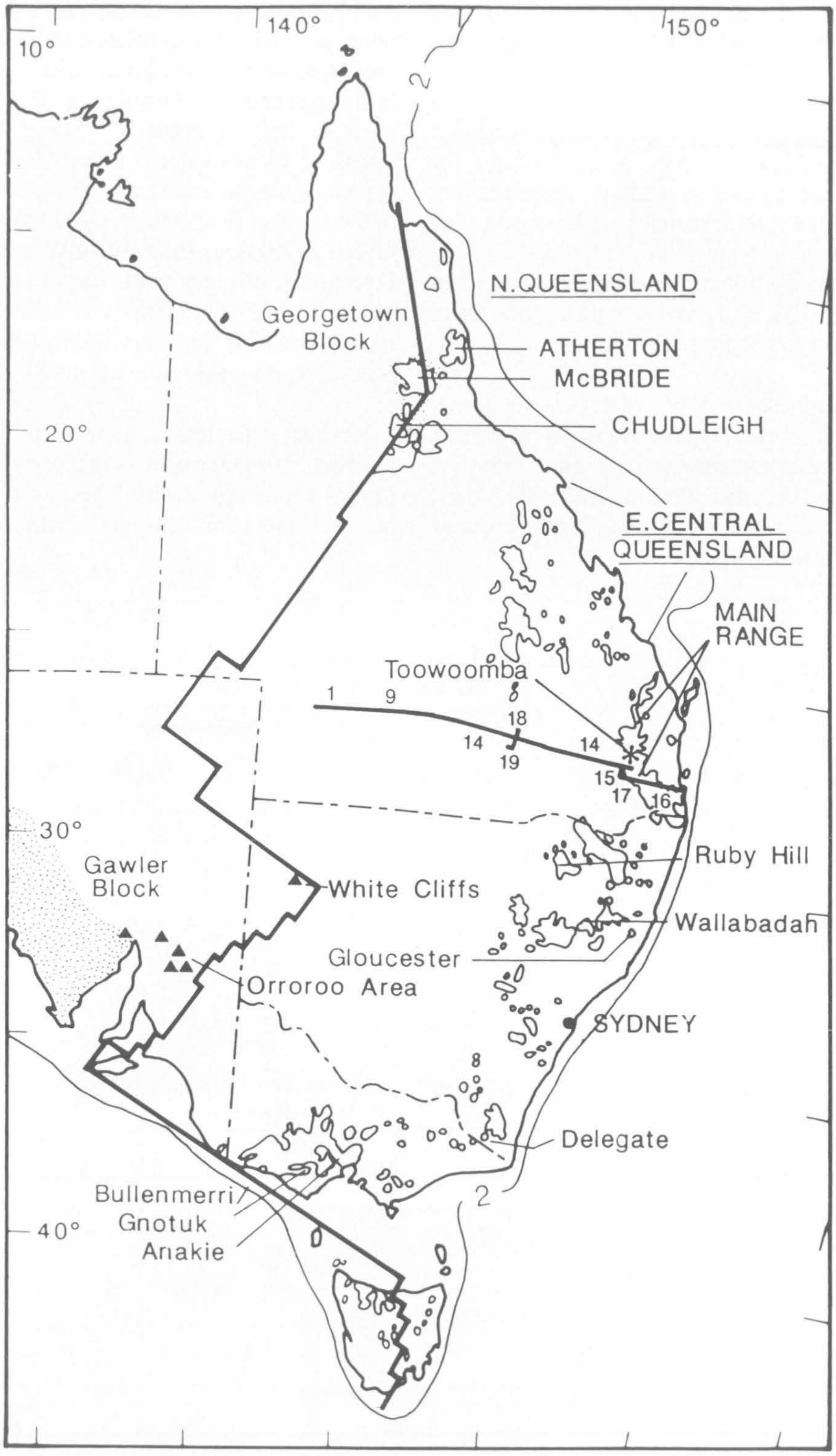

Fig. 1 Locality map showing: key xenolith localities in eastern Australia; distribution of basaltic provinces in eastern Australia (irregular enclosed areas); craton boundary (heavy jagged line); and the location of the Eromanga Transect line (from BMR data, 1988).

uniformity of the eastern Australian Phanerozoic terrane at the time of activity of each basaltic episode, and the difference from the EMAC regime. This may be important to the change in seismic reflection characteristics noted for western regions of the Eromanga-Brisbane transect (Finlayson & others, this Bulletin).

Petrophysical measurements (e.g. density, magnetic intensity and sound wave velocity as an analogue for seismic wave velocity) can be made on representative high-pressure xenolith rock samples, under appropriate high-pressure and/or temperature conditions. Such measurements, especially compressional and shear wave velocities (V_P and V_S), then provide realistic parameters which can be used in geophysical models.

Laboratory measurements of V_P analogues for spinel lherzolite mantle wall-rock, typical of a major proportion of the mantle beneath eastern Australia, range from 7.9 to 8.4 km/sec at 10kb and 25°C (O'Reilly & others, 1990), with the lower values reflecting higher contents of pyroxenes and/or amphiboles.

Types of Evidence from Geophysical Data

Seismic reflection and refraction profiles provide a critical database for modelling lower-crust and upper mantle structures and lithologies when petrologic and other geophysical data (magnetic, thermal, gravity) are also available. The information provided by the Eromanga-Brisbane Geoscience Transect represents a quantum leap in available seismic data and will be discussed in more detail below. Previously, the most comprehensive seismic data were from the refraction line of the Lachlan Fold Belt reversed traverse (Finlayson & others, 1979) from southern NSW to northern Victoria. Fortunately, this line crossed a portion of the broad belt of basaltic activity, which is the source of xenolith rock type information, and so is of direct relevance to mantle regions represented in the xenoliths.

Excellent reflection data were provided by Mathur (1983a,b; 1984) for sites in the lower Bowen Basin and central Eromanga Basin. There is a good correlation between the refraction and reflection data for the

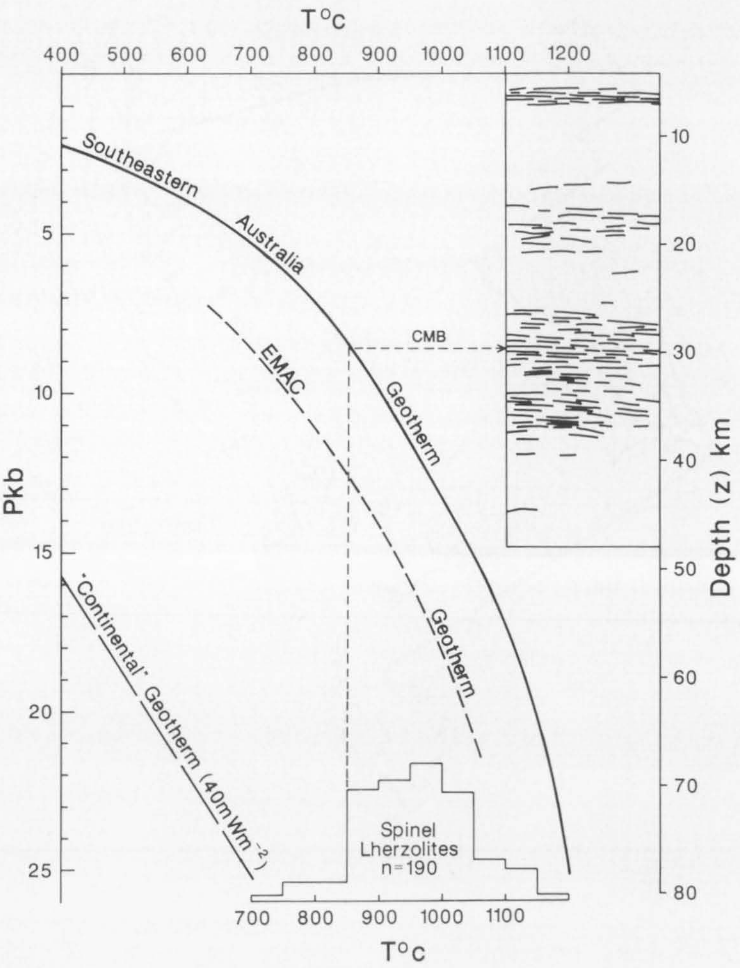

Fig. 2 The empirical, xenolith-derived southeastern Australia geotherm with a "continental" geotherm of 40 mW m^{-2} shown for comparison. Histogram is a compilation of T data for spinel lherzolite mantle rocks. Uppermost abundant occurrence of this rock type defines CMB (dashed line). Inset shows location of CMB in a sketch of a crustal column from site F in the Eromanga Basin based on Mathur (1983a).

northern Eromanga Basin sites (Finlayson, 1983; Finlayson & Mathur, 1984). The reflection profiles typically show, (a) a region of transparent upper crust underlying shallow sediments, (b) a mid-crustal region with a few horizontal reflectors, (c) a region at about 8-12 sec two-way travel time (24-36 km) characterized by numerous horizontal reflectors, each extending for several kilometers, and (d) beneath this latter zone (interpreted as lower crust by Mathur), an essentially transparent upper mantle region extending to 16-20 sec (maximum recording depth). This type of profile is similar to the eastern part of the Eromanga-Brisbane transect, but contrasts with that to the west as discussed further below and by Finlayson & Collins (1987).

Magnetic data provide additional information, which can be used to interpret the nature of the deep-seated environments. Mayhew & Johnson (1987) have constructed a magnetization map for Australia based on MAGSAT records (Figure 3). This long wave-length magnetic signature is not perturbed by small to meso-scale magnetic anomalies, but represents the collective magnetization of all rocks to the depth of the Curie point. The magnetic mineralogy of rock types representative of lower crust and upper mantle sequences has been determined worldwide (Wasilewski & others, 1979; Frost & Shrive, 1986) and for eastern Australia in particular (P. Wasilewski, personal communication, 1988). The results demonstrate that the measurable magnetization is overwhelmingly due to magnetite-series minerals and thus the relevant Curie point temperature is about 550°C. In western Victoria and east-central Queensland, this is about 13-15 km (well within the crust), by reference to the 550°C depth on the geotherm (Fig. 2). Because magnetic rock properties and the thermal regime are interdependent, and because the crustal regions of eastern Australia beneath the basaltic provinces are inferred to have similar structures and lithologies, this magnetization map probably reflects the topography of the Curie isotherm in these regions. Consequently, these magnetization contours also reflect present-day regional heat flow.

It is important to stress that such an interpretation is valid only for regions with crustal lithologic profiles similar to those beneath the broad band defined by the eastern Australian basaltic provinces (Fig. 1). Other regions which have not undergone Phanerozoic basaltic igneous activity may be characterized by quite different thermal regimes at depth and/or distinct rock types with different magnetic signatures.

The Crust-Mantle Boundary (CMB) beneath the Eastern Australian Basaltic Provinces

Full details of the nature of the lower crust, upper mantle and CMB for eastern Australia have been published in a series of papers (e.g. Griffin & O'Reilly 1987; Griffin & others, 1987; O'Reilly & others, 1988), so only a review is presented here.

The CMB is defined as the depth below which ultramafic rocks (spinel lherzolites as defined above) become volumetrically significant. This definition is petrological, and is independent of the seismically determined Moho, where P-wave velocities increase to (usually) more than 7.6 km/sec. The differences in depth to the CMB and Moho may carry significant geological information (Griffin & O'Reilly, 1987). Griffin & others (1984, 1987) showed that histograms of temperature estimates for spinel lherzolite typically show a well-defined "edge" at some low temperature(T) (850°C in Fig. 2). The depth derived by referring this temperature to the xenolith-derived geotherm (Fig. 2) is taken here as the CMB.

Beneath the basaltic provinces extending the length of eastern Australia (Fig. 1), the CMB is located at depths ranging from 25 to 40 km in different regions. Above the CMB, the dominant wall-rock consists of lower crustal mafic granulites. This is illustrated in Figure 4. The horizontal black lines above and below the CMB represent lenses and sills of basaltic origin formed by basaltic intrusions around the CMB due to rheological contrast and/or the stress distribution at the time of intrusion (Etheridge, 1988).

The seismic refraction profiles for the southern NSW reversed line (Finlayson, & others, 1979) suggest that the Moho discontinuity under the highest topography of the Lachlan Fold Belt is deep (maximum 52km). The petrologic and geothermobarometric evidence from xenoliths indicates that this seismic discontinuity does not represent the V_P change from crustal granulite to spinel lherzolite wall-rock; instead it defines the region where the metamorphic reaction of spinel lherzolite to garnet lherzolite takes place within the mantle. This change would occur over a restricted depth interval and could result in a V_P contrast of about 0.3 km/s (from about 55 to 65km) for the thermal conditions of the eastern Australian lithosphere. In the Eromanga and Bowen Basin areas, where high-quality reflection data show a zone of abundant horizontal reflectors at about 24-39 km , the Moho lies at the base of this package of reflectors (e.g. Mathur, 1983a; Finlayson & Collins, 1987). With this type of interpretation, the zone of reflectors is regarded as representing layered lower crust, with seismically transparent mantle below. This interpretation has been strengthened by several other studies showing that the refraction Moho coincides with the base of such layered zones (e.g. McGeary & Warner, 1985; Meissner, 1986).

However, it is not valid to then infer that the base of the layered zone (Moho) coincides with the CMB. Our model (Fig. 4) emphasises that the uppermost mantle consists of spinel lherzolite interleaved with subhorizontal lenses of mafic granulites and pyroxenites on a scale appropriate to reflecting seismic waves, and will therefore appear as a layered region on reflection profiles. It will also have a bulk density and V_P intermediate between "crust" and "mantle" values and will grade to higher values downwards as the incidence of mafic intrusions decreases away from the CMB. Thus, in our model, the CMB, as defined petrologically (geologically), lies somewhere within the layered zone.

This fits well with the xenolith data from east-central Queensland, which is a key locality as it is the best-characterized lithospheric section, in a petrologic sense, close to the eastern part of the Eromanga-Brisbane Transect. Spinel lherzolites in this area show a range of temperatures from 850-1050°C. By reference to the geotherm (Fig. 2), it can be seen that this corresponds to a depth range of about 28-50 km. Thus, the shallowest depth of abundant spinel lherzolite, which defines the depth to the CMB, is about 28 km.

Conclusions

There are three important concepts to emerge from the detailed studies of xenolith rock types in relation to lithospheric profiles beneath the basaltic provinces of eastern Australia:

(1) V_P's for mantle rocks are significantly lower than traditionally-accepted "mantle" V_P's for three reasons. Firstly, it is a hotter regime than a "continental" (= cratonic) environment. Secondly, the mantle wall-rock is spinel lherzolite (not dunite) and contains abundant pyroxenes ± amphibole ± apatite ± mica ± CO_2-filled volumes which result in lower V_P's. Thirdly, at least in the uppermost mantle, the mixture of lherzolite and mafic lenses (granulites or pyroxenites) lowers the bulk velocity.

(2) The lower crust is dominantly mafic, rather than intermediate or silicic.

(3) Abundant subhorizontal basaltic intrusives around the CMB, in the lower crust and uppermost mantle, correspond to zones of reflectors evident on seismic

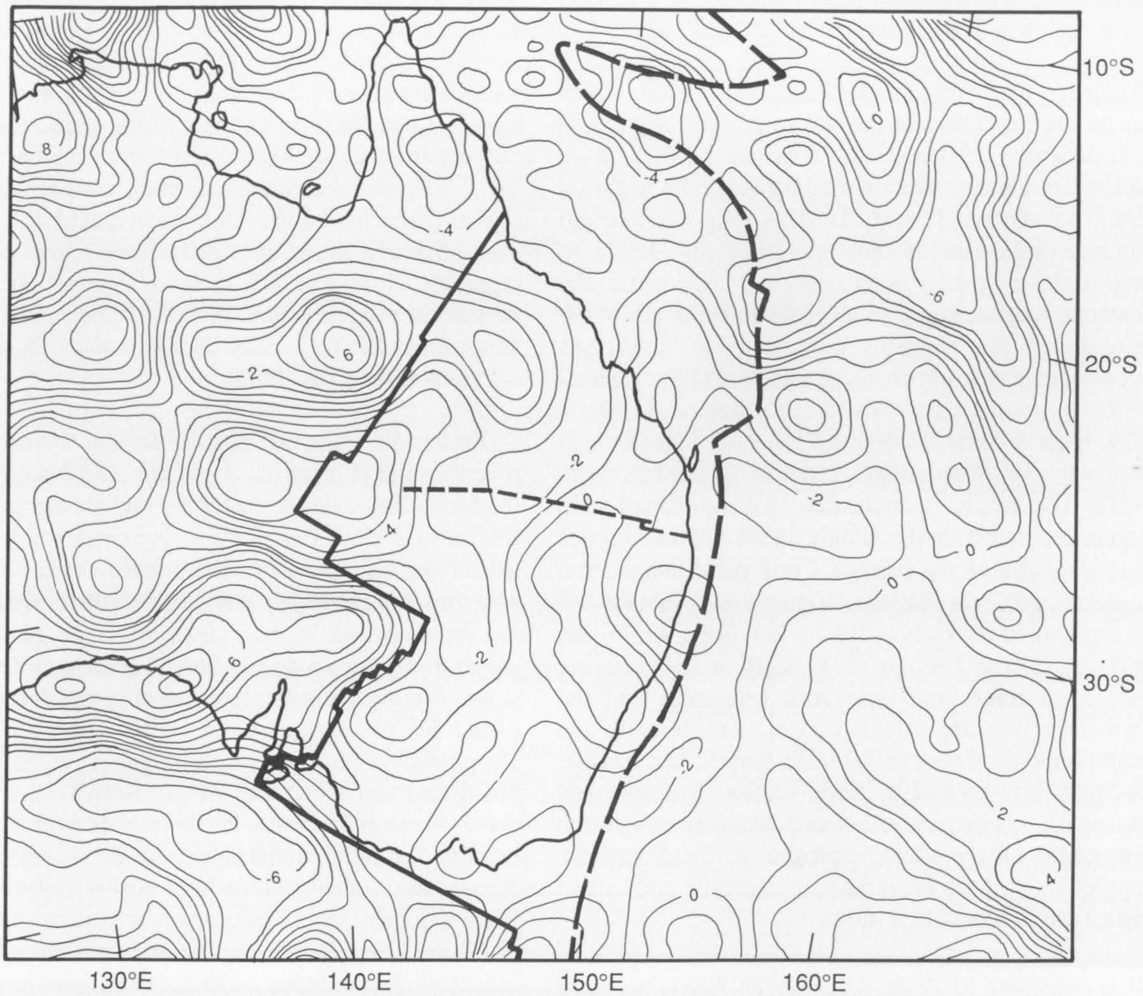

Fig. 3 Equivalent layer magnetization distribution obtained by inversion of MAGSAT total field anomaly data (from Mayhew & Johnson, 1987). Units are tenths of A/m, contour interval 0.1 A/m. Distribution represents apparent magnetization contrast in a layer of (arbitrary) thickness 40 km, the top of which is the earth's surface. Heavy line onshore is the Tasman Line; heavy line offshore is the ocean-continent boundary of Veevers (1984); line across southern Queensland is the Eromanga Transect locus.

sections. The CMB is located within (rather than at the base of) these zones.

RELEVANCE TO THE EROMANGA-BRISBANE TRANSECT

The features of the transect most pertinent to the interpretation of the deep-seated stratigraphy and the nature of the CMB are: (a) the location of reflectors, (b) the orientation of these reflectors, (c) the depth to the base of reflector zones, and (d) the presence of a band of dipping reflectors in the upper mantle beneath the Roma Shelf.

A package of reflectors, generally 6-10 km thick and located near the CMB, persists across most of the transect. The base of this reflecting zone shallows eastward from about 14 secs (two-way travel-time) beneath the Nebine Ridge to about 13 secs beneath the Taroom Trough, and to about 12 secs close to the coast. Toward the western margin of the section (beneath the Cooladdi and Westgate Troughs), the base of this zone apparently deepens to about 17 secs.

There is also a dramatic change in the transect profile style west of the Roma Shelf (e.g. Finlayson & others, this Bulletin; Finlayson & Collins, 1987). Below and west of the Nebine Ridge, the Moho discontinuity appears to be quite well defined and the package of reflectors around the CMB region appears to be tectonized. Instead of being subhorizontal as these reflectors are to the east, they show abrupt lateral changes and reversals in dip, superimposed on the general trend of the deepening of the base of this zone of reflectors westward. Coinciding with the change in orientation of reflectors is a change in their distribution through the profile. From the eastern end of the transect to about the western edge of the Roma Shelf, there are prominent reflectors through the whole crust. West of the Roma Shelf, the upper crust (beneath the uppermost sedimentary basins) is quite transparent, while the middle and lower crust contain prominent reflectors.

Although the Eromanga-Brisbane Transect is 100 km wide, there are no basaltic rocks with complete assemblages of deep-seated rock types outcropping directly within the transect region: it is necessary to extrapolate from relevant geological information outside the transect. There are some direct data available from two main xenolith localities near Toowoomba in the east (Ewart & Grenfell, 1985; Chen & O'Reilly, 1990). However, the closest comprehensive information comes from the Denison Trough area of the Bowen Basin in east-central Queensland, about 200 km north of the transect (Mathur, 1984; Griffin & others, 1987).

In east-central Queensland, the CMB, as inferred from xenolith information and Mathur's reflection profile from the Denison Trough of the Bowen Basin, lies at about 30 km. There is a zone of reflectors from about 28 to 36 km depth, which is interpreted as the zone of intrusion of basaltic magmas above and below the CMB. The Moho lies at the base of this zone, some 6 km below the crust-mantle boundary. This zone of reflectors appears analogous to that occurring at depth intervals ranging from 33-40 to 27-36 km east of the Nebine Ridge. A southerly extrapolation along the line of the Nebo, Clermont, Hay, Springsure, and Buckland basaltic provinces, as shown by Griffin & others (1987), intersects the transect around the junction of the Surat and Taroom Basins. Extrapolation of the Main Range province intersects the transect near Toowoomba in the Clarence-Moreton Basin.

The xenolith rock types from the Tertiary volcanics near Toowoomba comprise crustal granulites, spinel lherzolite mantle wall-rock, pyroxenites and very rare garnet websterites (with altered garnet). This confirms that the generalized lower crust/upper mantle lithologies are similar to those in other regions beneath the eastern Australian basaltic provinces and, in particular, to those from east-central Queensland.

The spinel lherzolites from the basaltic rocks near Toowoomba yield a temperature range from 900-1050°C (Chen & O'Reilly, 1990) at the time of their entrainment. The uppermost limit of observed spinel lherzolite occurrence corresponds to the lowest temperature(T) value (i.e. 900°C) and defines the maximum depth of the CMB. By reference to the eastern Australian geotherm (Fig. 2), this depth is 30-35 km. Beneath the Tertiary volcanics near Toowoomba, there are reflectors both above and below this depth. The xenolith data suggest that the base of the crust is located above these lowermost reflectors (12-13 secs two-way travel-time) as it is elsewhere below the eastern Australian Tertiary volcanic provinces, and that mafic sills extend into the uppermost mantle.

The low magnetization in the eastern part of the transect (Fig. 3) reflects the high heat flow implied by the xenolith data, as discussed above. The higher magnetic intensity of the central portion is consistent with a mafic lower crust, combined with a lower heat flow and correspondingly deeper Curie isotherm. The change in the character of deep reflections west of the Roma Shelf also coincides with the eastern edge of a large region of low magnetization, indicating either a hot or a weakly-magnetic crust (Fig. 3). Unfortunately, there are no xenoliths which give direct rock-type information for that lithospheric section. Silica geothermometry data (Swanberg & Morgan 1985) compiled for this region (F. Arnott, personal communication, 1989) indicate a low heat flow, so that the Curie-point isotherm for magnetite may lie near the CMB. The combination of low heat flow and low magnetization strongly suggests that the crustal column must consist largely of weakly- or non-magnetic rock types. These could be felsic rocks, amphibolites or eclogites rather than the magnetic mafic granulites characteristic of the hotter regions (Griffin & others, 1989). The low Bouguer anomalies along the western end of the transect are consistent with the occurrence of felsic rocks rather than eclogitic mafic rocks. One possible (but not unique) inference for the nature of the underlying lower crustal rock types is that they could be analogous to some of the older crustal blocks west of the "craton line" (Fig. 1), such as the Yilgarn Block in Western Australia. The Yilgarn Block also shows a relatively discrete Moho, which may coincide with the CMB, as well as low magnetization characteristics (Mayhew & Johnson, 1987).

The prominent dipping reflectors within the mantle, located below the Roma Shelf at about Kincora 13 well are located at the eastern limit of the apparently deformed crustal region. There is again no unique petrologic explanation for these. They could derive solely from tectonic interleaving of rock types with sufficiently contrasting seismic velocities or from rough Moho topography. However, if the mantle has undergone shearing this could result in a preferred orientation of minerals in the mantle rock to give a coherent fabric on a large scale. Such a deformed region could provide a sufficiently large seismic velocity contrast to produce seismic reflections (Fuchs, 1986). Such seismic anisotropy has been demonstrated by experimental determination of V_P analogues for spinel lherzolites showing preferred orientation of constituent minerals (Bezant, 1985). Anisotropy maxima within single natural samples reach 6%, representing a V_P differential of 0.5 km/sec for different fabric directions within the one sample. The presence of anisotropy, which could be analagous to that in the mantle beneath Kincora 13 well, has been demonstrated using SKS wave characteristics in some mantle sections recorded for nine stations in North America and Europe (Silver & Chan, 1988).

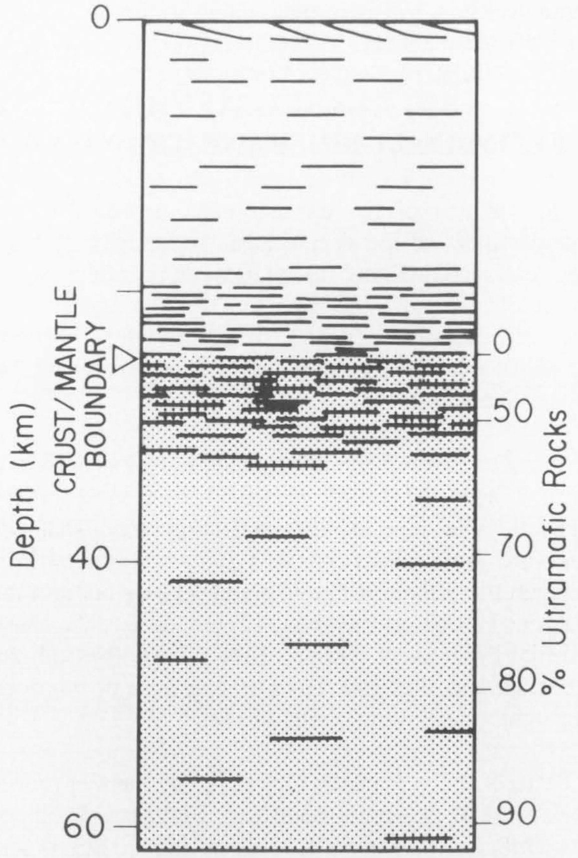

Fig. 4 Schematic representation of a generalized crust-mantle profile beneath eastern Australia basaltic provinces based on the model of Griffin & O'Reilly (1987). The dotted region depicts ultramafic mantle rocks. The black lines represent horizontal and subhorizontal basaltic intrusions.

IMPLICATIONS FOR CRUSTAL EVOLUTION

The features of the transect profile most relevant to the evolution of the lower crust are: (a) the distribution of reflectors through the crust in different profiles, (b) the change in depth of the base of the reflecting zone from about 36 km in the east and west to about 42 km maximum under the Nebine Ridge, and (c) the change in orientation of the package of reflectors around the CMB from west to east.

From the eastern end of the transect to about the western edge of the Roma Shelf, a band of subhorizontal reflectors persists around the CMB, and prominent reflections are evident through the whole crust. This part of the transect coincides approximately with the surface expression, to the north and south, of Tertiary to Recent basaltic volcanism. The eastern section is typical of the type of crust inferred from xenolith data for east-central Queensland and Toowoomba. It is consistent with relatively young crust which has undergone repeated basaltic injections. It shows high heat flow and medium to low MAGSAT anomalies (Mayhew & Johnson, 1987). The xenolith data indicate the presence of abundant basaltic rocks, especially around the CMB. The medium magnetic signatures here are due to high heat flow, and the resultant shallow Curie isotherm.

West of the Roma Shelf, the upper crust (beneath the uppermost sedimentary basins) is quite transparent while the middle and lower-crust contain prominent reflectors. The middle and lower crustal reflectors in the western section could also be derived from igneous activity, but are not likely to be basaltic considering the gravity, magnetic and heat flow data.

In the region of the Roma Shelf, the change in style and orientation of reflectors indicates possible tectonic activity. This is especially clear in lines 18 and 19 and suggests severe deformation and tectonic juxtaposition in the middle and lower crust. It may be significant that this area (Roma Shelf) overlies the region characterized by a zone of dipping reflectors in the upper mantle. As has been discussed above, a rough Moho topography or anisotropy of mantle rocks caused by deformation during tectonism could account for these mantle reflectors.

The apparently tectonized geometry of reflectors in the west of the transect suggests that any volcanism predated tectonic juxtaposition of crustal terranes forming the Cooladdi and Westgate Troughs, the Nebine Ridge, and the Roma Shelf. This is consistent with palaeogeographic reconstructions of this region through the Ordovician to Carboniferous in Veevers (1984). However, there is no evidence that any extensional (or collision) episodes were accompanied by large volumes of basaltic magmatism in that region.

It is clear from the transect profiles that the crust thickens significantly under, and west of the Nebine Ridge, consistent with increasing age of crustal stabilization. However, no xenolith data relevant to the region west of the Roma Shelf exist (there are no known Tertiary or Recent basalts there which could transport them), so the CMB cannot be as precisely located as it can farther east. The mid- to lower crustal column beneath the Westgate Trough shows seismic characteristics similar to those of a tectonized southern Bowen Basin section. However, there may be a contrast in the dominant type of igneous sill-like intrusions (in the mid to lower crust) from basaltic for the Bowen Basin section to more silicic in the western section. This is also consistent with the possible occurrence of large pluton-like bodies of silicic igneous rocks in the seismically transparent upper crust.

In summary, the combination of seismic reflection characteristics, heat flow, magnetic, and gravity data, integrated with interpretations from xenolith-derived petrologic data, indicates that the lower crust in the west is more silicic than that in the east. The more mafic component in the east is probably a result of the Tertiary and younger volcanic activity.

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XENOLITH SUITES FROM SOUTHERN QUEENSLAND: THE LITHOSPHERE BELOW VOLCANIC REGIONS

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ABSTRACT

Xenolith suites in extensive Cainozoic volcanics in southern Queensland include crustal and upper mantle assemblages. Over twenty suites occur in the Miocene Main Range volcanics, which cross the eastern end of the Eromanga-Brisbane Transect. Several suites are present in younger Miocene-Quaternary basalt fields to the north (Boyne River) and rare suites occur in basalts of the earliest Miocene Focal Peak volcano to the south.

Spinel lherzolite forms the common mantle rock. It is associated with a variety of metamorphic and igneous textured pyroxenites, wehrlites, granulites and gabbros (± accessory opaque oxides, garnet, amphibole, mica). The Gowrie Mountain and Cliffdale Quarry suites are two examples described in this study. Garnet-bearing assemblages are prominent in the Boyne basalts and at Mount Wyangapinni. Hydrous, amphibole-bearing suites occur in Boyne basalts and at Mount Mitchell, Owens Knob and Seven Mile Creek. Granulites and gabbros within the suites commonly contain calcic plagioclase, but sodic plagioclase varieties are known.

Temperature estimates for two-pyroxene, metamorphic textured assemblages range from 1000-1200°C for spinel lherzolite associations and 840-980°C for granulitic associations. Pressure-temperature estimates on Boyne garnet- two-pyroxene assemblages indicate a relatively hot geotherm, similar to or even hotter than the southeastern Australian geotherm. Projections of temperature estimates for the metamorphic assemblages onto the Boyne geotherm suggest that a granulitic lower crust lies between 22-33 km depth and that a spinel lherzolite mantle is dominant from 36-60 km depth. This matches lower crust, transition and mantle zones interpreted on the seismic profile of the Eromanga Transect adjacent to the Main Range volcanic section.

INTRODUCTION

Xenoliths and megacrysts from underlying crustal and mantle sections occur in some alkaline mafic rocks amongst the abundant Cainozoic extrusives of southern Queensland (Wass & Irving, 1976; Ewart and others, 1980; Hollis and others, 1983; Ewart & Grenfell, 1985; Robertson and others, 1985; Griffin and others, 1987; Sutherland and others, 1988). The inclusions provide a fragmentary record of lithospheric petrology at particular stages in the region's volcanic evolution. The lithospheric material here is believed (e.g. by Wellman & McDougall, 1974; Wellman, 1983; Sutherland, 1983, 1990) to have passed over 'hot spot' magma sources in late Oligocene to early Miocene time (28-21 Ma).

The 'hot spot' sources generated mafic and felsic magmas which were emplaced around and above the crust-mantle transition (Ewart & others, 1980). Lavas erupted from crustal reservoirs formed the Tweed, Focal Peak, Fassifern Valley, Coby Creek and Bunya Mountains central volcanoes. These rocks rarely contain high pressure xenoliths, but carry phenocrysts and/or show isotopic evidence of crustal fractionation and assimilation (Ewart & others, 1976, 1985; Ewart, 1982, 1985, 1987). This suggests considerable underplating and chambering of the crustal section. Later eruptives had the potential to sample such plutons and cumulate bodies.

XENOLITH SUITES

Southern Queensland basalts that contain xenolith/megacryst suites (Fig. 1) include the following:

(1) the Albert Basalt of the Focal Peak volcano (26-23 Ma; Ross, 1977).

(2) the Main Range Upper Formation, late alkaline rocks (22-18 Ma; Ewart & Grenfell, 1985).

(3) the Boyne Basalts (18-0.3 Ma; Sutherland and others, 1988).

The first suite samples lithosphere entering a 'hot spot' influence. The second reveals the lithospheric state at a late-stage along the 'hot spot' track. The last monitors lithosphere peripheral to and subsequent to the 'hot spot' passage.

The Main Range xenolith suites are the most relevant to the Eromanga-Brisbane Transect and lie on or within 50 km of BMR line 17. The Albert Basalt suite is 60 km south of BMR line 16 and the Boyne Basalt suites are 120 km north of BMR line 14.

The most extensive xenolith suites from the Main Range sequence come from a region that is only poorly recorded in the seismic section below the host volcanics. This, unfortunately, precludes close co-ordination between structure and petrology.

Each suite represents a relatively restricted part of a heterogeneously developed lithosphere. No definite material related to the garnet lherzolite zone is known, so that samples appear to come from depths of less than 55-65 km (based on identification of the spinel/garnet lherzolite transition in eastern Australia: Sutherland & others, 1984; O'Reilly & Griffin, 1985).

The range of xenolith types and associated megacryst species from sites discussed in this paper is given in Appendix 2, Table 1. Host-rock nomenclature follows that for east Australian volcanic rocks (Johnson & Associate editors, 1989).

ALBERT BASALT SUITES

Only two localities for inclusions are recorded (Ross, 1977). A plug of hawaiite near Palen Creek contains spinel lherzolite and orthopyroxene(opx) megacrysts (mineral analyses, Appendix 2, Tables 2a and 3a) and a basalt at Richmond Gap contains clinopyroxene(cpx) and orthopyroxene megacrysts (Appendix 2, Table 3a).

Palen Creek

The Palen Creek lherzolite is notable for more Fe-enriched clinopyroxene than usually found in such southern Queensland rocks (cf. Ewart & Grenfell, 1985, and other examples herein). This may reflect significant

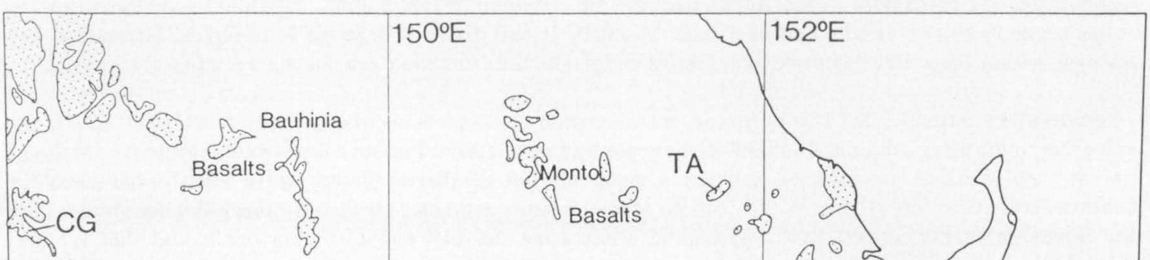

Fig. 1 Distribution of Cainozoic volcanic fields (stippled) and main xenolith localities (letters) described in the text in relation to the limits of the Eromanga-Brisbane transect. PC (Palen Creek - Richmond Gap), MW (Mt. Wyangapinni), GM (Gowrie Mountain), CQ (Cliffdale Quarry), LO (Lowood), DU (Durong Road), BA (Ballogie), BR (Brigooda), CG (Carnarvon Gorge), TA (Tararan).

disequilibrium in the assemblage due to metasomatic or other processes, but the relatively higher Cr/Cr + Al ratios in the Palen lherzolite mineralogy suggests a less depleted example of the mantle array (Arai, 1987). The Palen megacryst orthopyroxene is Fe-enriched compared to that of the lherzolite. It is probably a separate high temperature - high pressure phase, judging from its greater Ca content and similarity to near-liquidus orthopyroxene precipitated from transitional alkali basalt compositions at 1130-1230°C and 13-18 kbar (Green & Hibberson, 1970).

Richmond Gap

The Richmond Gap megacrysts are Fe and Ti-enriched over the Palen pyroxenes and most pyroxenes of southern Queensland websteritic xenoliths, if they are disaggregated material (cf. Ewart & Grenfell, 1985; and Appendix 2, Table 2). Providing that they represent co-existing phases of the host magma, the Richmond Gap pyroxenes would suggest a more advanced stage of liquidus or near-liquidus crystallization than does the Palen megacryst orthopyroxene.

MAIN RANGES SUITES

Out of twenty-two sites (Ewart & Grenfell, 1985; and this paper), twenty contain abundant spinel lherzolite up to 20 cm across and fifteen of these carry additional inclusions. Six suites contain lower crustal granulites, three show inclusions with common hydrous phases, and three include garnet in some assemblages. The more complex suites are confined to Gowrie Mountain hawaiite neck, Spring Bluff hawaiite flow, Mt. Wyangapinni nepheline hawaiite neck, Middle Branch Creek leucite basanite, and the Mt. Mitchell nepheline benmoreite.

Brief summaries of the salient features and context of these suites are given before discussing the thermal regimes represented within them.

Gowrie Suite

Over 95% of this suite consists of medium to coarse-grained spinel lherzolites. A feature is the occurrence of composite xenoliths up to 9 cm across showing:

(1) black pyroxenites which invade and enclose spinel lherzolite and contain xenocrysts of centimetre-size, green clinopyroxene with exsolution lamellae.

(2) medium-grained lherzolite in planar contact with fine grained green clinopyroxenites with <1% olivine.

(3) dunite enclosing narrow bands which contain up to 30% green clinopyroxene.

In addition to previously described xenoliths (Grenfell, 1984; Ewart & Grenfell, 1985), the authors also collected olivine websterite with poikilitic cumulate texture (up to 75% cpx, 25% ol, 15% opx), spinel metaclinopyroxenites (green clinopyroxene with exsolution lamellae), and crustal rocks such as amphibolite and aplite.

The Gowrie suite was erupted during the youngest Main Range event (18 Ma) and indicates a heterogeneous mantle-lower crust section, with only minor development of hydrous phases.

Cliffdale Quarry Suite

Coarse-grained spinel lherzolites up to 30 cm across are abundant here, within the columnar flow forming Spring Bluff. Some lherzolites show composite contacts with an olivine-poor variant or with veins of medium- to coarse-grained orthopyroxene-rich spinel metawebsterite. Other xenoliths included olivine pyroxenites which approach wehrlites, indistinctly banded two pyroxene granulites, and sandstone-siltstone partly fused to buchites.

The basalts within this part of the sequence are dated around 21.5 Ma (Ewart & Grenfell, 1985). Thus, the Cliffdale suite represents a slightly older, less complex mantle-crust sample than appears at Gowrie Mountain.

Mount Wyangapinni Suite

A range of inclusions forms up to 40% of the rock in this prominent neck near Pittsworth (Van der Zee & Macnish, 1979). Spinel lherzolite and clinopyroxene megacrysts predominate (Ewart & Grenfell, 1985), but megacrysts are more diverse and include greenish and brownish clinopyroxenes, garnet, ilmenite (± magnetite), anorthoclase, and zircon.

An anorthoclase megacryst was dated at 20.5 Ma (Appendix 2, Table 4) suggesting that Mount Wyangapinni belongs to the typical period of Main Range inclusion-bearing eruption (18.1-21.5 Ma; Ewart & Grenfell, 1985). The megacryst suite (analyses, Appendix 2, Table 3a) probably represents crustal crystallization, as pyroxene and garnet compositions are distinct to those in pyroxenites of mantle associations. A similar clinopyroxene and garnet-bearing suite occurs east of the Main Range in a dyke at Tarampa, near Lowood (authors' observations).

Middle Branch Creek Suite

This is characterised by the appearance of Ti-enriched micas, sodic pyroxenes and amphiboles amongst its members (Grenfell & Ewart, 1985). Abundant spinel lherzolites and coarse mantle clinopyroxenites occur, in conjunction with clinopyroxene-rich pyroxenites showing olivine-spinel reactions and sodic gabbros (Grenfell, 1984). This range of rock types suggests sampling took place across mantle-crustal levels. The coarse clinopyroxenites contain a green augite in which

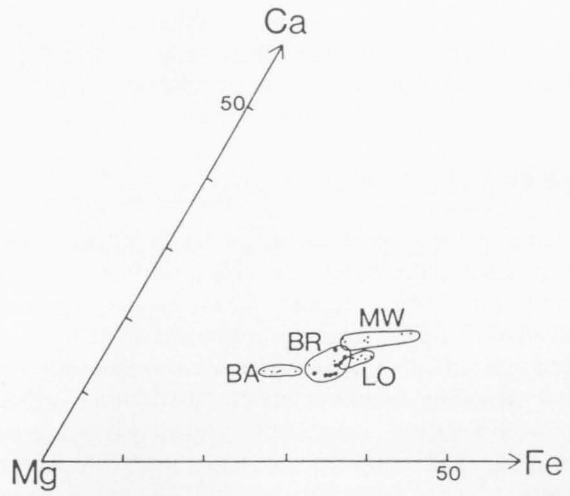

Fig. 2 Ca-Mg-Fe diagram showing fields of garnet compositions of S. Queensland megacrysts (smaller dots) and garnet from pyroxenite xenoliths (larger dots). BA (Ballogie), BR (Brigooda, megacrysts and xenoliths), MW (Mt. Wyangapinni), LO (Lowood).

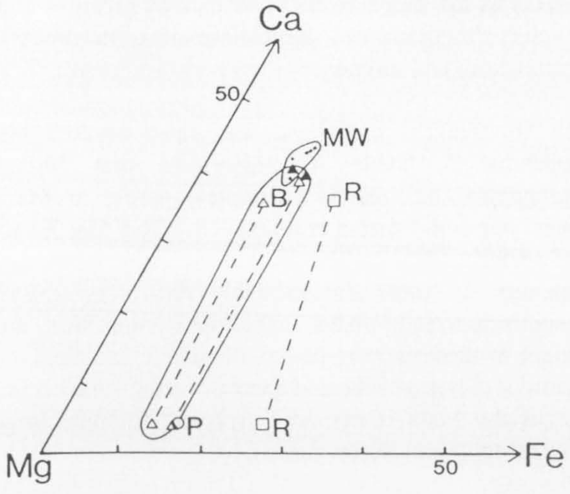

Fig. 3 Ca-Mg-Fe diagram showing fields of pyroxene compositions of S. Queensland megacrysts, B (Boyne Basalts; with Ballogie plots as open triangles, Brigooda plots filled triangles), MW (Mt Wyangapinni, dots), P (Palen Creek, open circle), R (Richmond Gap, open squares). Possible co-existing pyroxene megacryst pairs are joined by dashed lines.

orthopyroxene exsolution lamellae unmixed at temperatures around 1900-1210°C before re-equilibration.

The presence of hydrous alkali-rich minerals in the suite may be linked to the unusual composition of the host leucite basanite. Ti-magnetite (± apatite) and Mg-ilmenite inclusions are also confined to this host in the Main Range sequences and were related to fractional crystallization from the magma at different depths by Grenfell (1984).

Mount Mitchell Suite

This is another amphibole and mica-bearing suite, but one which illustrates both end-member fractionation and metasomatism in the mantle (Green & others, 1974; Grenfell, 1984; Ewart & Grenfell, 1985). Common spinel lherzolite is associated with 'pargasitized' olivine spinel clinopyroxenite, 'kaersutitized' clinopyroxenites (± spinel), hypersthene leucogabbro with reaction coronas, and hybridized garnet-bearing assemblages. Megacrysts include kaersutite, phlogopite, ilmenite (with Ti-magnetite reaction rims), and anorthoclase.

The suite incorporates products of crystallization and infiltration of highly fractionated 'nepheline benmoreite' magmas and fluids around the mantle-crust boundary (Grenfell, 1984). The origin of these evolved liquids through crystallization of hydrous phases such as kaersutitic amphibole (Green & others, 1974), or via melting of relatively Fe-rich mantle (Wilkinson, 1977) has been argued. Clearly, the Mount Mitchell xenoliths show some metasomatism prior to incorporation in the host nepheline benmoreite, but magmas generated from amphibole-enriched sources would also favour amphibole crystallization in their fractionation. Other suites bearing on this problem are found on the outskirts of the Main Range sequences (Owens Knob and Seven Mile Creek).

Owens Knob and Seven Mile Creek Suites

These both include hydrous phases, but contrast in their associations. Owens Knob contains mantle lherzolites and rare clinopyroxenites in relatively unfractionated olivine nephelinite. However, abundant anorthoclase and kaersutite megacrysts indicate disruption of previous crystallizations of melts and hydrous fluids.

The Seven Mile Creek camptonite in contrast lacks mantle inclusions, but carries abundant gabbros in which clinopyroxenes show similar Na and Al contents to those in granulites (see Grenfell, 1984). Abundant Ti amphibole also occurs in amphibole pyroxenties, as composite clots, phenocrysts and reaction rims on pyroxenes, and as groundmass grains which have the highest Ti contents. Thus, amphibole crystallization from this magma took place at crustal levels during its fractionation and eruption.

BOYNE BASALT SUITES

The Boyne Basalt field shows a much higher proportion of inclusion-bearing members than found in the previously described southern Queensland fields. Suites belonging to four main eruptive centres are recognized, the simplest being the Durong Road suite.

Durong Road Suite

Spinel lherzolites are common in the margins of a plug-like body of nepheline hawaiite here and are accompanied by Ti augite, spinel and anorthoclase megacrysts. The latter become more numerous towards the centre of the body, which grades into nepheline mugearite.

Ballogie Suite

Spinel lherzolite, and rare metapyroxenite and granulite, are found in a diatreme here, but are commonly altered. However, megacrysts up to 10cm are abundant. A statistical count from the breccia gave 78% clino- and orthopyroxenes, 8% garnet, 6% spinels, 5% kaersutite, 2% magnetite, and 1% ilmenite (Hollis & others, 1983). The suite preserved in a nepheline hawaiite dyke, in comparison, gave 65% spinel lherzolite, 21% pyroxenes, 3% anorthoclase, 1% kaersutite, and the remainder country rocks.

The breccia and inclusions were emplaced about 16 Ma ago, based on K-Ar dating of an anorthoclase megacryst (Sutherland, 1990). The suite probably represents disaggregated pyroxenite and garnet metapyroxenite bodies in spinel lherzolite mantle, associated with kaersutite and anorthoclase veining. The pressure-temperature conditions for such garnet metapyroxenites were estimated around 1000-1100°C and 14-15 kbar (Hollis & others, 1983).

North Brigooda Suite

This is similar to the Ballogie suite, but megacrysts are larger. They mostly range up to 14 cm across, but reach up to 30 cm for garnet. The inclusions occur in a marginal breccia of a neck intruding the Palaeozoic granitic basement of the area (Robertson & others, 1985). Altered spinel lherzolites are found with rare garnet metaclinopyroxenite (± orthopyroxene), notable for showing open cavities. Pieces of 2px-granulite and quenched kaersutite-basanite are found with common pieces of the country granite and basalt. Several K-rich alkaline lavas from the centre carry spinel lherzolite xenoliths and different assemblages of megacryst minerals. Zircon megacrysts from the breccia gave a fission track age around 16 Ma (author's data).

South Brigooda Suite

Related inclusions are found in breccias and lapilli tuffs within a crater-like structure to the south of the

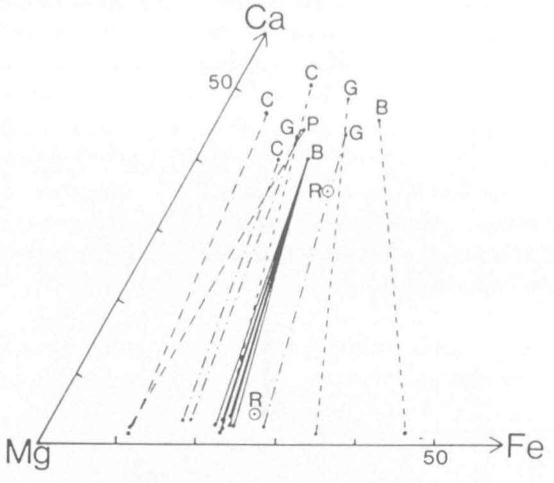

Fig. 4 Ca-Mg-Fe diagram showing plots of co-existing pyroxenes from some S. Queensland xenolith localities. C (Cliffdale Quarry), G (Gowrie Mountain Quarry), B (Brigooda Breccias), P (Palen Creek), R (Richmond Gap megacryst pair). Spinel lherzolites (long dashed tie lines), garnet pyroxenites (solid tie lines), websterites ± spinel, olivine (dash-dot tie lines), granulites (short dashed tie lines) and megacryst pair (circled dots).

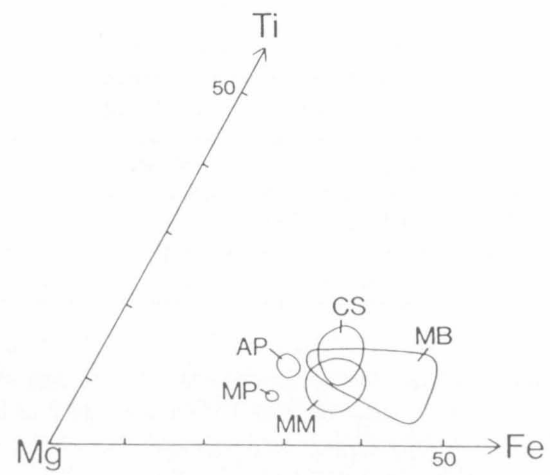

Fig. 5 Ti-Mg-Fe diagram showing plots of amphiboles from some S. Queensland xenolith-megacryst localities. Analyses taken from Hollis & others, 1983, Ewart & Grenfell, 1985, and this work. MP (metasomatized pyroxenite, Mt. Mitchell), AP (amphibole pyroxenites, Brigooda and Gowrie Mt.), MM (megacrysts, Main Range), CS (camptonite suite, Main Range), MB (megacrysts, Boyne Basalts). The metasomatized and amphibole pyroxenites are mantle suites and the camptonite suite is probably crustal.

Brigooda field (Robertson & others, 1985). Locally abundant blocks of a spinel lherzolite and kaersutite megacryst-rich nepheline mugearite occur here. Significantly, this rock also has rare garnet metapyroxenites. Larger examples are found loose in nearby soils. These range up to 10 cm across and are mostly garnet metaclinopyroxenites (± kaersutite, spinel, ilmenite, magnetite), but about one in ten recoveries are opx-bearing types. Unlike the garnet metapyroxenite from north Brigooda, cavities in the southern examples contain barium-bearing zeolites (F. L. Sutherland & K. Kinealy, unpublished data). This suggests late introductions of Ba-enriched alkaline silicate fluids from the host magmas.

Age dates on megacrysts from here outline a complex series of eruptive events. Kaersutite, separated from the

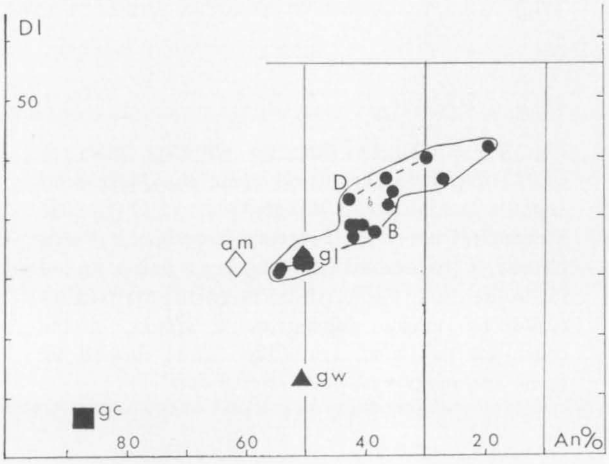

Fig. 6 Differentiation Index versus normative An% for Boyne Basalt xenoliths and volcanic rocks. gc (garnet clinopyroxenite), gw (garnet-opx clinopyroxenite), gl (granulite), am (amphibole basalt). Boyne Basalts are represented by solid circles and in addition to Brigooda rocks, include Ballogie (B) and Durong Road (D) basalts. Differentiated series are joined by dashed lines. The xenoliths are represented by different symbols to the solid circles for basalts.

blocks of nepheline mugearite gave a K-Ar age of 18 Ma, zircon megacrysts gave fission track ages of 8-12 Ma, and loose anorthoclase megacrysts gave K-Ar ages of 0.3-0.5 Ma (Robertson & others, 1985; F. L. Sutherland & A. D. Robertson, unpublished data).

Taken together, the Brigooda suites indicate an underlying spinel lherzolite mantle, interlayered with metapyroxenites and garnet metapyroxenites. The pyroxenites in particular were partly replaced by hydrous phases, such as amphibole (mineral analyses, Appendix 2, Tables 5 and 3b). This replacement started prior to 18 Ma ago.

DISCUSSION

Southern Queensland xenolith suites indicate a lherzolite-dominated mantle and granulitic lower crust, interspersed with a heterogeneous variety of metapyroxenite and igenous bodies with local hydrous and garnet-bearing developments. The depth of some of these rocks can be assessed by estimating temperature and pressure ranges of their formation. A critique of the methods used for these estimates, and the errors and uncertainties involved, are discussed in Appendix 1. Results using some of the methods are listed in Appendix 2, Table 6, and form a basis for subsequent discussion.

Mantle Lithologies

The lherzolite-metapyroxenite association characteristically carries Cr-enriched and Ti-impoverished pyroxenes and spinels. Other pyroxenites, and little modified cumulate igneous rocks, such as wehrlites, contain Cr-poor, Ti- rich compositions of these minerals. The first group is usually named the Group I (Cr-diopside) association, in contrast to the Group II (Ti-Al augute) association (Wilshire & Shervais, 1975). This grouping is not always clear-cut in Australian suites (Sutherland & Hollis, 1982; Ewart & Grenfell, 1985; Hollis, 1985; Morris, 1986), nor does it accommodate metasomatized Group I rocks (O'Reilly & Griffin, 1987). Three or more groups can be discerned on Cr and Ti contents of minerals in Southern Queensland suites (Appendix 2, Table 7).

Bulk analyses of spinel lherzolites are unavailable here, but presumably would match those from Carnarvon Gorge 150 km north of BMR line 14 of the Eromanga Transect (Example 1, Appendix 2, Table 8; from Griffin & others, 1987). Temperature estimates for lherzolites and associated metapyroxenites, based on different methods range from 900 to 1200°C. The maximum temperature range from any one section is up to 240°C

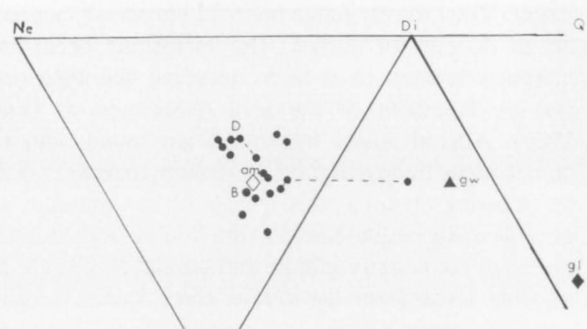

Fig. 7 Normative Ne-Ol-Di-Hy-Q plots of Boyne Basalt xenoliths and volcanic rocks. Symbols as in Figure 6.

within this interval. The highest temperature estimates for metapyroxenites within this interval include the low Cr, low Ti websterites at Gowrie Mountain, and the coarse clinopyroxenites with exsolution at Middle Branch Creek.

Much of the south Queensland mantle is probably comparable with that exposed in the Balmuccia peridotite in Italy (Sinigoi & others, 1983; Voshage & others, 1988). The lherzolite continuum (grading into harzburgite and subordinate dunite) is a residue of partial melting of originally undepleted mantle. Older dykes have been interpreted as fractionates of partial melts of the residue. They grade from Cr, Ti-poor assemblages into transitional assemblages with greater Cr and Ti. Younger intrusions represent distinctly Cr, Ti-enriched melts, believed to have been triggered by introductions of metasomatic fluids and/or liquids. Two types of metasomatism are thought to have operated in the Balmuccia peridotite, viz. mantle-derived volatiles which formed amphibole and fluids from crustal sources which formed phlogopite. Amphibole metasomatism is the prevailing type in southern Queensland, judging from xenoliths in which hydrous replacements are observed.

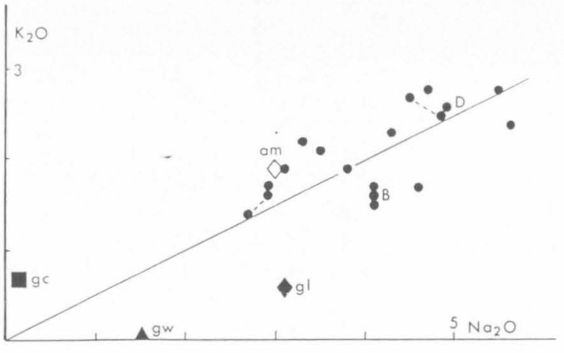

Fig. 8 K_2O versus Na_2O diagram, Boyne Basalt xenoliths and volcanic rocks. Plots below the dividing line are sodic types ($K_2O/Na_2O + K_2O$ <0.5) and plots above are K-rich types ($K_2O/Na_2O + K_2O$ >0.5). Symbols as for Fig. 6.

Different mantle is a feature in the Brigooda-Ballogie area, where garnet metapyroxenites appear. The more dominant garnet clinopyroxenites are distinct in bulk composition from the rarer orthopyroxene-bearing types, but both are hypersthene-normative rocks (Appendix 2, Table 8, Figs. 6-8). The former are more subject to amphibole replacement, but garnet compositions are similar in both rocks and also match the compositional range found in megacrysts (Fig. 2). However, pyroxenes in the rocks are more Fe-enriched than the range found in megacrysts (Figs. 3 and 4). This suggests a mantle veined by additional igneous pyroxenites. In this respect, the Ballogie pyroxene megacrysts form two distinct sets of clinopyroxenes and two distinct sets of orthopyroxenes. This provides two possible pairings for temperature estimates, if co-existing pairs are assumed. One pairing gives crystallization temperatures around 1150°C, the other around 1250°C (Hollis & others, 1983). The first temperature is similar to estimates for Brigooda garnet metapyroxenites (Appendix 2, Table 5), but the second is significantly higher, supporting the presence of more Mg-rich igneous pyroxenite veins or cumulates.

Crustal Lithologies

Granulites amongst Main Range suites are dominantly two pyroxene types and suggest a mafic lower crust containing calcic plagioclase as the common feldspar (Grenfell, 1984). Bulk chemistry is unavailable, but they resemble granulites from Carnarvon Gorge (Griffin & others, 1987, and Appendix 2, Table 8). Rare granulites with sodic plagioclase (± apatite) occur at Gowrie Mountain (Appendix 2, Table 2b) and north Brigooda (Griffin & others, 1987 and Appendix 2, Tables 5 and 8). The latter example also shows more Fe-enriched co-existing pyroxenes (Fig. 4). The sodic feldspar types are probably interspersed through the lower crust, as temperature estimates for them range across those estimated for the dominant calcic types (Appendix 2, Table 6). The Brigooda granulite is a quartz normative rock quite unrelated to the typical nepheline normative K-rich compositions of the Brigooda basalt field (Appendix 2, Table 8 and Figs. 7 and 8).

Some isotopic data is available for the Carnarvon and Brigooda granulites from the work of O'Reilly and others (1988), viz:

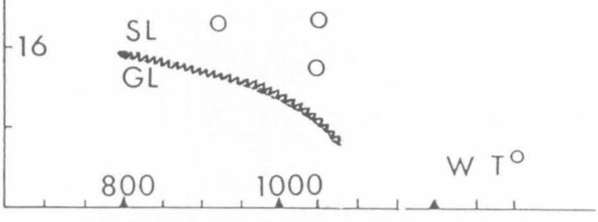

Fig. 9 Pressure-temperature plots of Brigooda gnt-2px xenoliths using Wells (1977) T and Nickel & Green (1985) P methods, in relation to the S.E. Australian Bullenmerri/Gnotuk field (BG, based on Wells Fe^{2+}T calculations appropriate to this field). The Brigooda results are shown for both Wells Fe^{2+}T (solid circles) and Wells Fe^{3+}T (open circles, appropriate for these rocks). SL and GL represent the spinel and garnet lherzolite transition zone (saw toothed line).

- a general coherence in Nd and Sm isotopes with the east Australian mantle array, but passing into 'enriched mantle' values.

- a positive correlation of 87Sr/86Sr and negative correlation of epsilon Nd with magnesium number (mg) related to an unusual mixing process between mafic magmas and crustal material.

- an ill-defined age of 450-500 Ma suggested by Nd isotopes for the Carnarvon granulite.

Some cumulate and other igneous textured rocks give temperature estimates within the low T end of the metamorphic mantle range, e.g. Cliffdale olivine clinopyroxenite and Gowrie Mountain wehrlite (Appendix 2, Table 6). These probably crystallized at lower crustal depths, as their pryoxenes show Na and Al contents comparable to those in granulites, but distinctly higher than is observed in crustal gabbros (cf. Appendix 2, Table 2; and Grenfell, 1984).

Hydrous Lithosphere

Amphibole (± mica) constitutes a significant phase in some suites, particularly within pyroxenites and as megacrysts (Mount Mitchell, Middle Branch Creek, Owens Knob, Seven Mile Creek, Ballogie-Brigooda). It is sometimes difficult to assign an igneous or metasomatic origin to these hydrous phases, due to limited evidence.

Pargasitic amphibole occurs in some rocks, but most amphiboles are Ti-enriched and kaersutitic (Ewart & Grenfell, 1985; and Fig. 5). Kaersutite in Brigooda garnet metapyroxenites is more magnesian than in Brigooda-Ballogie megacrysts (Fig. 5). This suggests that the megacrysts are derived from vein and/or cavity

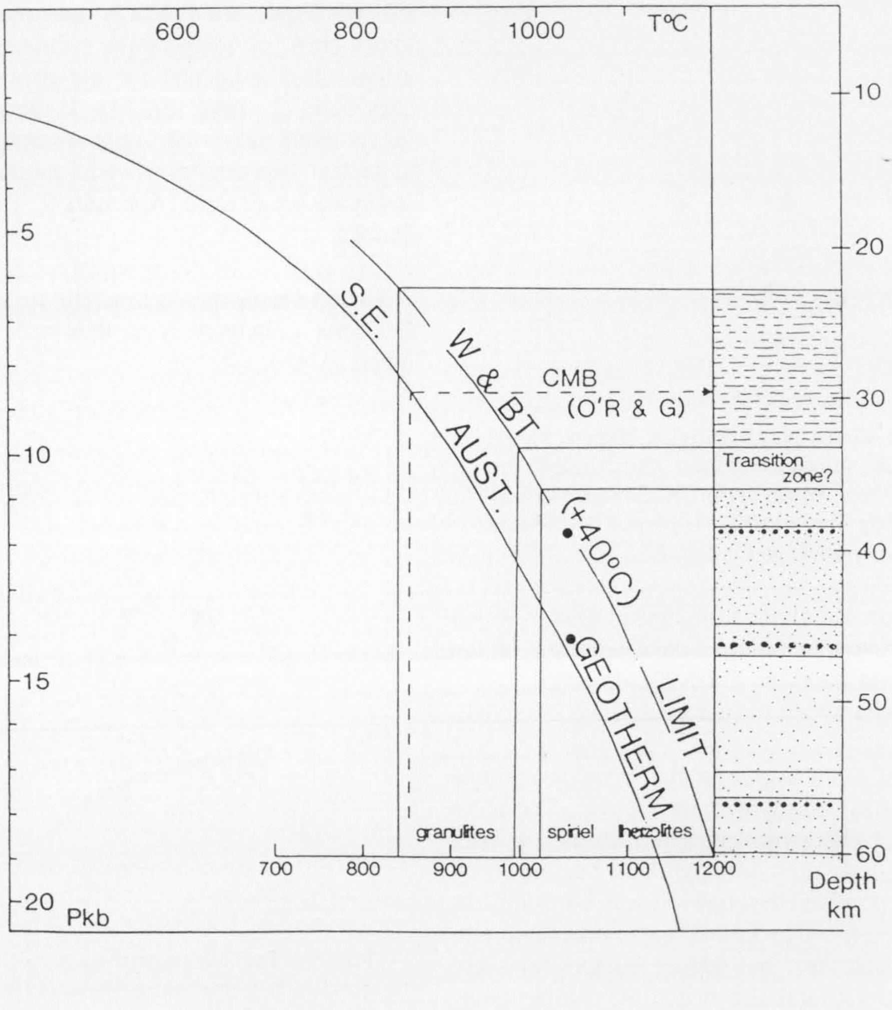

Fig. 10 Pressure-temperature diagram for correlating crust-mantle xenolith equilibration temperatures with the lithospheric section below southern Queensland. The southeastern Australian geotherm curve is based on O'Reilly & Griffin (1985) and the Wood & Banno (1973) T(+40°C) limit curve is taken through the maximum gnt-2px plot from Griffin & others (1987). The Wood & Banno T(+40)°C ranges for spinel lherzolites and granulites are projected (straight lines) on to the Wood & Banno T(+40)°C limit to obtain their respective lithospheric depths. Wood & Banno T(+40)°C levels for metapyroxenites (solid lines) and garnet metapyroxenites (dotted lines) are also shown in the section. Solid large dots are plots for Brigooda gnt pyroxenites using Wood & Banno T(+40)°C and Wood P; from Griffin & others, 1987).

crystallizations from basaltic melts, and are not necessarily the same as material replacing the pyroxenites. Rare finds of quenched 'basanitic' xenoliths with kaersutite micro-phenocrysts of similar composition to megacrysts, support such a melt origin.

The times of amphibole introductions into the lithosphere are imprecisely known. Some, such as the metasomatism and crystallization recorded at Mount Mitchell, forms part of the late stages of Main Range volcanism, around 20 Ma ago. Amphibole formation in the Boyne lithosphere, was present 18 Ma ago prior to the subsequent local volcanism. Ti-rich amphibole is also a component in inclusion suites from the crust-mantle section below the late Tertiary (3 Ma) basaltic field at Tararan (Robertson, 1985), farther north of the Eromanga-Brisbane transect (Fig. 1). These developments could resemble that described for amphibole-bearing garnet (± spinel) pyroxenites from the McBride province, north Queensland (Stolz, 1987). There, amphibole replacements of more pargasitic composition became more Fe and Ti-rich to form hornblendites composed of kaersutite similar in composition to megacrysts. This was ascribed to an influx of H_2O and CO_2-bearing fluids during the volcanic process, particularly within mantle levels.

SYNTHESIS

The two pyroxene xenoliths provide temperature estimates of crystallization or re-equilibration (Appendix 2, Table 6). However, only the Brigooda garnet-two pyroxene rocks allow pressure estimates for establishing more controlled depths (Appendix 1). The method used by Griffin & others (1984) and O'Reilly & Griffin (1985) for establishing the southeastern Australian geotherm gave erratic results for the Brigooda samples, but an indirect method gave 1025-1030°C and 11.9-14.2 kbar (Griffin & others, 1987). This is slightly lower than estimates determined with the method favoured by Carswell & Gibb (1987), which for two of the Brigooda xenoliths gave consistent results of 1043-1044°C and 15.2-16.5 kbar (Fig. 9).

The Brigooda results tend to lie on the high temperature side of the southeastern Australian results (Figs. 9, 10). Thus, the southeastern Australian geotherm can be taken as a maximum depth curve for projection of temperature estimates for re-equilibrated xenolith assemblages that lack garnet.

Projecting Wells (1977) or Wood & Banno (1973) temperature estimates (Appendix 2, Table 6) for south Queensland granulites and lherzolites onto the southeastern Australian geotherm gives a granulite section exceeding 28-40 km and a spinel lherzolite zone exceeding 43 km in depth (see Fig. 10). These values seem too deep in relation to the lower crust-mantle transition interpreted in the Eromanga seismic profile near the Main Range sequence. Alternatively, a maximum geothermal profile and projected intersections based on Wood & Banno's +40°C temperature estimates (after Griffin & others, 1987) gives a granulitic lower crust between 22-34 km and a spinel lherzolite zone between 36-60 km in depth (Appendix 2, Table 6; and Fig. 10).

Thus, the most consistent match between xenolith temperature estimates and the interpreted seismic crust-mantle section is achieved using Wood & Banno's T (+40°C) estimates. This lithospheric section is illustrated in Figure 10, with positions indicated for metapyroxenite horizons.

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APPENDIX 1

METHODS OF TEMPERATURE AND PRESSURE ESTIMATION

Commonly used methods for estimating temperatures of crystallization or re-equilibration for assemblages containing co-existing pyroxenes are those of Wells (1977) and Wood & Banno (1973). If the rocks are metamorphic, then the temperatures give some guide to depth of origin as temperatures generally increase down through the re-equilibrated lithosphere. Temperature estimates for igneous rocks only relate to the crystallizing melt and not its depth. These empirical thermometers are fraught with vagaries and often the Wells method gives lower results. Because of the errors inherent in the methods, values between the methods can differ by up to 120°C and results are often rounded off to the nearest 5 or 10°C. The Wells thermometer is considered more reliable at temperatures above 1000°C when these two methods, and others, such as the garnet-clinopyroxene thermometer of Ellis & Green (1979), are compared (Carswell & Gibbs, 1987).

The co-existence of garnet and two pyroxenes allows pressure estimation, utilizing various geothermometers and barometers. The combination of Wells' (1977) T and Nickel & Green's (1985) P was favoured by Carswell & Gibbs (1987). Many of the earlier Australian PT estimates used to establish geotherms were based on Ellis & Green's (1979) T and Wood's (1974) P (see Sutherland & others, 1984; Griffin & others, 1984; O'Reilly & Griffin, 1985; Griffin & others, 1987). The assignment of Fe as Fe^{2+} or Fe^{3+} in pyroxenes can cause significant variations in temperatures obtained (Griffin & others, 1984; O'Reilly & Griffin, 1985). In equating co-existing pyroxene results from garnet-free granulites with those from garnet granulites, Griffin & others (1987) used Wood & Banno T +40°C to give a satisfactory match for the groups.

Equilibration temperatures for metamorphic xenoliths were used to estimate depths of origin by projecting results on to the southeast Australian geotherm (Griffin & others, 1987; O'Reilly & others, 1989). This was used to interpret the depth of the crust-mantle boundary and the extent of transition zones. Some of the problems involved in this were raised by Sutherland & others (1989) in discussing the Tasmanian lithosphere. One is use of a standard Australian geotherm based on a single mantle section. Thus, projections of granulite temperatures in Tasmania were made onto a hotter Tasmanian geotherm established on both spinel and garnet lherzolite sections. This reduced depth estimates and gave crust-mantle depths consistent with known geophysical data.

Spinel lherzolite thermometry is another problem as this assemblage is not very responsive to accurate T estimates. A refined method, using a correction for spinel compositions (Sachtleben & Seck, 1981) has been used for T estimation and hence projection for P determination. However, the reliability of the method in some cases has been severely questioned by Cundari & others (1986), based on study of Australian lherzolites. For this reason, T estimates from this method were not applied in this study.

APPENDIX 2

ANALYSES OF XENOLITHS FROM SOUTHERN QUEENSLAND

TABLE 1

Summary of inclusion suites from South Queensland, discussed in paper.

Suite	Peridotites	Pyroxenites	Granulites	Crustal Rocks	Megacrysts
ALBERT BASALT					
Palen Creek	spl lherzolite^{A}				opx^{A}
Richmond Range				gabbro^{R}	opx^{C}, cpx^{C}
MAIN RANGE BASALTS					
Gowrie Mt.	spl lherzolite^{A}	cpxenite(±spl)^{A}	2px granulite^{S}	gabbro^{R}	cpx^{C}, opx^{C}
(Quarry)	wehrlite^{S}	websterite(±spl)^{C}		aplite^{R}	
	spl lherzolite^{R}	ol-websterite^{R}		amphibo-	
	(±amph.)	amph-pxenite^{R}		lite^{R}	
Cliffdale	spl lherzolite^{A}	spl websterite^{C}	2px granulite^{S}	sandstone-	opx^{C}
(Quarry)		ol-cpxenite^{R}		siltstone^{C}	
Mt. Wyangapinni	spl lherzolite^{C}	pxenite^{R}	granulite^{R}	sandstone^{C}	anorth^{R}, amph^{R}, zrc^{R}, cpx^{A}, gnt^{S}, il(±mgt)^{S},
Mt. Mitchell	spl lherzolite^{A}	cpxenite^{C}		gabbro^{S}	anorth^{C}, amph^{A}, mic^{S},
		(±ol, amph, spl)		meladiorie^{S}	il(+±mgt)^{S}.
Middle Branch	spl lherzolite^{A}	cpxenite^{C}		hy-gabbro^{S}	mgt(±apt)^{C}, il^{R}
Creek		(±opx, exsol.)		gabbro &	
		amph-ol-spl		gnt hybrids^{S}	
		cpxenite^{C}, amph-cpxenite(±spl)^{S}			
Owens Knob	spl lherzolite^{C}	cpxenite^{R}			amph^{A}, anorth^{C}
Seven Mile Ck		amph-pxenite^{S}		gabbro^{A}	amph^{A}
(Camptonite sill)		(±mica, Fe-Ti oxides)			
BOYNE BASALTS					
Durong Rd.	spl lherzolite^{C}				cpx^{C}, spl^{C}, anorth^{S}.
Ballogie	spl lherzolite^{C}	gnt pxenite^{R}	granulite^{R}	granite, etc^{A}	anorth^{S}, amph^{S}, cpx^{C}, opx^{S}, spl^{C},
					gnt^{S}, mgt^{R}, il^{S}.
Brigooda N.	spl lherzolite^{C}	gnt pxenite^{R}	2px granulite^{R}	granite, etc^{A}	cpx^{C}, spl^{S}, anorth^{S},
	websterite^{R}	(±opx, spl)		basalt, etc^{A}	il^{S}, mgt^{R}, zrc^{R}.
Brigooda S.	spl lherzolite^{C}	gnt cpxenite^{S} (±amph, spl, ilm)		granite, etc^{A}	cpx^{C}, spl^{S}, anorth^{S},
		gnt cpxenite^{R} (±opx, spl)		basalt, etc^{C}	opx^{R}, il^{R}, mgt^{R}, zrc^{R}.

^{A}=abundant, ^{C}=common, ^{S}=scarce, ^{R}=rare; pxenite=pyroxenite, cpxenite=clinopyroxenite, 2px=2 pyroxene, ol=olivine, cpx=clinopyroxene, opx=orthopyroxene, gnt=garnet, amph=amphibole, mic=mica, mgt=magnetite, il=ilmenite, anorth=anorthoclase, zrc=zircon, apt=apatite.

TABLE 2

Representative normalised microprobe analyses and cation contents of phases in xenoliths (a) Mt. Albert and (b) Main Range basalts.

(a) **PALEN CREEK** (from Ross, 1977) (b) **GOWRIE MOUNTAIN, TOOWOOMBA**

Spinel lherzolite **Ol-bearing spinel websterite**

	cpx	opx	spl	ol	cpx	opx	sp
SiO_2	51.19	55.35	0.28	40.80	47.08	50.36	0.04
TiO_2	0.81	0.06	nd	nd	2.14	0.41	1.01
Al_2O_3	3.44	4.34	58.39	nd	9.51	6.50	55.61
Cr_2O_3	1.02	0.32	9.40	nd	0.06	0.00	0.66
FeO	7.14	6.75	10.75	10.80	9.20	16.91	28.69
MnO	nd	0.08	nd	0.01	0.14	0.43	0.13
MgO	15.10	31.98	21.12	48.39	12.11	24.00	13.76
CaO	21.15	1.13	0.07	0.02	18.25	1.27	0.04
Na_2O	0.15	nd	0.00	nd	1.48	0.10	0.00
K_2O	nd	nd	nd	nd	0.05	0.03	0.05
Si^{4+}	1.894	1.915	0.058	1.003	1.753	1.837	0.009
Ti^{4+}	0.023	0.002	0.000	0.000	0.060	0.011	0.167
Al^{3+}	0.150	0.177	14.164	0.000	0.417	0.279	14.426
Cr^{3+}	0.030	0.009	1.530	0.000	0.002	0.000	0.116
Fe^{2+}	0.221	0.195	1.850	0.222	0.286	0.515	5.280
Mn^{2+}	0.001	0.002	0.000	0.002	0.004	0.013	0.025
Mg^{2+}	0.833	1.649	6.479	1.772	0.672	1.304	4.513
Ca^{2+}	0.838	0.043	0.015	0.001	0.728	0.049	0.009
Na^{1+}	0.011	0.000	0.000	0.000	0.107	0.007	0.000
K^{1+}	0.000	0.000	0.000	0.000	0.002	0.001	0.015
TOTAL	4.000	3.992	24.096	3.000	4.031	4.016	24.560
Ti							1.7
Cr			15.5				
Fe	11.7	10.3	18.8	11.1	17.0	27.6	53.0
Mg	44.0	87.4	65.7	88.8	39.8	69.8	45.3
Ca	44.3	2.3		0.1	43.2	2.6	
Wells (1977)							
(Fe^{3+})T°C		1140				990	
Wood & Banno (1973)							
(Fe^{3+})T°C		1150				980	

TABLE 2 (CONTINUED)

(b) GOWRIE MOUNTAIN, TOOWOOMBA (continued) - Analyst D. F. Hendry.

Olivine websterite | Two-pyroxene granulite

	cpx	opx	ol	host cpx	exsol opx	cpx	opx	plag
SiO_2	50.87	53.11	39.32	49.71	53.69	49.20	50.91	56.76
TiO_2	0.79	0.31	0.00	1.14	0.32	0.76	0.14	0.02
Al_2O_3	6.15	5.17	0.16	6.91	4.94	6.40	3.79	26.58
Cr_2O_3	0.31	0.17	0.03	0.33	0.15	0.20	0.02	0.00
FeO	6.65	10.98	17.27	5.89	11.07	8.53	21.57	0.00
MnO	0.19	0.23	0.15	0.19	0.09	0.13	0.44	0.00
MgO	16.74	28.10	42.95	14.58	28.45	11.82	22.54	0.00
CaO	17.06	1.80	0.13	19.74	1.13	21.82	0.57	9.60
Na_2O	1.18	0.15	0.00	1.46	0.14	1.17	0.05	6.37
K_2O	0.06	0.00	0.01	0.04	0.03	0.00	0.00	0.47
Si^{4+}	1.858	1.882	0.998	1.828	1.897	1.840	1.893	10.258
Ti^{4+}	0.022	0.008	0.000	0.031	0.008	0.021	0.004	0.003
Al^{3+}	0.265	0.216	0.005	0.300	0.205	0.282	0.166	5.642
Cr^{3+}	0.009	0.005	0.001	0.010	0.004	0.006	0.001	0.000
Fe^{2+}	0.203	0.325	0.366	0.181	0.327	0.267	0.671	0.000
Mn^{2+}	0.006	0.007	0.003	0.006	0.003	0.004	0.014	0.000
Mg^{2+}	0.911	1.484	1.624	0.799	1.497	0.659	1.248	0.000
Ca^{2+}	0.667	0.068	0.003	0.777	0.043	0.874	0.022	1.850
Na^{1+}	0.083	0.010	0.000	0.104	0.009	0.085	0.004	2.222
K^{1+}	0.003	0.000	0.000	0.002	0.001	0.000	0.000	0.107
TOTAL	4.027	4.005	3.000	4.042	3.994	4.038	4.023	20.082
Ti								
Cr								
Fe	11.3	17.3	18.4	10.3	17.5	14.8	34.5	
Mg	51.2	79.1	81.5	45.4	80.2	36.6	64.3	
Ca	37.5	3.6	0.2	44.3	2.3	48.6	1.1	44.2
Na								53.5
K								2.3
Wells								
(Fe^{3+})T°C		1170		990		780		
Wood & Banno								
(Fe^{3+})T°C		1160		1025		800		

TABLE 2 (CONTINUED)

(b) **CLIFFDALE QUARRY, TOOWOOMBA** Analyst D. F. Hendry.

	Websterite		Two-Pyroxene granulite			Spinel lherzolite		
	cpx	opx	cpx	opx	spl	cpx	opx	spl
SiO_2	51.63	53.69	48.89	51.47	0.20	51.40	54.11	0.09
TiO_2	0.55	0.23	1.00	0.24	0.09	0.66	0.19	0.17
Al_2O_3	3.84	4.56	8.48	6.65	61.50	7.00	5.23	59.04
Cr_2O_3	0.57	0.24	0.36	0.16	3.29	0.74	0.17	8.42
FeO	6.60	10.78	5.09	13.89	17.31	3.17	7.09	11.27
MnO	0.03	0.18	0.19	0.33	0.16	0.12	0.16	0.13
MgO	17.04	28.27	12.73	26.68	17.45	14.81	32.12	20.81
CaO	19.26	1.90	22.01	0.51	0.00	20.13	0.77	0.08
Na_2O	0.45	0.15	1.23	0.06	0.00	1.95	0.16	0.00
K_2O	0.02	0.00	0.01	0.02	0.00	0.04	0.00	0.00
Si^{4+}	1.894	1.900	1.801	1.844	0.041	1.866	1.878	0.018
Ti^{4+}	0.015	0.006	0.028	0.006	0.013	0.018	0.005	0.026
Al^{3+}	0.166	0.190	0.368	0.281	15.084	0.299	0.214	14.328
Cr^{3+}	0.017	0.007	0.010	0.004	0.541	0.021	0.005	1.372
Fe^{2+}	0.202	0.319	0.157	0.416	3.012	0.096	0.206	1.939
Mn^{2+}	0.001	0.005	0.006	0.010	0.028	0.004	0.005	0.023
Mg^{2+}	0.931	1.490	0.698	1.424	5.414	0.800	1.661	6.384
Ca^{2+}	0.757	0.072	0.869	0.019	0.000	0.783	0.029	0.017
Na^{1+}	0.032	0.010	0.088	0.004	0.000	0.137	0.011	0.000
K^{1+}	0.001	0.000	0.000	0.001	0.000	0.002	0.000	0.000
TOTAL	4.016	3.990	4.025	4.009	24.113	4.026	4.014	24.107
Ti								
Cr					6.3			14.2
Fe	10.7	17.0	9.1	22.4	34.6	5.7	10.9	20.0
Mg	49.3	79.2	40.5	76.6	59.1	47.6	87.6	65.8
Ca	40.1	3.8	50.4	1.0		46.6		1.5
Wells								
(Fe^{3+})T°C		1140		795		895		
Wood & Banno								
(Fe^{3+})T°C		1140		855		995		

Cationic formulae are based on 4 oxygens for olivine, 6 oxygens for pyroxenes, 32 oxygens for spinels and 32 oxygens for feldspars.

TABLE 3a

Representative Megacryst Analyses, Albert and Main Range Basalts

	PALEN CK.	RICHMOND GAP		MT. WYANGAPINNI				
	opx	cpx	opx	cpx(gn)	cpx(bn)	gnt	anorth	il
SiO_2	54.44	50.21	51.57	50.44	49.29	39.56	66.41	0.00
TiO_2	0.09	1.04	0.49	0.36	0.53	0.71	0.00	53.90
Al_2O_3	4.16	5.26	3.86	5.65	6.92	2.17	19.58	
Cr_2O_3	0.23	0.41	0.08	0.60	0.23	0.07	0.00	
FeO	9.75	10.86	16.19	7.00	7.34	16.15	0.21	37.54
MnO	0.04	0.07	0.21	0.20	0.22	0.51	0.00	0.46
MgO	29.00	15.28	24.83	16.74	15.07	14.36	0.00	6.13
CaO	2.01	16.33	1.93	18.79	20.09	6.77	0.98	0.10
Na_2O	0.20	0.94	0.43	-	-	-	8.25	1.79(V_2O_3)
K_2O	-	-	-	-	-	-	4.43	0.36(NiO)
TOTAL	99.92	100.02	99.57	99.78	99.69	99.30	99.86	99.83
	Analyst: J. D. Ross, 1977					Analyst: A. D. Robertson		
Si^{4+}	1.917	1.860	1.887	1.850	1.820	5.880	11.830	0.000
Ti^{4+}	0.002	0.029	0.013	0.010	0.015	0.079	0.000	1.960
Al^{3+}	0.173	0.230	0.167	0.245	0.301	3.711	4.112	0.000
Cr^{3+}	0.006	0.012	1.002	0.017	0.007	0.008	0.000	0.000
Fe^{3+}	-	-	-	0.009	0.023	0.321	0.031	0.000
Fe^{2+}	0.287	0.336	0.495	0.206	0.204	1.688	0.000	1.518
Mn^{2+}	0.001	0.002	0.006	0.006	0.007	0.064	0.000	0.019
Mg^{2+}	1.522	0.844	1.354	0.917	0.829	3.183	0.000	0.442
Ca^{2+}	0.076	0.648	0.076	0.739	0.795	1.079	0.187	0.005
Na^{1+}	0.014	0.060	0.030	0.000	0.000	0.000	2.851	0.062(V^{3+})
K^{1+}	0.000	0.000	0.000	0.000	0.000	0.000	1.007	0.005(Ni^{2+})
TOTAL	3.998	4.021	4.030	3.999	4.001	16.013	20.018	4.011
Ti								50.0
Fe	15.2	18.4	25.7	11.5	12.3	32.0		38.7
Mg	80.7	46.2	70.3	49.0	44.8	50.8		11.3
Ca	4.0	35.4	3.9	39.5	42.9	17.2	4.6	
Na							70.5	
K							24.9	

TABLE 3b

Representative comparative analyses, silicate megacrysts, Boyne basalts
[1] Brigooda breccia; [2] Ballogie Breccia)

	cpx^1	gnt^1	$amph^1$	opx^2	cpx^2	gnt^2	opx^2	$anorth^2$
SiO_2	49.84	41.16	39.41	54.36	51.16	41.03	54.09	67.03
TiO_2	0.94	0.44	5.15	0.14	0.45	0.46	0.28	0.00
Al_2O_3	9.23	22.82	14.14	4.78	6.70	23.55	5.76	20.47
Cr_2O_3	0.00	0.01	0.00	0.52	0.37	0.00	0.00	0.00
FeO	6.95	14.68	12.57	7.57	5.86	10.65	9.35	0.10
MnO	-	0.40	0.09	0.12	0.14	0.34	0.16	0.00
MgO	14.88	15.75	11.69	29.59	17.75	18.79	28.23	0.00
CaO	16.61	5.59	9.87	1.97	15.38	5.19	1.73	0.76
Na_2O	1.52	0.00	2.89	0.23	1.21	0.03	0.20	9.32
K_2O	-	0.00	2.02	-	0.00	0.00	-	2.56
TOTAL	**99.97**	**100.85**	**97.83**	**97.83**	**99.57**	**100.04**	**99.79**	**100.23**
Analysts*	1	2	2	3	3	3	3	3
Si^{4+}	1.810	5.972	5.877	1.909	1.874	5.890	1.900	11.798
Ti^{4+}	0.026	0.048	0.579	0.004	0.012	0.050	0.007	0.000
Al^{3+}	0.395	3.904	2.484	0.198	0.287	3.984	0.238	4.245
Cr^{3+}	-	0.001	0.000	0.014	0.011	0.000	0.000	0.000
Fe^{3+}	0.041	-	-	-	-	-	-	-
Fe^{2+}	0.170	1.781	1.568	0.222	0.178	1.279	0.275	0.014
Mn^{2+}	-	0.049	0.012	0.004	0.004	0.042	0.005	0.000
Mg^{2+}	0.805	3.406	2.595	1.549	0.960	4.022	1.479	0.000
Ca^{2+}	0.646	0.869	1.576	0.074	0.598	0.798	0.065	0.143
Na^{1+}	0.107	0.000	0.836	0.016	0.085	0.010	0.013	3.182
K^{1+}	-	0.000	0.383	-	0.000	0.000	-	0.574
TOTAL	**4.000**	**16.030**	**15.910**	**3.989**	**4.009**	**16.075**	**3.981**	**19.956**
Fe	12.7	29.4	27.3	12.0	10.3	21.0	15.1	
Mg	48.4	56.2	45.2	84.0	55.3	65.9	81.3	
Ca	38.9	14.3	27.5	4.0	34.4	13.1	3.6	14.7
Na								81.6
K								3.7

*Analysts: 1 - I. Jackson; 2 - L. R. Raynor; 3 - J. D. Hollis.

Cationic formulae are based on 6 oxygens for pyroxenes, 24 oxygens for garnet, 23 oxygens for amphibole and 32 oxygens for feldspar.

TABLE 4

Potassium-argon results, Mt. Wyangapinni

Sample	%K	$^{40}Ar^{*}$(x10^{-10} moles/g)	$^{40}Ar^{*}/^{40}Ar$	Total Age^{+}
anorthoclase	2.56	0.9104	0.771	20.5±0.2
megacryst	2.56			

* Denotes radiogenic 40Ar.

$^{+}$ Denotes age in Ma with error limits given for the analytical uncertainty at one standard deviation.

Constants ^{40}K=0.01167 atom% $\lambda_{\beta} = 4.962 \times 10^{-10}y^{-1}$ $\lambda_{\varepsilon} = 0.581 \times 10^{-10}y^{-1}$.

Determination by A. Webb, AMDEL, Frewville, South Australia.

TABLE 5

Representative microprobe analyses, Brigooda xenoliths

	Gnt-opx-clinopyroxenite			Gnt-amph-clinopyroxenite			2px-granulite*		
	cpx	opx	gnt	cpx	amph	gnt	cpx	opx	ilm
SiO_2	50.85	52.13	40.32	48.82	40.52	40.26	50.20	50.13	0.10
TiO_2	0.96	0.18	0.20	1.30	4.55	0.33	0.32	0.10	45.54
Al_2O_3	7.68	4.75	22.72	9.00	15.09	22.50	2.62	1.48	0.57
Cr_2O_3	0.27	0.12	0.13	0.00	0.00	0.00	0.00	0.00	0.10
FeO	7.58	14.75	15.64	8.25	10.16	14.82	12.11	28.18	49.50
MnO	0.20	0.30	0.52	0.00	0.00	0.40	0.33	0.71	0.35
MgO	13.52	26.27	15.60	13.56	13.90	15.47	11.89	18.31	3.00
CaO	16.66	1.06	5.08	17.21	10.13	5.11	21.62	0.57	0.00
Na_2O	2.51	0.63	0.00	2.25	2.74	0.00	0.64	0.00	0.00
K_2O	0.00	0.00	0.00	0.00	1.98	0.00	0.00	0.00	0.00
TOTAL	100.23	100.19	100.22	100.38	99.08	98.91	99.73	99.48	99.16
Analyst:		F. L. Sutherland			L. R. Raynor			W. L. Griffin	
Si^{4+}	1.861	1.878	2.960	1.794	6.132	2.981	1.913	1.939	0.005
Ti^{4+}	0.026	0.005	0.011	0.036	0.518	0.018	0.009	0.003	1.772
Al^{3+}	0.331	0.202	1.966	0.390	0.823	1.964	0.118	0.067	0.035
Cr^{3+}	0.008	0.003	0.008	0.000	0.000	0.000	0.000	0.000	0.004
Fe^{2+}	0.232	0.444	0.960	0.254	1.286	0.918	0.386	0.912	2.142
Mn^{2+}	0.006	0.009	0.032	0.000	0.000	0.025	0.011	0.023	0.015
Mg^{2+}	0.732	1.411	1.707	0.743	3.135	1.707	0.675	1.056	0.231
Ca^{2+}	0.653	0.041	0.400	0.678	1.650	0.405	0.883	0.024	0.000
Na^{1+}	0.178	0.044	0.000	0.160	0.804	0.000	0.047	0.000	0.000
K^{1+}	0.000	0.000	0.000	0.000	0.382	0.000	0.000	0.000	0.000
TOTAL	4.032	4.037	8.044	4.055	14.730	8.018	4.042	4.024	4.204
Ti									42.8
Fe	14.3	23.4	31.3	15.2	21.2	30.3	19.9	45.8	51.7
Mg	45.4	74.4	55.7	44.3	51.6	56.3	34.7	53.0	5.6
Ca	40.3	2.2	13.0	40.5	27.2	13.4	45.4	1.3	
Wells									
$(Fe^{3+})T°C$			1140					895	
Wood & Banno									
$(Fe^{3+})T°C$			1140					995	

Cationic formulae are based on 6 oxygens for pyroxenes and ilmenite, 12 oxygens for garnet and 24 oxygens for amphibole. *(plagioclase An 37-39%)

TABLE 6

Summary of temperature estimates, southern Queensland xenoliths.

Suite	Wells T°C	Wood & Banno T°C	W & B (T+40)°C
Albert Basalt			
Palen Creek: spl lherzolite	1140	1150	1190
Richmond Gap: cpx, opx	990	980	1020
Main Range Basalts			
Gowrie Mountain:			
spl lherzolites	910-1160	930-1170	970-1210
& metawebsterites			
wehrlites	1110-1240	1120-1160	1160-1200
ol-spl websterite	1170	1160	1200
(exsolution)	990	1020	1060
granulites	780- 860	800- 830	840- 870
Cliffdale:			
spl lherzolites	900-1140	990-1140	1030-1180
& metapyroxenites			
granulite	800	860	920
Middle Branch Creek:			
clinopyroxenites			
(exsolution)	1190-1210	1190-1210	1230-1250
Mount Mitchell:			
spl lherzolites	950- 990	900-1110	940-1150
& hy-gabbro			
Boyne Basalts			
Brigooda:			
spl lherzolites	920-1040	970-1110	1010-1100
& gnt pyroxenites			
granulite	840	810	850

Results based on electron microprobe of pyroxenes analyses in xenoliths, taken from Grenfell (1984) and this paper.

TABLE 7

Compositional parameters for co-existing pyroxene phases in mantle assemblages, S. Queensland

	Cr-enriched,Ti-poor group		Transitional Group		Cr-poor,Ti-enriched Group	
	cpx	opx	cpx	opx	cpx	opx
Cr_2O_3	>0.4	>0.2	0.2-0.4	0.1-0.2	<0.2	<0.1
TiO_2	<0.9	<0.3	0.5-1.0	0.1-0.4	>0.5	<0.1
Al_2O_3	<7.0	<6.0	3.0-8.0	4.0-6.0	>7.0	>6.0
e.g.	spl lherzolite; Gowrie Mt., Harlaxton, Palen Ck gnt-opx cpxenite; opx cpxenite; Gowrie Mt. websterite; Cliffdale websterite; Cliffdale		ol websterite; Gowrie Mt. gnt-opx cpxenite; Brigooda		ol-spl websterite; spl websterite; Gowrie Mt. gnt-spl cpxenite; Brigooda	

Based on analyses in Grenfell (1984); Ewart & Grenfell (1985) and this paper (Tables 2 and 5)

TABLE 8

Representative analyses, range of mantle-lower crust xenolithic rocks, South Queensland

Analysis	1	2	3	4	5	6
SiO_2	44.17	41.55	47.40	51.16	46.83	49.02
TiO_2	0.11	0.90	0.26	0.73	2.84	2.88
Al_2O_3	3.19	19.42	12.82	7.30	18.06	15.60
Fe_2O_3	-	2.19	-	1.41	-	-
FeO	8.79	10.97	5.73	7.91	10.76	11.61
MnO	0.16	0.21	0.15	0.19	0.19	0.24
MgO	37.95	14.93	21.93	15.89	6.06	5.05
CaO	2.85	7.91	9.07	12.86	8.77	8.43
Na_2O	0.37	0.64	1.11	1.50	3.03	3.11
K_2O	0.01	0.09	0.07	0.01	0.30	0.64
P_2O_5	0.02	0.07	0.03	0.05	0.12	1.08
L.O.I.	1.78	0.27	1.67	1.00	2.20	2.04
TOTAL	**99.40**	**98.95**	**100.24**	**100.13**	**99.16**	**99.70**
Rb ppm	<1	-	2	-	1	4
Ba	3	997	47	61	118	370
Sr	8	106	120	86	748	564
Zr	5	78	16	31	14	319
Y	4	47	13	14	15	67
V	70	275	189	263	393	204
Ni	2225	211	823	400	51	46
Cr	2700	5	4275	1211	24	80
Zn	52	90	56	67	84	126
Cu	35	46	15	94	178	36
S.G.	3.34	-	3.30	-	3.02	3.02

1. Spinel lherzolite Boowinda Creek (averaged from Griffin et al. 1987).
2. Garnet clinopyroxenite, Brigooda (L.R. Raynor, analyst).
3. Garnet websterite, Boowinda Creek (from Griffin et al.1987).
4. Garnet opx-bearing clinopyroxenite, Brigooda (Griffin et al. 1987).
5. 2px granulite, Boowinda Creek, (averaged from Griffin et al. 1987).
6. 2px granulite, Brigooda (from Griffin et al. 1987).

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GEOCHEMISTRY OF METABASALTS FROM THE GYMPIE PROVINCE: - IMPLICATIONS FOR CRUSTAL DEVELOPMENT AND THE ORIGIN OF SUSPECT TERRANES AT THE GONDWANA RIM

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ABSTRACT

Near the eastern end of the Eromanga-Brisbane Geoscience Transect, geochemical analyses of Gympie Province metabasalts indicate some likely Carboniferous events affecting the palaeo-Pacific margin of Gondwanaland prior to the formation of the major Permo-Triassic Bowen-Gunnedah-Sydney Basin system.

Five geochemically and petrographically distinct tholeiitic metabasalt suites are recognised within several discrete volcano-sedimentary sequences that comprise part of the Gympie Province, southeast Queensland. Markedly differing source contributions from (a) asthenospheric mantle, (b) sub-continental lithospheric mantle, and (c) subduction-related (slab-derived) components for the various magma suites imply their generation in contrasting tectonic regimes, despite the fact that they all show some affinities with eruptives from convergent plate margin settings.

Basalts from the Highbury Volcanics of the Gympie Group (Gympie basalts) are island arc tholeiites formed during an immature stage of eruptive activity in an intra-oceanic arc. They are unique among the analysed metabasites in possessing high Zr/Nb (Zr/Nb ≈ 30) and very high Ba/Nb ratios, as well as low Ti/V, Ti/Zr and Ti/Y ratios. These features are explicable in terms of contributions from a depleted, convecting upper mantle source, in addition to large-ion lithophile element (Ba, K and Rb)-enriched fluids ascending from subducted oceanic crust.

The Cedarton Volcanics display high Ba/Nb, low Ti/V and Ti/Y ratios, but have very low Zr/Nb ratios (Zr/Nb ≈ 5), and higher Zr/Y and Ti/Zr ratios than the Gympie basalts. In contrast to these latter basalts, the involvement of a substantial sub-continental lithospheric mantle component in the genesis of these rocks implies their generation in a continental margin subduction-related setting. Compositional variation within both suites is mainly due to fractionation of ferromagnesian phases (olivine + clinopyroxene), typical of volcanic arc basalts.

Sub-continental lithospheric mantle contributions are also required to account for geochemical features of metabasalts from the Rocksberg Greenstone and Amamoor beds. Two compositionally distinct suites of Amamoor basalts (Types 1 and 2) are identified. Zr/Nb ratios for Rocksberg and Amamoor Type 1 basalts (Zr/Nb ≈ 10) are intermediate between those of the Cedarton and Gympie basalts. Their Ba/Nb ratios are generally low. The majority of the Rocksberg and Amamoor Type 1 basalts have (high) Ti/V and Ti/Y ratios, as well as Ti/Cr - Ni (and some other trace element) distributions similar to those of mid-ocean ridge basalts (MORB). Relatively low Al_2O_3 contents and Al_2O_3/TiO_2 ratios for these rocks imply that low-pressure, plagioclase-dominated fractionation controlled their early-stage magmatic evolution, as often inferred for rift magmas. It is concluded that the Rocksberg and Amamoor Type 1 basalts originated in a locally extensional setting within an overall continental margin, subduction-related environment. Inter- or intra-arc rifting facilitated the eruption of a high proportion of relatively primitive (high $Mg^{\#}$) magmas. Compared to sources proposed for the Cedarton Volcanics, sub-continental lithospheric mantle and slab-derived source contributions were reduced, due to a greater input from convecting upper mantle. Amamoor Type 2 basalts show geochemical affinities with both the Cedarton and Amamoor Type 1 basalts, as well as strong depletions in a wide range of incompatible elements (Ce, P, Zr, Ti and Y). They may represent eruptive products of incipient rifting.

The diverse tectonic settings inferred for the origin of the Gympie Province metabasalt suites is consistent with the proposal that their host sequences of metasedimentary and metavolcanic rocks comprise discrete suspect terranes. Moreover, in the easternmost Gympie Province, the Gympie Group (which includes the Early Permian Gympie basalts of intra-oceanic island arc affinity) is almost certainly exotic with respect to the other (in part older) sequences of varied, but overall Cordilleran-margin affinities to the west. The latter include the Cedarton Volcanics (part of a volcanic arc proximal to, or comprising a portion of a continental margin), as

well as the Rocksberg Greenstone and Amamoor beds (of supra-subduction zone rift origin). The continental margin affinities of these units suggest their closer original relationship to Gondwanaland than proposed for the Gympie Group.

INTRODUCTION

Time-transgressive lateral accretion models for continental growth, which involve addition to continental lithosphere of terranes formed at active plate margins, have been applied to numerous circum-Pacific regions (e.g. Seely & others, 1974; Monger & others, 1982; Marriner & Millward, 1984; Avè Lallemant & Oldow, 1988). Active margins are often marked by both compressional and transcurrent structures along the boundary between plates (McKenzie & Morgan, 1969; Avè Lallemant & Oldow, 1988). Oblique plate convergence can cause terranes of diverse origin (e.g. magmatic arc, fore-arc, or subduction complex) in continental margin settings to behave semi-independently and migrate as coherent structural bodies parallel to the subduction zone along arc-parallel strike-slip faults. Some of the resulting juxtaposed suspect terranes may always have been peripheral to the continent. Others may represent exotic terranes that collided with the continent subsequent to their transport from a distal source (e.g., remnants of intra-oceanic arcs rafted by seafloor spreading), to be then displaced along the continental margin by transform movements (Saleeby, 1977; Avè Lellemant & Oldow, 1988).

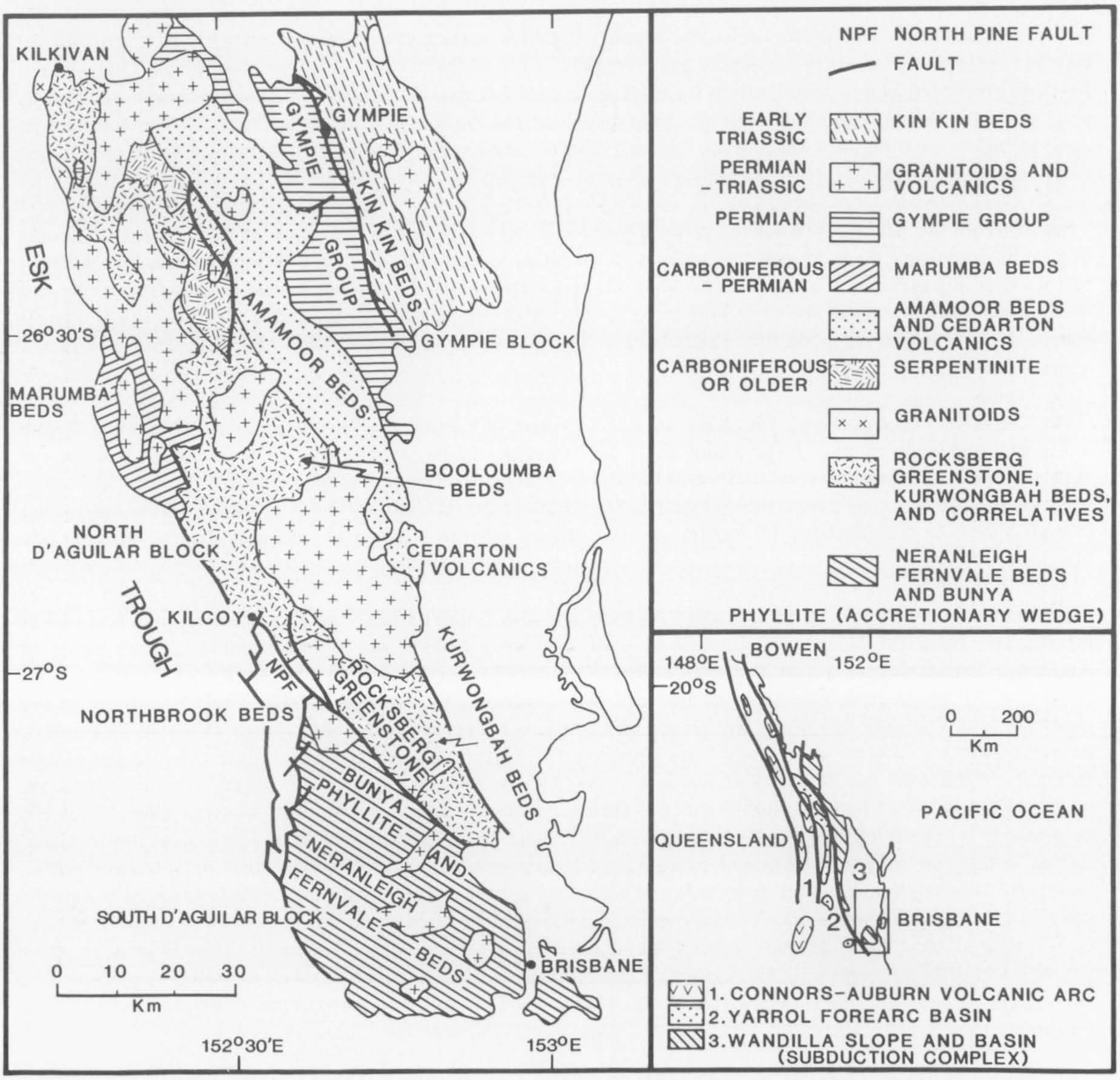

Fig. 1 Simplified geological map showing pre-Middle Triassic tectonostratigraphic terranes of southeast Queensland (modified after Murray, 1988). Inset shows inferred tripartite stratotectonic make-up of the Late Palaeozoic Yarrol Orogen.

TABLE 1 - Some lithological units from pre-middle Triassic terranes hosting metavolcanic suites in Gympie Province, southeast Queensland.

Terrane	Unit	Age	Lithology	Thickness	Depositional environment
"Gympie terrane" (exotic with respect to terranes listed below)	Kin Kin beds	Early Triassic	mudstone, shale, arenite, pebble conglomerate	several kms.	shallow marine (fore-arc)
	Gympie Group				
	Rammutt Fm.	Early Permian	shale, argillite, greywacke, volcanic conglomerate, dacitic volcanics.	760 m	volcanic arc
	Highbury Volcanics	Early Permian	mainly basaltic tuff-breccia	365 m	shallow marine, immature intra-oceanic island arc.
Suspect terranes of western Gympie Province*	Cedarton Volcanics	Probably Early Permian	mainly basaltic volcanics and feldspathic sandstone	unknown	immature arc in vicinity of (or proximal to) continental margin
	Amamoor beds (most likely a composite terrane)	Early Carboniferous-Early Permian	shale, pebbly mudstone, feldspathic sandstone, chert and jasper, amygdaloidal basalt, manganiferous lenses	probably several kms.	marine; inter- or intra-arc rift; in part Early Carboniferous seafloor proximal to continental margin
	North D'Aguilar Block				
	Kurwongbah beds	Early Carboniferous or older	phyllite, slate, basaltic andesite volcanics		marine
	Rocksberg Greenstone	Early Carboniferous or older	tholeiitic metabasalt (greenschists and transitional blueschists), minor pelitic and quartzofeldspathic schist	3 - 4 km	inter- or intra-arc rift in continental margin setting.

* Always proximal to, or comprising part of a Cordilleran-type continental margin (in contrast to "Gympie terrane"). See text for explanation.

This paper discusses the easternmost on-shore sector of the Eromanga - Brisbane transect of Late Palaeozoic Gondwanaland bordering the palaeo-Pacific Ocean. The Pacific margin of Gondwanaland records a complex history of mainly subduction-related magmatism from Late Devonian to Early Triassic time. In this paper, we outline petrogenetic relationships among metavolcanic suites from the Gympie Province, which comprises several suspect or exotic terranes in the easternmost part of the southeast Queensland (New England Orogen) segment of the Gondwana margin. Distinctive magmatic evolutions of these mafic volcanic suites imply the origin of their host terranes in diverse tectonic settings. We examine the relation of the various terranes to subduction and consequent crustal growth at the Gondwana rim, and suggest possible terrane interactions.

GEOLOGICAL SETTING - OUTLINE OF TERRANES

The Yarrol Province, the northern portion of the New England Fold Belt in eastern Australia, includes three meridionally trending belts of Late Devonian-Early Carboniferous rocks representing an Andean-style (Connors-Auburn) volcanic arc in the west, an unstable median (Yarrol) fore-arc basin, and a subduction complex to the east (Wandilla slope and basin) (Fig. 1). The tripartite stratotectonic framework of Yarrol Province implies that for much of the Late Palaeozoic, the Pacific margin of Gondwanaland was an active convergent plate margin above a west-dipping subduction zone (Day & others, 1978; Murray & others, 1987).

To the east of the Yarrol Province, in southeast Queensland, lies the Gympie Province, which does not fit the overall palaeogeographic pattern of the Yarrol Province (Day & others, 1978) and may represent a displaced block or terrane (Harrington, 1983; Waterhouse & Sivell, 1987). Murray (1988) and Murray & others (1989) consider the Gympie Province to comprise at least three discrete suspect terranes which may be exotic to the remainder of the New England Fold Belt and also to each other. These include the North D'Aguilar Block (mafic metavolcanics and fine-grained metasediments of the Rocksberg Greenstone, Kurwongbah beds and correlatives); the Good Night beds of slate, phyllite, and lenses of arenite, chert, limestone and mafic volcanics (not discussed here); and in the eastern part of the Gympie Province, varied sequences of sediments and volcanics of the Amamoor beds, Cedarton Volcanics and Gympie Group (Permian and older), together with the Kin Kin beds and Brooweena Formation (Early Triassic) (see Table 1).

In the North D'Aguilar Block, at the boundary of the Yarrol and Gympie Provinces (Fig. 1), the Rocksberg Greenstone consists of a thick pile of mafic tuffs and lavas and interbedded fine-grained sediments that are metamorphosed to greenschist and locally transitional blueschist (glaucophanitic greenschist) assemblages. The Kurwongbah beds, overlying the Rocksberg Greenstone, consist of phyllite, slate, mafic metavolcanics, minor chert and rare siltstone (Murphy & others, 1979). The sequence shows evidence of at least three generations of folding, the first and major generation forming a large antiform cored by the Rocksberg Greenstone (Donchak, 1976). There is no conclusive evidence for the age of the Rocksberg Greenstone protolith. However, its (pre-Middle Carboniferous) minimum age is constrained by sparse marine fossils of Early Carboniferous to Early Permian age from the Marumba beds, which unconformably overlie correlatives of the Rocksberg Greenstone and Kurwongbah beds, and also by 298 and 320 Ma S-type granitoid intrusions (Green, 1973; Murphy & others, 1976; Murray, 1988). Metasomatically altered serpentinite masses, folded and metamorphosed at the end of the Carboniferous along with the remainder of the North D'Aguilar Block (Murray, 1988), form the discernible part of the boundary of this terrane with the Amamoor beds to the east.

The Amamoor beds include shales, pebbly mudstone, feldspathic sandstone, greywackes and impure limestone, as well as basaltic flows, banded chert and jasper associated with manganese mineralisation (Murray, 1987). The unit was considered to be entirely of Permian age, because of the presence of abundant Early Permian bivalves in local coarse volcaniclastics (Murray & others, 1979). However, the subsequent discovery of Early Carboniferous radiolarians from chert lenses in both the Amamoor beds (Murray & others, 1989) and the southern part of the area mapped as Gympie Group in Figure 1 suggests that part of the sequence is older and unrelated. The boundaries of this older sequence have not yet been delineated. No structures typical of the North D'Aguilar Block occur. However, the Amamoor beds are more complexly deformed than the Gympie Group to the east and both S1 slaty cleavage and S2 crenulation may be developed. To the south of the Amamoor beds, and separated from them by only a thin cover of Tertiary basalts, the Cedarton Volcanics include mainly basaltic tuffs and tuff-breccias. A Permian marine fauna, similar to that in the Amamoor beds (Murray & others, 1979; Waterhouse & Balfe, 1987) has been recognised.

The Early Permian submarine volcanic sequence of the Gympie Group comprises mainly tholeiitic basaltic tuff-breccias and subordinate lavas at the base of the pile (Highbury Volcanics), with dacitic tuffs predominating higher in the succession (Rammutt Formation) (Sivell & Waterhouse, 1988). A prominent limestone horizon overlies the volcanics. The Gympie Group is in turn overlain by the Early Triassic Kin Kin beds (and Brooweena Formation to the north) of cleaved shales interbedded with arenite and pebbly conglomerate. The shallow-marine association of the Gympie Group-Kin Kin beds constitutes a discrete suspect terrane of island arc affinity showing some similarities to Permian magmatic arc-forearc basin sequences in New Zealand

TABLE 2 - Major and trace element analyses of Rocksberg metabasites

Analysis No.	1	2	3	4	5	6	7	8
Sample	11	54	29	46	1B	53	TC2	STP1
SiO_2	48.66	45.71	51.51	47.90	50.18	43.64	49.96	55.03
TiO_2	0.80	1.12	0.70	0.86	0.67	1.20	0.77	1.93
Al_2O_3	11.33	14.55	10.34	12.49	9.42	15.53	12.02	13.11
Fe_2O_3	6.25	7.68	5.38	3.17	5.08	5.19	4.42	7.63
FeO	3.69	2.95	3.10	7.06	5.56	6.66	4.56	5.11
MnO	0.13	0.20	0.10	0.15	0.15	0.10	0.14	0.14
MgO	11.32	4.88	7.90	11.68	12.28	12.18	9.00	3.13
CaO	8.92	10.00	10.40	10.50	11.05	3.93	13.86	3.56
Na_2O	2.05	2.29	2.10	2.32	2.04	4.14	1.94	6.46
K_2O	1.80	2.36	1.99	0.10	0.07	0.03	0.74	0.17
P_2O_5	0.15	0.12	0.13	0.15	0.14	0.18	0.14	0.13
H_2O^+	3.26	3.07	2.38	3.56	2.83	5.73	2.18	2.66
H_2O^-	0.04	0.05	0.05	0.04	0.19	0.04	0.17	0.73
CO_2	1.20	4.80	3.51	0.02	0.05	1.33	0.20	0.02
Σ	99.59	99.79	99.58	99.99	99.73	99.88	100.09	99.81
100 $Mg/(Mg + Fe^{2+})$	71.2	50.0	66.9	70.6	71.1	68.6	67.9	34.2
Trace elements								
Cr	1291	316	373	647	819	440	446	0
Ni	354	139	124	213	143	234	145	4
V	266	201	264	259	242	291	289	402
Rb	34	50	48	1	1	1	17	7
Sr	285	333	395	242	304	119	548	52
Ba	179	325	331	5	5	1	147	155
Zr	45	60	28	49	38	66	54	126
Nb	7	5	4	7	4	6	5	4
Y	16	19	16	18	15	22	20	53
Th	2	1	1	1	1	1	1	3
La	7.0	6.6	9.0	8.4	4.5	8.2	5.3	6.4
Ce	12.8	21.4	23.9	8.5	10.3	13.0	20.1	15.3
Nd	1.2	8.6	8.3	6.2	4.8	7.0	5.5	8.6

Analyses 1-3 = transitional blueschists; 4-6 = greenschists; 7 = unrecrystallized basalt; 8 = incompletely recrystallized stilpnomelane-bearing metabasite. $Mg/(Mg + Fe^{2+})$ ratio calculated assuming initial oxidation ratio for iron of Fe_2O_3/FeO = 0.15. Major oxides and trace elements were analysed by X-ray fluorescence in the School of Earth Sciences, Macquarie University, FeO by HF-vanadate procedure, and H_2O^+ and CO_2 by gravimetry.

TABLE 3 - Major and trace element analyses of metabasalts from the Cedarton Volcanics and Amamoor Beds

Analysis No.	1	2	3	4	5	6	7	8	9	10
Sample	1XB	38B	37A	31	16	14	10	16B	9B	42
SiO_2	48.63	47.51	47.45	50.86	50.82	46.80	46.61	52.33	46.93	47.23
TiO_2	1.08	0.72	0.75	0.76	1.31	0.99	1.69	1.13	0.84	0.67
Al_2O_3	18.57	18.43	16.08	15.58	14.64	16.31	14.72	13.46	18.67	16.50
Fe_2O_3	12.78	10.44	10.26	12.86	9.77	8.28	10.12	11.08	8.26	7.71
MnO	0.11	0.17	0.16	0.17	0.16	0.09	0.15	0.16	0.15	0.12
MgO	3.46	7.73	8.90	4.66	6.13	7.39	7.58	4.76	6.73	8.04
CaO	7.85	11.05	9.95	7.94	10.06	9.79	9.16	9.53	10.17	9.97
Na_2O	2.93	2.16	1.92	3.24	3.59	3.80	2.10	3.24	2.98	1.74
K_2O	1.40	0.14	0.28	0.75	0.40	0.20	2.30	0.17	0.58	2.15
P_2O_5	0.23	0.09	0.11	0.12	0.12	0.10	0.26	0.11	0.07	0.06
S	0.03	0.04	0.04	0.03	0.03	0.03	0.03	0.03	0.03	0.03
LOI	2.60	1.82	3.91	2.56	2.82	6.12	5.40	3.40	4.41	5.62
Σ	99.66	100.29	99.81	99.63	99.86	99.89	100.12	99.40	99.83	99.83
100 $Mg/(Mg + Fe^{2+})$	37.9	62.5	66.2	45.0	58.5	66.7	62.8	49.2	64.7	58.4
Trace elements										
Cr	48	219	554	260	313	581	195	281	244	394
Ni	32	87	181	148	72	216	56	102	70	117
V	397	308	273	347	291	185	289	253	175	173
Rb	29.8	3.9	8.4	23.1	7.3	2.8	23.7	6.0	6.4	41.0
Sr	268	147	349	296	69	80	182	192	177	51
Ba	287	100	164	177	94	13	102	63	147	117
Zr	59.8	39	52	50	81	65	121	94	42	32
Nb	8.3	9.0	7.5	9.8	8.7	8.7	24.0	7.9	6.3	8.2
Y	40	17	18	21	33	20	26	31	19	18
Th	4.2	3.3	2.0	2.8	1.8	2.3	2.8	3.3	1.3	1.7
La	9.6	4.0	4.8	6.8	2.4	1.9	6.3	4.9	1.5	2.8
Ce	28.1	13.9	18.6	18.4	15.1	8.3	30.6	18.9	4.8	5.8

Analyses 1-4 = Cedarton Volcanics metabasalts; 5-8 = Amamoor beds Type 1 metabasalts; 9-10 = Amamoor beds Type 2 metabasalts. $Mg/(Mg + Fe^{2+})$ as in Table 1. Major oxides were analysed by X-ray flourescence in the Department of Geology, The University of Sydney. LOl = loss on ignition. Geochemical analyses of metabasalts from the Highbury Volcanics of the Gympie Group (not included here) are reported in Sivell & Waterhouse (1988, Tables 1 & 2).

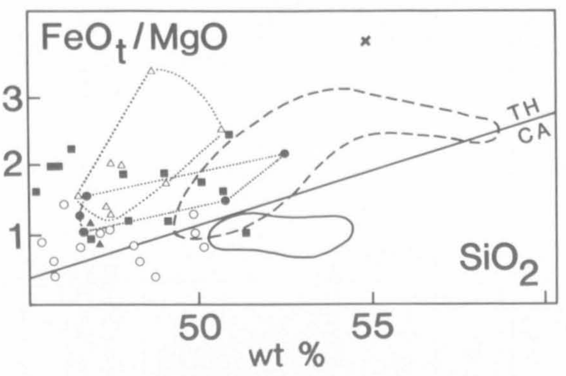

Fig. 2 FeO$_t$/MgO vs. SiO$_2$ variation diagram (after Miyashiro, 1974) for Late Palaeozoic southeast Queensland metabasalt suites. Symbols: open circles - Rocksberg greenschists; solid squares - Rocksberg transitional blueschists (glaucophanitic greenschists); cross - stilpnomelane-bearing Rocksberg metabasite; solid circles - Amamoor Type 1 basalts; solid triangles - Amamoor Type 2 basalts; open triangles - Cedarton Volcanics. Dashed balloon shows field of metabasalts from the Highbury Volcanics of the Gympie Group (Gympie basalts of Sivell & Waterhouse, 1988) and Permian island arc tholeiites from New Zealand which have a similar distribution (Sivell & Rankin, 1983). Solid balloon is field of Gympie and New Zealand tholeiitic ankaramites (with MgO > 10 wt. % due to olivine + clinopyroxene accumulation; Sivell & Rankin, 1983; Sivell & Waterhouse, 1988). Dotted balloons emphasise fields for Cedarton and Amamoor Type 1 basalts. TH/CA indicates tholeiitic/calc-alkaline divide.

(Sivell & Waterhouse, 1988; Waterhouse & Sivell, 1987).

PETROGRAPHY

The porphyritic basalts which comprise the tuff-breccias and lavas that dominate the Highbury Volcanics of the Gympie Group (referred to here as the Gympie basalts) contain prominent phenocrysts of clinopyroxene (up to 8 mm in diameter), together with subordinate olivine and plagioclase, in a groundmass of pyroxene, plagioclase, magnetite and glass. Total phenocryst content is typically in the range 20-30 mode %. Basalts from the Cedarton Volcanics are more highly porphyritic than the Gympie rocks (total phenocryst content often exceeds 40 mode %), and plagioclase (with subordinate clinopyroxene and olivine) dominates the phenocryst assemblages. Clinopyroxene crystals are generally 2 to 4 mm in diameter. The glassy or aphanitic groundmass may be rich in Fe-Ti oxides. In these partly recrystallised Gympie and Cedarton volcanics, plagioclase is albitised, and olivine is pseudomorphed by chlorite. Clinopyroxene is the only unaltered primary mineral. Abundant amygdales are infilled by pumpellyite, epidote, chlorite, albite, quartz and calcite. In some more extensively recrystallised Cedarton basalts, actinolite may replace ferromagnesian phenocryst and groundmass phases.

There are two petrographically distinct groups of Amamoor basalts. Amygdaloidal, variolitic Amamoor Type 1 basalts may be aphyric or include sparse (< 5%) phenocrysts of plagioclase (up to 2 mm in length) and lesser olivine (including a high proportion of skeletal microphenocrysts). Groundmass minerals include plagioclase, clinopyroxene and Fe-Ti oxides, with textures ranging from intersertal to intergranular. Amygdales contain pumpellyite, quartz, and calcite, and epidote and quartz veins are common. Amamoor Type 2 basalts include (a) medium to coarse-grained rocks in which clinopyroxene ophitically encloses plagioclase; and (b) porphyritic basalts with ≈ 15 mode % phenocrysts of predominant plagioclase (up to 8 mm in length) and (chloritised) olivine (< 3mm). The phenocrysts occur in a hypocrystalline groundmass of plagioclase and columnar, skeletal clinopyroxene. Pumpellyite veins are common.

The majority of metabasites from the Rocksberg Greenstone are fine- to medium- grained, slightly to extensively recrystallised rocks in which greenschist

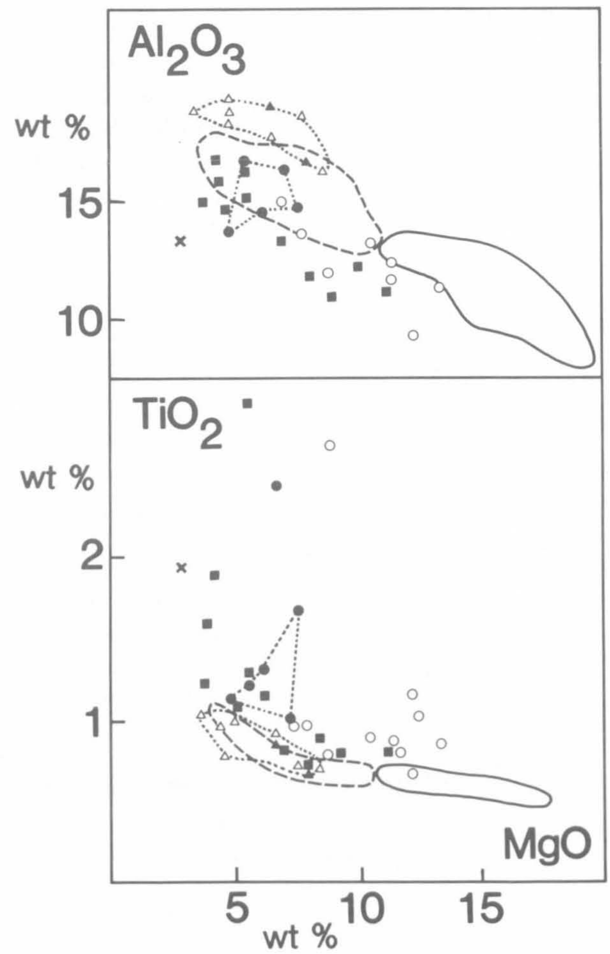

Fig. 3 Al$_2$O$_3$ and TiO$_2$ vs. MgO variation diagrams for southeast Queensland metabasalts. Symbols as in Figure 2.

facies assemblages are developed (e.g. albite + actinolite + epidote (Ps_{10}) + white mica + chlorite + quartz + magnetite $\pm$ calcite). Actinolite may partly or completely pseudomorph very coarse clinopyroxene 'megacrysts' (> 1cm in diameter) in these rocks. Late-stage (? post-tectonic) grain-coarsening and growth of porphyroblastic albite overprint earlier developed foliations in the most completely recrystallised rocks (cf. Wilson, 1972). Transitional blueschist-greenschist facies parageneses are developed in the most Fe_2O_3-rich Rocksberg metabasites (glaucophanitic greenschists). Early-formed, syn-tectonic assemblages in these rocks include sodic amphibole $\pm$ calcic amphibole + epidote (Ps_{20}) + quartz + albite + white mica + chlorite + sphene $\pm$ rutile $\pm$ hematite. The Na-amphibole-bearing rocks do not contain lawsonite, jadeite or aragonite. Glaucophane- and crossite-bearing assemblages may be partly to completely overprinted by actinolite-bearing (greenschist) assemblages (Sivell, in press). Some weakly recrystallised Rocksberg basalts (and basaltic andesites) in the upper part of the Rocksberg Greenstone sequence contain stilpnomelane.

GEOCHEMISTRY

Alteration Effects and Assessment of Geochemical Data

Metavolcanics comprising the Rocksberg Greenstone, Cedarton Volcanics, Amamoor beds, and Highbury Volcanics of the Gympie Group are mainly tholeiitic basalts metamorphosed to (locally glaucophane-bearing) greenschist facies or lower grade assemblages. The basalts range from only slightly recrystallised to extensively or completely recrystallised.

Studies of the mobilities of elements suggest that abundances of some major and trace elements (especially alkalies) are susceptible to alteration during low-grade metamorphism (e.g. Pearce & Cann, 1973; Grapes, 1976). Any assessment of magmatic processes that took place prior to metamorphism of the Gympie Province basalts is dependent upon the degree to which metamorphism was isochemical. Sivell & Waterhouse (1988) suggested that introduction of Na_2O accompanied complete albitisation of plagioclase in Highbury basalts from the Gympie Group, and that a wide spread of K_2O contents in these rocks (showing no relation to any parameter of igneous differentiation) implied some mobility of this element. However, these authors also concluded that the abundances of ferromagnesian elements were little affected by low-grade metamorphism, with the Gympie basalts maintaining their sub-alkaline character as demonstrated for spilitic suites elsewhere (e.g. Cann, 1969).

As observed for the Gympie basalts, systematic relationships are evident between the abundances and ratios of many elements in samples from the Rocksberg Greenstone, Cedarton Volcanics and Amamoor beds metavolcanic suites. Although they possess a wide range of major and trace element concentrations, samples from

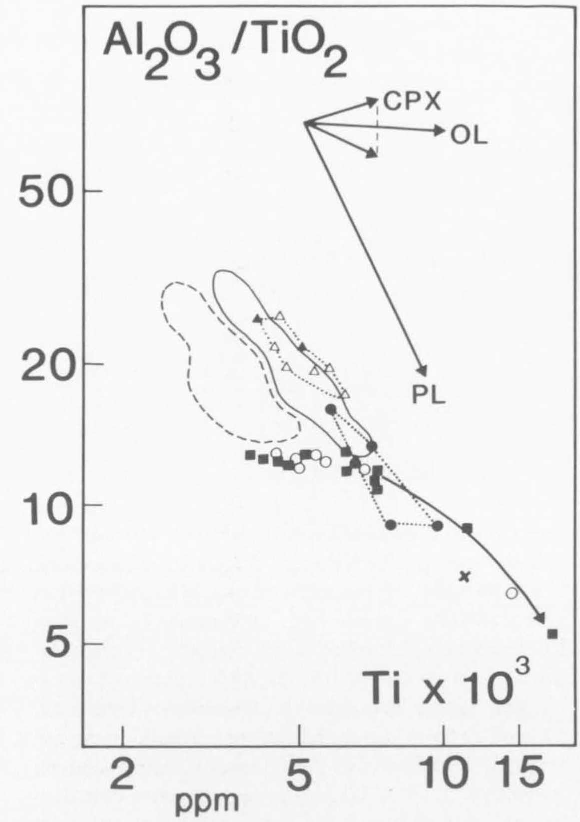

Fig. 4 Al_2O_3/TiO_2 vs. Ti (ppm) diagram for southeast Queensland metabasalts. Fractionation vectors represent 50% crystallisation of plagioclase (PL), olivine (OL), and clinopyroxene (CPX). Arrow shows trend of Rocksberg samples. Other symbols as in Figure 2.

these individual metabasalt associations show good correlations between a number of elements (e.g. Al, Mg, Ti, V, Zr, Nb, and Y), suggesting that metamorphic transformations did not appreciably affect the abundances of these elements. The continuous compositional trends displayed by individual metavolcanic suites on major and trace element plots correlate with those expected for igneous suites. For example, in the Gympie and Cedarton basalts, MgO level decreases (and Al_2O_3 content increases) with increasing abundance of Zr, which is immobile under most metamorphic conditions, and provides a good indicator of the degree of igneous fractionation (e.g. Winchester & Floyd, 1977). Relatively constant trace element ratios (e.g. Ti/Y, Zr/Nb and Ti/V) for samples from single suites accord well with basaltic series related by crystal fractionation. Moreover, fractionating mineral assemblages inferred on the basis of chemical trends for the various volcanic suites correspond to phenocryst assemblages actually recognised in the least recrystallised basalts.

In this study, except where specifically stated, it is assumed that systematic geochemical variation reflects primary igneous processes. Significant differences in inter-element ratios between different suites are considered to result from contrasting sources for their primary magmas (cf. Wood, 1979). This is particularly the case for ratios involving elements such as Zr, Nb, Ti, Cr, and Y, which are typically regarded as immobile during low-grade metamorphism (Pearce & Cann, 1973; Winchester & Floyd, 1977). However, in some instances, there is clear evidence that element mobility has accompanied metamorphism. Comparison of the chemical compositions of glaucophane-bearing Rocksberg metabasites with slightly recrystallised Rocksberg basalts indicates that the sodic amphibole-forming metamorphism was near-isochemical. By contrast, (later) metamorphism that led to formation of greenschist assemblages (including porphyroblastic albite) in the Rocksberg basalts was accompanied by leaching of K, Ba, Rb and Sr (Sivell, in press), so that concentrations of these elements in the greenschists may not represent original levels in the magmas.

Major Elements

The tholeiitic character of the Gympie Province metabasites is demonstrated by their elevated FeO_t/MgO ratios (Fig.2) and trends of increasing Fe, Ti and V with advancing fractionation (e.g., Fig. 3; see also Fig. 10). Highest FeO_t/MgO values (up to 3) are attained by the Cedarton and Gympie basalts, whilst Amamoor and most Rocksberg samples show $FeO_t/MgO < 2$. Metabasalts from the Rocksberg Greenstone, Gympie Group and Cedarton Volcanics define sub-parallel trends of increasing Al_2O_3 with decreasing MgO, each of these suites showing a different average alumina content for a given MgO (Fig. 3). The Cedarton basalts have the highest Al_2O_3 levels (16-19 wt%). Such high Al_2O_3 contents are often displayed by subduction-related magmas. Because of their hydrous nature, arc magmas typically do not undergo extensive plagioclase fractionation until they are quite evolved, leading to enhanced Al contents in the more differentiated magmas (Perfit & others, 1980). Rather, olivine and clinopyroxene are the early phases to crystallise from these melts. Although the Gympie basalts do not attain the very high Al_2O_3 levels of the Cedarton volcanics, they possess higher alumina contents than the majority of the Rocksberg metabasites over a wide range of MgO. Sivell & Waterhouse (1988) interpreted the Gympie basalts as island arc tholeiites (IAT). Like the Rocksberg samples, the majority of metabasalts from the Amamoor beds (Amamoor Type 1 basalts) also show rather low Al_2O_3, overlapping with the fields of fractionated Rocksberg and Gympie basalts. Two Amamoor samples (Amamoor Type 2 basalts 9B and 42) show "anomalously" high Al_2O_3 and plot in the region occupied by the Cedarton Volcanics.

Distinctive trends for the different volcanic suites are also evident on the TiO_2 - MgO plot (Fig. 3). In this diagram, the Gympie rocks adhere closely to a low-TiO_2 trend, TiO_2 increasing as MgO decreases. The Cedarton basalts overlap the low MgO (more evolved) portion of this well-defined trend. Rocksberg and Amamoor Type 1 metabasalts have distinctly higher TiO_2 contents than the Gympie and Cedarton volcanics at all MgO values. Again, Amamoor Type 2 basalts are geochemically distinct from their Type 1 counterparts, and fall within the field of the low-TiO_2 Gympie and Cedarton Volcanics.

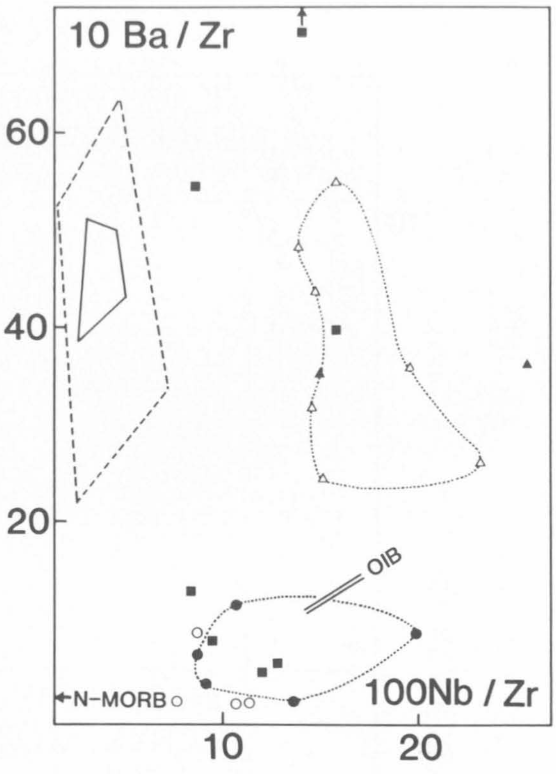

Fig. 5 Ba/Zr and Nb/Zr variation in southeast Queensland metabasalts. The field of Gympie basalts (solid-line quadrilateral) is encompassed by the field of New Zealand IAT (dashed quadrilateral). N-MORB = mid-ocean ridge basalts; OIB = ocean island basalts. Other symbols as in Figure 2.

In Figure 4, the Rocksberg metabasalts plot along a well-defined trend characterised by distinctly lower Al_2O_3/TiO_2 ratios than the Gympie basalts, implying substantial early-stage plagioclase fractionation in the Rocksberg magmas. Some of the lower-Ti Rocksberg basalts (which also have MgO > 10 wt%) may have accumulated mafic phenocrysts, as proposed for ankaramitic New Zealand IAT by Sivell & Rankin (1983). This is consistent with the presence of a high proportion of relict ferromagnesian phenocrysts, and pseudomorphs after large euhedral olivine and pyroxene crystals in the slightly to extensively recrystallised Rocksberg metabasites. The Cedarton (and Amamoor Type 2) basalts plot mainly in the elongate field of the Gympie eruptives. Amamoor Type 1 samples straddle the low Al_2O_3/TiO_2 Rocksberg trend.

Trace Elements

Figure 5 depicts Ba/Zr and Nb/Zr variation in the southeast Queensland metavolcanic suites. Highest

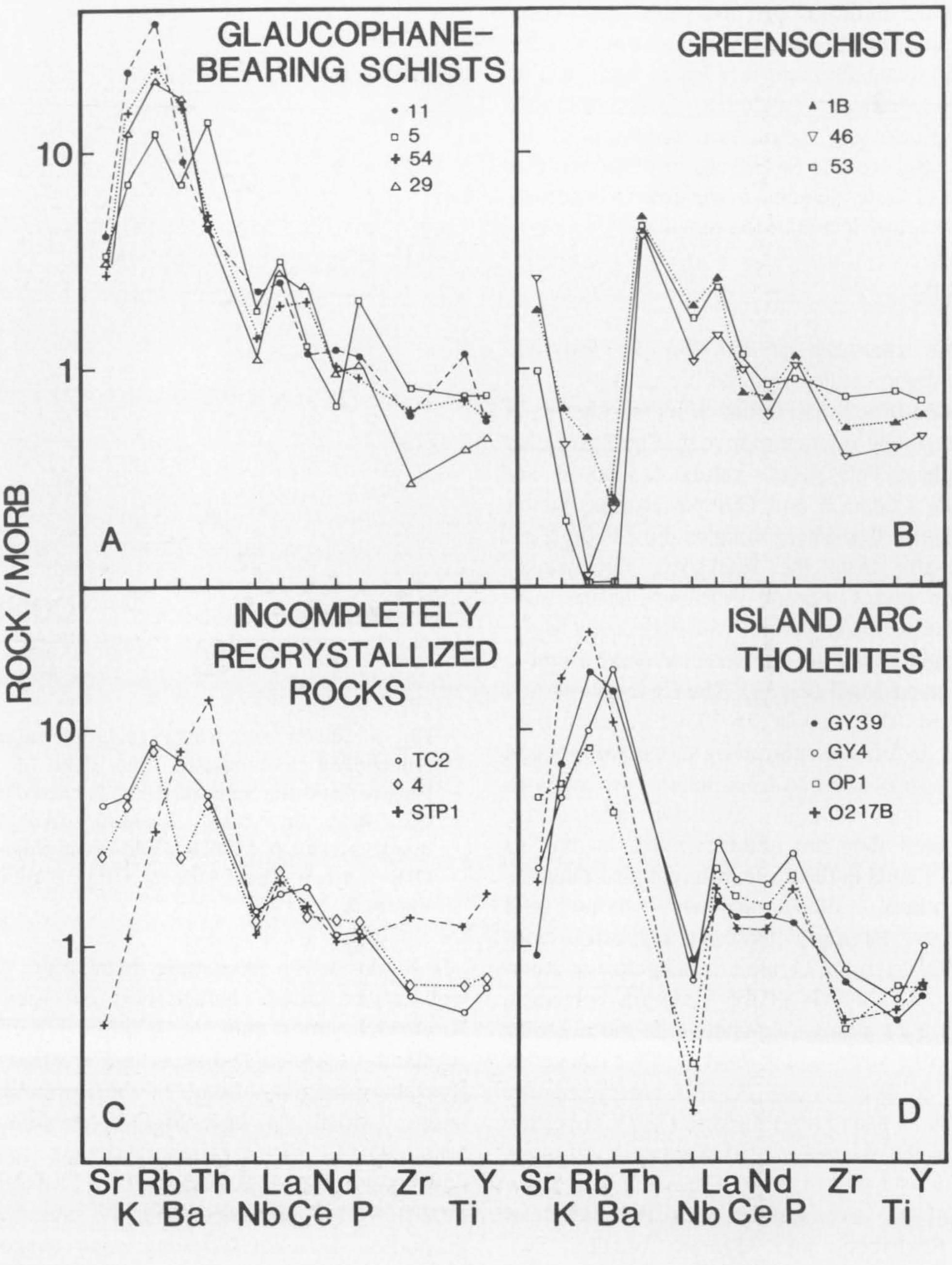

Fig. 6 Incompatible element 'spidergrams' for Rocksberg metabasites including: (A) - glaucophane-bearing schists; (B) - greenschists (highly mobile Ba, K, Rb and Th have been leached from these samples during pervasive greenschist facies overprint on early-formed transitional blueschist assemblages - Sivell, in press); and (C) - unrecrystallised or partly recrystallised basalts. Metabasalts from the Gympie Group (GY4, GY39), and Permian IAT from New Zealand (OP1, 0217B) are shown in Figure 6D.

Ba/Nb ratios (average Ba/Nb = 100) are shown by the Gympie basalts. These plot entirely within the field of New Zealand IAT and in a small area, which suggests that their high Ba/Nb values are not merely due to the mobility of Ba during metamorphism. The Cedarton Volcanics and Amamoor Type 2 basalts possess lower Ba/Nb ratios in the range 11 to 35. Amamoor Type 1 basalts and the majority of the Rocksberg metabasites (including both greenschists and transitional blueschists) have low (MORB-like) Ba/Nb < 10, although there is considerable scatter in the Rocksberg data.

MORB-normalised incompatible element 'spidergrams' for Rocksberg and Gympie basalts (Fig. 6) and Amamoor and Cedarton volcanics (Fig. 7) reveal a number of overall geochemical similarities between the different suites. These include low Zr and Y levels, and relatively high abundances of large ion lithophile elements (K, Rb and Ba), except in Rocksberg margin settings. However, the markedly contrasting Nb levels in basalts from these two suites imply substantially differing source components (see below).

Compared to the Gympie and Cedarton volcanics, the Amamoor Type 1 basalts have distinctly lower La_N/Y_N ratios (0.7-2.4), as well as lower Sr, La/Ce and La/Nb, and higher (MORB-like) Ti/Y. In particular, the P to Y portions of their incompatible element profiles do not have the trough-like shapes which characterise the "tails" of the trace element patterns for the Cedarton rocks. Some features of the Rocksberg metabasalts (e.g. their generally high Ti/Y ratios) are similar to Amamoor Type 1 samples. In other ways (e.g. high Sr and La_N/Y_N), they are more like the Cedarton Volcanics.

Amamoor Type 2 basalts have distinctive patterns characterised by significant depletions in elements from Ce through to Y, with flat patterns over this range. They have low La_N/Y_N (0.8-1.5) and La/Nb (< 0.3) ratios, and

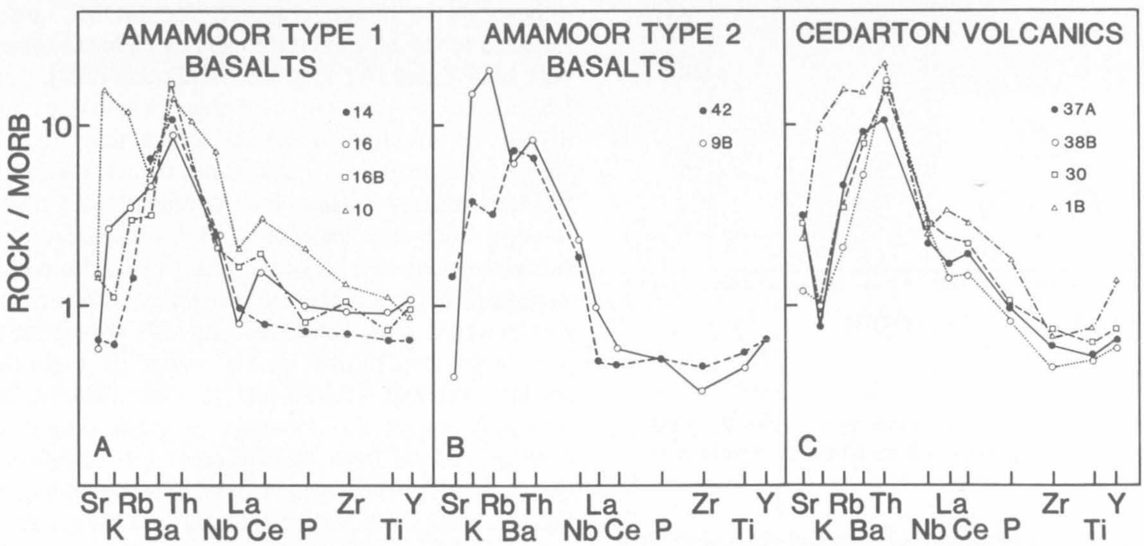

Fig. 7 Incompatible element 'spidergrams' for Amamoor and Cedarton basalts.

greenschists. Together with overall high Ba/Zr ratios (relative to MORB), these are features typical of subduction-related magmas (Perfit & others, 1980; Saunders & others, 1980).

Systematic differences also occur between incompatible element patterns for the various metabasite groups. Most importantly, the Rocksberg, Amamoor and Cedarton metabasalts do not possess the pronounced negative Nb anomalies of the Gympie rocks. The samples which have trace element patterns most closely resembling the Gympie IAT are the Cedarton Volcanics. Like the Gympie basalts, they possess high Ba/Nb, Ba/Zr and Sr, as well as particularly low Zr and Ti (and Ti/Y) that result in trough-shaped "tails" (i.e. P to Y segments) of their trace element profiles. In these respects, the Cedarton Volcanics and Gympie basalts share geochemical affinities with basalts from convergent low Sr contents like the Amamoor Type 1 basalts, but in other respects (higher Ba/Nb, Ba/Zr and La/Ce, and low Ti/Y) they more closely resemble the Cedarton volcanics.

Differences in a number of key incompatible element ratios for the Gympie Province metavolcanic suites are highlighted in Figure 8. On such biaxial plots, series derived from common parental magmas, and related by crystal fractionation, tend to maintain trends with relatively constant inter-element ratios (Treuil & Varet, 1973). Significant differences in observed ratios result in liquids produced by different degrees of partial melting, or by melting which involves variable source components (Wood, 1979). Importantly, the Gympie basalts possess lower Nb/Zr, Ti/Zr, Zr/P_2O_5 and Zr/Y ratios than most analysed samples from the other suites. These differences are explicable in terms of varying mantle source contributions to magmas generated in contrasting tectonic settings (see following section). On the Nb/Zr plot, Amamoor Type 2 samples and Cedarton basalts show distinctly higher Nb/Zr ratios than the Rocksberg and Amamoor Type 1 basalts.

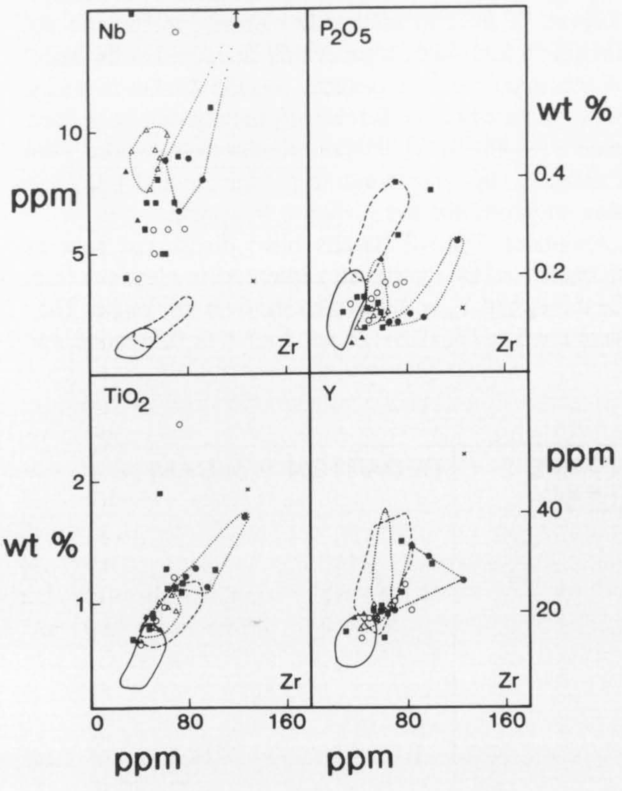

Fig. 8 Trace (and minor) element versus Zr plots for southeast Queensland metabasalts. Symbols as in Figure 2.

It is significant that the Cedarton and Amamoor Type 2 samples, although differing most from the Gympie rocks in terms of their Nb/Zr ratios, possess the most pronounced arc-like features (in common with the Gympie basalts) on trace element plots which discriminate between basalts from different tectonic settings. In the Ti/Cr - Ni diagram (Fig. 9), there is a clear distinction between the majority of Rocksberg and Amamoor Type 1 basalts, which plot in the field of ocean floor tholeiites, and most Gympie, Cedarton and Amamoor Type 2 basalts which fall in the IAT field. Similarly, all Amamoor Type 1 basalts, and many Rocksberg samples, possess the relatively high Ti/V ratios that characterise mid-ocean ridge basalts (MORB; Fig. 10). The Cedarton Volcanics closely overlap the region occupied by the Gympie basalts in the IAT field. Amamoor Type 2 samples show transitional character by plotting in the MORB field along with the Type 1 lavas.

SIGNIFICANCE OF GEOCHEMICAL RESULTS

Each of the Gympie Province metavolcanic suites shows some geochemical affinities with convergent plate margin basalts (overall high Ba/Zr, low Zr, Y, Ti/Y, etc.). These "arc signatures" are most pronounced for the highly porphyritic Gympie and Cedarton volcanics. Major differences between these two suites (particularly their different Nb/Zr, Zr/Y, Ti/Zr and Zr/P_2O_5 ratios) reflect different mantle source contributions to their subduction-related parent magmas, which were generated in contrasting tectonic regimes.

Beneath intra-oceanic island arcs, which may be far removed from any (sub-) continental lithospheric influence, the dominant magma source components will comprise (1) convecting upper mantle above the Benioff Zone; and (2) a large ion lithophile element (LILE)-enriched slab-derived contribution. The convecting upper mantle end-member is similar to the depleted asthenospheric source proposed for MORB, and is reflected in the depleted trace element characteristics of both MORB and IAT (e.g. low Nb/Zr and Zr/Y). Such features indicate the origin of these magmas by high degrees of fusion of a mantle source that has been strongly depleted in incompatible elements during a previous melting event. In convergent plate margin settings, LILE-enriched hydrous fluids, ascending from subducted oceanic lithosphere, may cause selective and variable incompatible element enrichment of the depleted mantle wedge above the subducting slab. However, the prior depletions of the mantle wedge in high field strength elements (Ti, Zr and Nb) are retained, and ultimately impart this signature to subduction-related magmas derived from these sources (cf. Saunders & others, 1980). The Gympie basalts, alone among the volcanic suites studied here, display both high Zr/Nb ratios (Zr/Nb $\approx$ 30) and pronounced negative Nb anomalies on incompatible element "spidergrams". They have been interpreted previously as IAT, generated during an immature stage of eruptive activity in an Early Permian intra-oceanic island arc (Waterhouse & Sivell, 1987; Sivell & Waterhouse, 1988). Geochemical features of these rocks are explicable essentially in terms of involvement of the two above-mentioned source components.

By contrast, the mantle wedge, which overlies the subducting oceanic plate beneath the magmatic arcs of active Cordilleran margins, may comprise both (asthenospheric) convecting upper mantle and geochemically distinct sub-continental lithospheric mantle (Pearce, 1984). Both of these mantle regions (in addition to a slab-derived subduction component) may contribute to the geochemical compositions of continental margin volcanic arc magmas. A significant sub-continental lithospheric mantle component is required to account for the low Zr/Nb ratios (Zr/Nb $\approx$ 5) observed for the Cedarton and Amamoor Type 2 basalts. This source component also influences Zr/Y, Ti/Zr and

Zr/P_2O_5 ratios which are high in these rocks compared to the Gympie IAT (cf. Pearce, 1984). Such sub-continental lithospheric mantle involvement indicates a continental margin subduction regime for the Cedarton Volcanics, in contrast to the Gympie rocks. For both the Gympie and Cedarton basalts, the bulk (60-90%) of their highly incompatible LILE content (including K, Rb and Ba) originates from a substantial slab-component. The asthenospheric mantle source end-member for both the Gympie and Cedarton volcanics was presumably a similarly depleted, convecting upper mantle region beneath the palaeo-Pacific basin.

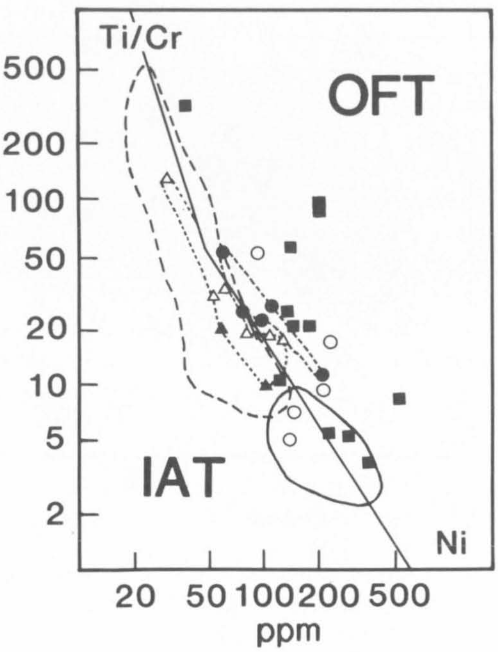

Fig. 9 Ti/Cr - Ni plot for southeast Queensland metabasites. Symbols as for Figure 2. Fields are OFT = ocean-floor tholeiites, and IAT = island arc tholeiites (after Beccaluva & others, 1979).

A diminished, although still significant sub-continental lithospheric mantle contribution is also required to explain the Zr/Nb ratios (Zr/Nb ≈ 10) of the Rocksberg and Amamoor Type 1 basalts, which are intermediate between those of the Cedarton and Gympie rocks. In addition, geochemical features of the Rocksberg and Amamoor basalts imply their evolution in more complex tectonic settings than the simple subduction systems advocated for the Gympie and Cedarton eruptives. Together with their apparent (but less pronounced) affinities with magmas from active continental margins, the Rocksberg, and particularly the Amamoor Type 1 basalts, exhibit a number of geochemical characteristics transitional to those of MORB. For a given MgO content, they possess lower La/Nb and Ba/Nb, as well as higher Ti/Y and Ti/V, than the Cedarton Volcanics. Furthermore, their lower Al_2O_3 contents and Al_2O_3/TiO_2 ratios imply significant plagioclase-controlled fractionation at an early stage of magma evolution. Such fractionation is common in magmas from rift-related tectonomagmatic environments (e.g. seafloor spreading ridges). In the case of the plagioclase+olivine - phyric Amamoor basalts, it is confirmed by observed phenocryst assemblages. Thus, geochemical and petrographic features of the Rocksberg and Amamoor basalts indicate that their generation accompanied local rifting within an overall continental margin subduction regime. This also explains the varying proportions of end-member source components inferred from the differing Zr/Nb and Ba/Nb ratios of the various metabasite suites. The initiation of local extensional tectonics within a regionally developed subduction system (e.g. inter- or intra-arc rifting) would facilitate the ascent to shallow depths of convecting asthenospheric mantle, thereby diminishing the opportunity for involvement in magma genesis of (high Nb/Zr, Zr/Y) sub-continental lithospheric mantle. As expected, the subduction component (high Ba/Nb, Ba/Zr) is also lowest in the rift-related Rocksberg and Amamoor Type 1 magmas. Enhanced flux of upwelling asthenosphere in the vicinity of the developing rift would reduce the amount of interaction of a given mantle volume with slab-derived fluids. High rate of magma generation would also favour the eruption of more primitive (i.e. less differentiated) magmas in the rift settings. We have already noted that the Amamoor and most Rocksberg basalts do not attain the high FeO_t /MgO ratios, which are typical of the evolved Cedarton and Gympie volcanics.

Finally, we draw attention to the fact that geochemical features of the Amamoor Type 2 basalts are in many respects transitional between the Cedarton and Amamoor Type 1 eruptives. Such features would appear to be most consistent with the generation of the Amamoor Type 2 magmas during an incipient stage of rifting in the overall subduction setting. Renewed partial fusion of mantle that had already yielded Cedarton-type melts may have been involved, in order to account for the severe depletions of elements like Ce, P, Zr, Ti and Y in these rocks. At this stage, their genesis remains somewhat equivocal.

IMPLICATIONS FOR SUSPECT TERRANES OF THE GYMPIE PROVINCE

Geochemical data presented in this paper imply substantially differing tectonic settings for the origin of a number of metavolcanic suites from the Gympie Province. Contrasting tectonic evolutions are inferred for their host terranes, subsequently juxtaposed at the Gondwana rim. The combined rift and convergent plate margin chemical affinities of the Rocksberg and Amamoor metabasalts imply their origin during episodes of rift-related magmatism in an overall Cordilleran-style subduction environment. There is no conclusive evidence for the age of the Rocksberg Greenstone protolith, but its minimum age is probably constrained to Early Carboniferous (Murray, 1988). In addition, radiolarians from chert lenses imply that the Amamoor beds and part of the Gympie Group in part comprise an Early Carboniferous seafloor sequence (Murray & others, 1989). The Amamoor beds basalts along the Sunday Creek road are most probably Permian. They appear to be part of a continuous sequence which contains Early Permian fossils at Cambroon. Both the Amamoor beds and North D'Aguilar Block (including the Rocksberg Greenstone) are suspect terranes which in part represent pre-Middle Carboniferous elements of a continental margin (probably Gondwanaland), but perhaps displaced from their original positions by strike-slip faulting along the continental edge, and affected by later crustal shortening within a Late Palaeozoic transpressional regime.

Metasomatised serpentinite masses form part of the boundary between the Amamoor beds and correlatives of the Rocksberg Greenstone in the North D'Aguilar Block. The ultramafic rocks were folded and metamorphosed at the end of the Carboniferous, together with the remainder of the North D'Aguilar Block. They were emplaced prior to collision of the Permian-Early Triassic portion of the Gympie Province (Gympie Group-Kin Kin beds; see below) with the continental margin in the Middle Triassic (cf. Harrington & Korsch, 1985). We suggest that the Early Carboniferous part of the Amamoor beds may already have been adjacent to the North D'Aguilar Block (in the vicinity of the Gondwanaland margin) by the end of the Carboniferous. The precise tectonic configuration of these two terranes, and their palaeo-geographic relation to the Gondwana rim are unknown, but geochemical features of the analysed metavolcanics indicate that they formed part of the evolving system of rift and arc segments that comprised the Cordilleran-type margin.

A continental margin subduction regime is also invoked for the origin of the Cedarton Volcanics. These crop out as an apparent southward continuation of the Amamoor beds and contain a (probable Early Permian) marine fauna similar to that from parts of the Amamoor beds. However, the dominant arc affinities of the Cedarton basalts imply a different tectonic setting to that proposed for the Amamoor succession. The Cedarton Volcanics are most likely the products of early-stage eruptive activity in a segment of a volcanic arc, proximal to or forming part of a continental margin. They may represent a discrete suspect terrane from the Amamoor beds.

Major geochemical differences exist between metabasalts from the Gympie Group and those from each of the other analysed metavolcanic suites, which share continental margin features. The mafic volcanic succession of the Gympie Group accumulated in an island arc environment (Sivell & Waterhouse, 1988), with predominant tuff-breccias representing an early submarine stage of volcanism. In contrast to the Rocksberg, Amamoor and Cedarton volcanics, the Gympie basalts were generated in an intra-oceanic region, remote from any continental influence upon their magma-generating mantle source. The shallow marine association of the Gympie Group-Kin Kin beds constitutes a discrete exotic terrane, which accreted to the Gondwanaland margin in Triassic time (cf. Harrington, 1983). Prior to docking of the Gympie 'arc' assemblage with the Early Permian and older terranes in the western part of Gympie Province, these suspect terranes may already have undergone considerable displacement along the continental edge. In contrast to the Gympie rocks, however, they were always proximal to the Gondwana rim.

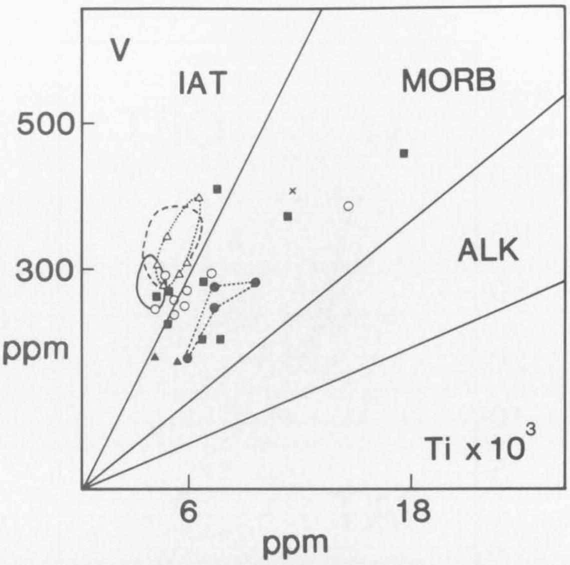

Fig. 10 Ti versus V plot for southeast Queensland metabasalts. Symbols as in Figure 2. Fields are IAT = island arc tholeiites; MORB = mid-ocean ridge basalts; and ALK = alkalic ocean island basalts (after Shervais, 1982).

The Gympie arc may have been partly coeval with the Camboon Volcanic Arc, which developed on the site of the former Connors-Auburn Volcanic Arc along the Gondwana margin from the Late Carboniferous to the Early Permian. Volcanics in the Camboon Volcanic Arc have given isotopic ages as young as 270 and 280 Ma (cf. Webb & McDougall, 1968). Thus, in the Early Permian, there may have existed two partly contemporaneous arcs: (1) the continental margin Camboon Volcanic Arc, and (2) the Gympie island arc. Tectonic translation of the Gympie arc towards the Gondwana rim, due to either subduction or strike-slip faulting, may have telescoped the facies of an inter-arc basin to the rear of the island arc. Geochemical data suggests that the Amamoor beds were deposited in a comparable setting, close to a continental margin. If the Amamoor accumulated in a marginal basin behind the Gympie arc, then this could explain the tectonic juxtaposition of older (Early Carboniferous) seafloor fragments with younger (Early Permian) sequences within this composite terrane (c.f. Murray & others, 1989).

The probable Early Permian fauna recognised in some strata from the Amamoor beds and Cedarton Volcanics indicates that these units at least partly overlap the Gympie Group in age (Murray & others, 1979). However, Amamoor and Cedarton metavolcanics are chemically distinct from the Gympie basalts, and share Cordilleran-margin affinities with the Rocksberg Greenstone. Also, whilst structurally dissimilar to the North D'Aguilar Block, the Amamoor beds are more complexly deformed than the Gympie Group (Murray, 1988). Neither the Amamoor beds nor the Cedarton Volcanics can be regarded as facies equivalents of the Gympie Group.

In conclusion, we highlight the fundamental geochemical distinction between Late Palaeozoic metavolcanic suites in the eastern and western parts of the Gympie Province, and the implications for the origin of their discrete, host suspect terranes. Each of the Early Permian or older metavolcanic sequences of the North D'Aguilar Block, Amamoor beds and Cedarton Volcanics, in the western part of Gympie Province, possesses affinities with Cordilleran-style continental margins. Their host terranes formed in close proximity to the Gondwana rim, although their inferred tectonic settings differ in detail (i.e. possess arc or rift characteristics), and they may have undergone substantial strike-slip displacement along the continental margin prior to Triassic docking of the Gympie arc. In contrast, the Gympie Group - Kin Kin beds sequence in the eastern Gympie Province, is part of an exotic terrane of intra-oceanic origin, requiring tectonic translation towards the Gondwanaland margin, with possible subsequent strike-slip displacement along the continental edge.

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BASIN AND CRUSTAL EVOLUTION ALONG THE EROMANGA-BRISBANE GEOSCIENCE TRANSECT: - PRECIS AND ANALOGUES

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"Our understanding of the way sedimentary basins form has advanced...(through) a new interdisciplinary approach to basin analysis, which recognises that processes in the lithosphere are ultimately responsible for basin subsidence. No single piece of evidence is diagnostic of a process,...it is only through a synthesis of what may initially appear to be disparate data that progress in our understanding is made."

C. Beaumont & A. J. Tankard, in "Sedimentary Basins and Basin-forming Mechanisms", Canadian Society of Petroleum Geologists Memoir 12, 1987.

INTRODUCTION

An important objective of the Eromanga-Brisbane Geoscience Transect across southern Queensland is to provide an impression of the processes within the crust throughout the geological history of the region which have contributed to the evolution of the region's sedimentary basins. Deep seismic profiling along a 1100 km transect across southern Queensland during 1980-86 has enabled, for the first time, a realistic three-dimensional interpretation of the crustal architecture relative to the near-surface geology. A transect study, such as this, emphasizes the depth extent of the events involved, rather than the palaeogeography. However, this paper summarises interpretations presented in BMR Bulletin 232 which certainly consider the regional geology of the whole of southern Queensland and northern New South Wales as a matter of course. Lateral displacements, strike-slip movements, uplift and subsidence are all appraised in the context of processes at depth throughout the crust. This paper gives a interpretational sketch of probable events within the crust down to depths of 50-60 km, and endeavours to identify modern-day analogues for the processes applying at any one time.

The Phanerozoic development of continental crust across southern Queensland is representative of that across the palaeo-Pacific margin of Gondwanaland. Basement geology in this region is only poorly understood because it is largely obscured by the extensive Mesozoic cover rocks of the Eromanga, Surat and Clarence-Moreton Basins (Fig. 1). Beneath the platform cover, the transect crosses, from west to east:

(1) the central Eromanga Basin and its infra-basins (middle Palaeozoic - Mesozoic) overlying the Thomson Fold Belt (early - middle Palaeozoic).

(2) the Surat Basin and Taroom Trough of the Bowen Basin (late Palaeozoic - Mesozoic) overlying northernmost Lachlan Fold Belt (early - middle Palaeozoic).

(3) the Esk Trough and Ipswich and Clarence-Moreton Basins (late Permian - Mesozoic) overlying the New England Fold Belt (middle Palaeozoic - Mesozoic).

In general, orogenesis became younger from west to east along the transect, reflecting progressive development of continental crust since the early Palaeozoic breakup of the palaeo-Pacific margin of the Australian craton established during the Precambrian. Figure 2 shows a development of the region in a space-time diagram (Murray, 1990) and Figure 3 the regional gravity trends and geological province boundaries (Wellman, 1990). The three major provinces listed above are considered as they developed through geological time. The processes involved in basin formation must be considered at all times and Figure 4 gives a summary of the lateral stresses likely in any basin-forming episodes (Gibbs, 1986). Figures 5 and 6 show, respectively, (a) the crustal architecture derived from seismic reflection profiling (Finlayson & others, 1990a), and (b) a cartoon of conceptual crustal developments during the Palaeozoic - early Mesozoic.

BASINS OVERLYING THE THOMSON FOLD BELT

The **Thomson Fold Belt** is believed to comprise mainly Cambrian and Ordovician quartzose turbidites deposited on thinned continental crust (marginal sea? quasi-continental crust?) southeast of prominent gravity and aeromagnetic lineaments, which mark the boundary with the thick (~50 km) Proterozoic crust of the North Australian Craton (Murray, 1990). Such developments were part of wider, late Proterozoic to early Palaeozoic

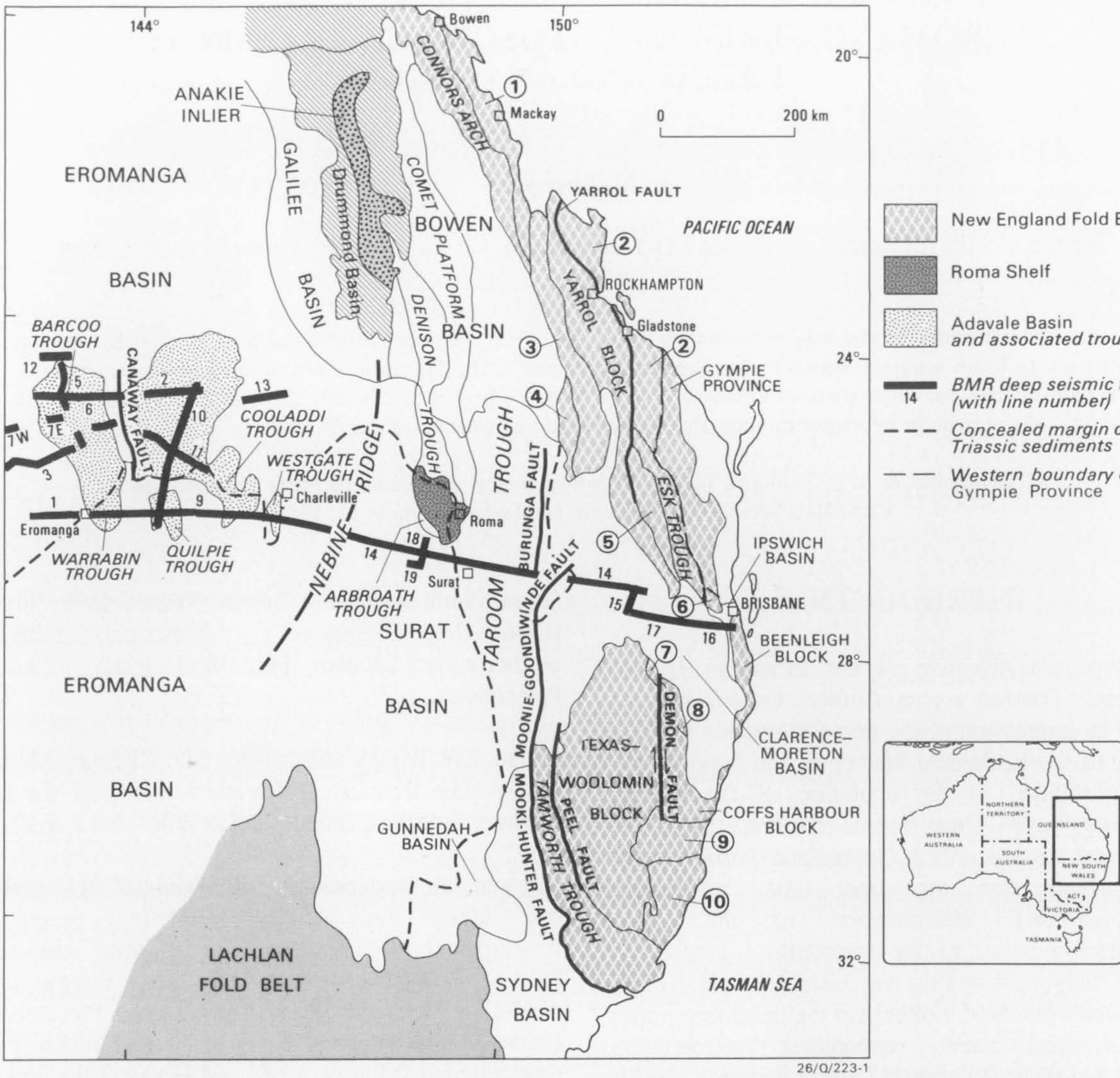

Fig. 1 Regional geology and seismic reflection profiles in southern Queensland. Numbered geological provinces are as follows: (1) Campwyn Block, (2) Coastal Block, (3) Gogango Overfolded Zone, (4) Auburn Arch, (5) Yarraman Block, (6) South D'Aguilar Block, (7) Silverwood Block, (8) Emu Creek Block, (9) Numbucca Block, and (10) Hastings Block.

rifting/breakup events postulated for eastern Australia. The detail of events during the fold belt's history is not well known, but the broad crustal framework was established in a period of over 150 Ma during the early Palaeozoic. Orogenic development terminated after Late Ordovician-to-Silurian plutonism, with deformation and uplift close to sea level by the beginning of Devonian time.

The processes affecting the Thomson Orogen were probably the same as those in younger (and present-day) orogens studied in more detail at the surface. It is thought that the general pattern of a seismically non-reflective upper-crustal fold belt overlying a reflective, more mafic, lower crust was probably established during the Late Ordovician - Silurian (Finlayson & others, 1990b). Compositional layering, and hence rheological zoning, at mid-crustal depths probably set the boundaries for major detachment surfaces during later upper crustal thrusting and ramping (Kusznir & others, 1987). Analogues of the crustal developments during this time might include the present-day North Atlantic continental margins of Europe and Canada. There, one generally sees a non-reflective upper crust overlying a reflective lower crust across much of the continental shelf, quasi-continental platforms (e.g. the Rockall Plateau), crustal attenuation (under the Viking Graben), isolated older massifs within the whole margin (e.g. the London-Brabant Massif), faults/shear zones extending through the crust (e.g. the Outer Isles Thrust), and other fault systems interpreted to detach at various levels within the crust (e.g. the Orphan Basin, eastern Canadian margin).

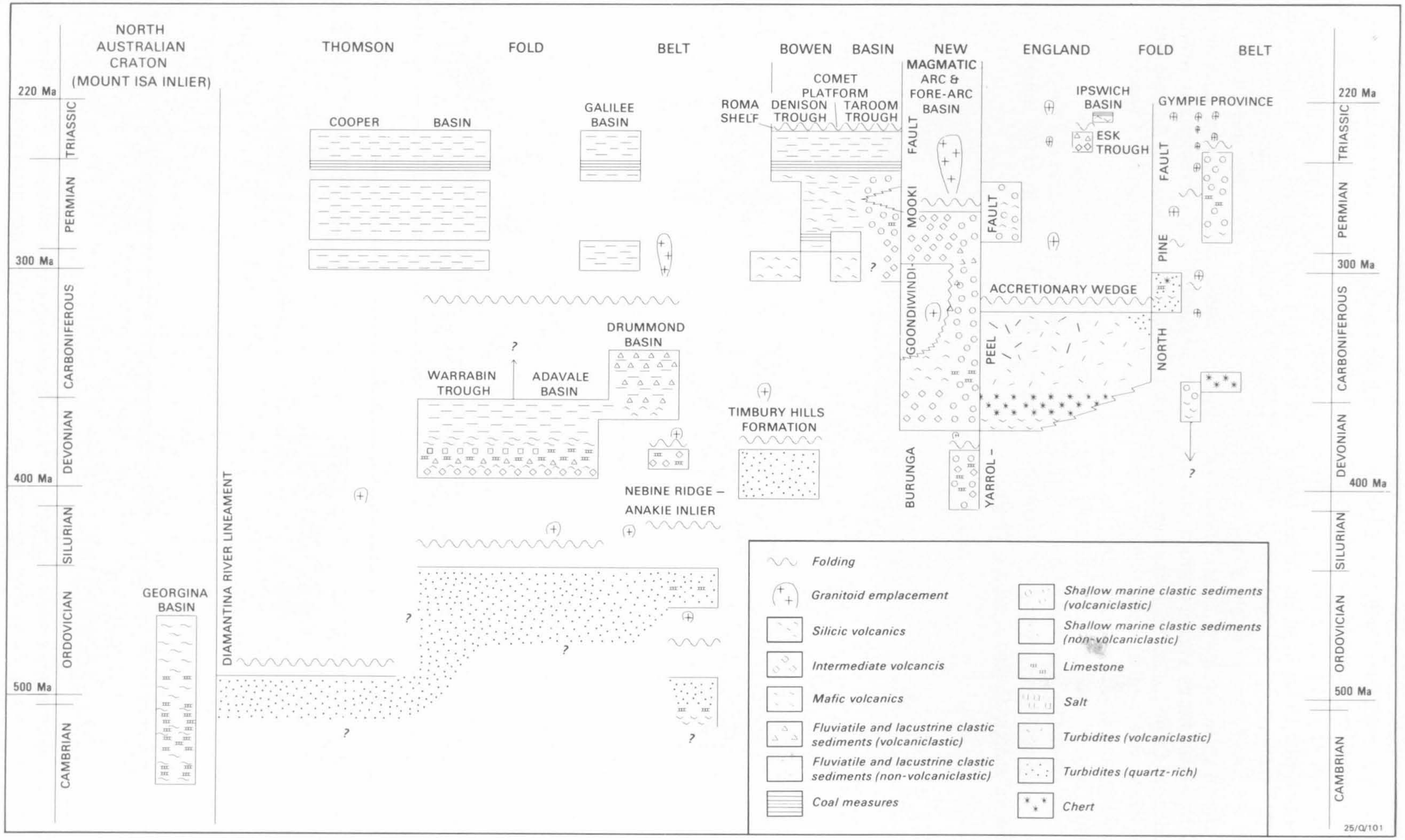

Fig. 2 Palaeozoic - early Mesozoic development of the major geological units along the Eromanga-Brisbane Geoscience Transect (from Murray, 1990).

The **Adavale Basin** is interpreted as an asymmetric rift-and-sag basin, which developed within the Thomson Orogen during Early Devonian intra-continental transtension involving pre-existing orthogonal fault systems (Evans & others, 1990). The Early Devonian tectonic regime has similarities with the Cainozoic transtensional rift corridors identified along the Indian - SE Asian plate boundary. An unconformity below the mid-Devonian Cooladdi Dolomite is interpreted to be a marker for the termination of transtension. Mafic intrusion into the lower crust probably added to early Palaeozoic reflectivity of the lower crust. There were Middle-Late Devonian facies changes which indicate restricted marine and fluvio-lacustrine environments over a wide area. Three stages of foreland thrusting, with increasing intensity, deformed the basin during the Late Devonian and Early to Middle Carboniferous. The evolution of the basin and its subsequent deformation are attributed to N-S sinistral shear along the Tasman Fold belt/paleao-Pacific margin (Evans & others, 1990).

The last transpressional event substantially deformed the Devonian Basins during what is called here the **Quilpie Orogeny** (Finlayson & others, 1990a). This was the last major orogenic event to affect the central Thomson Fold Belt. This event was contemporaneous with the Kanimblan Orogeny in southeastern Australian and the Alice Springs Orogeny in central Australia, all probably associated with the rapid poleward movement of eastern Gondwanaland at that time. In southern Queensland, upper crustal movements were detached from the lower crust by mid-crustal decollement surfaces; lateral relative movements of several tens of kilometres and vertical movements of several kilometres are interpreted in the southern Adavale Basin - Quilpie Trough area during the Quilpie Orogeny.

The mid-crustal ramping played a major part in uplifting basement structures and the deformation of the Adavale Basin into a series of troughs, the Quilpie,

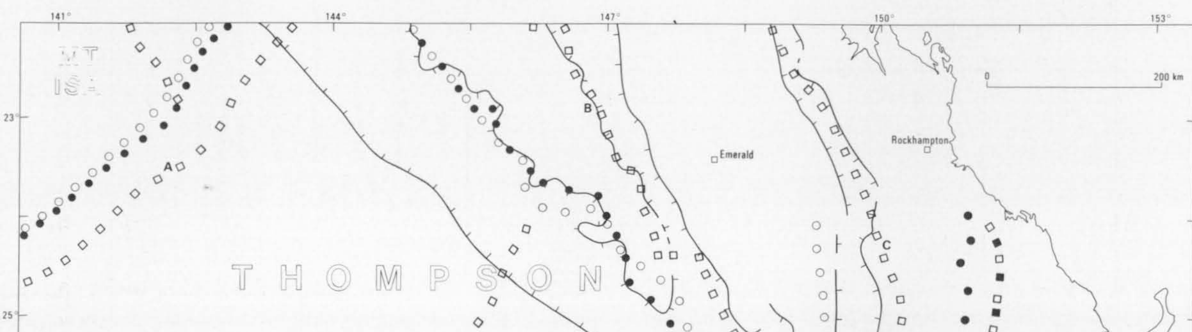

Fig. 3 Major crustal boundaries based on gravity and magnetic anomaly trends (Wellman, 1990). Thin lines show the extent of gravity anomaly highs. Rectangles delineate inferred block boundaries (open rectangles from gravity data, solid rectangles from magnetic data), circles show inferred margins of reworking (open circles form gravity data, solid circles from magnetic data). Medium thickness lines with ticks give the position of prominent magnetic gradients, or magnetic lows. The block boundaries are given letters, that are referred to by Wellman (1990). Numbers are points on the main east-west seismic reflection profile referred to by Wellman (1990). The line from 1 to 9 is the BMR seismic reflection traverse.

Cooladdi, Westgate and Warrabin Troughs, adjacent to the present-day Adavale Basin (Leven & others, 1990). Strike-slip motion is required to accommodate the envisaged deformation but, at present, its extent has not been clearly defined. The crustal thickness is now 36-39 km, determined by wide-angle seismic data and the termination of lower-crustal reflections above a non-reflective upper mantle. The rough topography, observed on the Moho at places under the Thomson Fold Belt, is considered to date from the Quiplie Orogeny (Finlayson & others, 1990b).

Analogues of the Early Devonian basin development might include the Basin and Range province of the Western Cordillera in western U.S.A. In this province, extension/trans-tension is thought to be related to episodic volcanism during the Cainozoic, which may be older than the current tectonic regime at the Pacific plate boundary some distance to the west. Magmatism and extension are considered to be intimately related, and lateral flow/ductile shearing in the lower crust is probably decoupled from events in the upper crust. The subsequent deformation of the Devonian sequences in central Queensland may have analogues in the Laramide Orogeny within the Western Cordillera, which corresponds roughly with the termination of subduction of the Farallon Plate on the U.S. Pacific margin, and is related to A-subduction and uplift (Bally & Oldow, 1984). It has been shown that much of the late Cretaceous - early Tertiary uplift is related to an orthogonal set of listric faults, which decouple at different levels in the crust. In the Wind River Mountains of Wyoming, horizontal and vertical displacements of over 20 and 10 km, respectively, are interpreted and elements of Precambrian basement have been brought to the surface.

The **Nebine Ridge** lies towards the southeast margin of the Thomson Fold Belt and has different seismic characteristics from those under the central Eromanga Basin to the west (Collins & Lock, 1990; Finlayson & others, 1990c). The crust under the ridge is the thickest in the region (maximum 44 km) and its velocity structure is not as clearly subdivided as that farther west. The non-reflective upper crust thins to a minimum of 6 km across the ridge and this is interpreted to indicate uplift and erosion of Thomson Fold Belt rocks. Geophysical data indicate plutonic subcrop under much of the Surat Basin cover rocks along the transect. On its eastern side, the non-reflective upper crust thickens (greater than 12 km) and this corresponds to a zone of crustal reworking defined from magnetic data. Basement rocks in this region comprise mainly highly deformed Devonian(?) metasediments of the Timbury Hills Formation intruded by Carboniferous granites (Roma Granites).

The crustal architecture of the Nebine Ridge contrasts it with the Thomson Fold Belt farther west and it may incorporate elements of Precambrian craton. It was uplifted by the Late Devonian - Early Carboniferous

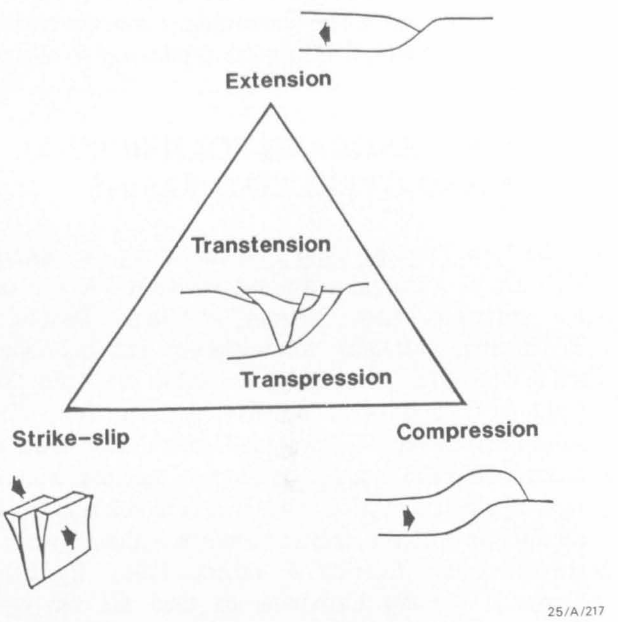

Fig. 4 Likely lateral stresses involved in basin-forming processes (from A. D. Gibbs, Structural Geology Course Notes, 1986).

crustal shortening events, which deformed the Devonian Adavale and Drummond Basins, the Quilpie Orogeny. The shallowing of intra-crustal reflections from west to east suggests that the crustal thickening under the ridge may have occurred by ramp-related movements (Finlayson & others, 1990b). The major orogenic events affecting the Thomson Fold Belt ceased by mid-Carboniferous times.

The eastern limit of the Thomson Fold Belt is identified by the termination of the upper crustal non-reflective region just south of the Arbroath Trough and a series of prominent west/southwest dipping reflectors extending to the lower crust, the **Foyleview Geosuture** (Finlayson & others, 1990b). The reflectors extend westward under the Nebine Ridge for about 100 kilometres. This geosuture corresponds, at its eastern end where it is shallowest, to major gravity and magnetic province boundaries. East of the geosuture, there is a significant increase in seismic P-wave velocity under the reflectors and also a change in the character of reflections within the crust.

The **Moho** under the Thomson Fold belt is defined by wide-angle reflection and refraction seismic data and corresponds to the extinction of prominent lower-crustal reflectors at depths greater than 36 to 44 km (increasing from west to east). There are cross-cutting reflections at the Moho giving the impression of a rugged topography; there are no large short-wavelength changes in Moho depth (constrained by gravity data) but it is thought that small-scale topograhy is present, a possible relic of Early to Middle Carboniferous Quilpie Orogeny. This feature probably implies that there has been no major post-orogenic geothermal event under the Thomson Fold Belt which would have re-mobilised the Moho since mid-Carboniferous times, contrasting it with the region east of the Foyleview Geosuture (Finlayson & others, 1990a).

NEW ENGLAND FOLD BELT AND OVERLYING BASINS

The **New England Fold Belt** lies along the eastern seaboard of Australia and is the youngest fold belt on the continent. Late Silurian - Middle Devonian calc-alkaline volcanics, volcaniclastics and limestones are interpreted as remnants of a volcanic island arc, **the Calliope Island Arc**, possibly separate from the Australian craton. In Devonian-Carboniferous time, a convergent plate margin developed, together with a west-dipping subduction zone; a western volcanic arc, a central fore-arc basin, and an eastern accretionary wedge are interpreted (Korsch & others, 1990). By Late Devonian - Early Carboniferous time the arc was considered to be developing on thin continental crust, so that by the Late Carboniferous an Andean-style margin had evolved (Sivell & others, 1990). Subduction had ceased by Late Carboniferous time and much of the arc-related tectonism had moved farther to the east by the Early Permian. The western bounds of the fold belt are now defined by the **Burunga-Mooki Fault System**, which separates it from the Taroom Trough of the Bowen Basin (Fig. 1).

The identification of a large double orocline/megafold in the central part of the fold belt, the **Texas - Coffs Harbour Oroclines**, has led to the interpretation of strike-slip faulting in the Late Carboniferous or Early Permian above a detachment surface within the crust (Murray, 1990). Subduction-related volcanism resumed in the northern part of the fold belt in Early Permian time, but not in the central or southern parts, which were in an extensional/trans-tensional back-arc setting with related volcanism. During the Permian and Triasssic, voluminous silicic magmatic activity produced granitoids and comagmatic continental volcanics; there was uplift of the fold belt at this time. Late Permian - Triassic basins developed within, or on the margins of the double orocline, the **Esk Tough, Ipswich Basin, and Clarence-Moreton Basin** (O'Brien & others, 1990). The final phase of orogenesis, which folded strata of the Mesozoic Maryborough Basin along the eastern extremity of the fold belt, was of Cretaceous age.

Deep seismic profiles cross the deformed fore-arc sequence just north of the Texas Orocline, the overlying transtensional Late Permian - Triassic basins within the orocline, and end on the Beenleigh Block, a possible exotic terrane. There are mid-crustal reflection horizons and velocity features at about 20 km depth interpreted to be decoupling surfaces for oroclinal bending, the **Texas and Brisbane Mid-crustal Detachments** (Finlayson & others, 1990b). An analogue of oroclinal deformation on a similar scale is evident in the megafold structures across the Straits of Gibraltar, the Alboran Sea/Gibraltar orocline.

In the upper/middle crust under the Texas Orocline, there are reflection features which may be a detachment surface related to Early - Middle Triassic compression/transpression on the Moonie Fault (Finlayson & Fielding, 1990). Above the Brisbane Mid-Crustal Detachment, near the eastern end of the seismic line, a prominent series of west-dipping reflectors are interpreted to be an imbricate thrust stack of sedimentary rocks (Korsch & others, 1986), the **Greenbank Deep Layered Sequence** (Finlayson & others, 1990b), underlying the Beenleigh Block and Ipswich Basin. Possible analogues include the deep layered sequences identified under Vancouver Island resulting from subduction processes.

In the central part of the fold belt, the seismic data image the structures under the Esk Trough, South Moreton Anticline, and Ipswich Basin. The basin architecture is interpreted to indicate a trans-tensional origin for these basins followed by a sag phase. Also in the central part of the fold belt, the seismic line crosses the Tertiary **Main Range Volcanics**, which are a source of xenoliths from the crust and upper mantle (O'Reilly & Griffin, 1990; Sutherland & others, 1990). Lower crustal rocks comprise dominantly mafic granulites, and upper mantle rocks mainly spinel lherzolites. Abundant, subhorizontal basaltic intrusives are interpreted in the lower crust, suggesting a geochemical "crust/mantle boundary" at depths shallower than the seismic Moho (O'Reilly & Griffin, 1990).

The **Moho** across the central part of the New England Fold Belt is defined from seismic data at 36 km depth. There is a band of reflectors at this depth right across the fold belt. The lack of Moho topography suggests that post-orogenic geothermal events reset the Moho at its present level (Finlayson & others, 1990b). In this respect, it may have analogues in the Basin and Range Province of western U.S.A.

SURAT BASIN AND TAROOM TROUGH OF THE BOWEN BASIN

The **Bowen-Gunnedah-Sydney Basin System** extends along the entire western margin of the New England Fold Belt. The Bowen Basin has three main structural

Fig. 5 (foldout page opposite) Interpreted cross-section to 60 km depth along the Eromanga-Brisbane Geoscience Transect based on integrated geophysical and geological data. The section is curved to the proportional Earth's radius.

EROMANGA–BRISBANE GEOSCIENCE TRANSECT

SEISMIC PROFILES: Digitized line diagram of significant reflectors

SEISMIC PROFILES: Migrated line diagram of reflectors; velocity/depth interpretations

INTERPRETED CRUSTAL SECTION

Scale $\frac{V}{H}$ = 1 approximately

25/A/218

0 100 200 Kilometres

Lambert Conformal Conic Projection with Standard Parallels at 25°40'S and 28°20'S

Seismic Profiles: Explanatory Notes

The deep seismic reflection profiles illustrated on this mapsheet are plots of digitized reflection events identified on full-scale seismic sections. All significant reflection events are plotted as a time section in the uppermost diagram, thus providing a good representation of the original seismic data. These data were migrated using a velocity of 6 km/s (an average crustal velocity) and re-plotted in the lower seismic profile to give an indication of the true position of dipping horizons on a depth section. This migration process assumes that the dipping events arise from geological features that are within a vertical plane under the seismic traverse. In many cases this assumption may not be valid, with reflecting horizons within basement on either side of the traverse contributing to the recorded signal. Reflection profiles were commonly recorded as 6-fold common mid-point data using explosive sources.

The velocity/depth functions illustrated in the lower seimic profile are derived from wide-angle reflection and refraction data. In all cases, except in the Taroom Trough, data were recorded coincident with, or intersecting, the reflection traverses. In the Taroom Trough, velocity/depth data are from an along-strike profile about 250 km farther north, still within the trough. Data were commonly recorded at 5 km intervals out to distances of 150-300 km from the sources; both travel-times and signal amplitudes were modelled to derive velocity/depth functions.

Interpreted Crustal Section: Explanatory Notes

The interpreted section illustrates the major geological features considered to be significant at depths down to 60 km. It is based on interpretations of seismic reflection and velocity data, regional gravity and magnetic trends and models, and the tectonic processes considered to apply throughout southern Queensland based on the structural interpretation of near-surface geology. Mesozoic cover sequences (from west to east) of the Eromanga, Surat and Clarence-Moreton Basins conceal most of the basement features along the transect. The structural geology of the major rock units within the underlying Thomson, Lachlan and New England Fold Belts has therefore been inferred from drillcore and geophysical data. The interpreted crustal section integrates available geological and geophysical information to provide an interpretational sketch of the crust across Phanerozoic eastern Australia, including ideas on the possible processes applying during its geological history.

Details of the interpretation are contained in papers contributed to Australian Bureau of Mineral Resources, Geology and Geophysics Bulletin 232.

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CANBERRA, AUSTRALIA
1989

provinces: the **Denison Trough** in the northwest, the **Taroom Trough** in the east, and the central **Comet Platform** (Fig. 1). The basin is interpreted in the general sense as a foreland basin to the New England Fold Belt, with extensional episodes during its early history, possibly associated with strike-slip faulting in a transtensional environment (Fielding & others, 1990). The significance of the Meandarra Gravity Ridge down the centre of the Taroom Trough has yet to be fully explored, but mafic intrusion into the crust is throught likely during a limited, predominantly extensional episode early in the basin's evolution. The history of the Denison and Arbroath Troughs on the western side of the Bowen Basin indicates an Early Permian extensional episode. There is evidence of a thick Carboniferous volcaniclastic/sedimentary sequence underlying the Permian near the Moonie Fault, but the pre-Permian history is not well known; the fault/fold strutures in that area, however, strongly suggest an early sinistral strike-slip movement controlling later deposition. The Bowen Basin as a whole consists of a series of depocentres in which subsidence varied with time, and which received thick sequences of continental and marine sediments, including enormous coal resources, from the beginning of the Permian to Middle Triassic time. Intensity of deformation throughout the Bowen Basin is variable, but, in general, increases from west to east.

Deep seismic profiles cross the **Taroom Trough** of the Bowen Basin from just south of the Arbroath Trough to the Moonie Fault. The western boundary is taken to be the Foyleview Geosuture at the eastern margin of the Thomson Fold Belt. The intra-crustal reflectors and velocity structure differs from that under the Thomson and New England Fold Belts (Finlayson & others, 1990b). There are no prominent horizons identified with velocity changes that might form significant intra-crustal boundaries. Extension is thought to have occurred by a distributed shear mechanism in the lower crust, rather than on a single detachment surface. There is a prominent Moho band of reflectors similar to that under the New England Fold Belt. The Moho east of the Foyleview Geosuture (36-39 km) undulates only slightly at about the same depth as that under the fold belt with no offsets, suggesting that it too has been reset by post-orogenic (Cainozoic?) geothermal events (underplating?).

The eastern margin of the basin was deformed and uplifted along the **Burunga-Mooki Fault System (Geosuture),** during resurgence of volcanism on the palaeo-Pacific margin beginning in the Late Permian and terminating in Middle Triassic times (Korsch & others, 1990). This phase is interpreted as a compressional/transpressional episode which involved

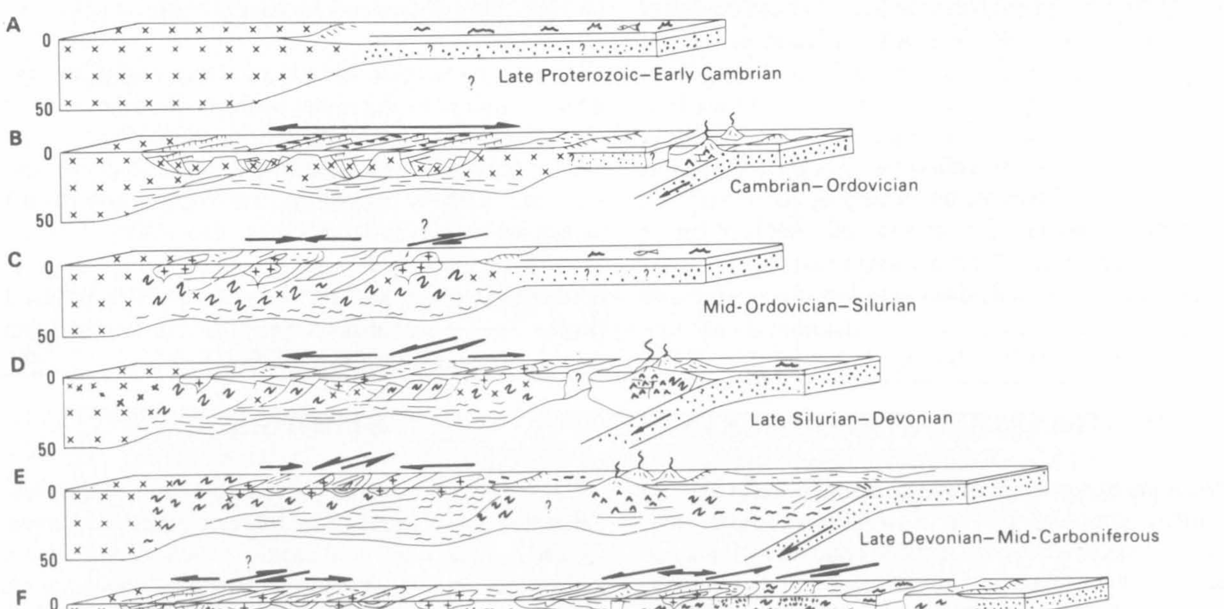

Fig. 6 Interpretational sketch of possible crustal developments along the Eromanga-Brisbane Geoscience Transect in southern Queensland (Precambrian to Middle Triassic).

crustal loading and basin subsidence (Murray, 1990). There are differences in the seismic fabric within the crust across this Burunga-Mooki Geosuture, indicating the juxtaposition of major crustal terranes.

"Most models dealing with convergent-boundary tectonism are two-dimensional and commonly neglect the strike-slip component of relative plate-motions observed in virtually all convergent boundaries. Assessment of the nature and distribution of structures associated with strike-slip displacements in the upper plate are necessary for a realistic understanding of the tectonism of these important regimes" (Bally & Oldow, 1984).

Overall, one should not be surprised by extensional structures in the back-arc environment near convergent boundary settings. In the Bowen Basin region there must have been a significant shear stress, at least during the Carboniferous development of oroclines in the adjacent New England Fold Belt, which would have affected the continental interior.

Present-day convergent plate margins (e.g. North and South America, New Zealand-Kermadec region, Summatra-Java region) indicate that compressional/trans-pressional, strike-slip, and extensional/transtensional features can all exist on the craton side of the margin, depending on the rate and obliquity of convergence. If a general foreland model of basin development is considered, analogues of the Taroom Trough might include the peri-sutural basins in the American continent (Bally & Oldow, 1984), e.g. Alberta Basin in western Canada adjacent to the Rocky Mountains fold-thrust belt; foreland basins eastward of the Andean fold-thrust belt. All are on the craton side of megasutures associated with active ocean/continent collisional margins. However, there are rifting antecedents underlying the Palaeozoic sequences in western Canada and other foreland basins (Appalachian Basin). Such analogues may, therefore, be oversimplifications and other models involving a number of mechanisms should not be discounted for the Bowen-Gunedah-Sydney Basins system.

POST-TRIASSIC DEVELOPMENTS

With the exception of the Maryborough Basin and the northern extremity of the New England Fold Belt, eastern Australia has been a stable cratonic region since the end of the Triassic. At this time the **Eromanga, Surat, and Clarence-Moreton Basins** developed as broad intra-cratonic sags (Gallagher, 1990). Strata of these basins are essentially undeformed and cover basement sequences along all but the easternmost extremity of the transect (Fig. 1). A model for this areally extensive subsidence relates it to a deep thermally based mechanism (sub-lithospheric convection related to subduction at the plate margins?). A rapid subsidence phase in the Cretaceous is attributed to an excess surface load resulting from a large influx of volcanogenically derived detritus from the east at a time of high sea level. **Tertiary reactivation of older fault systems** affected large parts of Australia (associated with the northward movement of the continent towards southeast Asia), and is evident in the Mesozoic sequences in southern Queensland (Finlayson & others, 1988).

The most significant Tertiary tectonism was east of the transect area with the formation of narrow, relatively deep graben along the eastern margin of the continent during opening of the **Tasman and Coral Sea basins** in the Paleocene and Eocene. This rifting can be separated into four phases: (1) pre-rift, (2) rift development, (3) seafloor spreading, and (4) quiescent phase (Shaw, 1990). During the Jurassic, the proto-Pacific-Australian plate boundary was located east of the present Australian margin and extended from New Guinea to the Transantarctic Mountains. Rifting of the Tasman Sea commenced at 140 M.a.; seafloor spreading began at about 85 M.a. in the south and 63 M.a. near the transect; spreading ceased about 55 M.a.(early Eocene). East of the transect the rifted margin is now recognised in the Dampier Ridge and Lord Howe Rise.

In mid- to late Cainozoic time large volumes of continental flood basalts, including the Main Range Volcanics, were erupted over the New England Fold Belt and extended as far west as the Anakie Inlier. Eruption of these basalts has been attributed to passage of the Australian continent over hot spots in the mantle during northward movement away from Antarctica (O'Reilly & Griffin, 1990; Sutherland & others, 1990).

Present-day tectonic activity affecting basins and the adjacent fold belts is evident in the earthquake data for southeast Queensland (Cuthbertson, 1990). Much of the current activity is associated with the New England Fold Belt and offshore region, but the Wunger (basement) Ridge under the Surat Basin is also active. Seismic activity is associated with the Esk Trough and its northern extension, and also the western New England Block. Focal mechanisms suggest reverse faulting as a result of northeast-southwest stress.

SYNOPSIS

The basins of southern Queensland formed in mid-Palaeozoic to Mesozoic times following the deformation of early Palaeozoic fold belts, which (?)may have anticedents in an attenuated/extended Precambrian Australian craton. They developed on the continental side of a predominantly convergent palaeo-Pacific margin with the rate and obliquity of convergence at various times affecting the style and extent of basin evolution. Basement structures in the Thomson, Lachlan and New England Fold Belts, the deep crust, and lithosphere had a major influence on the depositional trends and structures developed in the basins. Movement on both shallow-angle and high-angle boundaries/faults in the crust and (?)deeper lithosphere have played a major role in determining the present-day basin architecture. Tectonic activity in the region continues to the present-day.

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EROMANGA–BRISBANE GEOSCIENCE TRANSECT

ICL Global Geoscience Transect No. 7

MAP 1: GEOLOGY

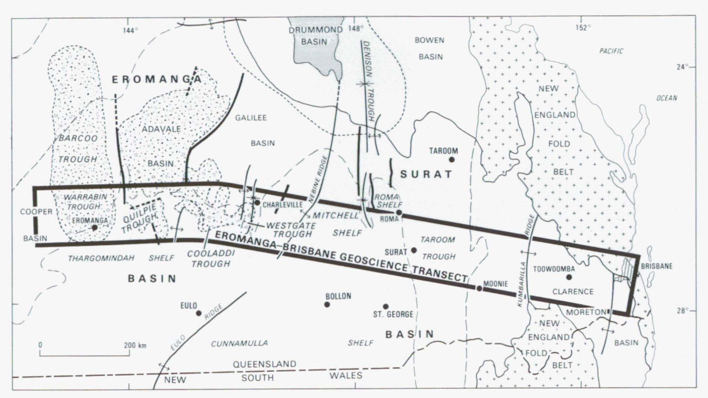
EASTERN AUSTRALIA

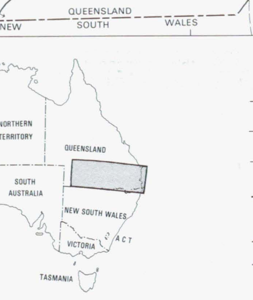
Scale 1:1 000 000

0 100 200 Kilometres

Lambert Conformal Conic Projection with Standard Parallels at 25°40′S and 28°20′S

Locality Maps





Margin of Eromanga, Surat and Clarence — Moreton Basins (Jurassic — Cretaceous)

Concealed margin of Cooper, Galilee and Bowen Basins (Permo — Triassic)

Concealed margin of Drummond and Adavale Basins and associated troughs. (Devonian–Carboniferous)

Fault (where location of fault is approximate, line is broken; where concealed, faults are shown as short dashes)

Anticline

Syncline

Limits of Permian basin sequences

Limits of Devonian basin sequences

Sub-surface unit boundary (?-interpreted from geophysics)

Normal fault. Square on downthrown block

Thrust fault. Triangle on upthrown block

Fault showing relative displacement

Geological boundary. Post-Triassic

RJbw Stratigraphic letter symbol. Post-Triassic

Geological boundary. Pre-Jurassic

PRn4 Stratigraphic letter symbol. Pre-Jurrasic

Orient 1 Drill hole

Geology: Explanatory Notes

The surface geology at 1:1 000 000 scale within the area 25° to 29° South, 142° to 154° East has been digitized and compiled from four mapped regions; the Eromanga Basin (Senior & others, 1978), the Surat Basin (Exon, 1976), the Clarence-Moreton Basin (Whitaker & Green, 1980) and the Maryborough mapsheet area (Queensland Department of Resource Industries, 1989). The references for these four mapped regions (listed below) should be consulted for detailed information. In some places, recent mapping information has been incorporated, and there has been some simplification of 1:250 000 and 1:500 000 scale mapping. Surface geology over much of the mapsheet area is dominated by the Jurassic-Cretaceous basin sequences, Tertiary volcanic sequences and Quaternary sedimentary deposits. Since the main purpose of this map is to highlight basement geological controls on basin evolution, much of the surface geology is shown only as grey-tone background information. Only selected surface geological units and formations have been annotated, as indicated in the legend.

Pre-Jurassic rock units crop out in the eastern part of the mapsheet area; these units are highlighted in bold line-work. Sub-surface basement geology is derived from hydrocarbon and water exploration drillholes. Not all drillholes penetrated basement; where they did, some indication of geology is shown according to the attached legend. Because there are numerous drillholes in some areas prospective for oil and gas, only selected holes are shown on this mapsheet; unnamed drillholes are usually water bores. The structural information under cover rocks is derived from extensive seismic reflection profiling

Many geological references are contained in papers in the Australian Bureau of Mineral Resources, Geology & Geophysics Bulletin 232, which this mapsheet accompanies. General references include papers by Finlayson & others (1988) and Elliott & Brown (1988)

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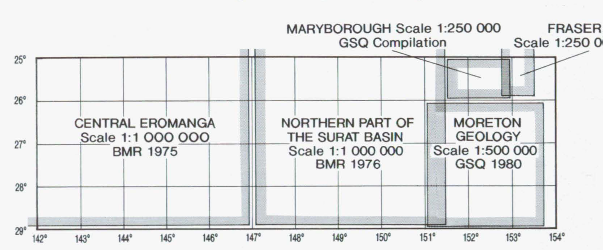
Geological Map Coverage



CAINOZOIC

QUATERNARY

- Q — *Alluvium of older flood plains, gravel, soil*
- Qb — *Beach systems, dunes, ridges, quartz sand and heavy mineral sands*
- Qa — *Alluvium of present-day flood plains*
- Qt — *Estuarine flood plain tide delta deposites*
- Qpb — BARAMBAH BASALT.
- Cz — *Quartzose sandstone, conglomerate, sandstone, siltstone and shale*
- Czs — *Poorly consolidated sandstone, conglomerate, siltstone, mudstone*

TERTIARY

- Tc — CHINCHILLA SAND: *Sandstone*
- Tg — GLENDOWER FORMATION: *Quartzose sandstone, sandy conglomerate, quartz pebble conglomerate, minor siltstone, silcrete*
- Ts — *Sandstone, conglomerate*
- Tx — OXLEY GROUP: *Claystone, sandstone, shale*
- Te — ELLIOT FORMATION: *Sandstone, conglomerate*
- Tmb — LAMINGTON GROUP/MAIN RANGE VOLCANICS: *Basalt flows, pyroclastics, basalt, rhyolite, agglomerate*
- Tob — *Basalt flows, minor trachyte and pyroclastics*
- Tb — *Basalt flows, minor pyroclastics*
- Tt — TAKURA BEDS: *Conglomerate and breccia*
- T — *Gabbro intrusion*

MESOZOIC

CRETACEOUS

- Kw — WINTON FORMATION: *Sandstone, siltstone*
- Klm — MACKUNDA FORMATION: *Sandstone, siltstone, mudstone, conquinite*
- Klmg — MOOGA SANDSTONE
- Klg — GRIMAN CREEK FORMATION: *Sandstone, siltstone*
- Kls — SURAT SILTSTONE
- Kla — ALLURA MUDSTONE
- Klc — COREENA MEMBER (WALLUMBILLA FORMATION): *Siltstone, sandstone*
- Kld — DONCASTER MEMBER (WALLUMBILLA FORMATION): *Carbonaceous mudstone*
- Kly — BUNGIL FORMATION: *Sandstone, siltstone*
- Kb — BURRUM COAL MEASURES: *Sandstone, siltstone*
- Km — MARYBOROUGH FORMATION: *Mudstone, siltstone*

JURASSIC-CRETACEOUS

- JKs — SOUTHLANDS FORMATION: *Sandstone*
- JKh — HOORAY SANDSTONE
- JKk — KUMBARILLA BEDS: *Sandstone*
- JKw — KANGAROO CREEK SANDSTONE/WOODENBONG BEDS: *Sandstone, siltstone*
- JKg — GRAFTON FORMATION: *Sandstone*
- JKi — *Quartz diorite, microdiorite*
- JKu — MOUNT URAH GRANODIORITE
- JKp — GRAHAMS CREEK FORMATION: *Intermediate to acid flows*

JURASSIC

- Jmc — MULGILDIE COAL MEASURES: *Mudstone, siltstone, coal*
- Juo — ORALLO FORMATION: *Sandstone*
- Jug — GUBBERAMUNDA SANDSTONE
- Ji — INJUNE CREEK GROUP (Undivided)
- Juw — WESTBOURNE FORMATION: *Siltstone* (INJUNE CREEK GROUP)
- Ja — ADORI SANDSTONE (INJUNE CREEK GROUP)
- Js — SPRINGBOK SANDSTONE (INJUNE CREEK GROUP)
- Jmb — BIRKHEAD FORMATION: *Carbonaceous mudstone* (INJUNE CREEK GROUP)
- Jw — WALLOON COAL MEASURES *Carbonaceous mudstone* (INJUNE CREEK GROUP)
- Jt — TIARO COAL MEASURES (DUCKINWILLA GROUP): *Sandstone, siltstone*
- Jp — PRECIPICE SANDSTONE
- Jlh — HUTTON SANDSTONE
- Jle — EVERGREEN FORMATION: *Siltstone, mudstone*
- Jlm — MARBURG FORMATION: *Sandstone*

MESOZOIC

TRIASSIC-JURASSIC

- RJb — BUNDAMBA GROUP (Undivided): *Sandstone, siltstone*
- RJbh — HELIDON SANDSTONE
- RJbw — WOOGAROO SUBGROUP: *Sandstone*
- RJl — LANDSBOROUGH SANDSTONE
- RJm — MYRTLE CREEK SANDSTONE (DUCKINWILLA GROUP)

TRIASSIC

- R — *Acid flows and pyroclastics*
- Rm — MOOLAYEMBER FORMATION: *Sandstone*
- Re — CLEMATIS GROUP: *Sandstone*
- Rlr — REWAN GROUP: *Rhyolite*
- Ra — ARANBANGA VOLCANIC GROUP: *Andesitic to rhyolitic flows*
- Rbc — BRISBANE TUFF/CHILLINGHAM VOLCANICS
- Rk — KIN KIN BEDS/KEEFTON FORMATION/TRAVESTON FORMATION: *Sandstone, phyllite*
- Ri — IPSWICH COAL MEASURES: *Shale, conglomerate*
- Rmb — MOUNT BYRON VOLCANICS: *Rhyolite, basalt, andesite*
- Rn — NORTH ARM VOLCANICS: *Andesite, tuff, rhyolite*
- Rr — TARONG BEDS: *Sandstone, shale*
- Rt — TOOGOOLAWAH GROUP (Undivided) (TOOGOOLAWAH GROUP)
- Rtb — BRYDEN FORMATION: *Conglomerate, shale* (TOOGOOLAWAH GROUP)
- Rte — ESK FORMATION: *Shale, sandstone* (TOOGOOLAWAH GROUP)
- Rtn — NEARA VOLCANICS (TOOGOOLAWAH GROUP)
- Rtg — GAYNDAH FORMATION: *Sandstone* (TOOGOOLAWAH GROUP)
- Rd — DEGILBO GRANITE
- Rf — BROOMFIELD GRANITE
- Rh — HOGBACK GRANITE
- Rgm — MUNGORE GRANITE
- Rs — MUSKET FLAT GRANITE
- Ru — BOOGOORAMUNYA GRANITE
- Rv — *Rhyolite, andesite, basalt*
- Rw — WONBAH GRANODIORITE
- Rz — TAWAH GRANITE
- Rg — *Granite, granodiorite*
- Rca — CALGOA DIORITE
- Rbf — BROOWEENA FORMATION: *Sandstone, siltstone*
- Rgb — BOONARA GRANODIORITE
- Rg1 — *Acid intrusives*
- Rg2 — STATION CREEK GRANODIORITE
- Rg3 — BRUXNER ADAMELLITE
- Rg4 — CULLENS CREEK GRANITE
- Rg5 — DAYBORO TONALITE
- Rg6 — EERWAHVALE TONALITE
- Rg7 — ENOGGERA GRANITE
- Rg8 — HERRIES ADAMELLITE
- Rg9 — GOOMBOORIAN DIORITE
- Rg10 — DJUAN TONALITE
- Rg11 — KARANA QUARTZ DIORITE
- Rg12 — GREYMARE GRANODIORITE
- Rg13 — MAIN CAMP SYENITE
- Rg14 — NEURUM TONALITE
- Rg15 — MOUNT SAMSON GRANODIORITE
- Rg16 — SOMERSET DAM IGNEOUS COMPLEX
- Rg17 — SAMFORD GRANODIORITE
- Rg18 — TOONDAHRA GRANITE
- Rg19 — TUNGI CREEK GRANODIORITE
- Rg20 — WOONDUM GRANITE
- Rg21 — MORGANS CREEK ADAMELLITE
- Rg22 — *Gabbro*
- Rms — MOUNT SAUL ADAMELLITE
- Rme — MOUNT EAGLE BEDS: *Rhyolitic and dacitic flows*
- Rmg — BOOLGAL GRANITE
- Rmc — CYNTHIA BEDS: *Boulder conglomerate, arenite*
- Rn1 — BLACK SNAKE CREEK GRANITE
- Rn2 — KOREELAN CREEK GRANODIORITE
- Rn3 — RIVERTREE GRANITE
- Rn4 — STANTHORPE ADAMELLITE
- Rn5 — UNDERCLIFFE FALLS ADAMELLITE
- Rn6 — RUBY CREEK ADAMELLITE
- Rn7 — GREENBANK QUARTZ DIORITE/NOUR NOUR GRANODIORITE
- Rn8 — NEW MOONTA DIORITE

MESOZOIC

PERMIAN-TRIASSIC

- PRg — GILLA ANDESITE
- PRca — BELLTHORPE ANDESITE
- PRa — *Andesite, rhyolite volcanics*
- PRb — BRIGGS GRANODIORITE
- PRw — WATERANGA GABBRO
- PRy — YENDA GRANODIORITE
- PRg1 — *Acid to intermediate intrusives, granite, granodiorite, diorite*
- PRg2 — BOONDOOMA IGNEOUS COMPLEX
- PRg3 — CROWS NEST GRANITE
- PRg4 — *Diorite, quartz diorite*
- PRg5 — ESKDALE GRANODIORITE
- PRg6 — KINGAHAM CREEK GRANODIORITE
- PRg7 — JENNY LIND GRANITE
- PRg8 — KIMBALA GRANODIORITE
- PRg9 — TAROMEO TONALITE
- PRn1 — BOOKOOKOORARA ADAMELLITE
- PRn2 — DUNDEE RHYODACITE
- PRn3 — MARYLAND GRANITE
- PRn4 — POSTMANS CREEK GRANODIORITE
- PRe — EIDSVOLD COMPLEX
- PRr — RAWBELLE BATHOLITH (Undivided)
- PRi — CULGRAIGE GRANITE (RAWBELLE BATHOLITH (Informal name))
- PRv — CRYSTAL VALE ADAMELLITE (RAWBELLE BATHOLITH (Informal name))
- PRwa — WINGFIELD ADAMELLITE (RAWBELLE BATHOLITH (Informal name))
- PRt — CHELTENHAM CREEK ADAMELLITE (RAWBELLE BATHOLITH (Informal name))
- PRc — COONAMBULA GRANODIORITE (RAWBELLE BATHOLITH (Informal name))
- PRd — DELUBRA QUARTZ GABBRO (RAWBELLE BATHOLITH (Informal name))
- PRag — CADARGA CREEK GRANODIORITE (RAWBELLE BATHOLITH (Informal name))
- PRh — HAWKWOOD GABBRO (RAWBELLE BATHOLITH (Informal name))

PALAEOZOIC

PERMIAN

- P — *Conglomerate, sandstone*
- Pc — CRESSBROOK CREEK GROUP: *Volcanics, mudstone, shale*
- Pd — DRAKE ANDESITE
- Pe — CEDARTON VOLCANICS/Unnamed volcanics and sediments
- Pg — GYMPIE GROUP: *Shale, mudstone, siltstone, volcanics*
- Ph — RHYOLITE RANGE BEDS: *Volcanics*
- Pi — GILGURRY MUDSTONE
- Pk — EIGHT MILE CREEK BEDS: *Conglomerate, sandstone*
- Pm — MARUMBA BEDS: *Mudstone, sandstone*
- Pn — NORTHBROOK BEDS: *Acid volcanics, mudstone*
- Pv — *Acid volcanics*
- Puw — BLACKWATER GROUP: *Shale, siltstone, sandstone*
- Pb — BACK CREEK GROUP: *Shale, siltstone, sandstone*
- Plt — SILVER SPUR BEDS: *Conglomerate, sandstone*
- Pln — CAMBOON ANDESITE
- Plo — NOGO BEDS: *Intermediate to basic volcanics, arenite*
- Pla — NARAYEN BEDS: *Intermediate to basic volcanics, arenite*
- Plb — BURNETT FORMATION: *Arenite, siltstone*
- Pgw — WIGTON GRANITE
- Pgm — MONSIDALE GRANODIORITE
- Pgo — WOOLSHED MOUNTAIN GRANODIORITE
- Po — WOLCA GRANITE
- Pt — TENNINGERING GRANODIORITE
- Pmg — MINGO GRANITE
- Pgc — CHOWEY GRANITE
- Pr — RAZORBACK CREEK MUDSTONE/PLUMBAGO CREEK MUDSTONE
- Pmc — CONDAMINE BEDS: *Mudstone, acidic tuff, sandstone*
- Pct — TERRICA BEDS: *Conglomerate, sandstone*
- Pa — ALUM ROCK CONGLOMERATE

CARBONIFEROUS-PERMIAN

- CPb — MOUNT BARNEY BEDS: *Sandstone, siltstone, volcanics*
- CPe — EMU CREEK FORMATION: *Sandstone, mudstone*
- CPx — *Serpentinite*

PALAEOZOIC

CARBONIFEROUS

- C — *Conglomerate*
- C1 — *Mudstone, chert, jasper, metavolcanics*
- Cd — BOOLOUMBA BEDS/DURUNDUR SHALE/Unnamed metamorphics: *Slate, phyllite, schist*
- Ck — KURWONGBAH BEDS: *Phyllite, slate*
- Cr — ROCKSBERG GREENSTONE: *Greenschist, glaucophane schist*
- Cgc — CLADDAGH GRANODIORITE/CALLANGOWAN GRANODIORITE
- Cpo — GOOD NIGHT BEDS: *Slate, phyllite*
- Cc — CURTIS ISLAND GROUP (Undivided): *Mica schist, gneiss*
- Ccs — SHOALWATER FORMATION: *Arenite*
- Ccw — WANDILLA FORMATION: *Mudstone, arenite, siltstone*
- Ct — TORSDALE BEDS: *Acid to intermediate volcanics*
- Cuo — BOILING CREEK GROUP: *Sandstone, siltstone*
- Cl — CASWELL CREEK GROUP/CRANA BEDS: *Arenite, mudstone, siltstone*
- Cld — DERRARABUNGY BEDS: *Mudstone, arenite, siltstone*
- Cug — KILBEGGAN ADAMELLITE/MOOCOOROOBA ADAMELLITE/GLANDORE GRANODIORITE/Undivided granodiorite and adamellite

DEVONIAN-CARBONIFEROUS

- DCa — AMAMOOR BEDS: *Mudstone, slate, basalt, chert*
- DCb — BUNYA PHYLLITE: *Phyllite, greywacke*
- DCm — MARONGHI CREEK BEDS: *Mudstone, slate, greywacke*
- DCn — NERANLEIGH-FERNVALE BEDS: *Mudstone, shale, greywacke*
- DCs — SUGARLOAF METAMORPHICS: *Slate, phyllite*
- DCt — TEXAS BEDS: *Greywacke, mudstone*
- DCy — *Schist, gneiss, granodiorite*

DEVONIAN

- D — PUMKIN HUT MUDSTONE/FIRST BRANCH CREEK SANDSTONE/PHILPOT LIMESTONE/DUNOLLIE BEDS/MUNBOOREE BEDS: *Andesitic, volcanics, sandstone, mudstone*
- Dm — *Granite, adamellite, pegmatite*
- Duc — CHANNER CREEK BEDS: *Arenite, intermediate volcanics*
- Ds — SILVERWOOD GROUP: *Volcanics, sandstone, mudstone*

PALAEOZOIC

- Pz — Undivided: *Schist, gneiss*

SUB-SURFACE BASEMENT GEOLOGY

- V1 — *Early Devonian basalt, acid volcanics, andesitic pyroclastics, interbedded volcaniclastic sediments* (basal volcanic unit of the Adavale Basin, GUMBARDO FORMATION)
- V2 — *Late Carboniferous to Early Permian calc-alkaline volcanics, rhyolitic ignimbrite, altered andesite and basalt* (COMBARNGO VOLCANICS)
- V3 — *Carboniferous acid volcanics, andesite, altered basalt and andesite, andesitic tuff, dacitic tuff, rhyolitic tuff, rhyolitic ignimbrite* (volcanic units of the informally-named KUTTUNG FORMATION)
- M1 — *Middle Cambrian to mid-Ordovician steeply-dipping, greenschist facies metasediments, fine to medium grained quartzoze sandstones, interbedded siltstone and mudstone*
- M2 — *Early to Middle Devonian(?) indurated and deformed sediments and metasediments, quartzose sandstones and siltstones interbedded with cleaved argillites* (TIMBURY HILLS FORMATION)
- M3 — *Carboniferous indurated clastics, volcaniclastic sandstones, siltstones, andesitic conglomerates* (sedimentary/metasedimentary units of the informally-named KUTTUNG FORMATION)
- G1 — *Silurian to Early Devonian S-type granite*
- G2 — *Late Devonian to Early Carboniferous muscovite/biotite granite (S-type)* (ROMA GRANITE)
- G3 — *Late Devonian biotite granite, adamellite, pegmatite*
- G4 — *Permian-Early Triassic granite, granodiorite*

Acknowledgements

The compiler and editor of BMR Bulletin 232 (D. M. Finlayson) wishes to acknowledge the major contributions and assistance rendered by many people in the compilation of the geological map. The contributions of the following persons should be particularly mentioned: the basement geology was compiled by Cec Murray (Queensland Department of Resource Industries); map digitising, compilation and drafting was by Terry Brown (BMR Cartographic Services Unit); colour design was by Rainer Swoboda (BMR Cartographic Services Unit). Kinta Hoffmann (Queensland Department of Resource Industries), Ray Shaw (Petroconsultants Australasia Pty. Ltd., Sydney), Joe Salomon (LASMO Oil Company Australia Pty. Ltd., Brisbane), Adrian Barwell (OGM Development Pty. Ltd., Brisbane), Phil O'Brien and Alan Wells (BMR Onshore Basins and Petroleum Geology Branch) contributed data and information on subsurface geological structures in various parts of the mapsheet area

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It is recommended that this map be referred to as:
Finlayson, D. M. (Compiler and Editor), 1990. The Eromanga-Brisbane Geoscience Transect: Map 1 - Geology. *Bureau of Mineral Resources, Geology and Geophysics, Australia, Bulletin 232*

Availability of geological data:
The digital geological database for this map was compiled on the BMR Intergraph system and, in some circumstances, may be suitable for transfer to other digital databases. Further information on data release conditions and costs can be obtained by writing to -

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EROMANGA–BRISBANE GEOSCIENCE TRANSECT

ICL Global Geoscience Transect No. 7

MAP 2: BOUGUER GRAVITY ANOMALIES

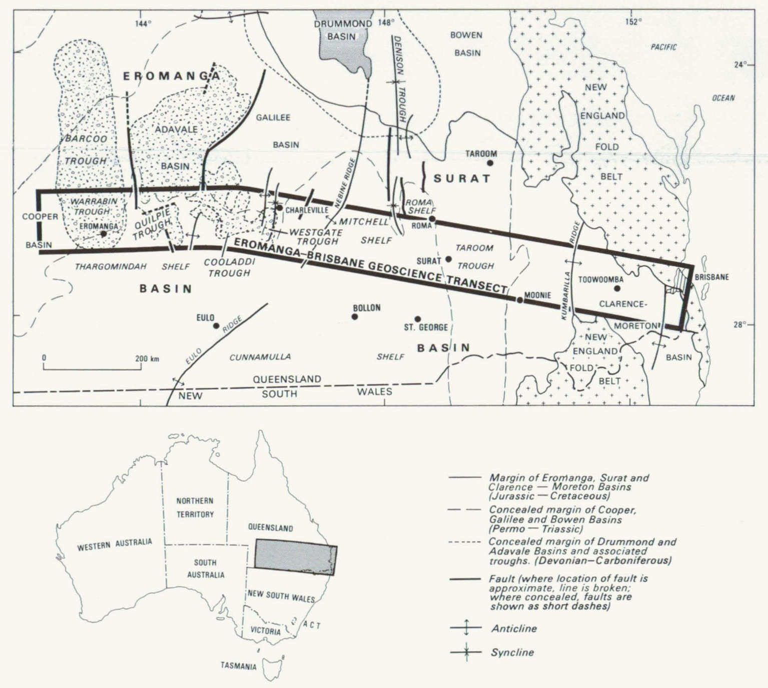
EASTERN AUSTRALIA

Scale 1:1 000 000

0 100 200 Kilometres

Lambert Conformal Conic Projection with Standard Parallels at 25°40'S and 28°20'S

Locality Maps



Open red rectangle	Inferred block boundary from gravity data
Open red circle	Inferred margin of crustal reworking from gravity data
Solid red rectangle	Inferred block boundary from magnetic data
Solid red dot	Inferred margin of crustal reworking from magnetic data
Red line with tick	Prominent magnetic gradients; ticks on low side
Red dashed line	Fault interpreted from offset gravity and magnetic data
7 (hatched line)	BMR deep seismic reflection profile with traverse number
+	Gravity anomaly — relative high
−	Gravity anomaly — relative low
100	Contour interval $20\,\mu m.s^{-2}$ (2 mgal)
Red	Gravity anomaly $> +100\,\mu m.s^{-2}$ (+10 mgal)
Blue	Gravity anomaly $< -300\,\mu m.s^{-2}$ (−30 mgal)

Bouguer Gravity Anomalies: Explanatory Notes

The Bouguer gravity anomaly data are reduced using a density of $2.67\ t.m^{-3}$. No terrain corrections have been applied. The 1967 International Gravity Formulae and the IGSN71 gravity datum modified to the Australian standard have been used. The gravity anomaly values of regional stations have an error (standard deviation) of about $12\ \mu m.s^{-2}$ (1.2 mgal). Dots indicate the positions of gravity stations. In some places, close-spaced stations appear as continuous lines, e.g. the main east-west seismic traverse has gravity stations at 500 m intervals.

The Bouguer gravity values are contoured at 20 and $100\ \mu m.s^{-2}$ (2 and 10 mgal) intervals, based on a 3 minute latitude and longitude grid derived from a two-dimensional surface spline fitted to the data. The data reduction procedures have been described by Anfiloff & others (1976). Bouguer anomalies less than $-300\ \mu m.s^{-2}$ (30 mgal) are highlighted in blue; anomalies greater than $+100\ \mu m.s^{-2}$ (+10 mgal) are highlighted in red.

Reference

Anfiloff, W., Barlow, B. C., Murray, A. S., Denham, D., & Sandford, R., 1976. Compilation and production of the 1976 1:5 000 000 gravity map of Australia. BMR Journal of Australian Geology and Geophysics, 1, 273-276

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Availability of gravity data:

Copies of the Australian National Gravity Data Base and the combined Digital Terrain and Gravity Models (gridded data) are available on magnetic tape. Further information on data release conditions and costs can be obtained by writing to -

Executive Director,
Bureau of Mineral Resources, Geology and Geophysics,
G.P.O. Box 378,
Canberra A.C.T. 2601
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The Minister for Primary Industries and Energy

This map accompanies a text publication:
Finlayson, D. M. (Compiler and Editor), 1990. The Eromanga-Brisbane Geoscience Transect: a guide to basin development across Phanerozoic Australia in southern Queensland. *Bureau of Mineral Resources, Geology and Geophysics, Australia, Bulletin 232*

It is recommended that this map be referred to as:
Finlayson, D. M. (Compiler and Editor), 1990. The Eromanga-Brisbane Geoscience Transect: Map 2 - Bouguer Gravity Anomalies. *Bureau of Mineral Resources, Geology and Geophysics, Australia, Bulletin 232*

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EROMANGA–BRISBANE GEOSCIENCE TRANSECT
MAP 2: BOUGUER GRAVITY ANOMALIES

EROMANGA–BRISBANE GEOSCIENCE TRANSECT

ICL Global Geoscience Transect No. 7

MAP 3: TOTAL MAGNETIC FIELD ANOMALIES

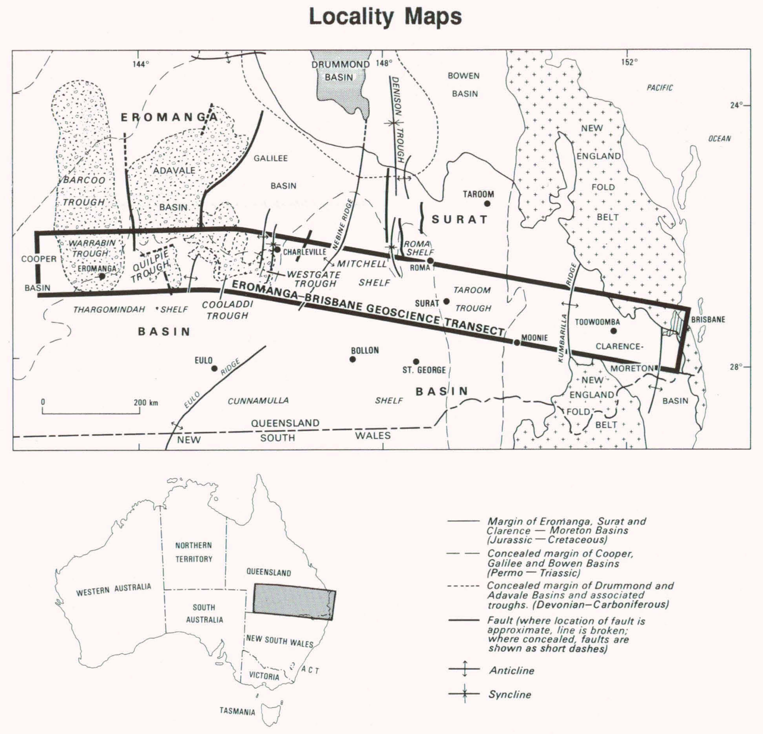
EASTERN AUSTRALIA

Scale 1:1 000 000

0 100 200 Kilometres

Lambert Conformal Conic Projection with Standard Parallels at 25°40′S and 28°20′S

Locality Maps



- Inferred block boundary from gravity data
- Inferred margin of crustal reworking from gravity data
- Inferred block boundary from magnetic data
- Inferred margin of crustal reworking from magnetic data
- Prominent magnetic gradients; ticks on low side
- Fault interpreted from offset gravity and magnetic data
- 7 BMR deep seismic reflection profile with traverse number
- 100 Relative magnetic contour with value (nT)
- Relative magnetic high
- Magnetic gradient; tick on low side
- Magnetic anomalies > 0nT

Total Magnetic Field Anomalies: Explanatory Notes

The anomalies shown on this map are of total magnetic field intensity derived from regional airborne surveys. The contours are produced from surveys conducted by the Australian Bureau of Mineral Resources, Geology and Geophysics, exploration companies operating under the Petroleum Search Subsidy Act, and the New South Wales Department of Mineral Resources. The regional field along the transect was removed from the data using the International Geomagnetic Reference Field (IGRF) as standard. Flight-line spacing varies, being 1.5 km over much of the mapped region but as much as 8 km in some areas in the west. Four data sets are incorporated in the map, with the boundaries between the data sets shown as breaks in contour lines. There are datum offsets between the various data sets.

For the data set from the western two-thirds of the transect area, the original contoured maps (based on analogue data) were digitised to derive 1.2 minute grids, the grids joined and upward continued to 3 km; the contour interval is 25 nT (25 gamma) (BMR, 1976). For the other three data sets in the transect area, the magnetic anomalies tend to be more extreme because of near-surface basement rocks. The digitally-recorded data were levelled, then used to calculate 0.25 minute grids and filtered to remove wavelengths shorter than 0.2°; the contour interval is 20 nT (20 gamma).

Reference

BMR, 1976. Magnetic map of Australia; residuals of total intensity, 1:2 500 000 scale. Australian Bureau of Mineral Resources, Geology & Geophysics.

Acknowledgements

The compiler and editor of BMR Bulletin 232 (D. M. Finlayson) wishes to acknowledge the major contributions and assistance rendered by many people in the compilation of the total magnetic field anomaly map. The contributions of the following persons should be particularly mentioned: aeromagnetic data integration, checking, and contouring by Mario Bacchin, Brian Minty and Peter Wellman of BMR Geophysical Observatories and Mapping Branch, and map compilation by Terry Brown, Rainer Swoboda and Jon Stirzaker of BMR Cartographic Office.

Availability of aeromagnetic data:

Copies of information contained in the BMR aeromagnetic database and gridded data within particular survey areas are available on magnetic tape. Further information on data release conditions and costs can be obtained by writing to -

Executive Director,
Bureau of Mineral Resources, Geology and Geophysics,
G.P.O. Box 378,
Canberra A.C.T. 2601
(Telex AA26109; Fax 062 488178)

This map is published by:
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Base map compiled by:
The BMR from digital data supplied by the Australian Surveying and Land Information Group, Department of Administrative Services

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The Minister for Primary Industries and Energy

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It is recommended that this map be referred to as:
Finlayson, D. M. (Compiler and Editor), 1990. The Eromanga-Brisbane Geoscience Transect: Map 3 - Total Magnetic Field Anomalies. *Bureau of Mineral Resources, Geology and Geophysics, Australia, Bulletin 232*

Printed by **Mercury-Walch, Hobart, Tasmania**

EROMANGA–BRISBANE GEOSCIENCE TRANSECT
MAP 3: TOTAL MAGNETIC FIELD ANOMALIES

EROMANGA–BRISBANE GEOSCIENCE TRANSECT

ICL Global Geoscience Transect No. 7

MAP 4: GEOSCIENCE COMPILATION

EASTERN AUSTRALIA

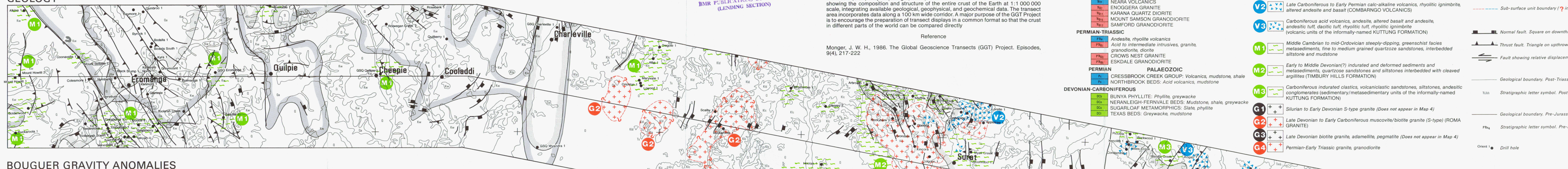
Geoscience Transect: Explanatory Notes

This geoscience compilation map is assembled broadly according to the guidelines established by the Inter-Union Commission on the Lithosphere (ICL) Global Geoscience Transects (GGT) Project (Monger, 1986). The term "transect" refers to a cross-section showing the composition and structure of the entire crust of the Earth at 1:1 000 000 scale, integrating available geological, geophysical, and geochemical data. The transect area incorporates data along a 100 km wide corridor. A major purpose of the GGT Project is to encourage the preparation of transect displays in a common format so that the crust in different parts of the world can be compared directly

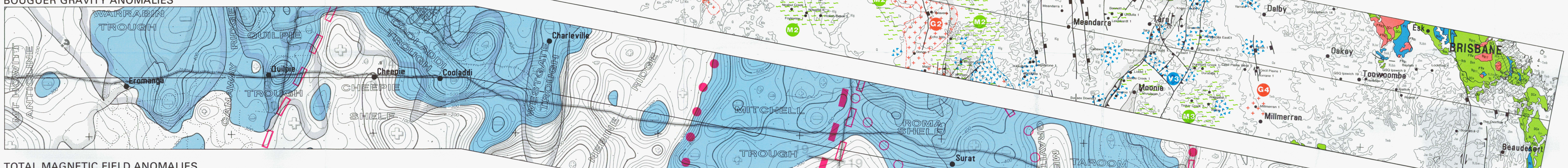
Reference

Monger, J. W. H., 1986. The Global Geoscience Transects (GGT) Project. Episodes, 9(4), 217-222

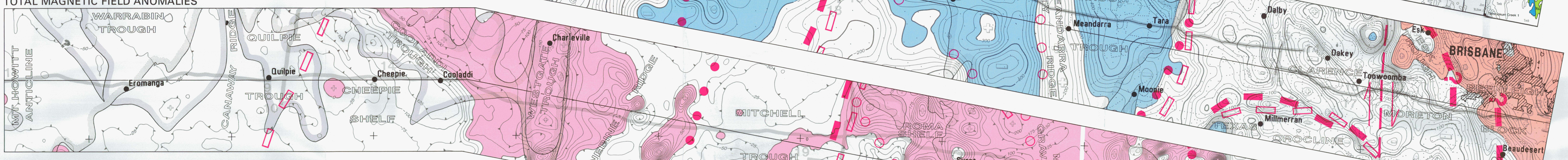
GEOLOGY



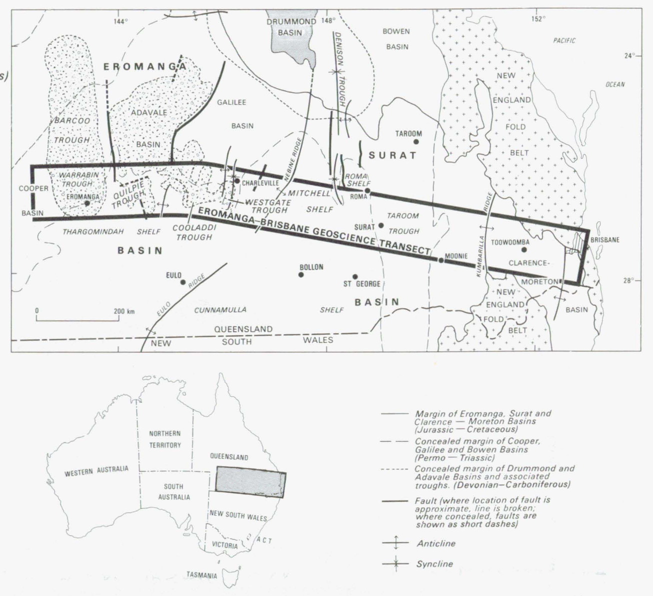
BOUGUER GRAVITY ANOMALIES

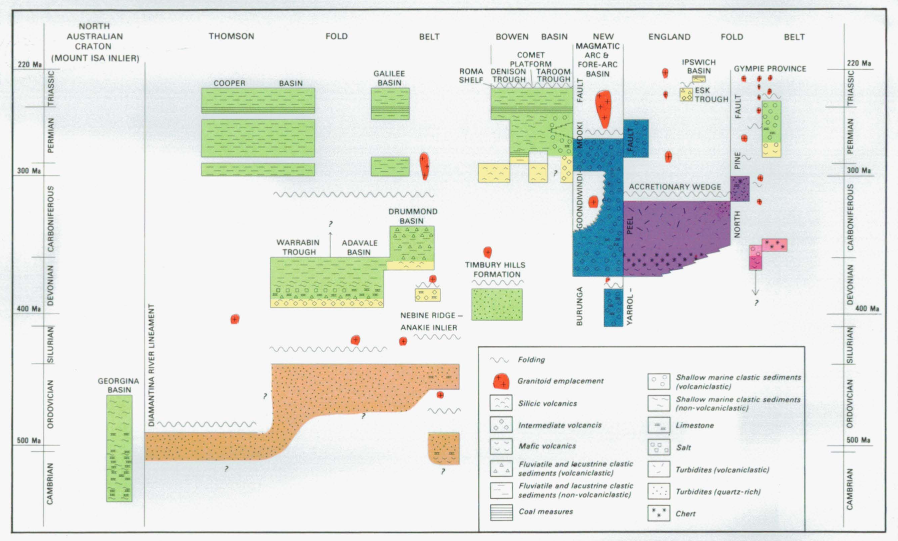


TOTAL MAGNETIC FIELD ANOMALIES



Locality Maps





Space Time Diagram: Explanatory Notes

The space-time diagram for the southern Queensland region indicates the principal geological events, based on data by Day & others (1983) and Murray (1986). Ages of folding events are derived from metamorphic isotopic dates and from observed angular unconformities. The inferred tectonic setting within which the geological units formed is indicated by colour according to the legend attached to the interpreted crustal section at the bottom of this mapsheet

References

Day, R.W., Whitaker, W.G., Murray, C.G., Wilson, I.H. & Grimes, K.G., 1983. Queensland Geology. Geological Survey of Queensland, Publication No. 383

Murray, C.G., 1986. Metallogenic and tectonic development of the Tasman Fold Belt system in Queensland. Ore Geology Reviews, 1, 315-400

Geology: Explanatory Notes

The surface geology at 1:1 000 000 scale within the area 25° to 29° South, 142° to 154° East has been digitized and compiled from four mapped regions; the Eromanga Basin (Senior & others, 1978), the Surat Basin (Exon, 1976), the Clarence-Moreton Basin (Whitaker & Green, 1980) and the Maryborough mapsheet area (Queensland Department of Resource Industries, 1989). The references for these four mapped regions (listed below) should be consulted for detailed information. In some places, recent mapping information has been incorporated, and there has been some simplification of 1:250 000 and 1:500 000 scale mapping. Surface geology over much of the mapsheet area is dominated by the Jurassic-Cretaceous basin sequences, Tertiary volcanic sequences and Quaternary sedimentary deposits. Since the main purpose of this map is to highlight basement geological controls on basin evolution, much of the surface geology is shown only as grey-tone background information. Only selected surface geological units and formations have been annotated, as indicated in the legend

Pre-Jurassic rock units crop out in the eastern part of the mapsheet area; these units are highlighted in bold line-work. Sub-surface basement geology is derived from hydrocarbon and water exploration drillholes. Not all drillholes penetrated basement; where they did, some indication of geology is shown according to the attached legend. Because there are numerous drillholes in some areas prospective for oil and gas, only selected holes are shown on this mapsheet; unnamed drillholes are usually water bores. The structural information under cover rocks is derived from extensive seismic reflection profiling

Many geological references are contained in papers in the Australian Bureau of Mineral Resources, Geology & Geophysics Bulletin 232, which this mapsheet accompanies. General references include papers by Finlayson & others (1988) and Elliott & Brown (1988)

ELLIOTT, L. G., & BROWN, R. S., 1988. The Surat and Bowen Basins, a historical review. In: Petroleum in Australia, the First Century, Australian Petroleum Exploration Association, 120-138
EXON, N. F., 1976. Geology of the Surat Basin in Queensland. Australian Bureau of Mineral Resources, Geology & Geophysics Bulletin, 166
FINLAYSON, D. M., LEVEN, J. H., & ETHERIDGE, M. A., 1988. Structural styles and basin evolution in the Eromanga region, eastern Australia. American Association of Petroleum Geologists Bulletin, 72, 33-48
MURRAY, C. G., 1986. Metallogenic and tectonic development of the Tasman Fold Belt in Queensland. Ore Geology Reviews, 1, 315-400
QUEENSLAND DEPARTMENT OF RESOURCE INDUSTRIES, 1989. Geology of the Maryborough 1:250 000 mapsheet area. Preliminary compilation
SENIOR, B. R., MOND, A., & HARRISON, P. L., 1978. Geology of the Eromanga Basin. Australian Bureau of Mineral Resources, Geology & Geophysics Bulletin, 167
WHITAKER, W. G., & GREEN, P. M., 1980. Moreton Geology, 1:500 000 scale. Geological Survey of Queensland, Queensland Department of Mines, Brisbane

Bouguer Gravity Anomalies: Explanatory Notes

The Bouguer gravity anomaly data are reduced using a density of 2.67 $t.m^{-3}$. No terrain corrections have been applied. The 1967 International Gravity Formulae and the IGSN71 gravity datum modified to the Australian standard have been used. The gravity anomaly values of regional stations have an error (standard deviation) of about 12 $\mu m.s^{-2}$ (1.2 mgal). Dots indicate the positions of gravity stations. In some places, close-spaced stations appear as continuous lines, e.g. the main east-west seismic traverse has gravity stations at 500 m intervals

The Bouguer gravity values are contoured at 20 and 100 $\mu m.s^{-2}$ (2 and 10mgal) intervals, based on a 3 minute latitude and longitude grid derived from a two-dimensional surface spline fitted to the data. The data reduction procedures have been described by Anfiloff & others (1976). Bouguer anomalies less than -300 $\mu m.s^{-2}$ (30 mgal) are highlighted in blue; anomalies greater than + 100 $\mu m.s^{-2}$ (+ 10 mgal) are highlighted in red

Reference

Anfiloff, W., Barlow, B. C., Murray, A. S., Denham, D., & Sandford, R., 1976. Compilation and production of the 1976 1:5 000 000 gravity map of Australia. BMR Journal of Australian Geology and Geophysics, 1, 273-276

Total Magnetic Field Anomalies: Explanatory Notes

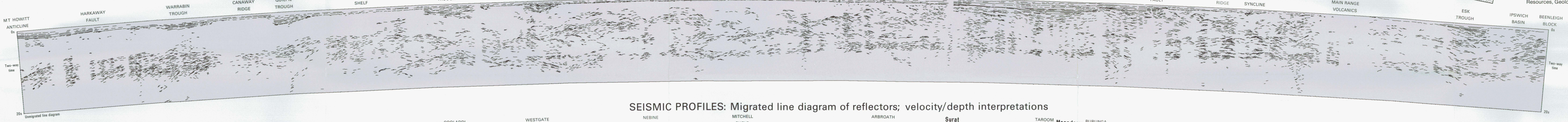
The anomalies shown on this map are of total magnetic field intensity derived from regional airborne surveys. The contours are produced from surveys conducted by the Australian Bureau of Mineral Resources, Geology and Geophysics, exploration companies operating under the Petroleum Search Subsidy Act, and the New South Wales Department of Mineral Resources. The regional field along the transect was removed from the data using the International Geomagnetic Reference Field (IGRF) as standard. Flight-line spacing varies, being 1.5 km over much of the mapped region but as much as 8 km in some areas in the west. Two data sets are incorporated in this geoscience compilation map, with the boundary between the data sets shown as breaks in contour lines. There is a datum offset between the two data sets

For the data set from the western two-thirds of the transect area, the original contoured maps (based on analogue data) were digitised to derive 1.2 minute grids, the grids joined and upward continued to 3 km; the contour interval is 25 nT (25 gamma) (BMR, 1976). For the eastern data set on this geoscience compilation map, the magnetic anomalies tend to be more extreme because of near-surface basement rocks. The digitally-recorded data were levelled, then used to calculate 0.25 minute grids and filtered to remove wavelengths shorter than 0.2°; the contour interval is 20 nT (20 gamma)

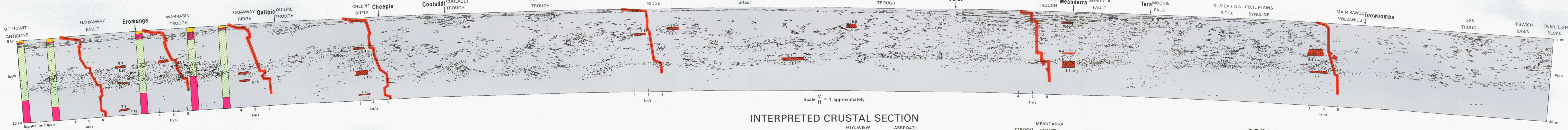
Reference

BMR, 1976. Magnetic map of Australia; residuals of total intensity, 1:2 500 000 scale. Australian Bureau of Mineral Resources, Geology & Geophysics

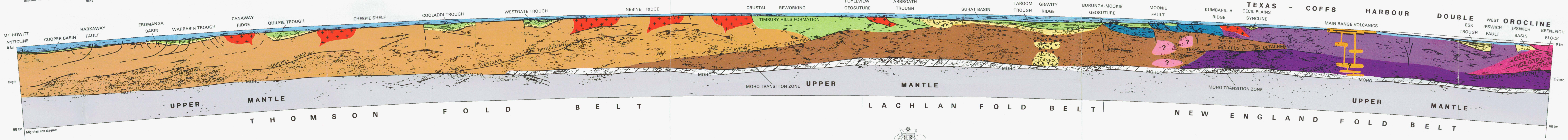
SEISMIC PROFILES: Digitized line diagram of significant reflectors



SEISMIC PROFILES: Migrated line diagram of reflectors; velocity/depth interpretations



INTERPRETED CRUSTAL SECTION



Seismic Profiles: Explanatory Notes

The deep seismic reflection profiles illustrated on this mapsheet are plots of digitized reflection events identified on full-scale seismic sections (Finlayson & others, 1990, Leven & Finlayson, 1990). All significant reflection events are plotted as a time section in the uppermost diagram, thus providing a good representation of the original seismic data. These data were migrated using a velocity of 6km/s (an average crustal velocity) and re-plotted in the lower seismic profile to give an indication of the true position of dipping horizons on a depth section. This migration process assumes that the dipping events arise from geological features that are within a vertical plane under the seismic traverse. In many cases this assumption may not be valid, with reflecting horizons within basement on either side of the traverse contributing to the recorded signal. Reflection profiles were commonly recorded as 6-fold common mid-point data using explosive sources.

The velocity/depth functions illustrated in the lower seimic profile are derived from wide-angle reflection and refraction data (Collins, 1988). In all cases, except in the Taroom Trough, data were recorded coincident with, or intersecting, the reflection traverses. In the Taroom Trough, velocity/depth data are from an along-strike profile about 250 km farther north, still within the trough. Data were commonly recorded at 5 km intervals out to distances of 150-300 km from the sources; both travel-times and signal amplitudes were modelled to derive velocity/depth functions

In the western part of the transect, apparent resistivity models are shown. These are derived from magnetotelluric soundings at sites along the seismic profiles (Spence & Finlayson, 1983)

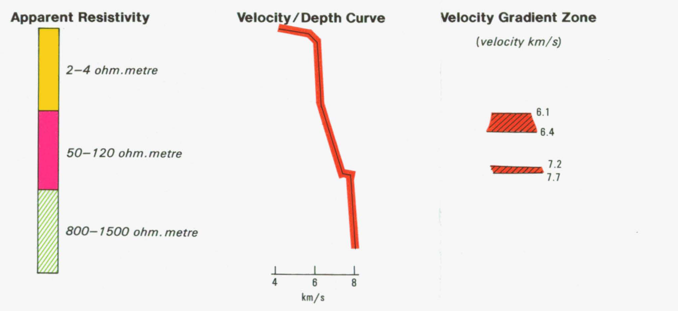
References

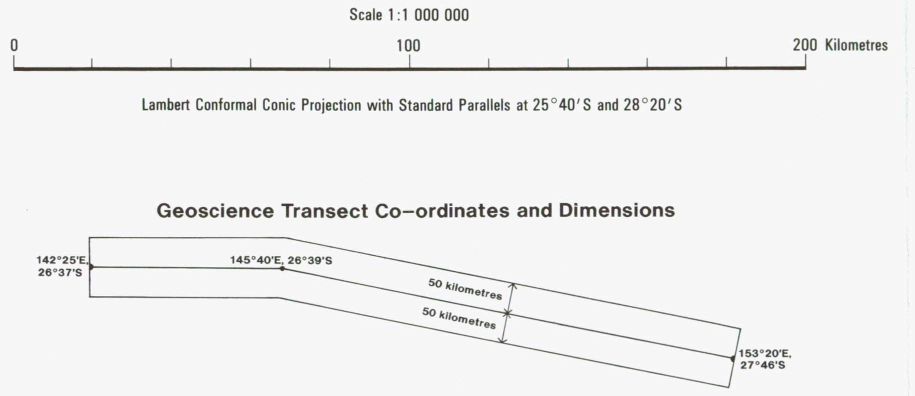
Collins, C. D. N., 1988. Seismic velocities in the crust and upper mantle of Australia. Bureau of Mineral Resources, Geology & Geophysics, Australia, Report, 277

Finlayson, D.M., Wake-Dyster, K. D., Leven, J. H., Johnstone, D. W., Murray, C. G., Harrington, H. J., Korsch, R. J. & Wellman, P., 1990. Seismic imaging of major tectonic features in the crust of Phanerozoic eastern Australia. Tectonophysics, 173, 211-230

Leven, J. H., Finlayson, D. M. & Wake-Dyster, K. D., 1990. Mid-crustal detachments controlling basin formation: ramp synclines and basins in southwestern Queensland. Tectonophysics, 173, 231-246

Spence, A. G., & Finlayson, D. M., 1983 The resistivity of the crust and upper mantle in the central Eromanga Basin, Queensland, using magnetotelluric techniques. Journal of the Geological Society of Australia, 30, 1-16





Scale 1:1 000 000

Lambert Conformal Conic Projection with Standard Parallels at 25°40'S and 28°20'S

Geoscience Transect Co-ordinates and Dimensions

Interpreted Crustal Section: Explanatory Notes

The interpreted section illustrates the major geological features considered to be significant at depths down to 60 km. It is based on interpretations of seismic reflection and velocity data, regional gravity and magnetic trends and models, and the tectonic processes considered to apply throughout southern Queensland from structural interpretations of near-surface geology. Only a minimum of geological inference is included within units incorporated in the interpreted section. In this respect the interpretation can be regarded as conservative, i.e. with little detail not closely allied to reasonably well known geological or geophysical constraints

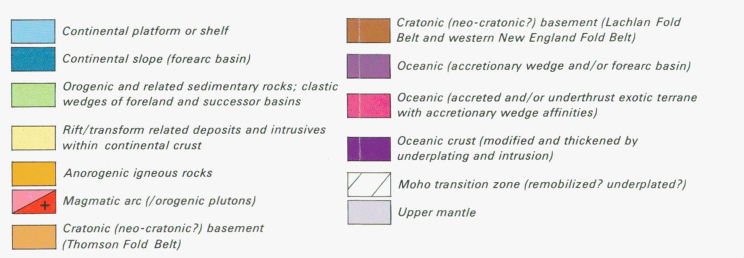
Mesozoic cover sequences (from west to east) of the Eromanga, Surat and Clarence-Moreton Basins conceal most of the basement features along the transect. The structural geology of the major rock units within the underlying Thomson, Lachlan and New England Fold Belts has therefore been inferred from drillcore and geophysical data. The interpreted crustal section integrates available geological and geophysical information to provide an interpretational sketch of the crust across Phanerozoic eastern Australia, including ideas of the possible processes applying during its geological history

Details of the interpretation are contained in papers contributed to Australian Bureau of Mineral Resources, Geology and Geophysics Bulletin 232. Units are differentiated according to the inferred tectonic setting within which units formed, not according to geological age. The guidelines for these tectonic settings are broadly those set out by the ICL Global Geoscience Transects (GGT) Project (Monger, 1986)

Reference

Monger, J. W. H., 1986. The Global Geoscience Transects (GGT) Project. Episodes, 9(4), 217-222

Inferred Tectonic Setting



Acknowledgements

The compiler and editor of BMR Bulletin 232 (D. M. Finlayson) wishes to acknowledge the major contributions by many people in the compilation of this map. The contributions of the following persons should be particularly mentioned: cartography by Terry Brown and Rainer Swoboda (BMR Cartographic Services Unit); contributed data and information on subsurface geological structures from Cec Murray and Kinta Hoffmann (Queensland Deparment of Resource Industries), Ray Shaw (Petroconsultants Australasia Pty. Ltd., Sydney), Joe Salomon (LASMO Oil Company Australia Pty. Ltd., Brisbane), Adrian Barwell (OGM Development Pty. Ltd., Brisbane), Phil O'Brien and Alan Wells (BMR Onshore Basins and Petroleum Geology Branch); gravity and aeromagnetic data integration, contouring and presentation from John Williams, Alan Murray, Peter Wellman, Mario Bacchin, and Brian Minty (BMR Geophysical Observatories and Mapping Branch); programming assistance by David Downie (BMR Systems Information Branch); seismic section production by Ray Bracewell, Jim Leven and Bruce Goleby (BMR Onshore Basins and Petroleum Geology Branch)

Co-operating Institutions

The compilation of data, information and interpretations along the Eromanga-Brisbane Geoscience Transect involved earth scientists from a number of institutions and companies in co-operation with those at the Australian Bureau of Mineral Resources, Geology and Geophysics, Canberra. These institutions were as follows:

Queensland Department of Resource Industries, Brisbane
Department of Geology & Mineralogy, University of Queensland, Brisbane
Department of Applied Geology, University of New South Wales, Sydney
School of Earth Sciences, Macquarie University, Sydney
Division of Exploration Geoscience, Commonwealth Scientific & Industrial Research Organisation, Sydney
The Australian Museum, Sydney
Department of Geology & Geophysics, University of Sydney
Department of Geology, University of Wollongong
Department of Geology and Geophysics, University of New England, Armidale
Petroconsultants Australasia Pty. Ltd., Sydney
LASMO Oil Co. Australia Pty., Ltd
Department of Geology, La Trobe University, Melbourne
School of Science & Technology, University of Western Sydney

This map is published by:
The Bureau of Mineral Resources, Geology and Geophysics, Department of Primary Industries and Energy

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Availability of computer compatible data:
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Bureau of Mineral Resources, Geology and Geophysics,
G.P.O. Box 378,
Canberra A.C.T. 2601
(Telex AA62109; Fax 06 2488178)

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