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Front cover:
Surface deformation, including crack in soil (inset), caused by the Tennant Creek earthquakes along the eastern Lake Surprise scarp (see paper by Jones & others in this issue). Vertical displacement of the scarp was 1.7m at this site. (Photo by J.R. Bowman.)

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Silurian and Late Carboniferous conodonts from the Charles Louis Range and central Birds Head, Irian Jaya, Indonesia

Robert S. Nicoll & G.M. Bladon

Conodonts have been recovered in samples from two stratigraphic units in Irian Jaya: the Modio Dolomite in the Charles Louis Range, and probably the Aimau Formation in the central Birds Head. The Modio Dolomite fauna includes *Panderodus* cf. *P. simplex* (Branson & Mehl, 1933), and probably has a Silurian age. Two float samples of conodonts were recovered from the Modio Dolomite in the Charles Louis Range (part of WAGHETE) and the Aimau Formation in the central Birds Head (in TAMINABUAN).

**Modio Dolomite**

**Geology and pertinent biostratigraphy**

Outcrop of the Modio Dolomite is confined to an area of less than 10 km² in the upper reaches of the Mapia River and its tributaries draining the southern fall of the Charles Louis Range between Timepa and Modio (Fig. 1), at the western end of the Irian Jaya central range. The Modio Dolomite occupies a belt of Palaeozoic and early Mesozoic, mainly siliciclastic rocks intruded by stocks of the largely Pliocene Timepa Monzonite (Panggabean & Pigram, 1989). These rocks are faulted to the south against younger Mesozoic siliciclastic rocks (Kembelangan Group) and mainly Palaeogene limestone (New Guinea Limestone Supergroup).

The Palaeozoic siliciclastic rocks in the Charles Louis Range are assigned to the Aiduna Formation, which overlies the Modio Dolomite conformably or paraconformably (Pigram & Panggabean, 1989; Fig. 2). The Aiduna Formation west of the Modio Dolomite outcrop has yielded an extensive brachiopod fauna, and a diverse flora, including several species of *Glossopteris* (Pigram & Panggabean, 1989, appendix 2). Archbold (1991, this issue) favours an early Artinskian (Aktastinian—early Baigendzhinian; Early Permian) age for the brachiopod fauna. J. Rigby (Geological Survey of Queensland, personal communication 1982, cited in Pigram & Panggabean, 1983, appendix 3) likened the flora to that of the Newcastle Coal Measures in eastern Australia, which are considered to be Tatarian (latest Permian). These are the only detailed biostratigraphic age determinations that have a direct bearing on the minimum age of the Modio Dolomite (viz., before early Artinskian).

The Modio Dolomite was sampled for conodont examination at two localities in the westernmost part of its outcrop area (Fig. 1). Two samples — 80CP515A and 80CP516A — were collected a few tens of metres from one another, within 100 m of the fault that truncates the outcrop of the dolomite near the junction of the Waga (also referred to as 'Wogec') and Pori Rivers, tributaries of the Mapia River. Sample 80CP515A, collected closer to the fault, is a grey recrystallised limestone; sample 80CP516A is a limestone breccia.

**Conodont fauna**

A fauna of 30 conodont elements was recovered from the two samples of the Modio Dolomite (Table 1). All of the elements are poorly preserved; most are broken and fractured, but not considered to have been derived from the Aimau Formation. The Aimau Formation yield included *Neognathodus* cf. *N. bassleri* (Harris & Hollingsworth, 1933) and *Hindeodus minutus* (Ellison, 1941), which indicate a Late Carboniferous age.

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2 Corporate Relations, Information, & Planning Branch, Bureau of Mineral Resources, Geology and Geophysics, GPO Box 378, Canberra, ACT 2601.
3 Names of 1:250 000 Sheet areas are printed in capitals.
The age of the Modio Dolomite can be only approximately restricted to the Silurian, where it ranges from the Llandovery to the Upper Setul Formation (Branson & Mehli, 1933), but determined from the recovered conodont fauna. The genus Panderodus has a range from late Early Ordovician to the Early Devonian (Rexroad, 1967) to the Ludlow rocks (Rexroad, 1967). The proximity of the samples to the Neogene Timea Formation (Rexroad, 1967) to the Ludlow Formation (Rexroad, 1967) does not suggest that the sections of the Modio Dolomite from which the two samples were collected are most probably of Silurian age, but they could be as young as the Early Devonian.

Regional relationships of the conodont fauna

A Silurian conodont fauna has been previously described from eastern Irian Jaya (van den Boogaard, 1990). It was recovered from a sample collected in early years of this century from river gravel, or boulders, along the Lorentz River (Martin, 1911) in eastern Irian Jaya. The fauna contains elements of Ozarkodina confluens (Branson & Mehli, 1933) and O. crispa (Walliser, 1964), and specimens assigned to Coryssognathus denatus Link & Druce, 1972, or Distomodus dubius (Rhodes, 1953). None of these species was found in the fauna recovered in our study. Coryssognathus denatus Link & Druce, 1972, is considered to be a junior synonym of Peneleosgnathus dubius Jeppson, 1972 (Nicoll, 1982).

Silurian conodont faunas from eastern Australia (e.g., Link & Druce, 1972; Bischoff, 1986) are common, and range in age from the Llandovery to the Pridoli. In Western Australia, Philip (1969) reported a Late Silurian conodont fauna from the Dirk Hartog Formation in the Carnarvon Basin. Silurian conodont faunas also have been reported from the Upper Setul Formation of Malaysia (Igo & Koike, 1968; Idris, 1989). None of these faunas contains elements similar to the elements of this study assigned to 'Genus & species indt. A', but both Link & Druce (1972) and Philip (1969) recorded Panderodus simplex in their faunas.

Aimau Formation

Geology and pertinent biostratigraphy

The Aimau Formation is the oldest of three conformable units assigned to the Aifat Group (Fig. 3). It has an outcrop area of about 200 km² confined to the southwestern fall of the central Birds Head mountains. The Aifam Group is separated by an angular unconformity from underlying siliciclastic rocks of the Silurian–Devonian Kemum Formation and its possible partial lateral equivalent, the apparently unfossiliferous Aiasjur Formation (Pigram & Sukanta, 1989; Fig. 4). It is conformably and disconformably overlain by a discontinuous succession of Mesozoic (Tipuma and Jass Formations) and Tertiary (New Guinean Limestone Supergroup) rocks (Sukanta & Pigram, 1989).

Mostly ex-situ rock assigned to the Aimau Formation, and samples from outcrop of the conformably overlying Aifat Mudstone, have yielded a wealth of diverse faunas. Some of the faunas collected from these two formations are quite similar, prompting Pigram & Sukanta (1982) to speculate that the Aimau Formation was a nearshore and partly older partial time equivalent of the Aifat Mudstone deposited farther offshore below wave-base. Alternatively, some of the biostratigraphically dated float samples thought to be derived from the Aimau Formation might have been shed instead from the Aifat Mudstone.

The Aimau Formation has yielded only one biostratigraphically dated in-situ sample: a calcareous quartzofeldspathic sandstone containing corals considered to reflect a Late Carbonif-
SILURIAN AND LATE CARBONIFEROUS CONODONTS, IRIAN JAYA

Ages inferred for the faunas of the Aifat Mudstone are: Sterlingian to Aktasian (Archbold, 1990) and late Artinskian (late Baigendzhinian) or early Kungurian (late Early Permian) for brachiopods (Archbold, 1981); possibly late Baigendzhinian but no younger for pelecypods (Dickins & Skwarko, 1981), Baigendzhinian for cephalopods (Glenister & others, 1983), and late Baigendzhinian for Bryozoa (Wass, 1989).

Two samples of float with lithologies typical of the outcropping Aimau Formation were collected for conodont examination. Sample 79CP199F is a brown marl collected in the outcrop area of the Triassic to Early Jurassic Tipuma Formation in the Aimau River a few kilometres downstream from the Aimau Formation outcrop, and sample 79SH63B is a grey bioclastic limestone collected from the Aifam River close to the Aimau Formation/Aisasjur Formation contact (i.e., near the base of the Aimau Formation; Fig. 4).
Conodont fauna

Eight conodont elements (Table 1) were recovered from the two samples considered to be derived from the Aimau Formation. This small fauna is useful for dating the unit because the two species identified in it provide some biostratigraphic control. *Hindeodus minutus* is restricted to the Late Carboniferous and Early Permian (von Bitter & Merrill, 1985). Merrill (1973) observed that older forms of *H. minutus*, 'pre-Missourian', tend to be larger than later forms, and the elements examined in this study are relatively large forms.

The second species identified is assigned to the genus *Neognathodus*, which has a distribution in the early part of the Pennsylvanian (Morrowan to Desmoinesian) or the late Namurian to late Westphalian. If the *Neognathodus* species is interpreted as *N. bassleri*, a late Morrowan (early Westphalian A) age would be indicated (Jones, 1991), but, if it represents one of several younger species, the age could be as young as Desmoinesian (Westphalian D). The broken nature of both specimens recovered prevents positive identification of the elements.

Thus both faunal elements support a Late Carboniferous age; the parts of the unit (probably Aimau Formation) from which the samples were shed are probably no older than Westphalian A, but possibly as young as Westphalian C or D.

![Figure 5. Conodonts from the Modio Dolomite.](image-url)

All figures × 145, except as noted. 1: M element, ?Genus & species indt., CPC 23137 (80CP516A), left element; a, outer lateral view; b, inner lateral view. 2: Sa element, *Scotopodus* sp. indt., CPC 23138 (80CP516A); a, lateral view; b, outer lateral view. 3: S element, Genus & species indt., CPC 23139 (80CP516A), right element; a, outer lateral view; b, inner lateral view. 4: S element, Genus & species indt., CPC 23140 (80CP516A), left element; a, inner lateral view; b, outer lateral view. 5: *Panderodus* cf. *P. simplex*, CPC 23141 (80CP516A), right element (x 130); a, inner lateral view; b, oral view; c, outer lateral view.
Regional relationships of the conodont fauna

Records of Late Carboniferous conodonts in Australia are limited to a report by Palmieri (1969) from Queensland. They include elements assigned to both Neognathodus (Gnathodus) bassleri and Hindeodus (Spathognathodus) minutus. No conodonts are known from potentially stratigraphically equivalent rocks in the Bonaparte Basin or Arafura Basin of northern Australia (Bradshaw & others, 1990).

Metcalfe (1980, 1983) reported on two Late Carboniferous conodont faunas from West Malaysia and Sumatra. The fauna from Sumatra, on the West Malaya Block, is of Visean age, and the fauna from West Malaysia, on the East Malaya Block, is of Namurian age. Both of these faunas are slightly older than the fauna in the samples considered to have been shed from the Aimau Formation.
Figure 7. Conodonts from the Aimau Formation.
All figures x150: 1: Neognathodus cf. N. bassleri, CPC 23148 (79SH63B); a, oral view; b, lateral view. 2: Neognathodus cf. N. bassleri, CPC 23149 (79SH63B); a, oral view; b, lateral view. 3: Hindeodus minutus, CPC 23150 (79CP199F); lateral view, posterior broken and missing, blade broken and restored. 4: Hindeodus minutus, CPC 23151 (79SH63B); a, lateral view; b, oral view; tips of five denticles over posterior part of basal cavity broken, but recovered in sample.

Systematic palaeontology

Genus Hindeodus Rexroad & Furnish, 1964

Type species. Spathognathodus cristulus Youngquist & Miller, 1949.

Hindeodus minutus (Ellison, 1941)
Figures 7.3, 7.4

Material studied. 2 Pa elements.

Description. Carminate Pa element with prominent cusp, 11 post-cusp denticles, three denticles on the anterior margin of the cusp and an enlarged basal attachment area.

Remarks. The single whole specimen recovered in this study conforms with the revised analysis of the species given by Merrill (1973).

Genus Neognathodus Dunn, 1970

Type species. Polygnathus bassleri Harris & Hollingsworth, 1933.

Neognathodus cf. N. bassleri
(Harris & Hollingsworth, 1933)
Figures 7.1, 7.2

Material studied. 2 Pa elements.

Description. Carminiscaphate element with long anterior blade and platform with lateral parapets. The nodes of the inner parapet are larger than those of the outer parapet. The basal attachment area is enlarged.

Remarks. Both elements are tentatively assigned to Neognathodus cf. N. bassleri because the morphology of the posterior portion of the platform, broken on both specimens, is critical to the identification of the various species of Neognathodus. Both elements appear to have a well developed inner lateral parapet, but the length of the parapet cannot be determined.

Genus Panderodus Ethington, 1959

Type species. Paltodus unicostatus Branson & Mehl, 1933.

Panderodus cf. P. simplex (Branson & Mehl, 1933)
Figure 5.5

Material studied. One element.

Remarks. The single specimen recovered in this study is tentatively considered to be an Sd element because it is highly asymmetrical. The element is robust with a longitudinal groove on one lateral face, indications of some fine striae on the lateral surfaces, and wrinkles at the basal margin. The element is adenate.

The specimen is similar to P. simplex, as it is a robust element with a relatively long base and short cusp. Other species of Panderodus tend to be more delicate or have a relatively greater length of cusp-to-base proportion.

Genus Scolopodus Pander, 1856

Type species. Scolopodus sublaevis Pander, 1856.

Scolopodus sp. indt.
Figure 5.2
Material studied. One element.

Description. Coniform element with smooth anterior face and well developed costae on the posterior face. The element is recurved posteriorly.

Remarks. The element is assigned to Scolopodus because it has well developed costae.

Genus & species indt. A
Figures 5.1, 6.1–6.6

Material studied. 19 elements.

Diagnosis. Multimembrate apparatus of at least four element types — Sa, Sc, Sd, and P elements — distinguished. An M element is also possibly present. The Sa element is alate with a denticulate posterior process; the Sc element is bipennate; the Sd element is digyrate; the possible M element is makellate; and the P element is angulate. All elements have denticulate processes.

Description. All elements included in this species have laterally compressed processes with small denticles fused to near their apex. The denticles of one of the Sd elements (Fig. 5.2) appear to be striate, but too little surface texture of the other elements is preserved to determine if this is true of all elements. The Sa element is alate with very short lateral processes and a moderately long posterior process. The Sc elements are bipennate with an inwardly bent anterolateral process. The Sd element has a large but short inner lateral process, and a long posterior process that is developed at a right-angle to the inner lateral process. The P elements are angulate with enlarged lateral flanges over the basal cavity and steeply downward-sloping anterior and posterior processes. Both processes appear to be denticulate but are poorly preserved.

Remarks. The elements recovered and assigned to this species are all poorly preserved, and no proper generic assignment can be made. The specimens would be assigned to Ozarkodina, except for the well developed posterior process of the Sa element. The Sd element is similar to elements of the form genus Plectospathodus, except for the bent posterior process. The processes are too laterally compressed and the denticles too short for the elements to be assigned to Oulodus. The two poorly preserved P elements are similar to the P elements of Plectodina or Aphelognathus, but the rest of the elements do not support this assignment.

Genus & species indt. B
Figures 5.3, 5.4

Material studied. 2 elements.

Description. Coniform element with smooth outer face and large carina on the inner face. Anterior and posterior margins sharp.

Remarks. These elements are too poorly preserved to assign to a genus.

Acknowledgements

The samples were collected by IGOMP field geologists C.J. Pigram (BMR) and A. Safri Hakim (Geological Research and Development Centre, Bandung). P.E. Pieters, C.J. Pigram, and J.H. Shergold read an early draft of the manuscript, and made suggestions regarding its improvement. I. Metcalfe provided a critical review of the manuscript. Gail Hill drew Figures 1 to 4, and A.T. Wilson photographed the conodont specimens depicted in Figures 5 to 7. E. Robinson keyed-in the manuscript to a word-processing document file, and transferred it on to a floppy disc, which was used to generate the computer-tyepsed printer’s copy.

References


Branson, E.B., & Mehl, M.G., 1933 — Conodonts from the Bainbridge (Silurian) of Missouri. The University of Missouri Studies, 8, 39–52.


Idrs, M.B., 1989 — Early Silurian multielement conodont assemblages from Pulau Tanjung Dendang, northeastern Langkawi Islands, Kedah. Warta Geologi, 15, 63–68.

Igo, H., & Koike, T., 1968 — Ordovician and Silurian conodonts from the Langkawi Islands, Malaya, part II. Geology and Palaeontology of Southeast Asia, 8, 1–21.


Early Permian Brachiopoda from Irian Jaya

N.W. Archbold

A new Early Permian (Aktastinian–early Baigendzhinian, early Arinskian) brachiopod fauna is described from the Aiduna Formation, which crops out in the upper reaches of the Mapia River, southern Irian Jaya. New taxa are Neochonetes (Sommeriella) irianensis sp. nov., Chonetinella aidunaensis sp. nov., Sulcataria skwarkoi sp. nov., and Aulosteges tenuispinosus sp. nov. Although material is fragmentary, the new fauna demonstrates significant links with Early Permian faunas of Western Australia and peninsula Thailand. The Carboniferous–Permian brachiopod biostratigraphy of Irian Jaya is reviewed, and known faunas are shown to range in age from probable Namurian to early Kungurian.

Introduction

The Irian Jaya Geological Mapping Project (IJGMP), for which fieldwork was carried out jointly by BMR and the Indonesian Geological Research and Development Centre between 1978 and 1981, yielded substantial new information on the late Palaeozoic geology of Irian Jaya. Work before the project started was summarised by Visser & Hermes (1962), who provided details of all previous palaeontological determinations from Palaeozoic rocks in Irian Jaya. Despite this earlier work, little precise information was available for the late Palaeozoic succession; ages assigned were invariably referred to the ‘Permian’.

New palaeontological data from the late Palaeozoic succession have been published for a number of groups: plants (Prasad, 1981), bivalves and gastropods (Dickins & Skwarko, 1981), brachiopods (Archbold, 1981a, 1981b, 1981c, 1991; Archbold & others, 1982), trilobites (Archbold, 1981c), cephalopods (Glenister & others, 1983), and bryozoans (Wass, 1989). Data on a new, moderately diverse Early Permian brachiopod fauna from the Aiduna Formation are presented herein.

Stratigraphy, locality, and preservation

The stratigraphy of western Irian Jaya is reviewed by Pieters & others (1983). In the northwest (Birds Head), a Late Carboniferous–Permian sequence constitutes the Aifam Group. This unit, from which previous brachiopod faunas have been described, is divided into three formations: Aimau Formation (oldest), Aifat Mudstone, and Anim Formation.

Farther east, partly contemporaneous rocks make up the Aiduna Formation. Though Pieters & others (1983) included this unit in the Aifam Group, more recent work has suggested that it be excluded (Pigram & Panggabean, 1989). The Aiduna Formation consists of well bedded feldspathic and micaceous lithic sandstones interbedded with carbonaceous shale and siltstone.

The new material described herein is from a float boulder from the Aiduna Formation. Though Pieters & others (1983) included this unit in the Aifam Group, more recent work has suggested that it be excluded (Pigram & Panggabean, 1989). The Aiduna Formation consists of well bedded feldspathic and micaceous lithic sandstones interbedded with carbonaceous shale and siltstone.

Age of assemblage

Determination of the age of the assemblage must be regarded as provisional, because many of the species are inadequately known (and left in open nomenclature). Nevertheless, in view of the remoteness of the source region, it is unlikely that further collections will be forthcoming in the foreseeable future. As discussed in detail under the appropriate systematic descriptions, the assemblage demonstrates significant links with late Sakmarian (Sterlitamakian) and early Artinskian (Aktastinian) faunas of Western Australia and peninsula Thailand.

Biostratigraphy of the Aiduna Formation and Aifam Group

The new assemblage adds to the knowledge of the biostratigraphy of the late Palaeozoic of Irian Jaya by providing data on an early Artinskian fauna. As with other recent data on the late Palaeozoic succession, fossils come from isolated localities within broadly defined sections. Nevertheless, four brachiopod faunal horizons appear to be recognisable within these correlative sections to date, as briefly summarised below.

- **Syringothyris irianensis Zone.** This assemblage zone from float material, apparently from the Aimau Formation, represents the oldest yet discovered, and includes Syringothyris irianensis Archbold, Derbyioid? sp., Beecheria? sp., and poorly known aulostegid and spiriferinid material. A Namurian age was favoured for the assemblage by Archbold (1991).

- **Taeniothaerus aifamensis Zone.** This poorly known assemblage from low in the Aifat Mudstone includes only the species Taeniothaerus aifamensis Archbold and undescribed...
pectinacean bivalves. The similarity of this species to forms in comparable faunas of Kashmir and the Pamirs suggested a Sterlitamakian age to Archbold (1991).

- **Neochonetes (Sommertella) irianensis Zone.** This zone is represented by the assemblage described herein, for which an early Artinskian (Aktastinian–early Baigendzhinian) age is preferred.

- **Stereochia irianensis Zone.** This assemblage zone, also from the Aifat Mudstone but from a level higher than the *Taeniothaerus aifamensis* Zone, is represented by the youngest fauna described to date. The assemblage, described by Archbold (1981a, b), shows significant links with the brachiopod fauna from the Rat Buri Limestone, peninsula Thailand (Waterhouse & Piyasin, 1970; Grant, 1976). It includes *Streptorhynchus* sp., *Rhipidomella* sp., *Chonetinella? ainiti* Archbold, *Stictozoster cf. leptus* Grant, *Stereochia irianensis* Archbold, *Linoproductus pigram* Archbold, *Cancrinella* sp., *Stenoscisma ratmani* Archbold, *Stenoscisma cf. tetricum* Grant, *Cruricella? sp.*, *Callispirina* sp., *Spiriferellina* sp., *Hustedia cf. ratburiensis* Waterhouse & Piyasin, *Cleiothyridina* sp., and, at a different locality, *Quinquenella magnifica* Archbold. On the basis of similarities with Thai and Western Australian faunas, a late or latest Baigendzhinian–early Kungurian age was favoured for the assemblage (Archbold, 1981a).

**Palaeogeographical importance of Permian faunas of Irian Jaya**

The Permian faunas of Irian Jaya occupy a pivotal palaeogeographical position between the Gondwanan, intracratonic faunas of Western Australia and the Shan–Thai (or Sibumasu) terrane faunas of peninsula Thailand. The late Early Permian *Stereochia irianensis* assemblage was shown to possess strong links with Thai faunas (Archbold, 1981a; Archbold & others, 1982) and yet retain links with Western Australian faunas. The somewhat older *Neochonetes (Sommertella) irianensis* assemblage, described herein, demonstrates a rather stronger Western Australian element, but still shows links with Thai faunas (and has distinctive components, such as *Sulcataria*, not known elsewhere in the Western Australian and Shan–Thai region).

Faunal relationships point to close palaeogeographical links between Western Australia, Irian Jaya, and peninsula Thailand during the Early Permian. The growing body of data on such links indicates that the concept of a Permian Tethys Ocean appears to be unrealistic, and that detailed continued unravelling of Asian geology is required in order to understand the timing of tectonic events in the region. Recent reviews of the role of palaeontological data for providing critical constraints on such tectonic models include those of Archbold (1987), Shi & Waterhouse (1990), and Bambach (1990).

**Systematic palaeontology**

All specimens are housed in the Indonesian Macropalaeontological Collections (IMC prefix), Geological Research and Development Centre, Bandung, Indonesia. All specimens were coated with ammonium chloride for the illustrations.
Type species. *Chonetes dominus* King, 1938.

**Diagnosis.** The diagnosis provided by Archbold (1981d, p. 113) is followed herein. Possible subgeneric groupings of species of *Neochonetes* are discussed in the same work, and the new species described here appears to fall within the subgenus *Sommeriella* Archbold (1982a; = *Sommeria* Archbold 1981d).

Subgenus *Sommeriella* Archbold, 1982a

**Type species.** *Sommeriella* sp. nov.

**Description.** Ventral valve relatively weakly convex, sulcus distinct, of variable depth, broadens widely anteriorly, usually arises close to umbo. Dorsal valve gently concave with low median fold corresponding to ventral sulcus. Interareas very low. External ornament of fine costellae, increasing in number by bifurcation and intercalation. Growth lines indistinct. Hinge spines distinct, curved, emerge at high angle (50°).

Cardinal process poorly known, appears low. Dorsal median septum distinct, half of valve length, arises anteriorly of deep alveolus as do the short lateral septa. Dorsal valve strongly concave with prominent median fold corresponding to ventral sulcus. Interareas very low. Ornament of fine costellae, increasing in number by bifurcation. Growth lines indistinct. Hinge spines distinct, curved, emerge at high angle (30°+).

Figure 2. A—S. Neochonetes (Sommeriella) irianensis sp. nov.

Figure 3. A–P. Chonetinella aidunaensis sp. nov.

**A, B.** Holotype, IMF 119, latex cast of dorsal valve external mould and external mould, x 5.

**C, D.** IMF 120, latex cast of dorsal valve external mould and external mould, x 5.

**E, F.** IMF 121, latex cast of dorsal valve internal mould, x 5.

**G.** IMF 122, latex cast of dorsal valve internal mould, x 4.5.

**H.** IMF 123, latex cast of dorsal valve internal mould, x 4.5.

**I.** IMF 127, latex cast of ventral valve external mould, x 5.

**J.** IMF 124, latex cast of dorsal valve internal mould, x 4.5.

**K.** IMF 125, latex cast of dorsal valve internal mould, x 4.5.

**L.** IMF 126, latex cast of dorsal valve internal mould, x 5.

**M.** IMF 129, latex cast of ventral valve internal mould, x 4.5.

**N.** IMF 130, latex cast of ventral valve internal mould, x 4.5.

**O.** IMF 131, latex cast of ventral valve internal mould, x 4.5.

**P.** IMF 132, latex cast of ventral valve internal mould, x 4.5.

**Discussion.** The small size, subquadrate outline and pronounced concavo-convexity of the new species readily distinguish it from Early Permian species from Thailand described by Grant (1976) and Waterhouse (1981).

*Chonetinella* aidunaensis Archbold (1981a, pl.1, figs. 5–9, 11–13) from the late Early Permian Aifat Mudstone, Irian Jaya, is a small species, less subquadrate and more deeply concavo-convex than the new species; perhaps the new species is ancestral.

The Kungurian *Chonetinella* sp. from the Coolkilya Sandstone, Carnarvon Basin, Western Australia (Archbold, 1981d, fig. 12D), has a similar outline to the new species, but is otherwise too poorly known for closer comparison.

*Chonetinella aidunaensis* sp. nov. has the general shell shape and internal details that resemble some species of the Late Permian genus *Waagenites* as described by Fang (1983) and Archbold (1988), but is distinguished from those species by its finer external ornament.

Subfamily Svalbardiinae Archbold, 1982b

**Diagnosis.** Externally smooth rugosochonetids. Small to medium size. Dorsal exterior pseudocapillate when worn. Hinge spines at low to moderate angle.

**Genus Sulcataria** Cooper & Grant, 1969

**Type species.** *Chonetina? rostrata* Dunbar & Condra, 1932.

**Diagnosis.** Small, smooth rugosochonetid with ventral valve deeply sulcate, dorsal valve flatly concave with strong median fold. Deep antero-median trough in dorsal valve.

**Discussion.** The basis for the recognition of the Svalbardiinae was discussed by Archbold (1982b). Afanasyeva (1988), in a major review of the order Chonetida, suppressed the Svalbardiinae and included the name genus within her Chalimochonetinae. Even if her classification scheme is accepted, the Svalbardiinae has priority over the Chalimochonetinae.

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*Chonetinella aidunaensis* sp. nov. has the general shell shape and internal details that resemble some species of the Late Permian genus *Waagenites* as described by Fang (1983) and Archbold (1988), but is distinguished from those species by its finer external ornament.

Subfamily Svalbardiinae Archbold, 1982b

**Diagnosis.** Externally smooth rugosochonetids. Small to medium size. Dorsal exterior pseudocapillate when worn. Hinge spines at low to moderate angle.
**Sulcataria skwarkoi** sp. nov.

**Fig. 4A–F**

**Etymology.** Named for Dr S.K. Skwarko, who collected many late Palaeozoic specimens of brachiopods during the IJGMP fieldwork.

**Holotype.** A dorsal valve internal mould, formed by leaching a preserved dorsal valve in rock (IMF 133).

**Material.** 2 ventral valve external moulds (IMF 134–135), 2 dorsal valve external moulds (IMF 136–137), and 1 dorsal valve internal mould, holotype (IMF 133).

**Measurements.** Holotype, IMF 133: maximum width, 8.4 mm; hinge width, 8.2 mm; height of valve, 5.5 mm. IMF 134 (ventral valve): maximum width, 8.5 mm; hinge width, 8.1 mm; height of valve, 5.7 mm.

**Description.** Shell outline subquadrate. Ventral valve strongly convex. Sulcus distinct, arises anteriorly of umbo and deepens markedly anteriorly of mid-length of valve. Dorsal valve flatish to gently concave with prominent dorsal fold. Interareas very low.

Exterior of valves smooth, traces of pseudocapillae on worn areas of dorsal valves adjacent to median fold. Growth lines visible on anterior of valve exteriors.

Dorsal interior unknown. Dorsal interior with minute cardinal process. Dorsal median septum short, developed anteriorly of small deep alveolus, fills in the posterior section of the median trough of dorsal interior. Lateral septa short, low. Sockets distinct with pronounced inner socket ridges and small sharp outer socket ridges. Brachial ridges weakly developed, but on distinct raised region of valve interior. Papillae developed in radial rows on central anterior region of interior. Posterior lateral flanks of valve interior smooth. Pronounced, deepanteromedian trough developed corresponding with dorsal fold anterior.

**Discussion.** Many of the small chonetid specimens within the new fauna are external or internal moulds of isolated valves. Hence it is by no means clear that all internal moulds of valves are correctly assigned between *Chonetinella* and *Sulcataria*. Nevertheless, the holotype dorsal valve possesses a distinctive interior when compared with those assigned to *Chonetinella*.

*Sulcataria rostrata* (Dunbar & Condra) as described by Cooper & Grant (1975, pl. 480, figs. 37–55) possesses a broader sulcus anteriorly than the new species, and a prominent dorsal median septum. *Sulcataria compacta* Cooper & Grant (1975) from the Neal Ranch Formation (Asselian) is closer to the new species; it has a similar shell outline, but a narrower dorsal fold and more strongly developed dorsal internal structures. *S. latisulcata* Cooper & Grant (1975) from the same formation and age has a much lower dorsal fold and weaker ventral sulcus than the new species.

Of considerable interest are the Late Permian records of *Sulcataria* from the Himalayas. *S. pentagonalis* from the Djuflan of Nepal (Waterhouse, 1978, pl. 1, figs. 13–16) is based on poorly preserved material, but possesses a broader fold than the new species. Specimens from the Djuflan of Kashmir (Waterhouse & Gupta, 1979, pl. 1, figs. 5, 7; pl. 2, figs. 1–5) are at times more alate with ears, but other specimens are subquadrate and hence comparable with the new species. Dorsal internal features are weakly developed as in *S. skwarkoi* sp. nov.

**Order Productida** Sarycheva & Sokolskaya, 1959

**Suborder Strophalosiida** Waterhouse, 1975

**Superfamily Strophalosiacea** Schuchert, 1913

**Family Strophalosiidae** Schuchert, 1913

**Subfamily Strophalosiinae** Schuchert, 1913

**Genus Heteralosia** King, 1938

**Type species.** *Heteralosia slocomi* King, 1938.

**Discussion.** *Heteralosia* has been reviewed by Archbold (1986), who restricted the genus to small Strophalosiinae with no dorsal spines, while agreeing with Waterhouse (1959) that care was needed with ontogenetic studies in the group.

**Heteralosia** sp.

**Fig. 5D, E**

**Material and measurements.** 1 dorsal valve external mould (IMF 138): height of valve, 4.6 mm; estimated width of valve, 5.0 mm. 1 ventral valve internal mould (IMF 139): height of valve, 7.8 mm; width of valve, 7.2 mm.

**Comments.** The two specimens are referred to *Heteralosia* because of their small size and general similarities to species described from the Early Permian of Thailand (Grant, 1976), Texas (Cooper & Grant, 1975) and Western Australia (Archbold, 1986). Material is inadequate for detailed comparison with described species.
Genus *Echinalosia* Waterhouse, 1967

**Type species.** *Strophalosia maxwelli* Waterhouse, 1964.

**Diagnosis.** Medium to large Strophalosiinae with circular to elongate shell outline. Dorsal valve with unthickened trail. Dorsal spines finer than ventral spines.

**Discussion.** Distinctions between *Echinalosia* and related genera with dorsal spines have been discussed by Archbold (1986) and Waterhouse (1986), and will not be repeated here.

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**Echinalosia sp.**

**Fig. 5A–C**

**Material and measurements.** 1 incomplete dorsal valve internal mould (IMF 140b) and corresponding incomplete external mould (IMF 140b); estimated valve width, 38 mm; estimated valve length, 35 mm.

**Comments.** The material is provisionally referred to *Echinalosia* on the basis of its moderate size, presence of dorsal spines (although only preserved on the anterior margin of the external mould), and lack of valve thickening. A striking feature of the specimen is the cardinal process, which does not project above the hinge line unlike other species of the genus.
Superfamily *Aulostegacea* Muir-Wood & Cooper, 1960  
Family *Aulostegidae* Muir-Wood & Cooper, 1960  
Subfamily *Aulosteginae* Muir-Wood & Cooper, 1960  
Genus *Aulosteges* von Helmersen, 1847

**Type species.** *Orthis wangenheimi* de Verneuil, 1845 (= *Aulosteges variabilis* von Helmersen, 1847).

**Discussion.** The type species was described from poor material by de Verneuil (1845, p. 194, pl. 11, fig. 5), and fully described, under a different name, by von Helmersen (1847; 1848, figs. 1–12). Von Helmersen (1853) subsequently clarified the confusion over the two specific names. More recent studies of the type species have included those by Likharev (1959, pl. 4, figs. 1–8, pl. 5, figs. 1–5), Muir-Wood & Cooper (1960, pl. 10, figs. 14–19), and Grigoryeva (1962, pl. 5, figs. 1–4).

*Aulosteges tenuispinosus* sp. nov.

**Holotype.** A complete dorsal valve external mould (IMF 141a) and incomplete corresponding dorsal valve internal mould (IMF 141b).

**Additional material.** 1 incomplete external mould of juvenile portion of dorsal valve (IMF 142).

**Measurements of holotype.** Valve width, 45.5 mm; valve length, 40 mm; length of dorsal septum, 21 mm.

**Description.** Medium-size subquadrate dorsal valve. Widest anterior of mid-length. Thin-shelled. Ventral valve unknown. Dorsal valve gently convex at juvenile stage, becoming flattish and then markedly upturned. Narrow dorsal fold distinct, arises close to umbo. Valve exterior covered in small, elongate dimples and numerous closely scattered fine spine bases. Dorsal interior with pronounced blade-like median septum anterior of deep alveolus on either side of which are pronounced ridges. Interior face of cardinal process not preserved, but exterior face apparently massively trilobate judging from outline trace preserved on external mould. Anterior of dorsal interior strongly papillose. Adductor scars dendritic, not depressed.

**Discussion.** Despite the absence of ventral valve data and the limited nature of the material, the species is named because of the distinctive, well preserved material to hand.

The species is only provisionally referred to *Aulosteges*, of which the type species lacks the distinct dimples on the dorsal valve exterior as indicated by Likharev (1959) and Grigoryeva (1962). *A. tenuispinosus* sp. nov. is close to Western Australian Early Permian species as described by Coleman (1957), such as the Sterlitamakian *A. baracoodensis* Etheridge (1903) and the Artinskian *A. ingens* Hosking (1931) and *A. lyndonensis* Coleman (1957). These species all appear to be closely related and, in turn, allied to *A. tenuispinosus* sp. nov., but possess less subquadrate dorsal valves than the new species.

Suborder *Productidina* Waagen, 1883  
Superfamily *Productellacea* Schuchert & Le Vene, 1929  
Family *Productellidae* Schuchert & Le Vene, 1929  
Genus *Stictozoster* Grant, 1976

**Type species.** *Stictozoster leptus* Grant, 1976.

**Diagnosis.** The diagnosis provided by Grant (1976, p. 96) is accepted.

**Discussion.** *Stictozoster* has been reviewed by Waterhouse (1981) and Archbold (1981a, 1984). The genus has a late Sakmarian (Sterlitamakian) to Kungurian range in the southeast Asian–Western Australian region, and a Kungurian–Kazanian range in the Arctic. *Stictozoster* has previously been recorded from the late Early Permian of Irian Jaya (Archbold, 1981a).

*Stictozoster* sp.

**Fig. 5L**

**Comments.** A single dorsal valve external mould (IMF 143, with estimated width of 33.4 mm, and valve length of 21 mm) reveals the distinctive finely dimpled exterior with fine spines and lamellose growth lines of the genus. The specimen is large for the genus, but additional material is required for detailed comparison with described species.

Superfamily *Linoproductacea* Stehli, 1954  
Family *Linoproductidae* Stehli, 1954  
Subfamily *Linoproductinae* Stehli, 1954  
Genus *Cancrinella* Fredericks, 1928

**Type species.** *Productus cancrini* de Verneuil, 1842 (in de Koninck, 1842).
Cancrinella sp.  
Fig. 5M

Comments. A single incomplete internal mould of a juvenile ventral valve (IMF 144, with valve length of 7.2 mm) indicates the presence of Cancrinella. The specimen is inadequate for detailed comparison with described species of the genus.

Order Spiriferida Waagen, 1883

Comments. The order Spiriferida is represented in the fauna by five incomplete specimens (Fig. 6). Neosphirifer (IMF 145) is present, and two different syringothyridacean species are indicated (IMF 146–147). The punctate Spiriferinidina are represented, and two different syringothyridacean species are indicated (IMF 148). The other has broad, widely spaced, flattened plications and a capillate micro-ornament (IMF 149).

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References

Archbold, N.W., 1988 — Studies on Western Australian Permian brachiopods 8. The Late Permian brachiopod fauna of the Kirkby Range Member, Canning Basin. Proceedings of the Royal Society of Victoria, 100, 21–32.

Cooper, G.A., & Grant, R.E., 1975 — Permian brachiopods of West Texas 111. Smithsonian Contributions to Paleobiology, 19, 795–1921.
Fang, R.-S., 1983 — The Early Permian Brachiopoda from Xiaoxinzhai of Gengma Yunnan and its geological significance. Contributions to the Geology of the Qinghai–Xizang (Tibet) Plateau, 11, 93–119.
Hosking, L.P.F., 1931 — Fossils from the Wooramelt District, Western Australia. Journal of the Royal Society of Western Australia, 17, 7–52.
Likharev, B.K., 1959 — Nekotorye navlyudeniya nad zhivkhepalocoziskimi brachiepodyami. Palaeontologicheskii Zhurnal, 1959, 82–90.


A record of the genus *Lockhartia* (Foraminiferida) from Misool Archipelago, Irian Jaya

D.J. Belford

The foraminiferal genus *Lockhartia* is recorded from the previously restricted occurrence in the western part of the Indo-Pacific region.

Occurrence and significance of fauna

In 1981, field geologists of the Irian Jaya Geological Mapping Project sent thin sections of rock samples to BMR for examination and age determination. Among them was one section of a foraminiferal-algal grainstone (sample 81UH57A), considerably extending its Indo-Pacific region.

The sample contains several specimens of a species of the foraminiferal genus *Lockhartia*, and one small specimen of *Discocyclina* sp. with a heavy calcitic overgrowth, probably algal. Algae are also common, including numerous fragments of the species *Distichoplax biserialis* (Dietrich). Unfortunately no further material is available, and there is no possibility of collecting additional material at this time.

The specimens lack the labyrinthine umbilical area of the *L. diversa* group; the chambers are wedge-shaped in axial section, not semi-lunar as in the *L. tipperi* group. The *L. conditi* group has very large pillars, a feature not shared by the Misool specimens. The observed specimens are not referable to any described species. Figured specimens are deposited in the Commonwealth Palaeontological Collection, Canberra, Australia, under numbers 24927 to 24932.

The genus *Lockhartia* ranges from Late Paleocene to Middle Eocene (Adams, 1970), and *Distichoplax biserialis* occurs in the Late Paleocene and Early Eocene (see, for example, Keij, 1963). The sample therefore has a Late Paleocene to Early Eocene age. The fauna suggests a shallow-water back-reef or lagoonal deposit, and the preservation of some individual fossils indicates a period of abrasion and overgrowth by other marine organisms before lithification. This is the first record from Irian Jaya of carbonate sediments of this age. Earlier records of Late Paleocene deposits (Visser & Hermes, 1961) were based on planktonic deep-water faunas; beds of earliest Eocene age have not been identified definitely from Irian Jaya.

1. 45 Village Road, South Durras, NSW 2536.
The main interest in this fauna is the occurrence of *Lockhartia* sp. Smout (1954) stated that this genus is known with certainty only from India, east Africa, Arabia, and Iraq; Adams (1970) recorded it only from the western part of the Indo-Pacific region. Ruten (1948) recorded *Lockhartia* (as redeposited specimens) from northern Borneo; after examining the material, however, Smout (1954) considered that these specimens were not *Lockhartia*, but possible poorly preserved forms of *Rotalia mexicana* Nuttall and of *Pellatispira* Liechti & others (1960) recorded *Lockhartia* sp. from the Belaga Formation of central Sarawak, but did not present any figures. Belford (1974) recorded genus indet. cf. *Lockhartia* from the Nassau Range, Irian Jaya.

This record of the genus *Lockhartia* from the Misool Archipelago considerably extends the palaeogeographic distribution of the genus, and indicates west-to-east migration from its presumed centre of origin in the western Indo-Pacific region.

**Discussion**

*Lockhartia* is known from neither the Mediterranean nor north Africa. Its absence from these regions could indicate one of two things: there was no connection between these regions and that containing the Ranikot faunas in Paleocene to Middle Eocene times, or conditions were not favourable for east-to-west migration. Davies (1940) considered that these regions were not connected in the Paleocene, but later (1949) postulated a connection between India and the Sahara (Africa) to account for the Palaeogene distribution of the group of cordate...
operculines. On the basis of more recent palaeontological work, Adams (1967) noted that the two regions had faunal elements in common, and concluded that a connection had existed, probably via the Persian Gulf. In a later paper, Adams (1983, fig. 14.1) indicated that no physical barriers to faunal migration existed during the later part of the Paleocene; thus, unfavourable environments or unsuitable ocean currents might have accounted for the absence of Lockhartia to the west. However, Berggren & Hollister (1974) showed an east-to-west current in the Mediterranean in the Cretaceous–Tertiary. The absence of Lockhartia from the Mediterranean is at present unexplained; it may be more apparent than real, reflecting that limestones containing Lockhartia have been not yet fully investigated.

Adams (in discussion in Adams, 1967) noted that evidence for west-to-east migration of larger foraminifera is stronger than that for east-to-west migration; he also noted that evidence suggests that movement from Tethys to the Indo-Pacific was easier than movement in the reverse direction. The occurrence of the genus Lockhartia well to the east of the region where it may be presumed to have originated, and its absence from areas to the west, support these conclusions of Adams. It is to be expected that Lockhartia will in future be recognised in the area between Irian Jaya and the western Indo-Pacific.

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References


Smout, A.H., 1954 — Lower Tertiary Foraminifera of the Qatar Peninsula. British Museum (Natural History), London.

Belemnite successions and faunal provinces in the southwest Pacific, and the belemnites of Gondwana

A.B. Challinor

Dinoflagellate zonation allows a reasonably precise correlation of some Jurassic-Cretaceous coleoid-bearing sequences in Indonesia, Papua New Guinea, and New Zealand. Macrofossil evidence provides a precise New Zealand -- New Caledonia correlation in the Middle Jurassic, and general correlation allows Australian belemnite-bearing sequences to be compared with those of other regions.

Comparisons of Indonesian and Papua New Guinean with New Zealand sequences show that the time-distribution of Belemnopsis is partly disjunct. Hibolitides and Conodicoelites is mostly disjunct, and Dicoelites is completely disjunct. Xiphoteuthids and Belemnitina are known only from New Zealand and New Caledonia, Duvalia and Chaladorbulaeae only from Indonesian and the Dimitobilidae only from New Zealand, Australia, and Papua New Guinea. New Zealand Middle and Late Jurassic Belemnopsis are morphologically distinct from those of Indonesia and Papua New Guinea. New Zealand Hibolitides are a morphologically coherent group different from the more diverse Indonesian and Papua New Guinean members of the genus, and Dicoelites and Conodicoelites of the two regions also differ. The much discussed and widely recognised "uhiagi-complex" is incorrect in concept and has little correlative value.

A Jurassic and Early Cretaceous Tethyan Province populated by morphologically similar Hibolitides, and by Duvalia, Dicoelites, and Conodicoelites extended eastward from the Mediterranean region along the coast of Gondwana to at least Papua New Guinea. Within that province Late Jurassic Ethiopian and Indo-Tethyan Subprovinces can be recognised, each based on a distinctive Belemnopsis lineage. A South Pacific Province based on Belemnitina and Belemnopsisina extended from southern South America to New Zealand in the Middle and Late Jurassic, and possibly to Papua New Guinea in the Early Cretaceous. The belemnite assemblages of other former Gondwana continents support this division. Although the Tethyan and South Pacific Provinces remained distinct from Middle Jurassic probably to the Early Cretaceous, some trans-Gondwana migration from Kimmeridgian time, mainly from Tethys to the South Pacific, occurred along the developing seaway between Africa--South America and India--Antarctica.

Introduction

This paper examines the geographic and stratigraphic distribution of belemnites and other coleoids in the southwest Pacific region (here defined as including eastern Indonesia, Papua New Guinea, New Caledonia, Australia, and New Zealand). Much of it is based on my own investigation of Jurassic--Cretaceous faunas of eastern Indonesia, Papua New Guinea, New Caledonia, and New Zealand. Comparisons with other Gondwanan assemblages and with New Zealand Cretaceous taxa are based on the works of others, mostly published during the last 25 years. The philosophy adopted here and in earlier publications is that information on, and the interpretation of, belemnite morphology should be based on large samples wherever possible (i.e., population-based). It uses a statistical approach for easily measured parameters of the guard, and detailed description for non-measurable features, and it makes due allowance for random variation and ontogenetic and possible evolutionary changes (see Challinor & Skwarko, 1982). The approach differs from that of many belemnite researchers.

Two belemnite provinces

2 can be recognised in the southwest Pacific region from mid-Jurassic to Early Cretaceous. One is centred on eastern Indonesia, and includes Irian Jaya, Papua New Guinea, and parts of northern and Western Australia. The other includes New Zealand and most of Australia. New Caledonia, which appears to contain elements of both provinces, cannot be placed firmly in either one.

The classification adopted is that of Jelletzky (1966, 1980), who recognised three suborders within the Order Belemnida Zittel, 1895.

(1) Belemnitina Zittel, 1895, includes the geologically older and apparently most primitive belemnites (family Belenmitidae d'Orbigny, 1845, largely confined to the Early Jurassic and early Middle Jurassic), and younger representatives such as Cylindroteuthididae Stolley (1919) which extend into the Early Cretaceous. Belemnitina are not numerically important in the southwest Pacific, but do occur in New Zealand, New Caledonia, and South America.

(2) Belemnopsisina Jelletzky, 1965, comprises mostly geologically younger belemnites (Middle Jurassic to Cretaceous) that are considered to be the evolutionarily more advanced members of the Order. The families Belemnopsinae Naef, 1922, emend. Jelletzky, 1946, and Dimitobilidae Whitehouse, 1924, are extremely important in the southwest Pacific, and the families Dicoelitidae (Sachs & Nalinaeva, 1967, emend. Jelletzky, 1982) and Duvaliidae Pavlov, 1914, are prominent.

(3) Diplobelina Jelletzky, 1965, is an aberrant and rare suborder of latest Jurassic and Early Cretaceous age containing a single family with a member in Irian Jaya.

Correlation of regional sequences

The regional sequences relevant to this work (Fig. 1) have been correlated primarily by the dinoflagellate and spore-pollen zonation of Helby & others (1987). The zonation was originally developed for the Australian Mesozoic. It is based on well samples from most of the relevant Australian sedimentary basins, together with more limited samples from the Papuan basin. Many of its zones have been recently identified in Indonesia (in the Misool Archipelago by Helby & Hasibuan, 1988; and in the Sula Islands and Irian Jaya by R.W. Helby, Palynological Consultant, personal communication, 1989) and New Zealand (Helby & others, 1988).

An alternative dinoflagellate zonation for the mid-Callovian to early Aptian of Papua New Guinea based on detailed sampling in the Strickland River gorge of central Papua New Guinea has been proposed by Davey (1987). Although his sequence of zones and their age interpretation differ in part from those of Helby & others (1987), the two can be readily correlated (Davey, 1987), and the zonation for Papua New Guinea used in Figure 1 has been established by substituting the relevant zones and ages of Helby & others (1987) for those of Davey (1987).
Figure 1. Correlation of Jurassic and Early Cretaceous regional sequences in the southwest Pacific using the dinoflagellate and spore-pollen zonation of Helby & others (1987) and selected macrofossil data.

Sample numbers for New Zealand column from Helby & others (1988); Misool from Helby & Hasibuan (1988); Sula Islands from Sato & others (1978); and Irian Jaya from Challinor (1989a and new data).
Macrofossils provide limited control for correlating between New Zealand and Indonesia/Papua New Guinea (Fig. 1). The concurrent range-zone of the bivalves *Buchia marayomaorica* (Krumbeck) and *Retroceramus haasti* (Hochstetter; or *R. cf. subhaasti*) immediately above *R. galoi* (Boehn) is known from New Zealand (Helby & others, 1988), Papua New Guinea (Challinor, 1990), the Misool Archipelago (Challinor, 1989c), and the Sula Islands (Sato & others, 1978). The zone of *Buchia* in the late Tithonian. *Buchia* occurs within the *Dingodinium swanense* zone of Helby & others (1987), and reinforces the time-equivalence indicated by the dinoflagellates themselves (Fig. 1). It occurs immediately above the *Wanaea clathrata* zone in Misool, but its relationship to adjacent dinoflagellate zones is not known in the Sula Islands.

A less precise correlation is provided by the bivalve genus *Buchia* in the late Tithonian. *Buchia* occurs within the *Dingodinium jurassicum* dinoflagellate zone in Papua New Guinea (G. Francis, Geological Survey of Papua New Guinea, GSPNG, personal communication, 1988), in the *Omatia montgomeryi* to *D. jurassicum* interval in Misool (F. Hasibuan, University of Auckland, personal communication, 1988), and in the late Tithonian of New Zealand (Li & Grant-Mackie, 1988) and the Sula Islands (Sato & others, 1978). Its relationship to the *O. montgomeryi* and *D. jurassicum* zones in the Sula Islands and New Zealand is not directly known, but the zonal succession suggests that it is close (Fig. 1).

The zone of *Macrocephalites* provides a further macrofossil tie-point between New Zealand, Papua New Guinea, and the Sula Islands (Fig. 1). Although the macrocephalitids were originally regarded as indicators of basal Callovian, those of Papua New Guinea and the Sula Islands range down into the late and possibly middle Bathonian (Westermann & Callomon, 1988). The New Zealand occurrences are generally regarded as Callovian (Stevens & Speden, 1978), but a broader Bathonian–early Callovian correlation is adopted here.

All elements of the New Caledonian Permian–Jurassic biota (except the belemnites) are also known from New Zealand (Avias, 1953; Paris, 1981; Campbell & others, 1985; Pharo, 1967; MacFarlan, 1985; Grant–Mackie, 1985), and correlation between the New Zealand and New Caledonian Early and Middle Jurassic (Fig. 1) is a multi-taxon one. At this time, there are no known taxa which directly correlate the Early Jurassic and most of the Middle Jurassic of New Zealand and New Caledonia with the Indonesian and Papua New Guinean sequences studied here.

The dinoflagellate and macrofossil ages implied in Figure 1 are used only to establish tie-points, and to facilitate comparison of belemnite time distributions. Differences in age interpretations exist, particularly for the Late Jurassic dinoflagellate zones (Davey, 1987; Helby & others, 1987), and the ranges of the correlating macrofossils are far from certain (Crame, 1982, 1983; Helby & others, 1988). The ages indicated in Figure 1 are used for correlation purposes only, and their use does not necessarily imply acceptance. Slightly different belemnite age and range interpretations have been used elsewhere (Challinor, 1989c).

**Stratigraphic occurrences and distributions**

**New Zealand** (Fig. 2; Fig. 7, col. 2)


Most New Zealand Jurassic belemnites have been found in the Kawhia and Southland Regional Synclines (Stevens, 1965; Suggate & others, 1978). The few fragmentary specimens known from the Torlesse Supergroup (Suggate & others, 1978) of the axial ranges to the north and east of the regional synclines have been little studied. Early Cretaceous strata are absent from most of New Zealand. Cretaceous belemnites occur in mid-Albian to Maastrichtian beds, mostly in North Canterbury, Marlborough, and the Wairarapa district. Scattered occurrences are known from the Northland, East Cape, and eastern Otago districts.

The only major work on New Zealand assemblages is by Stevens (1965), who compared Jurassic and Cretaceous belemnites with those of the Indo-Pacific region. Cretaceous faunas have not been re-examined since. Hudson (1983) re-examined the distribution of Toarcian–Callovian taxa in the Kawhia Regional Syncline, and Challinor (1977) described a new *Belemnopsis* later found at the base of several Bajocian sections (Hudson, 1983, figs. 4:1–4:4). Late Jurassic taxa were extensively re-studied in a series of papers by Challinor (1974, 1975a, b, 1979a, b, 1980), based on large collections of new material. Knowledge of stratigraphic distributions was refined, and new interpretations proposed.

Ages and ranges discussed in the publications cited have been re-interpreted (where necessary) in terms of the dinoflagellate/ spore–pollen zonation of Helby & others (1987) and Helby & others (1988), and may differ from those of the original authors.
Early and Middle Jurassic. The earliest belemnoids known from New Zealand appear near the Triassic–Jurassic boundary. *Aulacoceras* and *Prograyphularia*, both members of Aulacoceratidae Mojsisovics, 1882, occur in latest Triassic and earliest Jurassic beds (Trechmann, 1918; Jeletzky & Zapfe, 1967), and *Atractites* (Xiphoteuthididae Naef, 1922) have been recently found in the basal Jurassic of southwest Auckland (Challinor, new data).

The true belemnites (Suborders Belemnitina and Belemnopseina) appear in the Early to Middle Jurassic. *Cylindroteuthis* (or a homeomorph) occurs in the Toarcian–Bathonian, *Brachybelus* and *Belemnopsis* in the Bajocian–Bathonian, and *Hibolithes* in the Bathonian–Callovian (Hudson, 1983, fig. 4:1).

Late Callovian and Oxfordian strata were not recognised in New Zealand until Hudson & others (1987) recorded late Callovian dinoflagellates (*Wanaea spectabilis* zone, confirmed by Helby & others, 1988) and the Oxfordian ammonite *Epinayaites* in a 20-m interval in the mid-Oraka Sandstone at Kawhia Harbour. The same interval had earlier yielded the early Kimmeridgian *Epicephalites* and *Subneumayria* (Fleming & Kear, 1960) and the belemnites *Conodicoelites* and *Belemnopsis*. Both belemnites occur within the upper 4m of the 20-m interval (stratigraphic data from Stevens, 1965), but their stratigraphic relationship to the other fossils is uncertain (Hudson & others, 1987). Stevens (1965:196) considered the *Conodicoelites* to be early Kimmeridgian, but the genus is Bathonian–Callovian in Indonesia and Papua New Guinea (Challinor & Skwarko, 1982; Challinor, 1989c, 1990) and Bajocian–?Oxfordian in India and Europe (Stevens, 1965). New Zealand *Conodicoelites* are regarded here as appearing near the Callovian–Oxfordian boundary.

Late Jurassic. Belemnites are abundant in the Late Jurassic, mostly in the Kawhia Regional Syncline, but a few have been recorded from the Torlesse Supergroup (Stevens, 1965). Although inadequately known, all Torlesse forms seem conspecific with Kawhia taxa. Only Belemnopseina are present.

Oxfordian to Kimmeridgian beds (*Wanaea spectabilis* to *Dingodinium swanense* zones) contain a number of short-ranging *Belemnopsis* (Challinor, 1979b, b, 1980), and *Dicoelites* appears very briefly in the Kimmeridgian (Challinor, 1980; mid–late *D. swanense* zone). Abundant *Belemnopsis* continues into the early Tithonian (Stevens, 1965; *Cribroperidiatum perforans* zone), and are replaced in the early–mid-Tithonian (*C. perforans*-to-*O. mongremonu* interval) by a rich zone of *Hibolithes* (Stevens, 1965; Challinor, 1974, 1975a, b). New data show this zone to extend somewhat lower than previously known, and to be more diverse in species. *Belemnopsis* reappears above the *Hibolithes* zone (Stevens, 1965; probably in the *Dingodinium jurassicum* zone), and is the youngest Jurassic genus known from New Zealand.

Cretaceous. Cretaceous belemnites appear in the mid-Albian, above a regional unconformity which has removed Early Cretaceous beds over much of New Zealand, and extend into the Campanian–Maastrichtian. All are Belemnopseina, family Dimitobelidae Whitehouse, 1924, and are placed in *Dimitobelus* (*Dimitobelus*) Whitehouse or *Dimitobelus* (*Dimitocamax*) Doyle, 1988.

New Caledonia (Fig. 3; Fig. 7, col. 1)

The first belemnites from New Caledonia were described by Avias (1953), and large collections were made later by Pharo (1967). Pharo’s and Avias’s collections and better preserved and more diverse material collected by members of ICGP Project 5 (New Zealand–New Caledonian chronostratigraphic correlation) were studied by Challinor & Grant–Mackie (1989), and most information is from that source. Most collections are from the southwest coast of New Caledonia (Iles Puen, Testard, Mathieu, and Nduke, Uitoe Peninsula, and Baie de St Vincent), but a few specimens are from near Moindou and Guipin in west and central New Caledonia. The material is mostly of Middle Jurassic age, except the Guipin specimens, which are Late Jurassic.

The earliest belemnoid known is a single specimen of the xiphoteuthid aulacocerid *Ausseites* Flower, 1944, of Sinemurian age. The true belemnites follow with a new genus of Belemnitidae d’Orbigny, 1845 (Suborder Belemnitina), appearing in the early Bathonian. The Suborder Belemnopseina is represented by two species of *Dicoelites* (early to late Bathonian), one species of *Hibolites* (Bathonian–Callovian), and two species of *Belemnopsis* (Callovian). Rare *Hibolites* are known from the Kimmeridgian–Tithonian.

Papua New Guinea (Fig. 4; Fig. 7, col. 4)

Early records include *Belemnopsis* and *Parahibolites* from the headwaters of the Wok Fennag (Fly River) in western Papua New Guinea. Glassner (1945, 1958) discussed *Tetrabelas* (later revised to *Dimitobelus*) from the Purari River farther east, and Banner & others (1961) recorded *Dimitobelus* and cf. *Duvalia* from western Papua New Guinea. Large new collections from the western highlands (in the Miannin, SB/54–5, and Ok Tedi, SB/54–7, 1:250 000 Sheet areas) held in the Commonwealth Palaeontological Collection maintained by BMR have been studied recently (Challinor, 1990), and most of the following information is from that source.

*Conodicoelites* (mid-Bathonian associated with *Macrocephalites*; Westermann & Callomon, 1988), *Belemnopsis* (Bathonian–late Tithonian, Berriasian), *Hibolites* (late Callovian, late Tithonian, Berriasian–Valanginian, Aptian–mid-Albian), *Parahibolites* (Albian), and *Dimitobelus* (Cenomanian) are present, most with significant stratigraphic ranges. Davey’s (1987) dinoflagellate zonation does not extend into the late Aptian, and beds of this age cannot be directly correlated with the Helby & others (1987) scheme. Ages adopted for the post-mid-Albian sequence are those currently used by GSPNG.

The *Conodicoelites*, Oxfordian–late Tithonian *Belemnopsis* and *Hibolites*, and Neocomian *Belemnopsis* and *Hibolites* are identical at species level with those of Indonesia. *Parahibolites* and the Aptian–mid-Albian *Hibolites* are not known elsewhere in the Indonesia–Papua New Guinea region, but *Dimitobelus* occurs widely in New Zealand and Australia. *Duvalia* is not present in the new collections. Doyle (1988) has recorded *Parahibolites* in the Albian–Cenomanian, but this appears to be based on an incorrect interpretation of Glassner’s (1945) record of *Parahibolites*, now confirmed in that genus (Challinor, 1990).

Indonesia (Fig. 5; Fig. 7, cols. 5–7)

Research on Indonesian Belemnitidae commenced near the turn of the century. By 1970, nearly 50 papers dealing partly or wholly with the Order had been published (Skwarko & Yusuf, 1982). Some of these are major works dealing only with belemnites (Stolley, 1929, 1935; Kruizinga, 1920; Stevens, 1963c, 1964a, b, 1965). Despite this extensive literature, stratigraphic and taxonomic relationships remained largely unresolved, and the taxonomy is characterised by multiple synonyms (Challinor, 1989c).
In recent years, two of the classical Indonesian Mesozoic regions were revisited. The Sula Islands were re-examined in 1976 (Sato & others, 1978; Westermann & others, 1978), and Misool Archipelago was remapped by Pigram & others (1982). The belemnites of the Sula Islands were described by Challinor & Skwarko (1982), and those of Misool by Challinor (1989c, 1991). Smaller but important collections from Irian Jaya also have been studied (Challinor, 1989a).

The Jurassic sequence of the Sula Islands, largely calcareous shale with minor sandstone and conglomerate, is separated by an unconformity from underlying ?late Palaeozoic-?early Mesozoic rocks. It ranges in age from Toarcian to late Tithonian. Cretaceous beds are present but have not been examined in detail (Sato & others, 1978).

The belemnite-bearing part of the Jurassic–Cretaceous sequence of Misool consists of a basal quartz grit overlain by shale, calcareous sandstone, and sandy and pure limestones. It overlies an erosion surface on Triassic rocks (Pigram & others, 1982), and ranges in age from ?Bathonian-Callovian to Hauterivian (Challinor, 1989a).

None of the Indonesian sequences has been fully zoned in terms of the dinoflagellate zonation of Helby & others (1987), but the Misool column has been partly correlated (Helby & Hasibuan, 1988), and a limited number of zones have been recognised from the Sula Islands and Irian Jaya (R.W.Helby, personal communication, 1989). Additional tie-points are provided in the Misool and Sula sequences by Buchia, the Retroceramus haasti–Malayomaorica zone, and Macrocephalites. A number of other macrofossils provide reasonably accurate ages in Misool (Challinor 1989c; F. Hasibuan, personal communication, 1988).

In most parts of Indonesia only Belemnopseina (Belemnopsectidae and Dicoelitidae) are known, but there is one record of the Diplobelina in Irian Jaya.

Knowledge of well dated Middle Jurassic taxa is derived almost entirely from a few specimens from the Sula Islands. They include Dicoelites (late Bajocian, Callovian, and early Oxfordian), distinctive Conodicoelites (mid-Bathonian; Challinor & Skwarko, 1982; ages revised from Westermann & Callomon, 1988), and late Bathonian Belemnopsis. Stratigraphic ranges of these forms are uncertain; most occurrences are isolated and dated by associated ammonites. Several Conodicoelites, Dicoelites, Belemnopsis, and Hibolithes described by Boehm (1912) and Kruizinga (1920) from Sula are also apparently of Middle Jurassic age, but their stratigraphic horizons are uncertain. A single well dated Oxfordian Dicoelites, abundant Kimmeridgian–Tithonian Belemnopsis, and scattered Tithonian Hibolithes also have been described from the Sula Islands (Challinor & Skwarko, 1982).
Belemnites are abundant in Misool. The ?Bathonian–Callovian contains rich assemblages of Belemnopsis, Dicoelites, and Conodicoelites, and the Oxfordian contains Belemnopsis, diverse Hibolites, occasional Dicoelites, and a new belemnopseid genus. The Oxfordian–Kimmeridgian contains abundant Belemnopsis; the Tithonian, Belemnopsis and rare Hibolites; and the Berriasian–?Hauterivian, diverse Hibolites, Duvalia, and a Belemnopsis (Challinor, 1989a).

Correlation between Misool and Sula, based on an evolving lineage of Belemnopsis (Challinor, 1989b), is excellent in the Kimmeridgian–Tithonian. Belemnites from the Neocomian (if any) of the Sula Islands are unknown. Dating of the earliest belemnite-bearing beds in Misool (?Bathonian–early Callovian) is imprecise, and correlation between Misool and Sula in the ?Bathonian–Oxfordian is poor (Challinor, 1989a).

Scattered occurrences of Belemnopsis, Hibolites, and Conodicoelites are known from the Enarotali, Waghete, and Timika 1:250 000 Sheet areas of mainland Irian Jaya (Fig. 5). Little detailed work has been done on this material, all of which was collected from Kembelangan Group rocks (Pigram & Panggabean, 1983), some as float samples. The Conodicoelites are apparently Middle Jurassic, and the Belemnopsis and Hibolites Early Cretaceous. Meagre dinoflagellate suites associated with fragmentary juvenile Belemnopsis and Hibolites from the Kipare River (Waghete Sheet area) and Sintabu River (Timika Sheet area) indicate a late Valanginian–early Hauterivian age (Phoberocysta burgeri zone; R.W. Helby, personal communication, 1989).

The only material from mainland Irian Jaya studied in detail was described by Challinor (1989a). Abundant Belemnopsis and Hibolites were collected from localities on the Ainim and Mios Rivers in the central Birds Head (Fig. 5). These specimens occur in the Jass Formation, which unconformably overlies the non-marine Triassic–Jurassic Tipuma Formation (Pigram & Sukanta, 1989). The collections are particularly interesting because they contain a single specimen of Chalalabelus, the only representative of the Suborder Diplobelina known from the southwest Pacific region. Although Duvalia is missing from the mainland Irian Jaya collections, the same species of Belemnopsis and Hibolites are associated with Duvalia in Misool.

The assemblage was originally thought to be approximately Hauterivian in age (Challinor, 1989a). This notion was based on the presence of associated ammonites, on the stratigraphic
Belemnites are abundant elsewhere in Indonesia. Most of the taxa discussed above have been recorded from Timor, Roti, Jamdena, and elsewhere, although ages and stratigraphic relationships are uncertain (Challinor, 1989c). A comprehensive zonal scheme for the Bajocian to Hauterivian of Indonesia has been proposed (Challinor, 1991). Aulacocerids are also known from Indonesia, mostly from Timor, where the genera Calliconites (Xiphoteuthididae), Prographularia, Aulacoceras, and Buelowiteuthis (Aulacoceratidae) are said to be common in the Late Triassic (von Bulow, 1915; Doyle, 1990).

**Australia** (Fig. 6; Fig. 7, col. 3)

The Australian belemnite record is poor, and stratigraphic distributions are uncertain, particularly in the Jurassic. Cretaceous occurrences are more numerous and widespread, but restricted to the Aptian–Cenomanian.

Jurassic marine strata are limited (Brown & others, 1968). Middle Jurassic Belemnopsis (or possibly Dicoelites; see Challinor, 1989c) are known from the Perth Basin near Geraldton (Whitehouse, 1924), and Late Jurassic Belemnopsis (probably identical with those of eastern Indonesia; Challinor, 1989c) occur in the Canning Basin near Broome (Teichert, 1940; Brunschwicker, 1960).

Neocomian Hibolites have been reported from the Canning Basin (Brunschwicker, 1960), but Stevens (1965) suggested an early Tithonian age for this occurrence. Hibolites of Neocomian age occur in the Styx Basin near Rockhampton in eastern Queensland (Whitehouse, 1946). The Dimitobelidae (Peratobelus and Dimitobelus) occur widely in the Aptian–Albian of the Great Artesian Basin of Queensland and northern South Australia (Stevens, 1965; Doyle, 1988), and include Dimitobelus in the Carnarvon Basin of Western Australia during the Cenomanian (Doyle, 1988). Parahibolites has been recorded from eastern and northern Australia in the Aptian, and Neohibolites in the Albian (Doyle, 1988).

**Summary of distributions** (Fig. 7)

Xiphoteuthids and Belemnitina are known only from New Zealand and New Caledonia in the Early and Middle Jurassic; Chalalabelus and Duvalia only from Indonesia in the Early Cretaceous; and the Dimitobelidae only from New Zealand,

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Australia, and Papua New Guinea in the middle and Late Cretaceous. The distribution of Belemnopsis in New Zealand and Indonesia is partly disjunct, that of Hibolithes and Conodicoelites largely so, and that of Dicoelites completely so. On this evidence, the belemnite assemblages of New Zealand and Indonesia—Papua New Guinea in the Pliensbachian to Hauterivian (as far as comparison is possible in the absence of Berriasian to mid-Aptian strata from New Zealand) represent parts of different biogeographic units. In the middle to Late Cretaceous, New Zealand, Australia, and Papua New Guinea appear to have been parts of a common province, within which Parahibolithes (?and Neohibolithes) were restricted to Australia and Papua New Guinea. Generic distributions in New Caledonia suggest links with both New Zealand and Indonesia in the Middle Jurassic.

The presence of the same species in Papua New Guinea, the Sula Islands, Misool, and mainland Irian Jaya indicates that the belemnites of those regions are part of a common assemblage. Different generic distributions in the Indonesia—Papua New Guinea region when compared with New Zealand are readily apparent.

**Morphologic comparisons (Figs. 8–10)**

Although Stevens (1965, 1973) recognised New Zealand and Indonesia as parts of an extensive Indo-Pacific province extending from northern India eastwards around Gondwana to South America, the stratigraphic distributions discussed above suggest that the belemnites of that region evince marked provincialism. Intragenic morphologic comparisons at species and group levels (Figs. 8–10) support this hypothesis.

Middle Jurassic Belemnopsis (B. mackayi Stevens and B. deborahae Challinor) from New Zealand are compressed in cross-section, and have short poorly defined dorsal alveolar grooves and long comparatively narrow deep ventral grooves (Stevens, 1965; Challinor, 1977). The only adequately described Middle Jurassic Belemnopsis from Indonesia are both depressed in cross-section: B. persulcata Stolley is small, with a long shallow narrow ventral groove and a short poorly defined dorsal groove; and B. wanneri Stolley is of medium size with a moderately developed ventral groove which terminates some distance from the apex (Challinor, 1989c). In New Caledonia, Middle Jurassic Belemnopsis (B. compressa Avias, and B. cf. compressa; Challinor & Grant—Mackie, 1989), like those of New Zealand, are laterally compressed with deep moderately narrow ventral grooves and short poorly defined dorsal grooves. Thus, New Zealand and New Caledonian forms are similar, whereas those of Indonesia are quite different.

Late Jurassic Belemnopsis from Indonesia are all members of the moluccana lineage (Challinor, 1989b), an evolving lineage which arose in the early Oxfordian and continued to the end of the Tithonian. This lineage contains five recognised species, together with numerous transitional forms. Failure to recognise the lineage as such, and to see relationships within it (although Stolley, 1929, and Stevens, 1965, both suspected them), led to extensive synonymy.

Late Jurassic Belemnopsis from New Zealand appear in the Oxfordian. Of the six or seven known species, the youngest persists into the late Tithonian. Most are isolated species with little evidence of close relationship to preceding or later taxa (Challinor, 1979a, b, 1980). Stevens (1965) suggested that two of the youngest (B. aucklandica trechmanni Stevens and B. a. aucklandica (Hochstetter), ?early–late Tithonian) are subspecies; however, there is little evidence to support this suggestion, apart from broad parallels in morphology and the fact that the two forms appear in stratigraphic succession (in which they are separated by a zone of several Hibolithes; Stevens, 1965).

Stevens (1965) recognised New Zealand Belemnopsis in Indonesia and conversely. Challinor & Skwarko (1982) suggested that Late Jurassic members of the genus from New Zealand and Indonesia were subspecifically related. These postulated relationships and occurrences are now known to be due to misidentifications resulting from close homeomorphy, and not to genetic interchange between the regions (Challinor, 1989c). One of the more obvious differences is the generally much larger size of Indonesian Belemnopsis (Table 1) and their relatively more prominent ventral grooves (compare Stevens, 1965, plates 2–13, with Challinor, 1989b, plates 1–6). Some juvenile New Zealand Tithonian Belemnopsis are particularly weakly grooved (Challinor, 1970), which has led to their misidentification as Hibolithes. Significant differences in morphology between New Zealand and Indonesian Belemnopsis from the Jurassic are clearly present.

![Table 1. Diametral statistics of Late Jurassic Belemnopsis](image)
The only known Middle Jurassic Hibolites from New Zealand — *H. calinensis* (Hector) — is approximately equidimensional in cross-section, and has a moderately long ventral groove (Stevens, 1965); its morphology could indicate that it is ancestral to Tithonian forms of the genus from New Zealand. The only reasonably well dated Indonesian forms of Hibolites of partly Middle Jurassic age (*Hibolites cf. ingrati Stolley, and *H. demeaneus* Challinor; late Callovian–Oxfordian; Challinor, 1989c) are robust, depressed, and somewhat *Belemnopsis*-like, with relatively strong grooves which terminate near midguard.

Nearly all Late Jurassic Hibolites from New Zealand are mid-late Tithonian in age. They are compressed in cross-section (including *H. marwickii marwickii* and *H. m. mangaroaensis* Stevens, which were originally described as circular in cross-section), and nearly all have long ventral grooves extending well into the apical region. They form a homogeneous group, and at least some are subspecifically related (Stevens, 1965; Challinor, 1975b). The only exceptions are rare, undescribed short-grooved forms from the Kimberdinadian and Tithonian (Challinor, new data).

Thirteen Late Jurassic or Early Cretaceous Hibolites have been described from eastern Indonesia. Four were recognised by Stolley (1929, 1955), and they, together with nine new species (some informally named), were described or redescribed by Challinor (1989c). They form a heterogeneous group with little evidence of close relationship and few similarities to New Zealand species. Six are depressed in cross-section, and seven compressed or circular. In seven, the ventral groove is short or non-existent; in four it terminates near midguard, but its posterior extension is weak. One taxon, *H. longiscissus* Stolley, is compressed in cross-section, and its ventral groove is highly variable, extending well into the apical region in some specimens, but confined to the alveolar region in others. In addition, the taxon exhibits a number of unusual and poorly understood features (Challinor, 1989c).

The only Indonesian Hibolites similar to New Zealand members of the genus is *H. sp. C*, a genuinely long-grooved compressed form. However, it is a rare taxon known only from a single specimen in the earliest Tithonian (Challinor, 1989c), a time when the genus is rare in Indonesia.

Three of the Indonesian species with significant stratigraphic ranges (*H. gamtaensis* Challinor, abundant; *H. miosensis* Challinor, common; *H. australis* Challinor, comparatively rare) are depressed in cross-section and have short ventral grooves. Two (*H. boloides* Stolley, common; *H. quadratus* Challinor, rare) are compressed, with apparently non-existent or short grooves. The aberrant *H. longiscissus* is moderately common, and present throughout the Oxfordian. The remaining seven species are either rare (four represented by single specimens) or are restricted to narrow stratigraphic horizons (Challinor, 1989c).

Thus there are few similarities between individual Indonesian and New Zealand Hibolites (except for *H. sp. C*). The heterogeneous Indonesian assemblages, numerically and stratigraphically dominated by compressed short-grooved forms, differ markedly from the morphologically constant (mainly Tithonian) New Zealand group. A number of (mainly Middle Jurassic) Indonesian Hibolites described by Boehm (1907), Kruizinga (1920), and Stolley (1929) are not included in this analysis because they are poorly known both taxonomically and stratigraphically, but illustrations by the cited authors suggest that they conform to the general picture described.

According to Stevens (1965) there are marked similarities between *Hibolites arkelii* Stevens (mid-Tithonian) from New Zealand and *H. longiscissus* (Oxfordian) from Indonesia, but the resemblances are superficial (Challinor, 1989c), and the two differ widely in age. He also suggested that *Belemnites lagoicus* Boehm (=*Hibolites lagoicus*) of Indonesia was similar to *H. m. marwicki* of New Zealand. Boehm’s specimens of *lagoicus* are lost (Stevens, 1965), and no specimens resembling *lagoicus* were recognised in the new collections studied by Challinor & Skwarko (1982) and Challinor (1989c). Stevens (1965) also saw resemblances between *H. m. marwicki* (and *H. m. mangaroaensis*) and the Indonesian *H. compressus* Stolley, but again the specimens of Stolley are apparently lost (Stevens, 1965) and few similarities were found in new material studied by Challinor (1989c); furthermore, *H. compressus* is restricted to a short interval of earliest Tithonian age in Misool, whereas *H. marwicki* s.l. is mid-late Tithonian on the ages adopted here.

Hibolites in the Late Jurassic of New Caledonia is known only from latex casts of the anterior regions of a single, short-grooved specimen. Middle Jurassic forms are better, but not well known. Latex casts and calcareous specimens indicate that the guards are laterally compressed and the ventral groove is short, though not as short as that of some Indonesian species. They resemble but are not identical with the New Zealand *H. calinensis* (Challinor & Grant–Mackie, 1989).

Conodicoelites Stevens contains taxa of diverse morphology (Jeleztsky, 1983), and may be polyphyletic. Two different groups occur in Indonesia. One is characterised by large robust to slender, fully conical elongate guards with long narrow dorsal and ventral grooves; this group includes *C. keewensis* (Boehm), and *C. kalepuensis* Challinor, which is also known from Papua New Guinea. The second group contains one known species, *C. abadi* Challinor, which has a robust conical to slightly hasteate guard with a deep wide ventral groove extending almost to the apex and a shorter shallower dorsal groove terminating close behind the protoconch.

New Zealand forms of *Conodicoelites* are of a moderate size, are subconical to slightly hasteate, and have moderately long shallow narrow ventral grooves and shorter dorsal grooves. Stevens (1965) recognised three species from the same stratigraphic interval, but all could well be included in a single variable species in which much of the variation is attributed to ontogenetic causes. Morphology suggests that New Zealand and Indonesian *Conodicoelites* are not closely related, and the same could be said for the two Indonesian groups. *Conodicoelites* has not been recorded from New Caledonia or Australia.

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**Figure 8. Selected representatives of the genus Belemnopsis, scale x 1; all ventral views.**

A. *B. deborahae* Challinor, C1240, Bajocian, New Zealand. B. *B. persulcata* Stolley, IMC560, Callovian, Misool. C. *B. wanneri* Stolley, IMC576, Callovian, Misool. D. *B. molluscana* (Boehm), IMC269, ‘Kimmeridgian, Sula Islands. E. *B. galoi* (Boehm), IMC521, Tithonian, Misool. F. *B. trolleyi* Stevens, IMC531, Tithonian, Misool. G. and F are large specimens, but specimens of greater diameter are known. G. *B. a. aucklandica* Stolley, IMC531, Tithonian, Misool; J. *B. jonkeri* Stevens, IMC227, Tithonian, Misool; E and F are large specimens, but specimens of greater diameter are known. J. *H. longiscissus* Stevens, IMC576, Tithonian, Misool. K. *B. compressus* Challinor, C1438, Oxfordian, New Zealand. L. *B. a. aucklandica* C1258e, Kimmeridgian, New Zealand. J. *B. aucklandica* trechmantti Stevens, C1430, Tithonian, New Zealand. K. *B. a. trechmantti* (Stevens, C1435, Tithonian, New Zealand. L. *B. a. aucklandica* (Boehm), C1439, Tithonian, New Zealand; a normal-size adult. M. *B. a. aucklandica* (Boehm), C1432, Tithonian, New Zealand; a very weakly grooved juvenile. N. *B. a. aucklandica* (Boehm), C1431, Tithonian, New Zealand; largest specimen known to the writer, about 25¢9, larger in diameter than most adults. Similarities between *B. a. trechmantti* (1, K), *B. a. aucklandica* (L–N), and *B. galoi* (E) have led to the incorrect identification of New Zealand Belemnopsis in Indonesia. Specimens with catalogue numbers prefixed by IMC are held in the Macropaleontology Laboratory, Geological Research and Development Centre, Bandung, Indonesia; those prefixed by CPC are at EMR, Canberra, Australia; and those prefixed by C are at the University of Auckland, New Zealand.
A large number of Dicoelites have been described from Indonesia (Boehm, 1906, 1912; Kruizinga, 1920; Stolley, 1929, 1935; Stevens, 1964a), but many are synonymous (Challinor, 1989c). Valid species appear to be D. sp. A, B, and C (Challinor & Skwarko, 1982), D. sp. D (Challinor, 1989c), D. rotundus Stolley, D. mihanus Boehm, D. longirostris (Stolley), and possibly D. cf. mihanus Kruizinga. Only D. rotundus is abundant, though no complete guards of it are known (Challinor, 1989c).

D. sp. A, B, and C are large elongate hastate forms, and D. B is particularly massive. D. rotundus is a smaller cylindrical or slightly hastate taxon with a long ventral groove and a highly variable (usually very weak) dorsal groove. D. longirostris is a slender very elongate depressed form with a long ventral and short dorsal groove. D. sp. D and D. longirostris were found at about the same stratigraphic interval in Misool, but the two seem to be distinct; both are known from single specimens.

Dicoelites is represented in New Zealand by a single species, D. kowhaiensis Challinor. The adult guard, known from one incomplete specimen, is laterally compressed and apparently extremely elongate (Challinor, 1980). Immature specimens are abundant, slender, very elongate, and compressed in cross-section, and the ventral groove of both adult and immature guards, although long, terminates some distance from the apex. The dorsal groove is not preserved in any specimen.

Two species of Dicoelites are known from New Caledonia (Challinor & Grant-Mackie, 1989). D. aviasi is slender, extremely elongate, and compressed in cross-section, and has a very prominent long ventral and short poorly developed dorsal groove; it resembles D. kowhaiensis of New Zealand. The second species is a more generalised form known only from latex casts, and is broadly similar to D. rotundus of Indonesia.

Thus most Indonesian Dicoelites are quite different from the single New Zealand species, but one poorly known New Caledonian form may be similar to the commonest Indonesian taxon. There are general similarities in the morphologically unusual D. kowhaiensis, D. aviasi, and D. longirostris, but the compressed New Zealand and New Caledonian taxa are distinct from the depressed Indonesian species. Dicoelites has not been recorded from Australia, although some of the Middle Jurassic Belemnopsis specimens described from Western Australia (Whitehouse, 1924) could well belong in Dicoelites (Challinor, 1989c).

On both stratigraphic and morphologic evidence, the belemnites of New Zealand and Indonesia appear to be parts of different biogeographic provinces. Middle Jurassic taxa from New Caledonia include species that are morphologically similar to New Zealand forms, and others that are similar to those of Indonesia, but their restricted stratigraphic distribution prevents a comprehensible comparison.

The ‘uhligi complex’

Although Stevens (1965) considered that there were strong parallels in the belemnite successions of New Zealand and Indonesia, and recognised species common to both regions, the evidence outlined above suggests that no such relationships exist. Restriction of some genera to one region, non-contemporaneous generic ranges, or the presence of distinctive species or lineages when genera were present at the same time, all suggest that these two regions are parts of different biogeographic provinces.

A major obstacle to the elucidation of southern hemisphere belemnite relationships has been the group of Belemnopsis known as the ‘uhligi complex’. First proposed by Stevens (1965), the ‘uhligi complex’ was considered by him to be a widely distributed group of closely related robust Belemnopsis of Late Jurassic age. It was named from Belemnopsis uhligi Stevens, identified from northern India and Indonesia, and the group was suggested as a valuable indicator of Kimmeridgian–Tithonian age. It is in fact no more than a loose association of broadly similar Belemnopsis, of which some are related closely and others only distantly so, and its so-called members and related forms range in age from Middle Jurassic to Early Cretaceous.

Uncritical acceptance of the ‘uhligi complex’ by later researchers (Challinor, 1970; Challinor & Skwarko, 1982; Jeleczky, 1983; Mutterlose, 1986) has resulted in the proposal of what are considered here to be incorrect phylogenetic relationships and migration patterns for southern hemisphere Belemnopsis.

Stevens (1965) recognised the ‘uhligi complex’ or related species in New Zealand, New Caledonia, Australia, Papua New Guinea, Indonesia, northern India, Arabia, Tanzania, and possibly South America. He compared 12 species (Stevens, 1965, table 20) which he regarded as either members of the complex or closely related, and he recognised nine of them from Indonesia. Of these nine species, B. moluccana (Boehm) and B. indica Kruizinga are synonymous; B. taliabutica (Boehm) and B. sularum (Boehm) are transitional members of the moluccana lineage provisionally synonymised with B. galoi (Boehm); the Indonesian specimens of B. uhligi are B. galoi, and B. affinis (Boehm) is closely related to both B. moluccana and B. galoi (Challinor, 1989b). All are members of the moluccana lineage, and Stevens’s ‘uhligi complex’ is thus partly equivalent. Of the remaining Indonesian species in Stevens’s table 20, B. suavis Stolley is synonymous with B. jonkeri Stolley (Early Cretaceous; Challinor, 1989c), and postdates the moluccana lineage, to which it is not closely related (Challinor, 1989b). The status of B. rumpfii (Kruizinga) is not clear, but it differs in morphology (see Kruizinga, 1920) from members of the moluccana lineage, and is probably geologically much older.

Three New Zealand taxa — B. aucklandica aucklandica, B. a. trechmanni, and B. keari Stevens — were considered by Stevens (1965, table 20) to be either members of the complex or closely related to it. The subspecies of B. aucklandica are close homeomorphs of B. galoi, but are not closely related to Indonesian Belemnopsis, and are not present in Indonesia (Challinor, 1989c). B. keari predates the subspecies of B. aucklandica in New Zealand, is not closely related to them, was a composite taxon (Challinor, 1979a) ranging in age from Oxfordian to Kimmeridgian or possibly Tithonian (vide Helby & others, 1988), and is not present in Indonesia. The Indonesian specimens figured as B. keari or compared with it (Stevens, 1965) and identified as B. indica by Stolley (1929), are geologically older members of B. moluccana (Challinor, 1989b).

The New Caledonian specimens identified by Stevens (1965) as B. aucklandica are of Middle Jurassic age and are now included in B. compressa Avias (Challinor & Grant–Mackie, 1989c).
1989), and the Australian and Papua New Guinean members of the ‘uhligli complex’ are part of the moluccana lineage (Challinor, 1989c). The status of the Indian members is not completely clear, but they are probably closely related to the Indonesian members. The holotype of B. uhligli was found in the Spiti Shales of northern India, and the Indonesian specimen figured by Stevens (1963c) is a B. galloi from the Lelinta Shale of Misool (Challinor, 1989c). The status of the Arabian, Tanzanian, and South American taxa is not known, but the South American forms are probably related to New Zealand Belemnopsis (Stevens, 1965).

In the sense of Stevens the ‘uhligli complex’ has little validity, but, as a geographically restricted group present in eastern Indonesia during the Oxfrodian–Tithonian and possibly India and Tibet in the Late Jurassic, it may have value.

The belemnites of Gondwana

If the belemnites of Indonesia and New Zealand represent different faunal provinces during much of the Jurassic and Early Cretaceous, questions concerning their origins and affinities arise. To investigate possible relationships, southwest Pacific assemblages are compared here with those of the Gondwanan continents. It has not been practical to examine collections from outside the immediate study area, so comparisons are based on published papers. There are a number of difficulties, some of which result from different methods and philosophies (Challinor & Skwarko, 1982), and the interpretations advanced here may differ from those of the original authors. Age control is not as accurate as that for the assemblages discussed above.

Africa

Stevens (1965) summarised earlier reports. He noted Belemnopsis gerardi (Oppel) and B. orientalis (Waagen) from east Africa, and the probable presence of the ‘uhligli complex’ in Arabia and South Africa. Earlier (1963b), he had reported the widespread occurrence of B. tangansensis (Futterer) in east Africa and southern Arabia. B. tangansensis and B. gerardi are subcylindrical taxa with elongate tapering apices, and are similar in some respects to B. wanneri of Indonesia. According to Stevens, the east African species range in age from late Oxfordian to Kimmeridgian, and are therefore younger than B. wanneri. Duvalia, Rhopaloteuthis, and Hibolites were present in Somalia and Tanzania in the Tithonian; Hibolites in the Early Cretaceous of Somalia; and Hibolites and Duvalia in the Early Cretaceous of Tanzania (Stevens, 1965).

Doyle (1987a) described Neohibolites and Peratobelus, and Jeletzky (1981) referred to Chalalabelus (Diplobelina), all from the late Aptonian of southern Mozambique. Duvalia and Hibolites (or Neohibolites) are probably present in southern Mozambique (Doyle, 1987a).

Although precise details are lacking, the belemnite succession of east Africa appears to be similar to that of Indonesia — that is, the Late Jurassic is dominated by Belemnopsis, and Hibolites and Duvalia are more prominent from near the Jurassic–Cretaceous boundary. Taxa not known from Indonesia (Neohibolites, Peratobelus) appear in the Early–mid-Cretaceous.

Malagasy Republic (Madagascar)

Diverse belemnite assemblages recorded from the Malagasy Republic by Besairie (1930, 1936), Besairie & Collignon (1956), and Nicolai (1950–1) contain a number of species originally thought to be similar to or identical with Indonesian forms, but more recent work throws doubt on most of these identifications.

Combeamorel (1988) recorded Belemnopsis (11 spp.), Hibolites (11 spp.), Duvalia (13 spp.), and less abundant Dicoelites (2 spp.), Rhopaloteuthis, and Neohibolites. The stratigraphic distributions of Belemnopsis, Hibolites, and Dicoelites (Combeamorel, 1988) strongly resemble those of Indonesia (Challinor, 1991). Duvalia has a much greater range in Malagasy (Bathonian to Turonian), but it undergoes a major diversification in the Early Cretaceous similar to that in Indonesia.

Neither of the Dicoelites species is identical with Indonesian forms of this genus, but D. challinori Combeamorel (Oxfordian, Malagasy) resembles D. sp. C (Oxfordian, Sula Islands). Combeamorel identified Belemnopsis alicafra in Malagasy, ranging from mid-Oxfordian to late Kimmeridgian; in view of its restricted range in Indonesia (late Kimmeridgian only), and its status as a briefly appearing and impersistent member of the moluccana lineage (Challinor, 1989b), its extended range in Malagasy is unexpected, and the identification should be regarded as provisional. No forms of Hibolites in the two regions are identical, but H. savornini Nicolai (early Kimmeridgian to late Tithonian) of Malagasy is similar to H. gamtaensis Challinor (Neocomian) of Indonesia. Earlier (1963b), Combeamorel identified Belemnopsis alicafra in Malagasy, resembling D. sp. C (Oxfordian, Sula Islands). Combeamorel (Universite Claude–Bernard, personal communication, 1987) considers Duvalia ceramensis Challinor of Misool to be conspecific with D. sakalava Besairie of Malagasy, but reasons for doubting identity are discussed elsewhere (Challinor, 1989c).

Although generic distributions are similar, strong endemism is apparent in Kimmeridgian–Valanginian Belemnopsis of both regions. Conversely, broadly similar species of Duvalia and Hibolites are known from Misool and Malagasy.

India and Pakistan

Jurassic and Cretaceous belemnites are known from northwest India and Pakistan, and Cretaceous forms from southern India and Sri Lanka.

According to Stevens (1965) the belemnite succession of Kutch includes the Callovian Belemnopsis callovensis (Oppel), B. gregoriais and B. gerardi (Oxfordian–Kimmeridgian), Hibolites (Kimmeridgian), and B. uhligli (mid-Tithonian). The Spill succession commences with Conodicoelites (‘Callovian: Stevens, 1964a) followed by Belemnopsis alicafra and B. alicafra (Late Jurassic, Uhlig, 1910; probably Kimmeridgian and Tithonian sensu Harland & others, 1982).

Belemnopsis, Hibolites, Conodicoelites, and Dicoelites are recorded from other regions of northwest India and Pakistan (Himalayas, Kalahag, Salt Range). The Belemnopsis species

range from Callovian to Kimmeridgian, the Conodicoelites are Callovian, and the Hibolithes forms are Late Jurassic, but the age of the Dicoelites is uncertain (Stevens, 1965).

Uhlig's (1910) Belemnopsis from the Spiti region includes both immature and adult specimens, and Stevens (1965c) included some in his new species B. uhligi, but excluded others. Uhlig's collection appears to represent a single taxon or closely related group. The adult guards are large robust, strongly grooved forms similar to members of the moluccana lineage of Indonesia, and the immature guards are morphologically related to the adults in the same way as immature guards of the moluccana lineage are related to their adult forms (see Challinor, 1989b). Indonesian and Papua New Guinean specimens included by Stevens (1965c) in B. uhligi are all members of the moluccana lineage (Challinor, 1989b), and it is clear that the Spiti Belemnopsis specimens strongly resemble those of the Late Jurassic of Indonesia.

Cretaceous belemnites of northern India include Hibolithes, Duvalia, and ?Belemnopsis (Stevens, 1965). Doyle (1988) reviewed early records and noted the presence of Tetradelphus, Parahibolithes, Neohibolithes, ?Tetradelphus, and ?Parahibolithes in southern India and Sri Lanka in the Albian–Cenomanian. Thus the Jurassic–Early Cretaceous succession of India is similar to that of Indonesia, but genera not recorded from Indonesia appear in the mid- Cretaceous of India.

Tibet

The belemnites of southern Tibet and Indonesia have a number of similarities, and a broad correlation based on population structure can be recognised. Identical species are not known with certainty, but several are possible.

Yang & Wu (1964) recognised two assemblages: the older, of late Oxfordian–early Tithonian age, consists of several Belemnopsis; the younger, of Early Cretaceous age, is dominated by Hibolithes. Belemnopsis of the older assemblage comprises mostly large strongly grooved forms. Belemnopsis dingriensis (Yang & Wu, 1964) resembles B. galoi (Boehm); B. tallabutica of Yang & Wu (1964) is similar to B. moluccana (Boehm); and B. gerardi of Yang & Wu (1964) resembles B. stolleyi Stevens. The younger assemblage contains Hibolithes subfusciformis of Yang & Wu (1964) and Belemnopsis sinensis Yang & Wu, similar to H. gamaeans (Challinor and B. jonkeri Stolley of the Early Cretaceous of Indonesia and Papua New Guinea. Also included in the Cretaceous group is H. parahisatus Yang & Wu, a large laterally compressed moderately long-grooved species which in some respects resembles Late Jurassic and Early Cretaceous Hibolithes of New Zealand and west Antarctica.

Chen (1982) described numerous belemnites from the northern slopes of the Himalayas, and recognised three groups: (1) a Belemnopsis gerardi (Oppel) assemblage (early Oxfordian–mid-Kimmeridgian); (2) a Belemnopsis uhligi Stevens assemblage (mid-Kimmeridgian–mid-Tithonian); and (3) a Belemnopsis sinensis Yang & Wu/Hibolithes jiabulensis Yen assemblage (late Tithonian–Berriasian). The B. gerardi and B. uhligi assemblages consist mostly of Belemnopsis, whereas the B. sinensis/H. jiabulensis assemblage contains Belemnopsis and many Hibolithes. B. uhligi of Chen (1982) is similar to B. galoi (Boehm), and B. sinensis strongly resembles B. jonkeri Stolley.

Thus the changes in population structure are similar to those of Indonesia. The B. gerardi and B. uhligi assemblages appear to correspond to the moluccana lineage of the Indonesian Late Jurassic, and — even though Duvalia is missing from it — the B. sinensis/H. jiabulensis assemblage appears to correspond to the Indonesian post-Tithonian assemblage.

Chen (1982) also recorded several species of Atractites. All appear to be based on phragmocones alone, and cannot be reliably separated from phragmocones of Aulacoceras, guards of which are also present in the collections.

South America

South American belemnites were discussed by Stevens (1965), who summarised the records of earlier writers and examined undescribed collections held at the University of La Plata, Buenos Aires. Most collections were from the Neuen, Santa Cruz, and Magallanes provinces of Chile and Argentina between about latitudes 38° and 58°S.

The Suborder Belemnitina is represented by Megateuthis, Passaloteuthis, Dactylioteuthis, Brachybelus, and Cylindroteuthis, suggesting that South America was a centre of development or dispersion for these ?Toarcian–Bajocian forms. The Belemnopsinae are represented by Belemnopsis in the ?Toarcian–Bajocian (closely resembling the New Zealand B. mackayi; Stevens, 1965) and Late Jurassic (resembling the New Zealand B. aucklandica trechmsani; Stevens, 1965). Dicoelites and Hibolithes occur in the Late Jurassic. The xiphoteuthid Atractites occurs widely in western South America in the Early Jurassic (Doyle, 1990).

Riccardi (1977) described Belemnopsis patagoniensis (Favre), cf. B. madagascariensis (Besairie), and Hibolithes aff. jaculum (Phillips) from the Berriasian of southern Patagonia. The two Belemnopsis (probably conspecific) are robust cylindro-conical forms with narrow ventral grooves. They strongly resemble B. madagascariensis (Besairie; see Comembolet, 1988), and consequently have affinities with the Belemnopsis forms of the Tithonian–Berriasian of the Malagasy Republic rather than those of New Zealand. Hibolithes aff. jaculum is a long-grooved form (Riccardi, 1977), and in this feature resembles New Zealand Hibolithes. However, the specimen seems likely to be a juvenile of the associated Belemnopsis, and its hastate form and weak ventral groove are consistent with that interpretation.

Urreta & Suarez (1985) identified B. madagascariensis from the Tithonian–Valanginian of Tierra del Fuego. All their specimens are fragments, and some have ventral groove characteristics consistent with the nominate taxon. The groove of others, however, is more like that of New Zealand forms of Belemnopsis.

Ferruglio (1936) identified Belemnopsis patagoniensis, Hibolithes aff. jaculum, and a new Hibolithes from the Early Cretaceous of the Argentine Cordillera. As far as can be determined from his illustrations, the Belemnopsis specimens resemble those of the Jurassic–Cretaceous in the Malagasy Republic, and at least some of the Hibolithes specimens are long-grooved forms similar to those of New Zealand. Ferruglio (1936) also recorded Neohibolithes and Parahibolithes (the latter as Belemnites sp. 1). His identification of Duvalia is incorrect, being based on indeterminate and badly deformed specimens (P. Howlett, Universitet Claude–Bernard, personal communication, 1990). Doyle (1988) discussed Parahibolithes and ?Dimitobelus from the Aiptian–Albian of southern South America.

Strong links between the Middle and Late Jurassic of southern South America and New Zealand are evident. The presence of Belemnitina, particularly Brachybelus and possibly
Cylindroteuthis in the Middle Jurassic and Dicoelites in the Late Jurassic of both regions, is significant. Middle and Late Jurassic Belemnopsis of both regions have strong affinities, and Atractites may indicate links in the Early Jurassic. These similarities are paralleled in other groups. Recent unpublished data indicate that a number of South American Jurassic bivalves and brachiopods (spiriferinids, inoceramids, Kalentera, and Campiochlamys) are closely related to or identical with New Zealand forms (J.A. Grant-Mackie, University of Auckland, personal communication, 1990).

Thus throughout the Middle Jurassic and much of the Late Jurassic, South American belemnites appear to be closely related to those of New Zealand. Genera typical of Tethys (east Africa and Malagasy Republic) appear in the Early Cretaceous, and by Aptian time an assemblage with mixed Tethyan and South Pacific affinities was present.

**Falkland Plateau**

Jeletzky (1983) described a collection of belemnites and other fossils from a borehole drilled at latitude 51°00.28′S, longitude 46°58.3′W on the eastern Falkland Plateau. He interpreted the stratigraphy as an ?Oxfordian–early Tithonian sequence and a Barremian–Albian sequence separated by a late Tithonian–Hauterivian unconformity, which has a similar position and extent to the Late Jurassic–Early Cretaceous unconformity in New Zealand.

The belemnites are difficult to interpret because most are immature and all are fragments. Six species of Belemnopsis occur in the ?Oxfordian–Tithonian interval, of which three (B. cf. keari, B. cf. stolleyi, and B. aff. spathi Stevens) bear relatively narrow grooves similar to those of New Zealand Belemnopsis. Two (B. cf. moluccana and B. sp. indet.) are strongly grooved, resembling Indonesian forms, although similar widely grooved species do occur rarely in New Zealand (Challinor, 1979a; Jeletzky, 1983). The remaining species (B. aff. orientalis) is broadly similar to the named Himalayan Tithonian–Hauterivian unconformity, which has a similar position and extent to the Late Jurassic–Early Cretaceous unconformity in New Zealand.

Jeletzky also recorded ?Hasrites sp. from the ?Oxfordian–Tithonian interval. This genus is not present in New Zealand as far as is known, but, according to Stevens (1965), occurs in the Late Jurassic of northern India. Although the evidence is hardly conclusive, it can be interpreted to suggest that some species in the Falkland Plateau ?Oxfordian–Tithonian assemblage have affinities with those of the South Pacific, whereas others resemble those of Tethys.

The only belemnites from the Early Cretaceous interval are two species of Hibolithes. While distinct from any known New Zealand Hibolithes, they are similarly long-grooved, and unlike most taxa from the Early Cretaceous of the Indonesian region.

**Antarctica**

Many belemnites have been described from the Jurassic–Cretaceous of west Antarctica. Stevens (1967) recorded Conodicoelites of ?early Kimmeridgian age from Lyon Nunataks in Ellsworth Land. They are morphologically quite distinct from the Himalayan and Indonesian members of the genus, and broadly similar to those of New Zealand, although not identical.

Willey (1972, 1973) discussed a large number of Jurassic and Cretaceous species from Alexander Island. Belemnopsis gladiatoris Willey and B. alexandri Willey (Berrisian) are probably conspecific, and strongly resemble B. jonkeri Stolley of the Indonesian Berrisian–Valanginian. They are similar to the latter in size, hastation, and ventral groove characteristics, but their guard cross-sections are less depressed than that of B. jonkeri. B. aff. uhligi (Tithonian–Berrisian) bears little resemblance to either the Himalayan B. uhligi Stevens or the morphologically similar B. galoii (Boehm) of Indonesia. It does, however, resemble a New Zealand specimen illustrated by Stevens (1965, plate 5, figs. 10–12) as B. alfuria, although the latter is probably a member of B. maccrawi (Challinor, 1979a).

**Hibolithes antarctica** Willey (Berrisian) is a moderately long-grooved strongly compressed form; H. belligerundi Willey is hastate, long-grooved, and depressed in cross-section; and H. sp. nov. appears compressed in cross-section, slightly hastate, and relatively short-grooved. All are similar in a number of features to H. arrelli arkelli (Stevens, 1965; Challinor, 1975a) or H. a. grantmackiei (Challinor, 1975b) of New Zealand, but are not identical. Hibolithes subfusciformis (Berrisian) is a depressed short-grooved species similar to H. galmassisi of Indonesia (Challinor, 1989c).

Most other species discussed by Willey are based on small numbers of usually fragmentary specimens, and little is known of their morphology and affinities. However, some of Willey’s species clearly have strong affinities with New Zealand taxa (H. antarctica, H. belligerundi, B. aff. uhligi), while others resemble those of the Malagasy Republic and Indonesia (B. gladiatoris, B. alexandri, H. subfusciformis).

Willey (1972) also recorded two species of Dimotobelus (similar to or identical with D. maggregori (Glaessner) known from New Zealand and Papua New Guinea), and three probably distinct species of Peratobelus, all with affinities to species described from Australia. All are of Abbian age.

Mutterlose (1986) described a number of belemnites from the Orville coast (southwestern Antarctic Peninsula) which closely resemble New Zealand species. Hibolithes aff. marwicki Stevens is strikingly similar to the named taxon, although shorter-grooved than most (but within its range of variation). Hibolithes aff. arrelli and Hibolithes sp. strongly resemble H. arrelli arrelli (Stevens) in general morphology, and H. a. grantmackiei Challinor in groove characteristics. Belemnopsis aff. keari Stevens is very like the nominate taxon in cross-section and groove characteristics, but is less hastate than many of the latter. All four species are of Kimmeridgian–Tithonian age, as are their New Zealand counterparts. Although they appear to be closely related to the New Zealand species, they may not be identical. It is difficult to evaluate them fully because they are either represented by few specimens or by more numerous but poorly preserved fragments.

Two further species described by Mutterlose do not have affinities with New Zealand taxa. Hibolithes aff. verbeeki Kruizinga is a moderately long-grooved form with a hastate depressed guard, similar in some features to Early Cretaceous Hibolithes from Indonesia. Prodorevalia aff. neryvensis (Favre) from the ?late Tithonian has no counterpart in New Zealand, but its presumed descendant, Duvalia, is common in the Berrisian–Hauterivian of Misool, and in the Late Jurassic–Early Cretaceous of the Mediterranean and Malagasy–west Africa regions.
Mutterlose considered that Antarctic faunal links with New Zealand existed in the Kimmeridgian–early Tithonian, but were replaced in the late Tithonian by links with Tethys.

Doyle (1985) identified *Parahibolites* from James Ross Island. Doyle & Zinsmeister (1988) recognised a new subgenus of *Dimitobelus*, *D. (Dimitocamax)*, from the Campanian–Maastrichtian of Seymour Island, and suggested that species of *D. (Dimitobelus)* from James Ross Island (Campanian) and New Zealand (Maastrichtian) should be included therein. *D. (Dimitocamax)* is apparently restricted to Antarctica and New Zealand.

Thus throughout most of the Late Jurassic, *Belemnopsis*, *Hibolites*, and *Conodicoelites* of Antarctica have strong affinities with those of New Zealand. Some Berriasian forms of *Hibolites* resemble New Zealand members of the genus, and others have features which suggest relationships with those of Tethys. Berriasian forms of *Belemnopsis* from Antarctica resemble species known from northern India and Indonesia. In Aptian–Maastrichtian time, both South Pacific elements (*Dimitiobelidae*) and Tethyan elements (*Parahibolites*) were present in the Antarctic region.

**Palaeobiogeography**

The evidence outlined above indicates that two different belemnite assemblages occupied the coasts of Gondwana in the Middle–Late Jurassic and Early Cretaceous. One group extended along the Tethyan coast from east Africa eastwards to Papua New Guinea and possibly at times to New Caledonia. The other was present in the South Pacific region (southern South America, west Antarctica, New Zealand, Australia and New Caledonia) and may have extended at times to Papua New Guinea. In terms of their belemnites the two regions are here regarded as different biogeographic provinces.

Boundaries between the two provinces were best defined in the Middle and early Late Jurassic. By Tithonian time, some intermigration of faunas had taken place between the South America–west Antarctica and Malagasy–India regions; by the Albian, extensive intermigration between the two provinces had occurred, apparently along a developing seaway between the Antarctica–Australia and Africa–South America plates (Mutterlose, 1986; Doyle, 1987a; Combemorel, 1988; Doyle & Howlett, 1989). If Jeletzky’s (1983) collections from the Falkland Plateau are correctly interpreted here, this seaway may have been initiated as early as Oxfordian time. The eastern boundary between the provinces apparently fluctuated back and forth across the Papua New Guinea–New Caledonia region, resulting in the presence at different times of groups with affinities of both provinces.

Endemic belemnites on the margins of Gondwana have been used to recognise several provinces within the Tethyan Realm, but the various groups and associations are morphologically more distinct and diverse than previously known, and a more comprehensive provincial subdivision is now possible. Stevens (1965, 1967, 1973) recognised three belemnite provinces within the Late Jurassic of the Tethyan Realm. They are:

- the Indo-Pacific province (northern India and south Tibet, eastern Indonesia, Papua New Guinea, northern and western Australia, New Zealand, west Antarctica, and southern South America).

A different subdivision of the Tethyan Realm based on the faunas inhabiting the Pacific and Tethyan coasts of Gondwana from Middle Jurassic to Early Cretaceous is proposed here. This paper updates terminology used earlier (Challinor, 1991), and assigns subprovincial status to the Mediterranean, Ethiopian, and Indo-Tethyan Provinces of that publication.

**South Pacific Province (Middle and Late Jurassic, Early Cretaceous; Figs. 11–13)**

The South Pacific Province includes the margins of Gondwana that extended along the Pacific coast from southern South America through west Antarctica to New Zealand and probably New Caledonia. It is recognised in the Middle Jurassic by Belemninitina (*Dactyloteuthis, Megateuthis, Passaloteuthis* in South America; *Cylindroteuthis* or a similar genus, and *Brachybelus*, in South America and New Zealand; and a new genus of Belemnitiidae in New Caledonia) and endemic *Belemnopsis* (*Belemnopsis* in South America, New Zealand, and New Caledonia; and moderately long-grooved *Hibolites* in New Zealand and New Caledonia). In the Late Jurassic and Early Cretaceous it is characterised by endemic *Belemnopsis* and *Hibolites* in New Zealand and west Antarctica, and by *Belemnopsis* similar to those of New Zealand in South America. *Dicoelitidae* are known from South America, Antarctica, and New Zealand in the Late Jurassic.

[Figure 11. Distribution of belemnite genera and some higher taxa in the Middle Jurassic Tethyan and South Pacific Provinces. Alternative symbols for *Belemnopsis* and *Hibolites* represent endemic regional populations or groups. Compiled from sources mentioned in text. Base-map from Smith & others (1981, map 40, modified).]
Belemnite Successions and Faunal Provinces, SW Pacific

(Dicoelites in the mid-late Bathonian and rare short-grooved Hibolithes in the Late Jurassic; Challinor & Grant-Mackie, 1989). Thus the boundary between the South Pacific and Tethyan Provinces may have fluctuated back and forth across the New Caledonian region, perhaps migrating eastwards as far as Papua New Guinea in the Aptian. No Cretaceous belemnites are known from New Caledonia.

Although Stevens (1963a, 1965) claims that the Passaloteuthinae (= Belemnitidae d'Orbigny, Suborder Belemnitina) occurred in Indonesia during the Bajocian-Bathonian, they are not represented in the large and stratigraphically well controlled Indonesian and Papua New Guinean collections studied by Challinor & Skwarko (1982) and Challinor (1989c, 1990). No Belemnitina were recorded from Indonesia by Stolley (1929, 1935), Boehm (1907, 1912), and Kruizinga (1920) or from India or Tibet by Uhlig (1910) or Yang & Wu (1964). However, Chen (1982) recorded them from the Sinemurian of Tibet, which represents the most easterly known occurrence of the assemblages which characterised Europe and the Mediterranean regions during the Liassic.

Figure 12. Distribution of belemnite genera in the South Pacific Province and Mediterranean, Ethiopian, and Indo-Tethyan Subprovinces during the Late Jurassic (approximately). Symbols and sources as for Figure 11. Base-map from Smith & others (1981, map 36, modified).

Some slight evidence suggests that the South Pacific Province may have continued into the late Early Cretaceous. Large compressed, moderately long-grooved Hibolithes which resemble those of New Zealand and west Antarctica are known as far eastwards as Papua New Guinea in the Aptian-Albian (Challinor, 1990).

Tethyan Province (Bajocian-early Oxfordian; Fig. 11)
The Tethyan Province includes central and southern Europe, the Caucasus, the Malagasy Republic, east and north Africa, northern India, Indonesia, and Papua New Guinea, and possibly New Caledonia at times. It includes the Middle Jurassic part (approximately) of the Himalayan province of Challinor & others (in press), and is best differentiated from the South Pacific Province by its Belemnopsis, Hibolithes, and Dicoelitidae. It is populated exclusively by Belemnopseina, and its most distinctive elements are abundant dicoelitids.

New Caledonia is only provisionally included within the province because its belemnites are difficult to interpret. Permian-Jurassic biotas of New Zealand and New Caledonia are virtually identical at specific level, except for their Belemnitida, of which (i) no identical species occurs, (ii) generic distributions are partly disjunct (Challinor & Grant-Mackie, 1989), and (iii) species morphology suggests that taxa similar to those of both the Tethyan and South Pacific Provinces were present at different times. The Belemnita present a biogeographically confusing picture: at times they suggest affinities with New Zealand (Belemnitina in the early Bathonian, and South Pacific-type Belemnopsis in the Callovian and rare moderately long-grooved compressed Hibolithes in the Bathonian-Callovian); and at other times they have affinities with those of Indonesia

Mediterranean Subprovince (Oxfordian-Tithonian; Fig. 12)
The Mediterranean Subprovince is essentially the Mediterranean province of Stevens (1965, 1973) with relict Belemnopsis in the Kimmeridgian, abundant Oxfordian-Tithonian Hibolithes, and minor short-ranging belemnopseids in the Oxfordian-early Tithonian (Stevens, 1965; Mutterlose, 1988). The few Conodicoelites present are here considered to be no younger than early Oxfordian. The rarity of Belemnopsis differentiates the Mediterranean from the Ethiopian and Indonesian Subprovinces.

Figure 13. Distribution of belemnite genera in the Early Cretaceous Tethyan and South Pacific Provinces. Symbols and sources as for Figure 11. Base-map from Smith & others (1981, map 32, modified).
Ethiopian Subprovince (Oxfordian–Valanginian; Fig. 12)

Stevens (1965, 1973) recognised that distinctive Belemnopsis (B. orientalis, B. tanganyensis, B. madagascariensis) occupied east Africa, the Malagasy Republic, and adjacent areas from about Callovian to Valanginian time. He proposed an Ethiopian province (Malagasy Republic, east Africa, Ethiopia, Somalia, parts of northwest India, and Pakistan) populated by endemic Belemnopsis from approximately mid-Oxfordian to middle or late Kimmeridgian time.

The same region is recognised here as the Ethiopian Subprovince, based on a number of distinctive very narrow-grooved Belemnopsis described from the Malagasy Republic by Combeamorel (1988). These species (B. solofoi, B. besairei, B. casterasi, B. ankomokensis, B. madagascariensis, B. sikilyensis) form what is apparently a lineage ranging from mid-Oxfordian to late Valanginian time. The Ethiopian Subprovince is temporarily co-extensive with the lineage. Short-grooved Hibolithes are present from the Oxfordian onwards, and Duvalia appears in the Valanginian but both continue into post-Valanginian time.

Indo-Tethyan Subprovince (Oxfordian–Tithonian; Fig. 12)

The Indo-Tethyan Subprovince is defined here as extending eastwards along the Tethyan coast of Gondwana from northern India and Tibet through eastern Indonesia to Papua New Guinea and possibly New Caledonia. It is characterised by Belemnopsis moluccana and its descendants in Indonesia, northern Australia, and Papua New Guinea, and by apparently closely related forms in India and Tibet. Hibolithes and an endemic genus are important in the Oxfordian of Indonesia. Very large, strongly grooved Belemnopsis continue into the Early Cretaceous of Indonesia and Tibet, but sub-provincialism is lost in post-Tithonian times owing to an influx of Mediterranean-type taxa.

Rare short-grooved Hibolithes occur in New Caledonia (Challinor & Grant-Mackie, 1989).

The Indo-Tethyan Subprovince is equivalent to part of the Indo-Pacific province of Stevens (1965, 1973), and includes the Late Jurassic part (approximately) of the Himalayan province of Challinor & others (in press). Like the Ethiopian, the Indo-Tethyan Subprovince is characterised by a distinctive Belemnopsis lineage, and Belemnopsis seems particularly prone to the development of endemic regional populations. Without these groups the Ethiopian and Indo-Tethyan Subprovinces would form part of an extended Tethyan Province ranging from Bajocian to Early Cretaceous, corresponding to the Middle Jurassic–Early Cretaceous South Pacific Province on the opposite coast of Gondwana.

Tethyan Province (Early Cretaceous; Fig. 13)

The Tethyan Province is re-established in the Berriasian with essentially the same geographic range as in the Middle Jurassic. It is characterised in Berriasian–Barremian time by Hibolithes and its descendants (Curtohibolites, Vaunagites, and Mesohibolites), and by Mesohibolites, Neohibolites, and Parahibolites in the Barremian–Aptian. Hibolithes, Neohibolites, and Parahibolites are pan-Tethyan; the remainder appear to be confined to the circum-Mediterranean (Mutterlose, 1988). The Duvaliidae, first appearing in the Bathonian, became abundant in the Berriasian–Valanginian of Europe, Indonesia, and Malagasy. The rare diplobelinid genus Chalalabelus is known from southern Britain and Mozambique in the Barremian–Aptian (Jeletzky, 1981), and from Indonesia in the Berriasian (Challinor, 1989a). Belemnopsis continues into the Berriasian–Valanginian of the eastern Tethyan Province.

Aptian–Late Cretaceous

The rich assemblages which characterised the Tethyan Province through the Middle and Late Jurassic and much of the Early Cretaceous declined as the Late Cretaceous approached. By Albian time the only hibolitid’s remaining were Neohibolites, Parahibolites and the regionally restricted Mesohibolites (Mutterlose, 1988, Challinor, 1990). Neohibolites and Parahibolites persisted into the Cenomanian (Doyle, 1988), and relict duvaliids entered the Turonian (Combeamorel, 1988). The Dimitobelidae (not shown in Fig. 13) first appeared in the early Aptian of Gondwana. Its genera (Dimitobelus, Tetraphelbus, Pterahibolus) reached their maximum development and distribution (?South America, Antarctica, Australia, New Zealand, southern India, southern Africa) in the Aptian–Albian (Doyle & Zinsmeister, 1988). By mid-Albian–Cenomanian, only Dimitobelus remained (in Antarctica, New Zealand, Australia, and Papua New Guinea), and persisted into the Turonian–Campanian in New Zealand. The subgenus D. (Dimitocamax) probably appeared in the Campanian of Antarctica, and became extinct in the Maastrichtian of New Zealand (Doyle & Zinsmeister, 1988).

Origin and distribution of assemblages

Stevens (1963a, 1965, 1973) proposed a Tethyan origin for Middle–Late Jurassic southwest Pacific belemnite faunas, followed by circum-Gondwana migration eastwards to the New Zealand–west Antarctic–South American region. Most later researchers (Mutterlose, 1986; Combeamorel, 1988) have accepted this. Some of them have interpreted migratory routes along postulated circum-Gondwana currents (Doyle & Howlett, 1989; Doyle, in press), and some trans-Gondwana migration between the African–South American and Antarctic–Australian landmasses in the Late Jurassic. Circum-Gondwana currents are to some extent hypothetical, but, even if they existed in the Early–Middle Jurassic, there is little faunal evidence for belemnite migrations in an eastward direction, except within the Tethyan Province, where both eastward and westward migrations of different genera might have occurred.

The hypothesis advanced here differs in proposing almost no eastward circum-Gondwana migration (except perhaps for Dicoelites to New Caledonia in the Barboian). Middle Jurassic assemblages of New Zealand are thought to have migrated westwards from Europe (if they did originate in that region) across the embryonic Atlantic seaway, between the Americas, down the west coast of South America, and into the west Antarctic–New Zealand region. Late Jurassic assemblages are considered to be in-situ developments from Middle Jurassic ancestors, perhaps with some Tethyan influence via the trans-Gondwana seaway.

Stevens (1965) envisaged a bidirectional migration of Liassic–Bathonian assemblages from Europe. Using a pre-mobilist model of Mesozoic continental distributions, he proposed a westward trans-Atlantic migration of Belennitina and Belemnopsis to South America, and, in the Bajocian–Bathonian, an eastward migration across Tethys. The westward migration foreshadows the hypothesis advanced here, but trans-Tethyan migration to the New Zealand region is not accepted.
The earliest belemnites appear in the Hettangian of Europe (Riegraf, 1980), and Sinemurian forms occur in west Antarctica, Pakistan, Canada, and Tibet (Doyle, 1987b). By Toarcian time, some division into Boreal and proto-Tethyan assemblages is evident (Doyle, 1987b). The important Tethyan genera, *Hibolites* and *Belemnopsis*, appear in the Aalenian–Bajocian of Europe (Riegraf, 1980; Stoyanova– Vergilova, 1982), Pakistan (Fatmi, 1972), and New Zealand (Stevens, 1973). Belemnmitina are present at the same time in Europe, New Zealand, and South America. The presence of both the Belemnopsinae and Belemnmitina in northern and southern hemispheres at the same time raises questions concerning their place of origin.

Stevens (1965, 1973) maintained that South Pacific mid-Jurassic genera originated in Europe. They seem to have been most abundant there, but this may simply reflect more extensive and detailed research compared with that in the southern hemisphere. Furthermore, to establish firm relationships between Aalenian–Bathonian belemnite faunas of Europe and the southern hemisphere requires detailed taxonomic and chronostratigraphic comparisons; as far as I am aware, this has not been done. Presumed relationships are based on generalised morphologic resemblances and approximate time equivalence.

Although it is not possible to state a certain place of origin for early Middle Jurassic populations, broad similarities between northern and southern hemisphere assemblages suggest they were part of a single cosmopolitan group. Migration between north and south seems highly probable, though the direction is uncertain. In view of the absence of well documented occurrences of the Belemnitina east of Pakistan, a circum-Gondwana route through eastern Tethys and Indonesia is considered unlikely.

**Origin of South Pacific Province assemblages**

A westward migration path between the Americas has been proposed by Westermann (1981) to explain cosmopolitan Middle Jurassic ammonite assemblages of Europe and western North America. He hypothesised intermittent shallow-marine connections via the proto-North Atlantic in the late Bajocian and Bathonian, and continuous connection throughout the remainder of the Middle Jurassic. This route is a probable path for the migration of Middle Jurassic Belemnitina and Belemnopsinae between Europe and the South Pacific region, although at this time no Middle or Late Jurassic belemnites are known from northern South America (Challinor & others, in press) and the few Middle Jurassic forms known from west Antarctica remain unstudied. Data in Doyle (1990) can be interpreted to indicate a westward migration path for xiphoteuthids between Europe, the western Americas, and New Zealand (?and New Caledonia; Challinor & Grant–Mackie, 1989) in the Early Jurassic.

Late Jurassic *Hibolites* of the South Pacific Province are considered to be probable descendants of moderately long-grooved forms, such as the New Zealand *Hibolites catlinensis*. Late Jurassic evolution seems likely in view of the close relationships between Late Jurassic members of the genus in New Zealand (Stevens, 1965; Challinor, 1975b), their time constraints, and their morphological uniformity. Circum- or trans-Gondwana migration seems unlikely in view of the absence, in central and eastern Tethys, of morphologically suitable ancestors, particularly as Late Jurassic *Hibolites* are not abundant in eastern Indonesia (through which any eastward migration must have passed).

The origin of Late Jurassic *Belemnopsis* from the South Pacific Province is more debatable. Species of the genus appear in New Zealand in a succession of sudden appearances (Stevens, 1965; Challinor, 1974, 1979a, 1980), and show little evidence that they are closely linearly related, in contrast to those of the same age in central and eastern Tethys. They may have evolved *in situ* from compressed narrow-grooved forms such as *Belemnopsis deboraha* or *B. mackayi* (New Zealand) or *B. avias* (New Caledonia), but their morphology does not support this view. Their appearance in New Zealand in a series of waves suggests migration from outside the immediate area, perhaps from the South American–west Antarctic region. Late Kimmeridgian and Tithonian forms might have entered the South Pacific region via the trans-Gondwana seaway, which — according to Martin & Hartnady (1986) — was not open before mid-Kimmeridgian time. Even so, there are few possible ancestors of suitable morphology known from Tethys, unless they arose from moderately broadly grooved precursors, such as ancestors of the Tithonian–Valanginian *Belemnopsis collignoni* Comemborel of the Malagasy Republic.

South Pacific dicoelitids are difficult to interpret biogeographically. New Zealand *Dicoelites* are completely isolated in time, and have no known possible relatives closer than the ?Kimmeridgian South American members of the genus, which are neither firmly identified nor accurately dated (Stevens, 1965). New Zealand and west Antarctic *Conodicoelites* (†late Callovian–early Oxfordian and ?Kimmeridgian respectively) might be members of a regional population, and more detailed work, particularly in west Antarctic and South America could reveal other related forms. When the age and relative abundance of *Dicoelites* from New Caledonia are considered, it seems possible that they are related to those of eastern Tethys, perhaps representing the most easterly migration point of that fauna.

The Berriasian belemnities of southern South America and west Antarctica include the first taxa of the South Pacific Province which have undoubted similarities to those of Tethys (*Belemnopsis gladiatoris*, *B. alexandri* Willey, *B. patagoniensis* Favre, *B. cf. madagascariensis*, *Hibolites* 'subfuniformis'). They are interpreted as north-to-south migrants through the trans-Gondwana seaway, perhaps preceded in the Kimmeridgian–Tithonian by *Hibolites* aff. verbeeki and *Prodavulia* aff. neyriivenis (Mutterlose, 1986) of west Antarctica.

The Dimitobelidae appear in the post-Neocomian, possibly evolving from originally Tethyan *Hibolites* or *Neohibolites* (Doyle, 1988) somewhere in the trans-Gondwana seaway. They occur first in west Antarctica and Australia in the Aptian, are widely distributed along the Pacific coast of Gondwana from South America to Papua New Guinea in the Albiano-Albian–Cenomanian, and are restricted to New Zealand and west Antarctica by the Campanian. This last restriction is thought to have resulted from southward movement of the Antarctic–Australian landmass (Doyle, in press), and isolation in shelf habitats unconnected with those of the remainder of Gondwana.

The presence of the Tethyan genera — *Prodavulia* (Oxfordian–Tithonian), *Neohibolites*, and *Parahibolites* (Aptian–Cenomanian) — in west Antarctica and South America is considered to result from trans-Gondwana migration. It is significant that, except for a doubtful occurrence of *Neohibolites* in eastern Australia (Doyle, 1988), they occur in the South Pacific Province only at the southern end of the trans-continent seaway. *Hibolites* of New Zealand affinities (Challinor, 1990) and *Parahibolites* are also present in Papua New Guinea.
suggested that this region marked the eastern boundary between the Tethyan and South Pacific Provinces in the Aptian–Albian.

**Origin of Tethyan Province assemblages**

Stevens (1963a, 1965, 1973) hypothesised an eastward migration for most Tethyan genera from a centre of origin in western Tethys, and this was accepted by most later workers. However, Combemorel (1988) suggested that migration westwards from Malagasy–east Africa towards western Tethys better fits the evidence provided by Bathonian Belemnopsis and Duvalia from Malagasy (although he accepts a European origin and southeast migration of Hibolithes).

Major genera of the Tethyan Province, like those of the Tethyan Realm, developed distinct regional populations, of which few species are present in more than one region (Combemorel, 1988; Challinor, 1989c). It is instructive to examine each genus for regional characteristics, which may indicate possible migration patterns. This discussion briefly examines known time relations and morphological variation within Tethyan genera. It is not comprehensive, and is based only on papers readily available to me. Morphological discussion is restricted to those characters which can be readily evaluated from figured specimens.

According to Mutterlose (1988), Belemnopsis first appeared in the Toarcian of the Mediterranean region, and had extended throughout Tethys by the Bajocian. It disappeared from Europe and the Mediterranean by about the late Oxfordian, but continued in central and eastern Tethys until the Valanginian–Hauterivian (Combemorel, 1988; Challinor, 1989c). Belemnopsis latesulcatus (d'Orbigny) occurs in western Europe in the Callovian, but appears in the Malagasy Republic in the late Bathonian; Combemorel (1988) suggested an east-west migration direction for this taxon. B. latesulcatus is broadly similar to the late Bathonian Belemnopsis sp. A (Challinor & Skwarko, 1982) and the ?late Callovian Belemnopsis wanneri (Challinor, 1989c) of Indonesia; all of them may be representatives of a group which extended throughout Tethys in the pre-Oxfordian. The distinctive Late Jurassic Belemnopsis lineages of the Indo-Tethyan and Ethio-Malian Subprovinces do not appear to be closely related to their Callovian or early Oxfordian predecessors, or to each other.

Temporal relationships suggest that Tethyan Belemnopsis of the Middle Jurassic originated in the Mediterranean, but the origin of Late Jurassic to Early Cretaceous taxa is less clear.

Hibolithes is said to have evolved from Belemnopsis in the Bajocian of the Mediterranean (Mutterlose, 1988); and it is confined to the western part of the Tethyan Province through the Bajocian–Bathonian (Stevens, 1965, 1973). From Callovian to Hauterivian time, it is widely distributed throughout Tethys, and persists sporadically until the Aptian–Albian (Mutterlose, 1988; Challinor, 1990). Ventral groove characteristics (the only feature readily compared in what is a morphologically fairly uniform taxon) do not support any particular migration direction. Hibolithes of northwest Europe (Mutterlose, 1978; Combemorel & Mariotti, 1986; Gayte, 1984) are all short- or medium-grooved forms; those of the Malagasy Republic (except for the medium-grooved early Callovian Hibolithes sp. A; Combemorel, 1988) are all short- or very short-grooved species; and those of Indonesia and Papua New Guinea are mostly short-grooved with a few medium- and long-grooved taxa. These limited morphological data do not support any particular region of origin. It could perhaps be argued that the morphologically more constant Malagasy assemblage represents the original population, of which early members remain undiscov- ered, and that the more diverse eastern and western Tethyan populations are their descendants.

The Duvaliidae first appeared in the Early Jurassic of Europe (Combemorel, 1973). They were rare until the Tithonian, when they diversified and extended throughout the Tethyan Province, and became common in the Early Cretaceous. They were extinct by the end of the Hauterivian, except for an early Turoonian species of Duvalia. They appeared in the late Bathonian in Malagasy and at the Jurassic–Cretaceous boundary in Indonesia. Duvalia of northern Europe (Mutterlose, 1979), Italy (Combemorel & Mariotti, 1986), France (Combemorel, 1973), and the Balkans (Stoyanova–Vergilova, 1965; Nazarishvili, 1972) appear to be mostly long-grooved, the dorsal groove extending past the protoconch and often well towards the guard apex. Those of Malagasy (Combemorel, 1988) are of variable morphology, comprising short-, medium-, and long-grooved forms, while those of Indonesia are consistently short-grooved (Challinor, 1989c).

Thus the time relationships of Duvalia favour appearance in Europe and a west-to-east migration pattern, while morphological gradation in the feature examined favours west-to-east or the reverse direction. However, Combemorel (1988) suggested that, in the Bathonian at least, east-to-west migration can be argued on the basis of a greater diversity in the Malagasy assemblage, although the genus can hardly be described as diverse in either region at that time.

The origin of the Dicoelitidae remains enigmatic, but central and eastern Tethys was a major centre of development. Dicoelites ranges in age from mid-Bajocian to early Oxfordian in Indonesia; late Callovian–early Oxfordian in Malagasy (Combemorel, 1988); ?Hettangian and Bajocian in New Caledonia (Challinor & Grant–Mackie, 1989; though Hettangian is unlikely in view of its age elsewhere and a general absence of Belemnopesiana from the Early Jurassic); ?Bajocian in Western Australia (the Geraldton occurrence identified as Belemnopsis by Whitehouse, 1924); Kimeridgian in New Zealand (Challinor, 1980); and ?Kimeridgian in the Himalayas and South America (Stevens, 1964a, 1965). Conodicoelites ranges from Bajocian to early Oxfordian in Indonesia and Papua New Guinea; Bajocian–Oxfordian in Europe (Stevens, 1965); Callovian–?early Kimeridgian in India (Stevens, 1965; the Kimeridgian doubtful); Callovian–Oxfordian in New Zealand; and early Kimeridgian in west Antarctica (Stevens, 1967).

Jeleztky (1980) suggested that Dicoelites and Conodicoelites originated in the Toarcian–Bajocian of northwest British Columbia, migrated down the coast of North America, then continued both eastwards between the Americas into Tethys and southwards into the west Antarctic–New Zealand region. Thus the New Zealand, South American, and west Antarctic Dicoelitidae may be relicts populations surviving later than their Tethyan counterparts. Those of New Caledonia may have had a similar origin and record maximum dispersion of the southward migration, although the apparent absence of Dicoelites from the New Zealand Middle Jurassic argues against this hypothesis.

Jeleztky's (1980) hypothesis of a North American origin for Dicoelites and Conodicoelites in the Toarcian became difficult to accept (Doyle, 1987b) when Toarcian representatives were subsequently reported from Tibet by Wu (1982). Instead, the Dicoelitidae could have originated in central Tethys, and migrated westward between the Americas, then north to British Columbia and south into the South Pacific Province. The taxonomic uncertainties which exist between and within
**Dicœlietes and Conodicœlietes** (Challinor, 1991) make relationships uncertain, and the origin and dispersal of both genera remain obscure.

At first sight, the presence of Conodicœlietes in the South Pacific and Tethyan Provinces simultaneously near the Cаллавий–


Doroephyzian boundary and possibly in the Kimeridian (though the Tethyan occurrence of the genus of this age is doubtful) does not support the provincial separation advanced here. However, the genus evinces great morphological variation, and could be polyphyletic. New Zealand and west Antarctic Conodicœlietes are morphologically distinct from those of Indonesia, Papua New Guinea, and India, and are more like (but not identical with) those of Europe. If the taxon is polyphyletic, arguments based on its distribution may be invalid.

The rare genus Chalalabelus is known from the Berriasian of the Tethyan occurrence of the genus of this age is doubtful (Jeletzky, 1981). These occurrences could be interpreted to indicate an origin in eastern Tethys followed by westward migration, although evidence from a genus with such a poorly known distribution should not be overemphasised.

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**References**


Brunenschweiler, R.O., 1960 — Marine fossils from the Upper Jurassic and the Lower Cretaceous of Dampier Peninsula, Western Aus


Faimi, A.N., 1972 — Stratigraphy of the Jurassic and Lower Cretaceous rocks and Jurassic ammonites from northern areas of west Pakistan. Bulletin of the British Museum of Natural History (Geology), 20, 290–308.


Stevens, G.R., 1967 - Upper Jurassic fossils from Ellsworth Land, west Antarctica, and notes on Upper Jurassic biogeography of the southwest Pacific region. New Zealand Journal of Geology & Geophysics, 10(2), 345-393.


Teichert, C., 1940 - Marine Jurassic of East-Indian affinities at Broome, northwestern Australia. Journal of the Royal Society of Western Australia, 26, 103-119.


Urreta, B.A., & Suarez, D., 1985 - Belemnites de una sucencia turbiditica volcanoclastica de la formacion Yahgan — Titonio — Cretaceo inferior del extremo sur de Chile. IV Congreso Geologico Chileno, 1, 1-16.


Wu, S.B., 1982 - Characteristics of Early Jurassic–Early Cretaceous belemnoid assemblages from southern Xizang (Tibet). Contributions to the Geology of Quinghai-Xizang Plateau, 10, 113-121.

Structure and hydrocarbon potential of the Bremer Basin, southwest Australia

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The Bremer Basin underlies part of the upper continental slope of offshore southwest Australia. It occupies an area of 9000 km², and contains a sedimentary pile probably 10 km thick in water depths of 200–3000 m. Though not tested by drilling, the basin is covered by a grid of seismic data. By analogy with the Eyre Sub-basin to the east, the Bremer Basin probably contains Late Jurassic to Barremian continental deposits overlain by Albion and Late Cretaceous marine deposits with a veneer of Tertiary open-marine carbonates of variable thickness.

The Bremer Basin formed during the period of continental extension that preceded the breakup of Australia and Antarctica in the mid-Cretaceous. However, Triassic (2nd older) extension and spreading events in the Perib Basin, a short distance to the west, are likely to have influenced its evolution. Basement structural trends in the basin indicate an old east–west-trending (?Palaeozoic) fabric that has been overprinted by north-northwesterly oriented Jurassic–Cretaceous extension and wrenching. The resultant structure is complex, particularly where the Palaeozoic and Mesozoic trends intersect.

The hydrocarbon potential of the Bremer Basin is currently unknown. However, by analogy with the Eyre Sub-basin, potential source and reservoir sections can be inferred to exist, although the presence of a regional seal and a tightflow regime adequate for the generation of hydrocarbons is less certain. Potential trapping mechanisms for hydrocarbons include wrench-induced anticoine, elastic aprons adjacent to boundary and transfer faults, and stratigraphic traps within dipping Neocomian rocks beneath a major angular unconformity.

Introduction

The Bremer Basin is an isolated continental-slope basin on the southern margin of Western Australia (Fig. 1). It formed in response to the Mesozoic extension between Australia and Antarctica that preceded breakup of that part of Gondwanaland. Despite its thick sedimentary pile (estimated at 10 000 m by Cooney, 1974), it probably remains the most poorly known basin on Australia’s southern margin.

The intention of this paper is to review all the available seismic data in the Bremer Basin, and to provide an interpretation of the structure and stratigraphy and a first assessment of its hydrocarbon potential.

Background

The Bremer Basin is the westernmost basin in what has been referred to as the ‘Southern Rift System’ (Stagg & others, 1990). This rift system, which extended for about 3000 km from near Broken Ridge (west of the Naturaliste Plateau; Fig. 1) to the South Tasman Rise, incorporates several other sedimentary basins — viz. the Great Australian Bight (GAB) Basin (including the Eyre, Recherche, and Ceduna Sub-basins), Duntroon Basin, Otway Basin (including the Robe and Penola Troughs), and Sorell Basin (including the King Island, Sandy Cape, and Strahan Sub-basins).

In this paper, the term ‘Bremer Basin’ is applied to the major sedimentary accumulation that is confined almost entirely to the continental slope between longitudes 117 and 121°E (Figs. 2 and 4). Onshore sedimentary rocks assigned to the Bremer Basin (e.g., Robertson & others, 1979) probably do not exceed 200 m in thickness, and are insignificant compared with the thick pile on the continental slope.

The main depocentre of the Bremer Basin (referred to as the Albany Sub-basin by Middleton, 1991) appears to be an extensional basin that formed during rifting of the Australian and Antarctic plates, and underwent sag-phase development following plate breakup and spreading, which are interpreted to


have taken place in the mid-Cretaceous (Cande & Mutter, 1982; Veevers & Eittreim, 1988; Stagg & Wilcox, in preparation). The basin has an area of about 9000 km², excluding any extension onto the continental rise; its size is thus similar to that of the better-known Eyre Sub-basin of the Great Australian Bight, to the east. These two basins share some common structural trends, and may well have similar geological histories.

The earliest seismic survey of the Bremer Basin was conducted by Teledyne in 1970, using a spark energy source with 12-fold CDP coverage; fair-quality data were recorded. The nearshore eastern flank of the basin was explored by Continental Oil in 1972, using an airgun source with 24-fold CDP coverage. The sedimentary cover in this area is thin, being no more than 1000 m at the edge of the continental shelf. Also in 1972, Shell Australia recorded portions of three high-quality 24-fold airgun-array lines across the basin, while BMR surveyed the entire area at a reconnaissance level during survey 19 of the continental margins survey (CGG, 1975); data quality from BMR survey 19 was poor.

The principal data set in the Bremer Basin is Esso’s Bremer (R74A) marine seismic survey, in which 2224 km of 24-fold seismic data were recorded in 1974 (Cooney, 1974; Cooney & others, 1975; Fig. 2). These data are of good quality for the vintage, though the regional line spacing is generally inadequate for correlating structures from line to line. Esso’s analysis of these seismic data led to the identification of some major anticlinal structures; however, the prognosis was not sufficiently encouraging for a drilling program to be mounted, and the acreage was eventually dropped. No further seismic data have been recorded since 1974, and the basin is now assumed to be unprospective, or is virtually unknown by most of the petroleum industry.

Studies of the Bremer Basin are rare. Cooney (1974) analysed the 1974 Esso data, and Cooney & others (1975) summarised some of the more noteworthy aspects of the data. More recently, Hocking (1990) summarised the broad structure of the Bremer Basin, concentrating on the thin onshore and continental-shelf sedimentary rocks. Middleton (1991) interpreted the tectonic history of the whole southern continental margin of Western Australia, including:

- sinistral transcurrent movement along the margin, and development of the Eyre and Albany (Bremer) Sub-basins, starting at about 150 Ma;
Figure 1. Structural elements of the Southern Rift System of Australia (after Wilcox, 1990).
The study area of this paper is outlined by the box off southwest Australia. Gross bathymetry is shown by the 300, 1000, 2000, 3000, and 4000-m isobaths. Sedimentary basins showing main extensional and transfer faults are represented by a coarse dot screen. J = Jerboa 1 well. Numbered lines on the oceanward side of the magnetic quiet zone (MQZ, hachured) are seafloor-spreading magnetic-anomaly traces interpreted by Cande & Mutter (1982; previous interpretation by Weissel & Hayes, 1972, shown in brackets). MT is the magnetic trough which defines the landward edge of the MQZ. Double-headed arrows show the sense of lithospheric extension (Wilcox, 1990); paired arrows along the southwest margin of the Ceduna Terrace and oceanward margin of the Otway Basin show predicted strike-slip tranpressional or transpressional zones during an extensional phase in the pre-Late Jurassic.
Figure 2. Bathymetry of the Bremer Basin area (contours in metres), and the tracks of multichannel seismic surveys used in the interpretation.
Though there are differences in detail, this interpretation is broadly consistent with observations that we have made in recent publications (Stagg & others, 1990; Willcox, 1990; Willcox & Stagg, 1990; Stagg & Willcox, 1991, in preparation).

Basin development

Bathymetry

The continental slope of the southwest Australian margin is highly variable in width and gradient, and is interrupted by several terraces (Fig. 2). From offshore Albany to the western GAB, gradients are up to 6°. Canyon development is extensive, particularly west of longitude 122°E, though individual canyons are poorly defined owing to the paucity of lines parallel to the slope. In the Bremer Basin area, in particular, canyon development appears to be structurally controlled, and many canyons have formed along fault-block boundaries (including transfer faults); this structural control has produced a number of major canyons oblique to, and in some instances parallel to, the continental slope. A small terrace at about 1000–2000 m water depth has developed above the central Bremer Basin between 119 and 120°30′E.

The base of the slope lies at about 4000 m depth in the Bremer Basin.-In the Bremer Basin area, in particular, canyon development appears to be structurally controlled, and many canyons have formed along fault-block boundaries (including transfer faults); this structural control has produced a number of major canyons oblique to, and in some instances parallel to, the continental slope. A small terrace at about 1000–2000 m water depth has developed above the central Bremer Basin between 119 and 120°30′E.

The base of the slope lies at about 4000 m depth in the Bremer Basin area. Here, the continental rise is composed of a sedimentary apron that is dammed on its southern side by the rugged topography of the Diamantina Zone.

Stratigraphy

The stratigraphy of the rocks assigned by Robertson & others (1979) to the onshore Bremer Basin is restricted to a marine transgressive sequence of Eocene age. This sequence constitutes the Plantagenet Group at the coast, and the Eudynie Group inland. These groups comprise limestone, spongolite, clay, siltstone, sandstone, lignite, and carbonaceous siltstone, and have no recognised hydrocarbon significance (Robertson & others, 1979).

Because no samples have been recovered offshore, the stratigraphy of the continental-slope accumulation of the Bremer Basin is unknown. It can only be inferred from the structural, lithological, and facies relationships with other basins in the Southern Rift System, most notably those of the Eyre Sub-basin (700 km to the east).

The stratigraphy of the Eyre Sub-basin is at present known only from Esso Jerboa 1 well (Bein & Taylor, 1981), and from dredge samples collected aboard RV Rig Seismic during BMR survey 66 (Davies & others, 1989).

Jerboa 1. Jerboa 1 was sited on the crest of a small fault-block within a half-graben in the western half of the Eyre Sub-basin. Its objective was to test a compaction-drape target in interpreted Cretaceous rocks above the fault-block (Fig. 3). The stratigraphic summary presented below comes from Powis & Partridge (1980), Bein & Taylor (1981), Stagg & others (1990), and Blevin (1991).

In the well, the basal sedimentary section, above Precambrian basement, consists of 410 m of the Berrisanian non-marine Loongana Sandstone. Poorly sorted sandstone at the base is interpreted to be a weathering product and debris derived locally from basement soon after basin initiation. The remainder of the basal section consists of sandstone with interbedded siltstone and shale deposited in fluvial and lacustrine environments. Eastwards-prograding foresets in the west of the basin in the upper part of this interval suggest deposition in a probably deep lacustrine environment.

The basal section is unformably overlain by a succession of dark grey to dark brown claystone and shale, with rare interbeds of siltstone, deposited in a fresh or brackish-water lacustrine environment. These beds, equivalents of the Neptune Formation of the eastern GAB (Stagg & others, 1990), are now interpreted to be early Valanginian to Barremian in age (Blevin, 1991).

Jerboa 1 lacks a Turonian to Early Eocene section, representing a gap of about 40 Ma in the sedimentary record. Though dredge samples show that at least some of this section is preserved in the structurally lower parts of the sub-basin, a major erosional event, probably combined with lengthy non-deposition, has apparently affected the region. Sedimentation resumed with the deposition of the Hampton Sandstone, 28 m thick, in the latest Early Eocene. This sandstone is succeeded by calcilutite and marlstone of the Wilson Bluff Limestone, and poorly consolidated open-marine prograding carbonate of the Nullarbor Limestone which dominate the remaining 335 m of the Jerboa 1 section.

Jerboa 1, which is centrally located in the Eyre Sub-basin, penetrated a representative Cretaceous–Tertiary section, but a condensed and possibly incomplete ?Jurassic–Neocomian section. The only section apparently missing from Jerboa 1 is a thin southwards-prograding unit which occurs between the Cenomanian and Eocene rocks in the northern half of the Eyre Sub-basin; this unit is presumed to consist of prograding shelf-edge sandstone of Paleocene or Early Eocene age.

Dredge samples. During BMR survey 66, three dredge-hauls (DR01–03; Davies & others, 1989) sampled rocks from the Eyre Sub-basin. Dredge DR02 recovered moderately altered and sheared granodiorite of presumed Precambrian age from the lower-slope basement scarp below the Eyre Terrace (Fig. 1). Similar rocks are expected to underlie the Eyre Sub-basin. Dredges DR01 and DR03 sampled the broad Eucla Canyon at the eastern end of the Eyre Sub-basin. This canyon is one of the rare locations in the central western GAB where pre-Tertiary rocks are accessible to dredging. Rocks recovered included Maastrichtian sandstone, siltstone, mudstone, and conglomerate; Paleocene phosphatic rock; and Tertiary fine-grained limestone and siliceous carbonate. Fragments of amygdaloid basaltic lava were also recovered. They contain an abundance of large amygdules, indicating eruption at a moderately shallow depth. Their occurrence with Maastrichtian and younger rocks suggests that they erupted no earlier than the Maastrichtian.
Figure 3. Seismic section through Esso Jerboa 1 well in the Eyre Sub-basin, showing interpreted stratigraphy (after Stagg & others, 1990).

Location of Jerboa 1 shown by 'I' in Figure 1.

**Basement structure**

The complexities of the Bremer Basin, combined with the reconnaissance nature of much of the data grid, preclude detailed seismic mapping. Consequently, only the gross tectonic elements shown in Figure 4, together with some key lines across the basin (Figs. 5–10), are discussed here. Many prominent individual structures are not shown in Figure 4, as they appear only on single lines and their trends cannot be determined.

In the Bremer Basin area, basement can only be unequivocally identified beneath the continental shelf. Here it is generally manifest as a moderately smooth or undulating surface at a depth of no more than 0.4 s two-way time (twt; ca 500 m) below seabed at the shelf edge (horizon 'b' in Figs. 7 and 8). The only indication of a sedimentary depocentre on the shelf is a narrow shallow half-graben trending west-southwesterly from the shelf edge east of Albany (Fig. 4); the seismic character suggests that this is an older sequence, probably of Palaeozoic age. Though the magnetic signature of most of the shallow basement is very subdued, isolated areas on the continental shelf south and west of Albany (Fig. 4) have associated intense magnetic anomalies; these anomalous areas are probably a function of variations in the basement composition or of later igneous intrusion.

The northern boundary of the Bremer Basin is typically a major high-angle fault scarp affecting the Precambrian basement (Figs. 6, 7, and 9); this boundary is the most prominent feature of the tectonic elements map (Fig. 4). The scarp shows changes in regional trend which broadly follow the coastline: from roughly easterly south of Albany, to east-northeasterly adjacent to the central part of the basin, and easterly again near its eastern end. The western limit of the basin is complex and ill-defined according to the seismic data, whereas the eastern termination at about longitude 121°E, associated with a pair of northwesterly trending transfer faults, is clear-cut (Fig. 10). Minor offsets in the basement scarp appear to be due to northwesterly trending transfer faults.

The moderately abrupt changes in trend of the basement scarp through the Bremer Basin and Eyre Sub-basins (from east-northeasterly to easterly and vice versa; see also fig. 40.1 in Stagg & others, 1990) are thought to be caused by the overprinting of a prominent, roughly northwesterly Jurassic–Cretaceous extensional direction on older (?Palaeozoic and Precambrian) easterly trending lines of crustal weakness. One of these older lineaments extends from the western end of the Eyre Sub-basin across the central GAB and through the Polda Trough. A second underlies the continental slope from about longitude 120°E to the eastern end of the Archipelago of the Recherche (longitude 124°E). A third, less well defined, may underlie the margin south of Albany. In an interpretation of the Polda Trough, Nelson & others (1986) suggested that these lineaments formed at the same time as the easterly oriented basins of central Australia (for example, the Ngaila and Amadeus Basins).

Beneath the Bremer Basin, basement is rarely evident because of the considerably thick sedimentary pile and the limited penetration displayed on the 1974 seismic data. It can be identified only with certainty within some elevated fault-blocks.
Figure 4. Tectonic elements of the Bremer Basin area (after Stagg & others, 1990).
Even so, it is inferred to be at moderately shallow depth beneath the lower continental slope (Fig. 4). Basement beneath the continental rise lies at considerable depth (2.5 s twt, or 3000-4000 m below the seabed; Stagg & others, 1990).

The main Bremer Basin depocentre can be broadly divided into four zones. From west to east, these are:

- a small syncline (Denmark trough) containing more than 3 s twt of sedimentary fill in water 1000-1300 m deep (Fig. 7); the complex bathymetry of this part of the margin appears to reflect the underlying structures, so that canyon development typically follows prominent basement fault traces; the Denmark trough is separated from the remainder of the Bremer Basin by a large canyon offshore from Albany;

- a zone of complex block-faulting in which throws on the major faults are up to 3 s twt (3000-5000 m); the disposition of the sedimentary horizons in relation to the fault planes indicates some compression due to wrenching (Fig. 8); and the trends of individual structures cannot be discerned with the 40-km line spacing;

- a zone of relatively mild structuring within a thick sedimentary pile in the central part of the basin underlying a narrow mid-slope terrace; and

- a zone of extensive high-angle faulting and wrench structuring in the east of the basin (Fig. 9), where some large-wavelength high-relief anticlines have been deeply eroded at the crests.

We suggest that complexities within the zones of complex block-faulting and extensive high-angle faulting and wrench structuring have resulted from interaction between Jurassic-Cretaceous extensional structures and easterly trending Palaeozoic and older lineaments.

**Sedimentary fill**

The sedimentary fill of the Bremer Basin is restricted to a narrow, highly structured belt beneath the continental slope between longitudes 117 and 121°E in water depths ranging from 200 to 3000+ m. The greatest sedimentary thickness, estimated to be at least 10 000 m by Cootey (1974), occurs in water depths of 600-1500 m.
Table 1. Characteristics of the seismic sequences in the Bremer Basin

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Upper boundary</th>
<th>Lower boundary</th>
<th>Internal configuration</th>
<th>Age</th>
<th>Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td>wb</td>
<td>erosional</td>
<td>onlap</td>
<td>moderate amplitude; low to moderate continuity</td>
<td>Miocene–Recent</td>
<td>open-marine carbonate</td>
</tr>
<tr>
<td>P</td>
<td>erosional</td>
<td>onlap</td>
<td>moderate amplitude; low to moderate continuity</td>
<td>Eocene</td>
<td>open-marine</td>
</tr>
<tr>
<td>Q</td>
<td>erosional</td>
<td>onlap; concordant</td>
<td>moderate amplitude; low to moderate continuity</td>
<td>late Neocomian–Late Cretaceous</td>
<td>base: fluvial to paralic; top: shallow-marine</td>
</tr>
<tr>
<td>R</td>
<td>erosional</td>
<td>onlap; concordant</td>
<td>moderate to high amplitude; moderate continuity</td>
<td>Berriasian–Valanginian</td>
<td>fluvial and lacustrine</td>
</tr>
<tr>
<td>S</td>
<td>erosional</td>
<td>obscured</td>
<td>low to moderate amplitude and continuity</td>
<td>?Late Jurassic</td>
<td>fluvial and lacustrine</td>
</tr>
<tr>
<td>b</td>
<td>erosional</td>
<td>acoustic basement</td>
<td></td>
<td>Precambrian</td>
<td>(crystalline)</td>
</tr>
</tbody>
</table>

Four sedimentary horizons (‘P’, ‘Q’, ‘R’, and ‘S’; Table 1) have been identified on the interpreted seismic sections (Figs. 6–10). Each is sufficiently characterised to facilitate line-to-line correlation. Though there are differences in detail between the interpretation presented here and that presented by Middleton (1991), these differences are probably due to the indifferent quality of the seismic data and the lack of stratigraphic information.

The most prominent unconformity within the Bremer Basin (excepting basement) is a prominent erosional surface generally 0.5–1.0 s twt (700–1500 m) below the seabed (horizon ‘R’). Near the margins of the basin, this unconformity truncates the underlying steeply dipping section, and has the overall characteristics of the Valanginian breakup unconformity in the south Perth Basin. This led Cooney (1974) to propose a Neocomian age for the unconformity. We agree that this is the most likely age, given the proximity of the Bremer Basin to the Perth Basin and the prominence of an interpreted Valanginian reflector elsewhere on the southern margin of Australia (Willcox & Stagg, 1990).

Horizon ‘S’ is a prominent reflector within or beneath the Early Cretaceous section, roughly 1 s twt below the Valanginian unconformity in the depocentres and truncated by the Valanginian unconformity at the outer basin margin. This horizon has high seismic amplitude and continuity, and is a mild unconformity in places. A top Late Jurassic or earliest Cretaceous age is proposed for this horizon.

By analogy with the Eyre Sub-basin, the section beneath the Valanginian unconformity in the Bremer Basin is expected to consist of clastic lacustrine or fluvial deposits, probably derived from exposed basement.

Figure 7. Part of Esso line R74A–9 from the western end of the Bremer Basin (‘Denmark trough’), showing a thick folded Cretaceous sedimentary section in a restricted trough between the shallow basement of the continental shelf and the outer margin high.
Figure 8. Part of Esso line R74A–10 along strike in the central Bremer Basin, showing an interpreted wrench anticline above tilted basement blocks.

Figure 9. Part of Esso line R74A–25 from the eastern end of the Bremer Basin, showing a wrench anticline with a deeply eroded crest.
Hydrocarbon potential

Since no stratigraphic, geochemical, or geothermal data exist for the offshore Bremer Basin, and the nearest basin in the Southern Rift System from which such information is available (Eyre Sub-basin) is 700 km distant, the hydrocarbon potential of the basin can be deduced only by analogy, inference, and seismic structural and stratigraphic analysis.

Although there are obvious structural and stratigraphic differences between the Eyre Sub-basin and the Bremer Basin (Stagg & others, 1990), other similarities allow us to draw some conclusions about the hydrocarbon potential of the Bremer Basin. In particular, the probable contemporaneity and the morphologic position of both basins (high on the continental slope seawards of a major basement fault) suggest that their thermal and sedimentary histories might be similar. If so, then the Bremer Basin probably has been moderately cool throughout its history (as indicated by lithospheric extension of only about 20%, and the drilling results at Jerboa 1 in the Eyre Sub-basin; Bein & Taylor, 1981), and hence its source rocks would be immature.

Even though this is a negative conclusion, there may be some cause for optimism: heatflow and maturity could have been enhanced by the greater thickness of sedimentary fill and more intense structuring in the Bremer Basin than in the Eyre Sub-basin. M.F. Middleton (Curtin University of Technology, personal communication, 1991) has pointed out that dyke swarms near Albany have been attributed to the Cretaceous breakup of the margin. If this interpretation is valid, then volcanics might occur in the sedimentary section in the Bremer Basin, and the volcanism could have generated a higher heatflow and perhaps increased maturity.

Source, reservoir, and seal

The interpretation of the Bremer Basin seismic data in this paper leads to the conclusion that much of the sedimentary section is composed of Late Jurassic and early Neocomian rocks with a capping of late Neocomian, Late Cretaceous, and Tertiary rocks of variable thickness. The interpreted similarities between the Bremer Basin and the Eyre Sub-basin allow some measure of assessment of the quality of reservoir, seal, and source in the undrilled Bremer Basin.

Although the only well drilled in the Eyre Sub-basin was dry, there were some positive outcomes with regard to hydrocarbon potential. A number of good-quality reservoirs were identified in the lacustrine and fluvial Berriasian-Valanginian section in Jerboa 1: average porosities were in the range 17–24%. Regional seals could be expected in the overlying Barremian.
claystone and shale. The thin prograding shelf-edge Palaeocene sands could have suitable porosities, but the quality of the overlying seal is likely to be poor.

TOC data from Jerboa 1 show that most of the section penetrated is moderately organic-rich (Stagg & others, 1990). Shaly beds throughout the well have moderately high TOC concentrations. TOC averages 0.94% in the Albien section, 1.05% in the Barremian section, and 1.84% in the Berriasian–Valanginian section. The most organically rich shales occur towards the base of the Berriasian, where TOC averages 2.88% and has a maximum of 5.46%. The kerogens in the shales are dominantly amorphous, and are rich in extractable hydrocarbons, suggesting that they have a high potential for liquid hydrocarbon generation. Unfortunately the shales are fairly thin in Jerboa 1, though they may be both thicker and more thermally mature in locations away from basement highs in flanking depocentres.

Play concepts

The basin contains several large structures which may have the potential for hydrocarbon entrapment. However, the regional nature of the seismic coverage makes it difficult to define the areal extent of structural closure. With this in mind, the following hydrocarbon play types can be envisaged.

- Wrench anticlines in the eastern half of the basin in water depths of 700 m and greater. These are relatively large structures, thought to be the product of reactivation of transform faults, affecting the synrift sedimentary fill. Fault-independent closure is typically about 0.3 s twt (ca 400 m), whereas fault-dependent closure is greater than 1 s twt (ca 1400 m). However, the anticlinal crests of some were deeply eroded by the Late Cretaceous, so that hydrocarbons might have been lost before a regional seal was deposited. If intraformational seals have developed within the Neocomian section, then reservoirs might be stacked at a single location.

- Clastic aprons adjacent to the northern boundary fault and orthogonal transfer faults. As with the corresponding play type in the GAB Basin, such clastic aprons might lie up-dip from potential Jurassic–Early Cretaceous source rocks, but they will also need to be sealed both vertically and against the fault plane.

- Stratigraphic traps within dipping Neocomian rocks below the Valanginian angular unconformity (Figs. 7 and 9). This play type requires sourcing from Jurassic or Early Cretaceous rocks, migration up-dip into Neocomian reservoirs, and sealing by the post-breakup Late Cretaceous section. Although there is some doubt about a seal having been deposited before hydrocarbons were generated, this play type does have the virtue of being at shallow depths below the seabed.

Conclusions

The Bremer Basin is a large feature on the Western Australian continental slope, and is part of the Southern Rift System. It is a frontier basin which, in recent years, has escaped the interest of the petroleum industry. It probably contains up to 10 000 m of Jurassic to Tertiary sedimentary rocks which are highly structured in places. Although the heatflow/maturity regime in Jerboa 1 was found to be low, the Bremer Basin is sufficiently far away from this well to warrant renewed consideration. Of the play types envisaged, the most obvious are two major wrench-related anticlines near the eastern end of the basin, in water depths of about 700 m.

Acknowledgements

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References


Three large intraplate earthquakes near Tennant Creek, Northern Territory, on 22 January 1988

T.D. Jones1,2, G. Gibson3, K.F. McCue1, D. Denham1, P.J. Gregson4, & J.R. Bowman5

Following a yearlong series of foreshocks, extensive surface faulting accompanied an extraordinary sequence of earthquakes in the Precambrian shield of central Australia on 22 January 1988. The mainshock sequence included earthquakes of magnitude Ms 6.3, 6.4, and 6.7. Hundreds of aftershocks were recorded in the epicentral area in the first few days after the mainshocks, and the sequence continued into 1991; the largest aftershock occurred nearly 9 hours after the third large mainshock.

The three largest earthquakes were felt over more than one-quarter of the land surface of Australia, and in high-rise buildings up to 2000 km from the epicentre. A buried natural-gas pipeline was shortened by 1 m at the surface rupture, and buildings in the town of Tennant Creek, 30 km to the north-northeast, sustained minor damage.

Focal mechanism solutions indicate thrusting for the first and third large earthquakes, and a combination of thrust and strike-slip motion for the second. Ground deformation and offsets along roads, fences, and the pipeline also indicate mainly thrust-faulting. The mean azimuth of the P-axes for the first three focal mechanisms is N31°E, in close agreement with the maximum principal stress direction determined from in-situ stress measurements made in a nearby gold mine.

Introduction

The 1988 Bicentennial Australia Day holiday weekend will be long remembered by residents of Tennant Creek, in the Northern Territory, for a series of strong earthquakes which began on Friday 22 January. Three mainshocks of magnitude Ms 6.3, 6.4, and 6.7 occurred on Friday at 10.06 a.m., 1.27 p.m., and 9.35 p.m. local time (or 0036, 0357, and 1205 UTC; Table 1). They were followed by a sequence of aftershocks, of which the largest — magnitude Ms 5.3 — happened at 6.24 a.m. local time the following day (or 2054 UTC on 22 January).

Table 1. Hypocentral details of the four largest earthquakes at Tennant Creek on 22 January 1988*

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>Latitude °S</th>
<th>Longitude °E</th>
<th>Depth (km)</th>
<th>Ms/ML</th>
</tr>
</thead>
<tbody>
<tr>
<td>0036 01.0</td>
<td>19.81 (± 0.02)</td>
<td>133.98 (± 0.06)</td>
<td>6 (± 4)</td>
<td>6.3</td>
</tr>
<tr>
<td>0357 28.5</td>
<td>19.83 (± 0.02)</td>
<td>133.98 (± 0.05)</td>
<td>4 (± 3)</td>
<td>6.4</td>
</tr>
<tr>
<td>1205 00.6</td>
<td>19.84 (± 0.02)</td>
<td>133.99 (± 0.03)</td>
<td>5 (± 3)</td>
<td>6.7</td>
</tr>
<tr>
<td>2054 05.6</td>
<td>19.89 (± 0.02)</td>
<td>134.08 (± 0.15)</td>
<td>3 (± 2)</td>
<td>5.3</td>
</tr>
</tbody>
</table>

* Hypocentres from the Australian Seismological Centre; magnitudes from the US Geological Survey.

There is no historical record of nearby activity before 11 February 1986, when an earthquake of magnitude ML 4.1 occurred. This was followed in January 1987 by a swarm of over 150 events, of which four caused minor damage in the town. The epicentres of the January 1987 events, located with a three-station field network and the Warramunga Seismic Array (WRA; only 40 km to the east), were about 35 km southwest of the town, in the same area as the 1988 events. Activity continued at a low level throughout 1987.

Surface faulting accompanying the 1988 earthquake sequence was evident on two main scarps trending east-southeast with a total length of about 32 km (Fig. 1). The southern block was thrust over the northern block, except in the area where the surface faulting was east-northeast; here the mechanism was reversed — that is, northern block over southern. Uplifts of the order of 1 m were common.

Within 24 hours of the mainshocks, three of the authors (GG, TJ, and JRB) began to deploy a ten-station field network of seismographs and accelerographs, and hundreds of aftershocks...
were recorded. Surveyors mapped the scarps and associated structures, and releveled lines surveyed in the early 1970s for BMR's gravity and magnetic traverses (Hone, 1974; Bullock, 1977).

Tectonic setting

Throughout Australia, there appears to be little or no correlation between epicentral distribution and the main tectonic elements (Fig. 2). However, there is an indication of a number of halo or doughnut effects, where the epicentres surround regions which appear to be aseismic. Several earthquakes approaching magnitude 7 have taken place in the last 100 years, particularly in the western part of the continent, and surface faulting is commonly associated with earthquakes exceeding magnitude 6 where the unconsolidated cover is thin.

Data from fault-plane solutions, borehole deformations, and in-situ stress measurements show that the crust of the continent is being compressed. However, the principal stress direction, mainly east–west, is not consistent throughout the continent (Lambeck & others, 1984).

In the Tennant Creek area, extensive superficial cover overlies basement that was cratonised during the early to middle Proterozoic. The main rock types are metamorphosed marine sedimentary rocks, volcanics, and granitic intrusives.

In the Warrego gold mine, 50 km northwest of Lake Surprise (Fig. 1), a comprehensive series of stress measurements was carried out in 1975 at depths of 245 m and 320 m (Worotnicki & Walton, 1979). The results showed that the vertical stress is lithostatic, and that the horizontal stresses (about 20 MPa) are about twice the vertical stress. The average azimuth of the maximum principal stress is N32°E.

The earthquake sequence

The foreshock sequence commenced on 5 January 1987. The six largest foreshocks occurred within the next five days, and ranged in magnitude from ML 4.8 to 5.4. The maximum intensity experienced in Tennant Creek from these earthquakes was MMVII. The instrumental epicentres for five of the six are within 1 km of the Lake Surprise scarp, and so is the epicentre for the largest earthquake of 1987, an event of magnitude ML 4.7 on 30 January. Their locations suggest that they occurred on the same fault surface as the 1988 Lake Surprise rupture at very shallow depths. The largest 1987 event has a hypocentre 3 km south of the scarp at a depth of 4 km (Fig. 1).

Hypocentres of the 30 or so earthquakes in January 1987 which exceeded Richter magnitude 3 were all shallow (less than 6 km). All were clustered around the central area of the 1988 surface rupture, some having occurred between the two surface scarps but none in the unconsolidated cover. The largest 1987 event was a magnitude 4 or larger, 11 of them in January.

The prelude to the dramatic 1988 sequence was a foreshock of magnitude ML 4.0 at 10.00 a.m. local time on 22 January. Six minutes later the first mainshock occurred. It was followed by thousands of earthquakes over the next few days. Seismologists working in the epicentral area were experiencing upwards of 50 earthquakes per day at first. In the last ten days of January, there were about 360 earthquakes of Richter magnitude 3 or larger, of which about 94 were of magnitude 4 or larger. Seismicity then gradually declined through 1988, except for a temporary increase in September. Cumulative 1988 totals were 836 events of magnitude 3 or larger, and 161 of magnitude 4 or larger. Aftershocks were still occurring in mid-1991.

To determine the earthquake hypocentres shown in Figure 1, we composed a travel-time model from the results of two crustal surveys in the region (Finlayson, 1981; Finlayson, 1982). The 1987 and 1988 mainshocks, and some other events from both years, were located using the 20-element WRA array and the 19-element Alice Springs array (420 km south). Except for the mainshocks, the precision obtained, corresponding approximately to a 95% confidence level, was typically ± 5 km east–west and vertically, and ± 2 km north–south. The computed errors in mainshock focal co-ordinates are listed in Table 1. The remaining hypocentres in Figure 1 were located using the two fixed arrays and field seismographs. The 1988 and 1989 aftershocks shown have location uncertainties of 5 km or less.

Hypocentres for the three 1988 mainshocks are grouped near the surface faulting, whereas the large aftershock at 0540 UTC occurred near the eastern extremity of the surface faulting. Most of the 1988 aftershocks that we located occurred in the central and southeastern portion of the epicentral area (Fig. 1), some apparently beyond the eastern extremity of the surface faulting. Nearly all the earthquakes that we have located, from 1987 and 1988, occurred down-dip from the trend of the surface faulting.

Damage and felt effects

Damage from the 1988 events is estimated at around $A1 million if lost production from local underground gold mines is included. At the surface rupture, a pipeline linking Darwin to gas fields west of Alice Springs was shortened by 0.97 m (S. Dykes, NTGAS, personal communication, 1988). Remarkably there was little structural damage in the town of Tennant Creek,
30 km from the surface rupture; walls were cracked in well constructed buildings, objects fell off shelves, and furniture was shifted.

The three largest earthquakes were felt over a large area of the Australian continent. The first two, at 10.06 a.m. and 1.27 p.m. local time, were felt in high-rise buildings in Perth (2200 km to the west of the epicentre), Adelaide (1600 km to the south), and Cairns (1250 km to the east). They were so close together that a combined isoseismal map was prepared (Fig. 3). The highest intensity reported was MMVIII at the western end of the surface faulting, where a full 150,000 L water tank was destroyed by one of the events. The MMIII isoseismal encompasses an area of 1.8 million km².

Unfortunately, responses for information on the third (largest) earthquake were not as numerous as for the first two, and its isoseismal map is not as well constrained. The radii of the isoseismals were significantly larger for the third event, 425 km and 625 km for the IV and V isoseismals respectively. These radii are similar to those of the 1941 Meeberrie earthquake, Western Australia, the largest known onshore Australian earthquake.

Focal mechanisms

Preliminary focal mechanisms (Fig. 4) were obtained from an inspection of regional long-period seismograms and from long-period P-wave arrivals listed in the Earthquake Data Report (USGS, 1988). We used the relative amplitude of the first P-wave arrival to indicate qualitatively which stations were near-nodal. Both the first and third earthquakes have simple thrust mechanisms with a southwest-dipping fault plane (chosen from field observations). The dip of this plane is constrained by the polarity of arrivals at MUN, NWAO, and ASP, and is greater for the first earthquake, 55°, than for the third, 36°. The second nodal plane for these earthquakes is poorly controlled, but is constrained for each one by the near-nodal arrival at CTAO.

Our mechanism for the second earthquake comprises approximately equal elements of thrust and lateral slip. The nodal plane striking N70°E is well constrained by data which differ from those of the first and third earthquakes. The first motion at PMG was up and near-nodal, station LEM was near-nodal, and CTAO was not. This nodal plane corresponds with the surface faulting observed along the western limb of the Lake Surprise scarp, which is a combination of left-lateral strike-slip and north-over-south thrusting. Other authors differ in their analysis of the mechanism for this earthquake. The moment-tensor and fault-plane solutions of the USGS (1988) are similar to our solution, but the Harvard (in USGS, 1988) and McCaffrey (1989) solutions have nearly identical thrust mechanisms for all three earthquakes, and Choy & Bowman (1990) assign the dominant moment release to slip on a southeast-dipping fault plane.

A fault-plane solution for the largest aftershock (at 2054 UTC) was determined from short-period data. It indicates a strike-slip mechanism with the left-lateral nodal plane paralleling the trend of faulting.

The three largest earthquakes have an average principal stress direction of N31°E, which is in close agreement with the measurements taken in 1975 at Warrego mine.

Crustal deformation and fault geometry

The surface expression of faulting comprised two main scarps (Fig. 1). Relative vertical movement of blocks adjacent to the scarps was apparent over an area of more than 100 km². Scarp morphology was typically a gradual step in the alluvium, extending 100 m or more across strike.

General compression of the crust in a north-northeasterly direction was indicated by offsets along fences and roads, and tensional cracks in soil, at this azimuth along all portions of the scarps; these indicators of compression had an en echelon...
arrangement on the western limb of the Lake Surprise fault. Hence, most of the surface faulting (all of the faulting trending east-southeasterly) was almost pure thrusting, with some evidence of minor left-lateral strike-slip; along the eastern limb of the Lake Surprise fault, the southern block was high. In contrast, along the east-northeasterly trending western limb of the Lake Surprise fault, the northern block was upthrown, and there was a much stronger left-lateral strike-slip component.

Two trenches, 3 m deep, were excavated across the scarp near the shortened pipeline, and the fault plane was clearly seen, apparently dipping south at 20°-25°. In shallow trenches dug elsewhere (A. Camacho, US Geological Survey, personal communication, 1988), the Kunanyanku fault segment was also seen to dip to the south-southwest. However, near the western end of the Lake Surprise fault, a trench showed a north-dipping fault plane.

Aftershock hypocentres confirm the fault geometry observed at the surface (Fig. 1; Bowman & others, 1990). We projected our best-located 1988 hypocentres (uncertainties less than 2 km and good azimuthal control) onto sections normal to the scarps: section AB with hypocentres between the two scarps, and section CD with hypocentres south of the Lake Surprise scarp. Section CD, in the east, depicts a fault surface dipping to the south-southwest at about 35° to a depth of 8 km. By contrast, section AB shows that the western limb has a fault surface with opposed dip direction, dipping to the north-northwest at 55°-60° to about the same depth. Bowman & others (1990) identified a surface dipping to the south-southwest at up to 45° to a depth of about 6 km on the Kunanyanku scarp.

The rupture at 0036 UTC appears to have begun near the cusp of the Lake Surprise fault, and propagated mainly northwest. The easterly extent of this propagation is limited by the account of the Lake Surprise fault, and propagated mainly northwest. The second mainshock, at 0357 UTC, originated near the first.

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The second mainshock, at 0357 UTC, originated near the first at a preferred depth of 3 or 4 km, which is similar to that of the largest 1987 earthquake. We suggest that this second earthquake fractured the western limb of the Lake Surprise fault, because its mechanism conforms with the observed faulting along this segment. We further suggest that the fracture was bilateral, extending to both limbs of the Lake Surprise fault and involving rupture on two fault surfaces of opposing dip— to the south-southwest and north-northwest. Therefore we regard the Lake Surprise scarp as essentially continuous, although complicated at its cusp. The complexity of this rupture would account for the variety of published focal mechanisms. It is not known if this fracture propagated across the pipeline.

The third (largest) event, at 1205 UTC, originated at a preferred depth of 5 km on the south-southwest dipping eastern limb near the first two earthquakes. We propose that the extent of its fracture included all of the southeastern fault surface; it almost certainly broke across the pipeline. If past Australian earthquakes are any guide, this one was large enough to have fractured or refractured the entire fault trace of 32 km. The similar-size 1968 Meckering earthquake (Fig. 2) had a surface fault 37 km long.

The surface faulting parallels isogals which define a local positive gravity anomaly, whose peak is between the Kunanyanku and western Lake Surprise scarps (Fig. 1). Bullock (1977) postulated a dense body at a depth of 1.2 km to model the peak of the anomaly, but conceded that a body with a lesser density contrast could reach the surface. The body responsible for the anomaly is clearly shallow enough to have affected the style of faulting, and might reach the surface in places. One such place is the westernmost 4 km of the Lake Surprise fault, coinciding with an exposed quartz ridge, where the fracture has reactivated a dislocation surface of unknown age presumably contained within the anomalous body.

**Implications for intraplate risk analysis**

Several attempts have been made to delineate earthquake risk in Australia (McEwin & others, 1976; others listed in Gault & others, 1990). These were based on the location of past earthquakes and on local geology to define the source zones. Two of the last six large-earthquake sequences since 1968 occurred outside recognised seismic zones: at Marryat Creek, South Australia, on 30 March 1986 (Ms 5.8; Fig. 2; McCue & others, 1987), and at Tennant Creek. When the whole continent is treated as a single-source zone, a statistical risk analysis results in negligibly small risk throughout. This is untenable, but, until the causes of intraplate seismicity are understood, it is impossible to predict whether the next large earthquake will also occur outside one of the mapped zones.

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