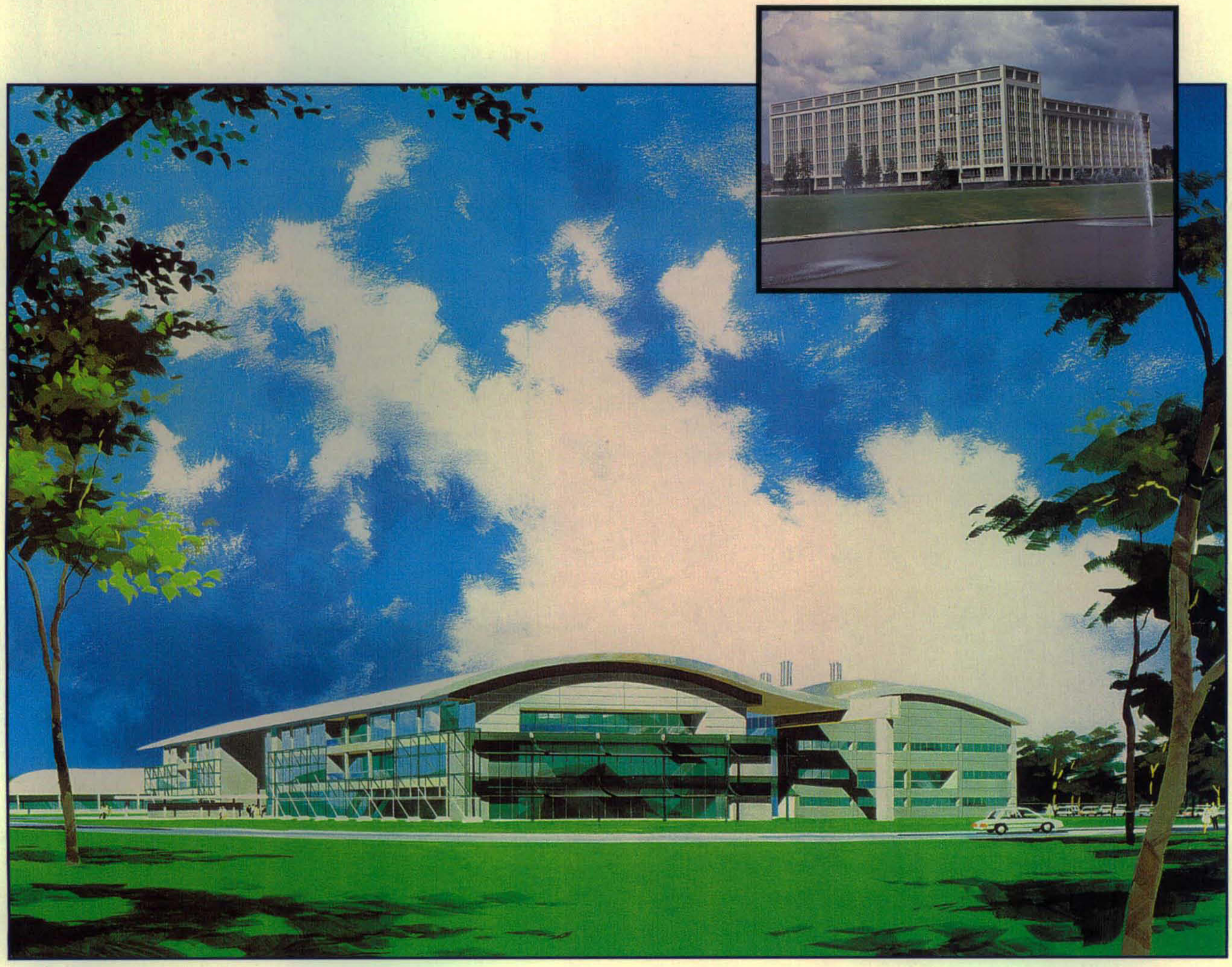
AUSTRALIA

AGSO JOURNAL
OF AUSTRALIAN GEOLOGY & GEOPHYSICS



AGSO JUBILEE 1946-96

VOLUME 17, NUMBER 1

AGSO Journal of Australian Geology & Geophysics

Information for contributors

From the end of volume 17, the *AGSO Journal of Australian Geology & Geophysics* will merge with the *Australian Journal of Earth Sciences* (*AJES*). The *AJES* is published by Blackwell Science for the Geological Society of Australia Inc. The editorial board of the *AGSO Journal of Australian Geology & Geophysics* is confident that the new combined journal, part of Blackwell's internationally recognised stable of quality scientific journals, will be well placed to keep Australian research at the leading edge of international geoscience.

As volume 17 is a series of special thematic issues, which are all fully committed, the Editor is unable to accept further unsolicited contributions. However, the Editor of the *Australian Journal of Earth Sciences* will be pleased to consider manuscripts for publication. Contributions should be sent to A.E. Cockbain, PO Box 8114, Angelo Street, South Perth, WA 6151, Australia (tel. and fax 09 367 7037; email: tcockbai@cyllene.uwa.edu.au).

AGSO JOURNAL

OF AUSTRALIAN GEOLOGY & GEOPHYSICS

VOLUME 17, NUMBER 1, 1997

ISSN 1320–1271

Copies of the AGSO Journal can be purchased from the Australian Geological Survey Organisation (GPO Box 378, Canberra ACT 2601; tel. 02 6249 9642, fax 02 6249 9982).

Other matters concerning the Journal should be sent to the Editor, AGSO Journal

Editors, AGSO Journal: Ian Hodgson & Bernadette Hince
Cover design and figures prepared by AGSO Spatial Information and Mapping Services, unless otherwise indicated
Prepared for publication by Lin Kay

Printed in Australia by National Capital Printing, Fyshwick, A.C.T. 2609

Month of issue, October 1997

***Front-cover illustration*:**

Nearing completion in Canberra, the Australian Geological Survey Organisation's new purpose-built home will finally bring all parts of the organisation together under one roof. Designed by Australian architects Eggleston Macdonald, it is being built by Baulderstone Hornibrook, with project management by Australia Pacific Projects. With nearly 30 state-of-the-art laboratories, including geochronology, organic geochemistry, microbiology, and isotope and mass spectrometry, the new AGSO building will showcase several innovative ESD (ecologically sustainable development) design features. These include geothermal heat-pump air-conditioning, which uses a ground loop heat exchange system, comprising 350 ground loops in bore holes 100 metres deep, making it the largest such installation in Australia. The inset shows AGSO's present building, which it has occupied since 1965.

AGSO Journal of Australian Geology & Geophysics, 17(1), 1-11

BMR's legacy and AGSO's mission: strategic influences on the future direction of Australia's national geological survey

T.G. Powell[1]

Rapid changes in the operating environment for national geological surveys are forcing an almost constant re-evaluation of their role and the Australian Geological Survey Organisation is no exception. The diversification of the Australian Geological Survey Organisation's client base to include national resource managers and policy makers, the changing demands of an increasingly competitive minerals and petroleum industry, and changes in government policy relating to program delivery have created the need for a sharp client focus on the organisation's work and output. This paper explores the present needs for geoscience information and draws some implications for the future direction of Australia's national geological survey.

Introduction

By the time the Australian Geological Survey Organisation (AGSO) entered its 50th year in 1996, the societal need for geoscience information had dramatically expanded beyond the traditional role of national development of the minerals and petroleum industries. The most important new area in which geoscience information is now needed is that of sustainable development. The need to abate and reverse environmental degradation will be the most serious challenge to society in the 21st century and reliable geoscience information on environmental issues will be essential. Whilst this change in societal requirements is widely recognised and is reflected in the changing agenda of national geological surveys (Price 1992, Cook 1995, Eaton 1995), each country has its own priorities based on its level of development, population, resources and environmental issues (Cook 1995). In Australia the following issues are particularly prominent:

- **land degradation,** which is threatening the sustainability of much agricultural production
- **groundwater** — 60% of Australia (by area) depends on groundwater, and an additional 20% is partly dependent, for water supply
- **land management and pollution** in the coastal zone as Australia's coastal urbanisation proceeds
- the **Australian Ocean Territory,** an additional 16 million square kilometres under Australia's jurisdiction for which basic information is lacking.

These issues, and the pace of change in government policies and in community expectations, require constant adjustment to AGSO's *modus operandi* and its role exemplified by the mission given to AGSO by government:

> *AGSO's primary mission is to build a vigorous, client-driven national geoscientific effort to encourage economically and environmentally sustainable management of Australia's minerals, energy, soil and water resources.*

AGSO needs to adapt constantly to meet the challenges of this rapidly changing external environment. The nature of a modern national geological survey has been aptly put by Price (1992):

> National geological surveys are in the geoscience information business. They exist ... to meet the geoscience information requirements of the nation, as defined and redefined from time to time by the government. Their welfare and their survival are contingent upon their success in identifying and satisfying the needs of their clients. ... In order to function effectively, a nation needs a source of impartial and scientifically trustworthy geoscience information and expertise. National geological surveys meet this need by conducting field research and related laboratory research; by compiling information available from State or provincial agencies, universities, industry and various other sources; and by using all of this to maintain a national geoscience knowledge base from which the needed geoscience information can be extracted as required. The national geoscience knowledge base is an important national resource, but it becomes depleted as science advances if it is not updated continually. In order to maintain the viability of a national geoscience knowledge base, the time required for the research, including the systematic geological surveys is longer than the lifetime of many national policy issues. Therefore, good strategic planning is crucial to the effective operation of national geological surveys.

This suggests that the future role of a national survey is assured. However, modern trends in government, the overall public and economic environment and technological change suggest that the geoscience information needs of society can be met in various ways. General acceptance of the strategic role of geological surveys, which underpins their existence, is not easy to attain with critical examination of the public and private interest in the provision of government services, and where the provision of services is being seen increasingly as a contracted activity with clear separation between purchaser and provider. Furthermore, government policies now relate more to the establishment of frameworks and decision-making processes than to the provision of services and management of resources, both of which are increasingly devolved and corporatised. In Australia, in particular, this is reinforced by the constitutional separation of responsibilities between State and Commonwealth Governments and by the enhanced interest of State Governments in the role of mineral and petroleum exploration in economic development.

This paper canvasses modern needs for geoscience information, policy trends which affect the provision of that information, and ways of providing the information to meet the needs of government and society. It outlines how these needs can be met by a national geological survey, and the levels of performance required.

Government policy

Resource development and management

The establishment of the Bureau of Mineral Resources in the post-war period occurred when national development was seen as a primary policy goal, which was to be achieved by a mixture of direct and indirect government support and regulation of the primary and energy industries. These industries now play a major role in the national economy. Present policy is to develop and maximise the economic benefits of these industries in an increasingly competitive and deregulated national and international market place, but also in an environment in which sustainable development and social justice considerations are of major importance.

Modern policy is made in a complex political and public environment. Government policy-making therefore emphasises

[1] Deputy Executive Director, Australian Geological Survey Organisation, GPO Box 378, Canberra, ACT 2601, Australia

the establishment of mechanisms and frameworks to facilitate consideration of issues in which a range of influences from across and outside government are brought together. This is particularly the case in the natural resources area, where there is frequently competition for land between developers, conservation interests, and indigenous peoples.

This policy complexity also extends to land management issues, sustainability of agriculture and social policy in rural Australia. Thus many policy issues are no longer the monopoly of a particular government portfolio and the capacity of particular interest groups to influence policy directions has risen commensurately. This concentration on complex policy issues provides further emphasis to the separate constitutional responsibilities of the States and Commonwealth Government where the former has the predominant responsibility for day to day management of land related issues. Only in the offshore does the Commonwealth have the primary legal responsibility for administration.

In recent times the States and Territories have renewed their interest in natural resource development. Many have refined their mining legislation and regulations to facilitate exploration and have invested significantly in geological survey work to enhance perceptions of prospectivity and attract exploration investment. In this process the State geological agencies have divested themselves of traditional geological functions outside the mineral and petroleum resources areas. Other areas of geological survey work are variously represented in other State agencies. such as water authorities.

Markets for science

The Industry Commission (1995) has recently reviewed the effects of research and development on innovation, industry competitiveness and economic growth, and the performance of government policies and programs that influence research and development and innovation in Australia. The review suggested guidelines for policy design relevant to the supply of geoscience information to meet community needs. The Industry Commission concluded that:

- diversity should be encouraged
- private incentives should be built on as much as possible
- research should be monitored and evaluated
- 'contestability' should have a major role in research funding
- a government role in sponsoring R&D should be clear and its requirements clearly articulated.

This approach to research and development echoes government policy relating to broader aspects of service delivery in government. These include:

- the identification of the public and private benefit in the provision of service, and the establishment of a framework for cost-sharing whereby the user pays or beneficiary compensates for the provision of the private-good component of the service;
- the separation of policy development from program delivery together with the devolution of resource management and the corporatisation/ commercialisation of program delivery, to allow government and community to obtain the benefits of competition and stakeholder involvement, and to provide improved focus on needs in the provision of services;
- a focus on evaluation of outcomes and performance in program delivery of agencies rather than on inputs and role and functions.

These trends have considerable implications for how the geoscience information needs of society are to be met and for the role and performance requirements of provider agencies.

Societal needs for geoscience information in Australia

Geoscience and sustainable development

Environmentally sound renewable and non-renewable resource

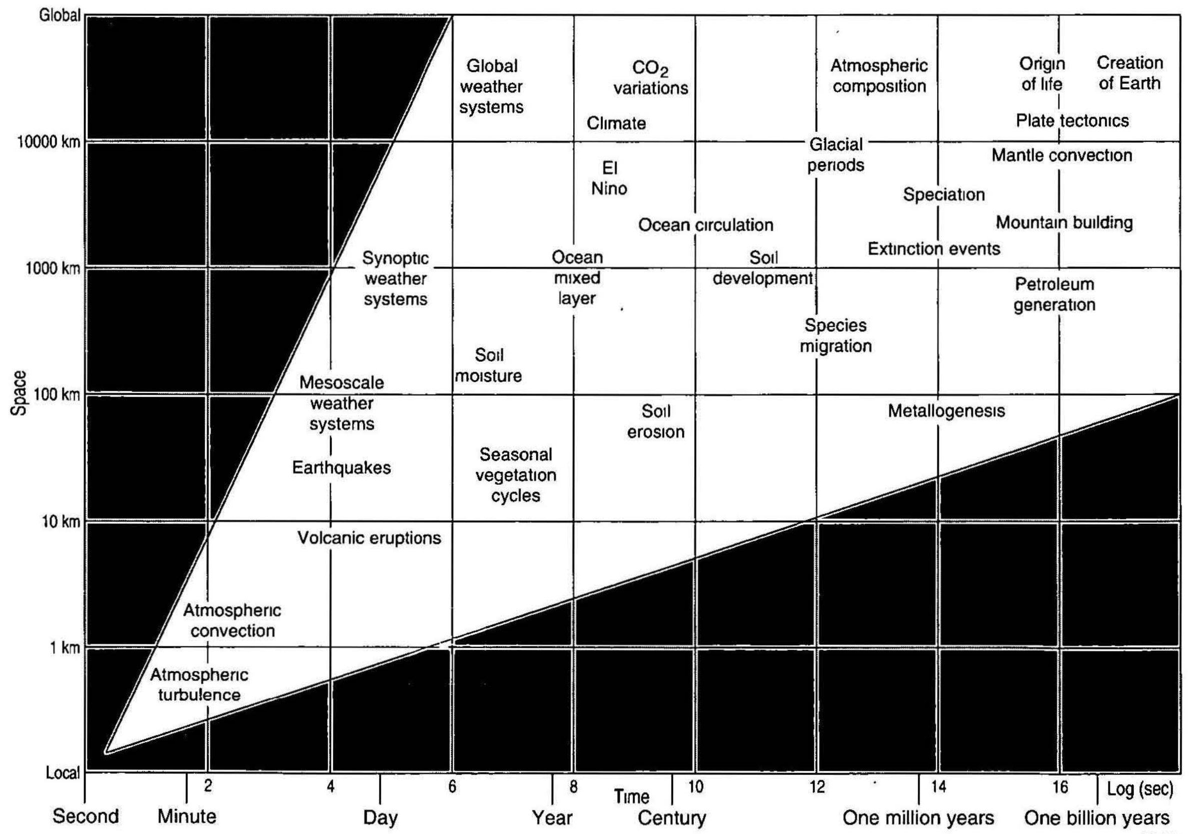


Figure 1. Characteristic space and time scales of Earth system processes. After NASA (1986).

development through sustainable development principles has been firmly embraced by the Australian Government. Under sustainable development, mankind's future on Earth depends on attaining a level of development commensurate with human aspirations whilst achieving a level of balance in the natural world — constancy of natural capital stock (Pearce et al. 1990) — which is not prejudicial to the needs of future generations.

The scientific basis for sustainable development therefore concerns the scientific knowledge concepts and hypotheses that provide a rational basis for planning mankind's future in a sustainable development framework. This science addresses the relationship between natural systems and processes and anthropogenic systems and processes. The critical aspects are those of rates and impacts; the former relate to, for example, the rate of exploitation versus the rate of renewal and the latter to the impact of mankind's activities on the natural world through the disturbance of the environment (Williams 1996).

The geoscience of sustainable development therefore concerns the science of the Earth systems that produce the resources to support development, the processes that affect human activities (e.g. biochemical processes), or that are accelerated by human activities (e.g. erosion which affects substrate control on biodiversity or results in soil loss) and that impact on mankind and its infrastructure (e.g. earthquakes). The time scales of these Earth systems and processes (Fig. 1) determine the nature of the geoscience required to support particular aspects of sustainable development (Williams 1996).

Before further considering societal needs it is worth considering the nature of geoscience information. 'Geoscience information and expertise are a special kind of national resource for any nation on Earth. Unlike many other kinds of scientific information and expertise, geoscience information and expertise have both local and universal significance. They pertain to a specific place in a specific country, as well as to the global corpus of scientific knowledge. They are part of the knowledge base concerning the nature and present state of the country' (Price 1992 p. 98). Irrespective of the relative priority of any other research, Australia has the primary responsibility for research into its own natural environment and its endowment of resources if it is to responsibly manage these resources in a sustainable manner.

The primary tool for the dissemination of geoscience information is a geological map and its associated databases of specialist geoscience information. At a regional level the information in a geological map has relevance to a wide range of societal needs. Geological maps contain descriptive information about the solid Earth. A geological map commonly identifies the spatial distribution of bedrock, weathered surface and transported materials and geological structures such as folds and faults. Importantly, they also provide an interpretation of how these materials and structures are related in time and space. The process of building a map is a basic field research activity involving a variety of techniques from the most mundane to the most sophisticated technology both in field and laboratory (Price 1992). A basic geological map provides background information for detailed investigations (Fig. 2 and Table 1).

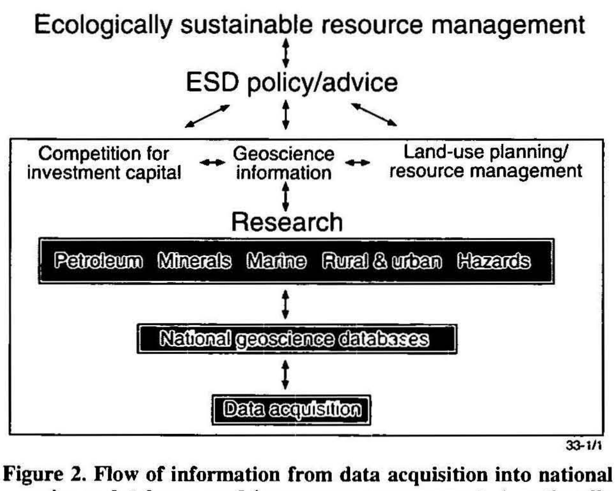


Figure 2. Flow of information from data acquisition into national geoscience databases used in resource management. A nationally accessible collection of coordinated and standardised geoscience databases would more effectively implement ecologically sustainable development (after Loutit 1995).

> The regional geologic map is universally recognised as the instrument of choice for planning and executing research and decisions that involve earth science information ... Its utility and value derive from the fact that the unique information content of a geologic map can be used to characterise the geologic setting of a specific site in the context of the surrounding region. Scientists, decision makers and managers can extrapolate the results of site specific investigations outward to adjacent sites or regions where investigations have not been conducted and thereby forecast or predict geologic conditions where data are limited. The regional geologic map forms a fundamental database for earth science applications that require a predictive capability (geohazards evaluation, resource assessment, environmental analysis (Bernknopf et al. 1993 p. 13).

However, published geological maps represent concepts, ideas and data available at the time of preparation and can become increasingly out-of-date as they age (Price 1992). First generation geological maps have been estimated to have a useful life of about 10 years before the need for revision as a result of changing ideas and the availability of new data. The rate of obsolescence varies enormously and the need for revision will vary from one area to another, depending on the complexity of the geology and the priority for up-to-date information and its application.

The collection, presentation and dissemination of geoscience information is therefore a long-term strategic issue for a nation (Price 1992). It took 30 years to complete reconnaissance geological maps of Australia involving Commonwealth, State and Territory geological surveys. It took 6 years to compile a complete hydrogeological map series for the Murray Darling Basin with the cooperation of State and Commonwealth agencies. This followed many years of local and regional geological and hydrogeological investigations by the parties.

The specific applications of geoscience information to different fields are presented below.

Geoscience information for mineral and petroleum resources

For much of this century the main role of geological surveys in Australia has been the basic mapping of the continent, and later its offshore areas, to provide fundamental knowledge for resource exploration and development. This was given impetus by the formation of the Bureau of Mineral Resources immediately post war, when the need for basic geological knowledge was recognised as a prerequisite for efficient resource exploration on a national scale (Williamson 1996). Basic geological mapping has been fundamental to the discovery of the mineral and fossil fuel wealth that underpins Australia's modern economic performance. Mineral and energy exports were estimated to have a value of more than $36 billion in 1996–97 (Waring 1997).

Following the completion of reconnaissance scale mapping, and given the success of mineral and petroleum exploration and development, is there a need for further mapping to support the exploration industry?

Minerals and petroleum are not renewable commodities on the human timescale — the rate of Earth processes is too slow.

Table 1. Societal needs, policy agenda and clients for geoscience information in Australia.

Geoscience information requirement	*Policy relationship*	*Institutional clientele*		*Industry and public clientele*
		National	*International*	
Minerals & petroleum	International competitiveness in exploration Regional development Land use	Energy and mineral policy departments (Commonwealth & State) Environment policy department (Commonwealth) CSIRO Cooperative Research Centres Universities	Research institutions Universities	Mineral exploration companies Petroleum exploration companies Community groups
Infrastructure development	Transport Regional development Urban development	State development agencies Regional & local planning boards & councils		Community groups and construction and waste management industries
Geohazards — earthquakes and landslides — nuclear monitoring	Risk mitigation Emergency management Comprehensive test ban treaty	Standards Australia Australian Dam Safety Committee Pipeline Authority State Emergency Services Australian Emergency Management Institute Planning authorities Universities Foreign Affairs Department	International Seismological Centre (UK) Information Centre, USA	Insurance industry Engineering companies Public and media
Groundwater and land degradation Crop productivity	Regional development Resource management Environmental degradation	Rural, water, forestry and environmental policy departments (Commonwealth and State) Murray–Darling Basin Commission Aboriginal and Torres Strait Islander Commission CSIRO Cooperative Research Centres (Rural) Research and Development Corporations	Research institutions Universities	Rural community groups
Earth's magnetic field		Australian Land Information Group Department of Defence Ionospheric Prediction Service Civil Aviation Authority	World data centres Research institutions	Mineral & petroleum exploration companies Electronics industry Surveyors Mariners
Coastal areas & marine zone	Regional development Resource management Environmental degradation Conservation	Petroleum & fisheries policy State conservation and land management agencies Local and regional planning authorities Environment policy department (Commonwealth) State departments of resource industries	Universities Research institutions	Community groups Tourist areas

Geoscience information requirement	*Policy relationship*	*Institutional clientele*		*Industry and public clientele*
		National	*International*	
		CSIRO		
		Universities		
National sovereignty	Australian jurisdictions	Foreign Affairs Department	Research institutions	
		Attorney General's Department	Universities	
		Antarctic Division, Environment Department		
		Australian Land Information Group		
		Universities		

The basis of a sustainable resources industry lies in replacing exploited deposits with newly discovered resources through exploration.

Mineral and petroleum exploration is a high-risk investment and the exploration industry seeks geoscience data information and knowledge to reduce that risk through more effective and efficient exploration. Successful discovery of ore bodies (Herriman 1989) requires knowledge of what to look for and where to look, access to the area of interest, and appropriate technology to make the discovery.

In Australia, geological surveys have played and continue to play an important role in exploration in determining what to look for and where to look. The importance of a knowledge base in successful mineral exploration is acknowledged by Woodall (1983 p. 41) who stated that 'ability to conceive useful conceptual models during the generative stage of exploration is limited only by the density of available, reliable geological, geochemical and geophysical data and our understanding of the physics and chemistry of earth processes ... The source of much of the data is ... the valuable national geoscience database assembled by the government geological surveys and universities'.

Perceptions of prospectivity held by mineral and petroleum exploration companies therefore depend on available geological knowledge, technical capability, geological ingenuity and climate of opinion. Mineral and petroleum potential assessment is very much in the eye of the beholder and forms the basis of competitive advantage in mineral and petroleum exploration — the company that makes the correct assessment first is most likely to discover the new deposit (Large 1992). All prospectivity assessments, however, depend fundamentally on geoscience data, concepts and knowledge which provide the framework for successful exploration and mineral and petroleum discovery. The absence of this framework reduces perceptions of prospectivity and increases the risk in exploration. A systematic program of geoscience mapping reduces risk and stimulates exploration investment by identifying new opportunities (Williams & Huleatt 1996). If a country wishes to actively develop and maintain its competitiveness for exploration investment, relevant and up-to-date information on prospectivity is essential (e.g. Lowder 1994, Offshore Strategy 1990). Increasingly sophisticated exploration techniques and concepts are required to discover deposits under surficial cover or in complex geological environments, and the information base and geological framework for mineral and petroleum exploration need to be progressively updated and refined.

National and State governments provide the geoscience infrastructure as an incentive to industry to invest in exploration for resources owned by the Australian people. Today, when capital moves readily from one country to another, Australia competes with other prospective countries for exploration capital. Many countries, having recognised the potential returns from resource development, provide attractive investment conditions through mining legislation and taxation incentives. Before this stage, investment by State and national governments in developing and maintaining the national geological framework is a powerful incentive for exploration companies to continue to consider exploration opportunities in Australia, and is essential in maintaining international competitiveness in the mineral and petroleum industry (e.g. Offshore Strategy 1990, Lowder 1994).

The primary targets for government geoscience information are the national and international mineral and petroleum exploration companies (Table 1). Government agencies and ministers responsible for this industry sector are stakeholders, for geoscience survey work represents part of the government input to maintaining a competitive mineral and petroleum exploration industry. Other users are scientific and technical groups in consulting firms, and academic or government research institutions with an interest in petroleum and mineral geoscience.

Information useful for those assessing opportunities for resource exploration can increasingly be used by governments as a basis for resource assessment (Fig. 2). Historically, resource assessments have been concerned with issues of self sufficiency and security of supply of key raw materials. This was the case particularly with petroleum, after the oil price shocks of the 1970s and early 1980s. Increasingly, emphasis in resource assessment has switched to decision-making on land use (e.g. Shoalwater Bay Military Training Area Resource Assessment 1993). As in all resource assessments the key requirement is geoscience information to allow sensible assessment of the likelihood of resource occurrence. This requirement is, however, somewhat different from traditional resource assessment. Areas nominated for resource assessment often arise out of consideration of conservation values relating to living resources, and have not necessarily been in the mainstream of thinking about resource potential until then. Resource assessment is frequently required in communities where opinions are strongly divided. In addition, nominated areas with resource potential might lack information crucial to a systematic assessment. The uncertainty which accompanies resource assessment is thus compounded. One way of coping with this is to identify areas of potential conflict in land use and undertake basic geoscience investigations before an actual conflict arises.

In addition to the clients listed above, government policy and decision-makers and community or sectoral interests are concerned with land use issues (Table 1).

The geoscience of the sustainable development of minerals and petroleum is the knowledge that relates to prospectivity and resource assessment on a geographic basis. Priorities can be established on a regional basis considering the state of development or knowledge, immediacy of the need (client driven) and application within the expected useful life of information generated.

Geoscience information and urban infrastructure development

Rapid urban development around Australia's coastal fringe, consolidation and infill of urban areas, development of residential shorelines and increased consciousness of land-use issues have all contributed to a growing need for geoscience information. This information is needed in assessing our vulnerability to natural hazards and in mitigating their impact; development of the national infrastructure; the construction and siting of major civil engineering structures; waste disposal; supplies of industrial minerals (particularly aggregate and sand) for construction; and management of groundwater to avoid contamination. The basis for consideration of these issues is the knowledge incorporated in regional geological maps.

In the case of waste repositories, geological information is required to assess hazards that may affect the integrity of the repository. These include earthquake faults, slope failures, ground subsidence, transmissivity potential and stability of underlying strata. Engineering requirements and extent of buffer zone have to be determined. Collocated geological resources, such as clay-sand-gravel resources and groundwater, need to be identified. Development of the resources might be either compromised or exploited in the context of the installation.

In any region the geological framework affects choices for transport corridors. Construction and maintenance costs are influenced by local variations in near-surface geology and topography, which also affect slope stability, either natural or created, along a corridor.

Natural hazards are rare, low-probability events that are largely unstoppable. A significant part of damage costs resulting from natural disasters has to be borne by governments. In Australia, insurance met only 38% of the estimated $178 million total damage costs resulting from the 1974 Brisbane floods. As Australia becomes progressively urbanised and continent-wide infrastructure is developed, the risk of damage from natural hazards also progressively increases. Although these hazards are not uniquely geological, there is a significant component of geological risk. There is a particular need to integrate information on geological hazards (e.g. earthquakes, landslides and tsunamis) with other forms of hazard (e.g. cyclones, severe storms, floods and bushfires) on a regional basis, to create vulnerability maps. These maps are required to provide a framework for emergency planning and risk mitigation.

Although Australia is seismically quieter than some parts of the globe, there is a significant seismic risk in Australia, as evidenced most recently by the Newcastle earthquake. Earthquake risk needs to be assessed on a continental scale and this can be used to develop building codes which, if adopted, would mitigate damage in earthquakes. Australia's position in the southern hemisphere means that the continent-wide earthquake monitoring network plays a crucial role in international seismicity records.

Assessment of landslide hazard risks can be made by evaluating geological materials for their inherent risk of failure. Such an assessment identifies the geological terranes with landslides in the recent geological past, and establishes the relationships between landslide-prone terranes and other geological structures (e.g. faults and folds) that may influence landslide occurrence.

Although not in the same category as the other geological hazards, the Earth's magnetic field is constantly changing. A national observatory network provides information on morphology and variations in magnetic field and allows the provision of numerical models of charts of the field and its secular variation over the Australian region (including offshore areas of interest to Australia). These are used for navigation and direction finding, reduction of aeromagnetic and marine magnetic survey data, levelling and correcting survey data, and delineating and modelling long-wavelength magnetic anomalies. As with earthquakes, its position in the southern hemisphere and its continent-wide magnetic observatory network give Australia a crucial role in the international record of geomagnetic variation and the establishment of the global geomagnetic reference field.

State planning authorities, emergency management agencies, industry authorities, planning boards, local council and to a lesser degree community groups use geoscience information relating to urban infrastructure and natural hazards. There is also a significant body of international users of data on the Earth's magnetic field.

Geoscience information and environment, and resource management

The continuing growth in national and global human populations increases the demand on energy, mineral, land and water resources. This growth is accompanied by increased food and water pollution and degradation in the overall environment. It is increasingly recognised that the natural environment and related biological habitats on the Earth are primarily determined by the interaction of geology and climate.

In Australia, groundwater is of particular importance for water supply. Sixty per cent of Australia depends totally on groundwater and an additional 20% is partly dependent on it. The trend in water and natural resource management at both State and Commonwealth levels is for community-generated and community-based management, facilitated through Landcare and Total Catchment Management. The success of community-based management depends on providing technical and scientific advice at all levels. Total Catchment Management also relies on all participants having fundamental knowledge of the key water/land interaction processes.

Increasingly, the water resource/land management process is moving towards the adaptive management cycle, which includes planning, implementation, monitoring, evaluation and reporting. In the Murray–Darling Basin groundwater project conducted by AGSO, issues relevant to total catchment management are:

- the fundamental science associated with physical processes;
- data gathering and interpretation aimed at defining the boundaries of the physical system.
- monitoring of baseline conditions and resource performance;
- information and knowledge transfer to the natural resource managers at both government and community levels.

It is significant that these phases of investigation were preceded by an extensive period of mapping, database development and investigation of the fundamental geology of the basin upon which the more specialist resource focused studies are based. It is a characteristic feature of Australian groundwater systems are characteristically very large (e.g. Murray Basin, Great Artesian Basin) and must be understood as totalities if local management is to succeed.

Groundwater quality (i.e. its acceptability in domestic, industrial, agricultural or environmental terms) is determined by both human and non-human activities, and can be assessed not only by factors such as salinity, but also by nutrient, toxic chemical and microbiological loads. Knowledge of the present status of groundwater quality and a clear understanding of the biogeochemical processes which determine it rely on geoscience. Our ability to predict the impact of management actions on the quality of groundwater resources depends heavily on knowledge of the interactions between water and earth materials and requires expert understanding of biogeochemical factors.

Groundwater resources underlying agricultural areas are commonly exploited for domestic and town water supplies as well as for irrigation; in salinised areas surplus groundwater is often pumped into adjacent surface waters for disposal. These activities can have far-reaching implications for health, the environment, and economic development, because of the risks of contamination from pesticides, herbicides and salt. Measurement of contaminant loads is required to a high degree of precision for all groundwater systems underlying high population areas or areas of intensive agriculture. This is essential in maintaining the quality of Australia's major groundwater resources, and in ensuring that baseline data for environmental protection and legal action purposes are established.

Soil acidification in crop-pasture systems, soil erosion and salinisation of agricultural and pastoral land (following loss of vegetation from practices such as overgrazing and till clearing) are widely recognised as Australia's paramount land degradation problems. Technologically the most difficult aspect of addressing land degradation is the absence of methods to measure rates of land degradation over large areas and short time periods. Development of efficient repeatable monitoring methods is essential in measuring the effectiveness of management strategies. Knowledge of the underlying geology, the interaction of groundwater with surficial processes and the weathered zone (regolith) and history of the landscape are critical in understanding many degradation processes. In addition, rapid geoscience survey techniques, such as aircraft (airborne radiometric or electromagnetic surveys) or spacecraft remote sensing, are required in the rapid mapping of land degradation. Land degradation is the outcome of complex interrelationships of land, water and vegetation, and a basic understanding of the underlying geology, groundwater and weathering patterns is crucial to the introduction of management strategies and risk assessment. Assessment of the potential for land degradation can be a significant tool for long-term policy development and management.

Recent work (Laffan 1995, Thwaites 1995) shows that forest productivity is related to the underlying geology as well as to the traditional controls of climate and aspect. The geoscience required to clearly establish the precise relationship requires clarification, but it includes in addition to the traditional bedrock mapping, geomorphological mapping and assessment of regolith development (weathered zone) and nutrient availability and leaching.

Just as land degradation and regolith characteristics affect the cultivation of crops and the productivity of forests, there is an emerging recognition of the relationship between the underlying geology, geomorphology and surficial geology on the distribution of the biota in both land and marine environments. Surficial sediments, the depth of weathering and land forms help determine habitats for flora and hence indirectly fauna (Cape York Land Use Study, unpublished results; E. Truswell, AGSO, pers. comm. September 1995). Similarly off shore the nature and topography of the sea floor profoundly affect the distribution of the biota (Edgvane & Burne 1995, Koslow & Exon 1995). The ramifications for the systematic mapping of habitats and the management of park areas both onshore and offshore are only just being recognised.

Environmental geochemistry is an emerging science that has yet to be widely applied in Australia. It is concerned with the sources, dispersion and distribution of elements in the surface environment, their chemical speciation and pathways into water, agricultural crops and animals and their effects on the health of plants, animals and humans. It deals with both natural variation and the effects of pollution. It is not widely recognised that naturally occurring concentrations of elements can adversely affect environmental quality and trace element concentrations in foodstuffs, and that depletion in certain elements due to natural variation can adversely affect cropping yields or cause clinical or subclinical trace element deficiencies in animals. Systematic geochemical surveys are now being used to study these effects in the United Kingdom (McGrath & Loveland 1992). In Australia relatively little of this work has been done. One example is the possible association of naturally occurring anomalous concentrations of uranium and thorium and other harmful elements in the southern Kakadu region (Wyborn & Needham 1990).

The coastal strip between Gladstone and Adelaide has Australia's highest population accumulation (75% of Australia's population lives within 50 km of the coast) and is an area subjected to high developmental pressure from industrial, residential and recreational demands affecting land forms, scenery, water bodies and vegetation. Major environmental problems of this zone are physical destruction, chemical pollution, overharvesting of resources, loss of habitat and introduction of alien species, onshore and offshore mining and possible consequences of global climate change (Resource Assessment Commission 1993). Increasingly the coastal zone will come under pressure from pollution (e.g. sewage and industrial wastes) and resource usage (e.g. nearshore mining of industrial materials — sand, gravel aggregate) to support onshore development in areas of high population density.

To address these issues in managing the coastal zone, we need fundamental knowledge of the underlying geoscience and associated processes and baseline conditions. The new discipline of coastal geoscience applies appropriate areas of engineering geology, geomorphology, sedimentology, marine geology, geochemistry, biogeology and remote sensing. In policy development, environmental protection, management strategy formulation and development projects, we need sound geoscientific data in formats acceptable to analysis using geographic information systems.

Human activities are affecting the climate through enhancement of the greenhouse effect. This is superimposed on the natural variation in climate, the implications of which are not fully understood either on a global or regional scale.

The modelling of future climate trends depends on the accurate representations of global and regional climates in the models and on a thorough understanding of natural variations in climate. If models are to predict future climate trends with any confidence, their ability to simulate present climates and those of the recent past must be demonstrated. The palaeoenvironmental record is the basis for understanding past climatic variability and can provide scenarios for reverse modelling to test the predictive capability of models. Elucidation of this record requires palaeoenvironmental analysis by biostratigraphy and age dating, analysis of marine and terrestrial cores, and ice cores.

Geoscience information relating to the environment is used by diverse disciplines and, because of the multidisciplinary nature of environmental issues, by other environmental scientists. In turn, the users of environmental science information are government policy and decision makers, regulators and community or sectoral industry interests.

The geoscience of sustainable development relates to risk assessment, management to mitigate impact, and reclamation. It comprises for example:

- assessment of the nature and extent of preconditions for land and water degradation through analysis of geological history and the technology and concepts for rapid assessment
- understanding the land degradation and coastal processes,

rates of these processes, and their translation into predictive models
- understanding the effects of human inputs to the system, e.g. agricultural and industrial pollutants and their impact in time and space
- development of decision support systems to assist in the management of the rural and coastal environment.

Similarly the geoscience of climate change is the development of understanding of the history of climate change, the Earth processes that affect it, and the provision of technical data and decision support systems focused on improving our abilities to predict its effects.

Geoscience information and sovereignty issues

The declaration of the United Nations Convention on the Law of the Sea and its ratification by Australia in 1994 mean that there is a particular need for geoscience information to establish the Australian claim for extending its offshore jurisdiction. Under the convention, countries have 10 years from the date of their ratification to define the outer limits of their legal continental shelf where it extends beyond the 200 nautical mile exclusive economic zone (EEZ) (Fig. 3). The latter concept involves natural prolongations of the continental margin which may extend beyond the EEZ. Such extensions of the margin involve defining sediment thickness and base of the continental slope morphology. Good systematic, bathymetric and sediment thickness data are essential to these determinations which will underpin the treaty negotiations undertaken by Australia.

An emerging issue is the development of a science base to manage Australia's offshore jurisdiction (Ocean Outlook 1994). In many ways our knowledge of Australia's offshore jurisdiction is similar to that of the land area a century ago and the mapping requirements reflect this state of knowledge. The geological substrate of the ocean is a dynamic environment of physical and chemical interactions with the overlying water column and its biota. Although many of the Earth processes that contribute to the establishment of the substrate fall outside of the human timescale (Fig. 1) the substrate can be extensively modified and its dynamic equilibrium with the biosphere and the ocean can be drastically affected by human activities.

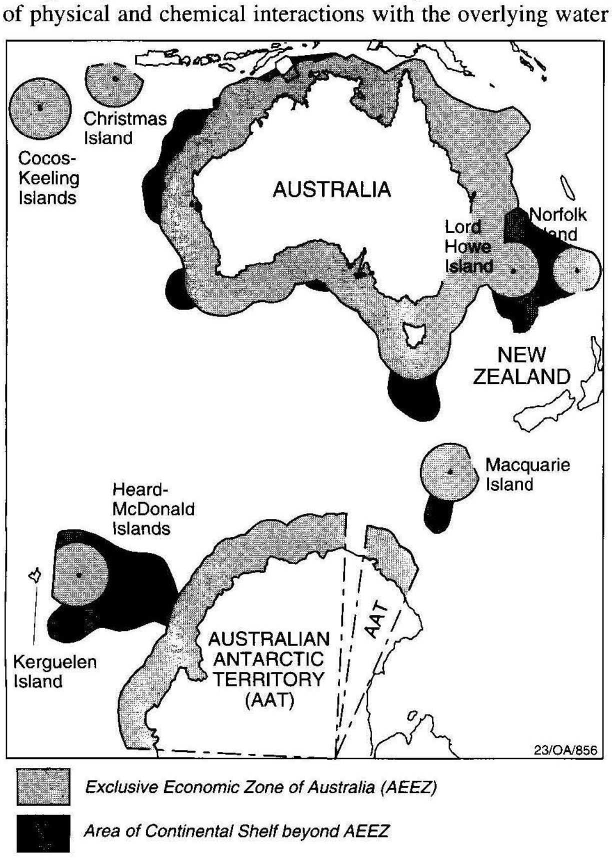


Figure 3. Australia's potential offshore jurisdiction.

The geoscience of sustainable development of the marine zone includes the definition of its full extent, and of the nature of the geological and geochemical process that affect the distribution of living and non-living resources, their exploitation and management. It includes:
- mapping of the extent and nature of the marine zone, including technology and concepts for rapid assessment.
- understanding the geological history that led to the substrate development and the application of this understanding to prediction of the substrate and characteristics of poorly known areas
- understanding the biogeochemical cycles active at the sea floor that affect the flux of carbon and nutrients in the sea
- understanding how the substrate controls biota and living resources
- understanding the effects of human inputs to the system, e.g. agricultural, sewage and industrial pollutants and their impacts in time and space
- the development of predictive models and decision support systems to help manage the marine zone, its biodiversity and its resources.

The users of marine zone geoscience are very similar to those identified for environmental geoscience.

Science plays an important part of Australian policy on Antarctica. The scientific research program of the Australian National Antarctic Research Expeditions is a tangible expression of Australia's commitment to an active role in Antarctic affairs and the maintenance of Antarctic treaties. A key component of this research is the geoscience component whereby Australia systematically maps and develops an understanding of the geology of the AAT. In this context, those using geoscience information are those with administrative and management responsibility — the Antarctic Division of the Department of Environment, Sports and Territories and other scientific and community groups with an interest in Antarctica.

Information requirements

The wide variety of applications of basic geological information means that geoscience information must be managed and made available in forms that facilitate its application for the wide variety of potential users (Fig. 2). The traditional way of presenting geological information has been in a published geological map and accompanying explanatory notes. Often a synthesis of the geology of an area incorporating a number of geological maps would be prepared and accompanying specialist information would be tabulated or perhaps presented in specialist maps. In these cases the user is dependent upon the interpretation of basic data by the author of the map, and much of the basic data upon which the map is based is not readily accessible.

The rapid developments in information technology and the availability of global positioning systems are dramatically changing the way geoscience information is presented and analysed. Modern maps are presented in digital form and use information from supporting disciplines, such as geophysics, geochemistry and geochronology, incorporated in accompanying databases or images. This has created a demand for topologically structured data which can be used in geographic information systems (Fig. 4). In modern information systems geophysical data, interpreted geology and point data stored in layers can be displayed and modelled through a GIS to provide analyses unique to the particular client. The key feature is that all aspects of the basic data are available to the potential user. If the geoscience information is properly referenced, the user is less dependent on the interpretations of an individual mapper. Furthermore, users can add their own information to the database and analyse this concurrently using the mapping package or apply modelling processes specific to the particular application.

This new approach allows the integration of disparate datasets and data types cost effectively, allowing information to be used in a variety of ways (Figs 2 & 4). Relational database management systems, image processing techniques and computer-aided map production can be routinely applied and the products incorporated into a geographic information system. The introduction of this modern technology has placed a demand for the capture of all publicly available geoscience information, whatever its source, and adequate descriptions of its quality and specifications or metadata available to enable its use. Such metadata needs to incorporate a description of the data source, and techniques involved in generation so that its suitability for a particular application can be readily determined.

There is an existing capital stock of information which can be exploited, so acquisition of new information needs to be carefully targeted to fill the information gap rather than collected for its own sake, as was required during the first reconnaissance phase of mapping of the continent.

Implications for performance requirements of AGSO

The geoscience requirements outlined above are not exclusively met by the national survey. Priorities are determined by the decisions of government and the needs of clients (Fig. 5). Nonetheless there has been a significant diversification in those using geoscience information and a significant change in the operating environment which must be met by any modern geological survey.

The traditional role of a geological survey has been as the main supplier of original spatially referenced geoscience data and information for geoscientists, who can search out the information relevant to them through a traditional academic method. Time frames for the supply of the data and information have in general been determined by the program of the supplier

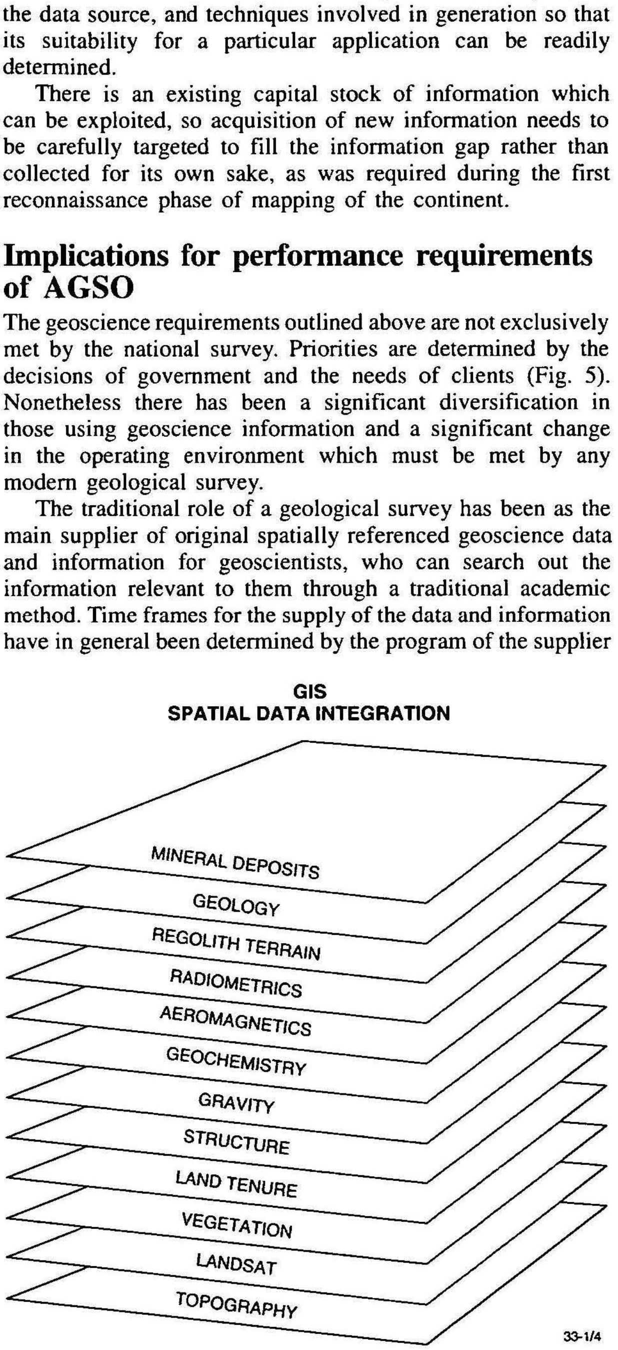


Figure 4. A geographic information system can integrate multiple datasets representing the geology of an area.

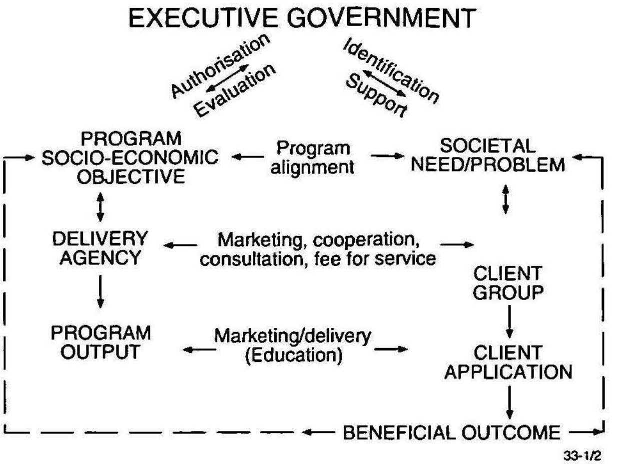


Figure 5. Schematic representation of relationships for a client-driven government program represented by a geological survey.

— the geological survey. The modern situation is quite different, with a diverse clientele (Table 1), each client having specific needs. Users have varying degrees of geological expertise and the provision of geoscience information as such might be insufficient or unusable without considerable assistance. Even in the traditional client areas of mineral and petroleum exploration the requirements have changed with the increased pace and commercial edge given by an increasingly competitive market both nationally and internationally and low commodity prices. The modern requirement for geoscience information is for targeted and relevant information packages backed with basic data and made available at an appropriate stage of the exploration cycle.

Increasingly, program and project managers must take responsibility for effective interaction with clients to ensure the scientific output can be effectively used. The model for a client driven organisation requires a clear understanding of the clients' needs and an approach that is committed to meeting these needs according to the clients' requirements for timeliness, cost and quality. (Fig. 5). The need to focus on the client use of geoscience information has been accompanied by a more rigorous accountability structure in government (Fig. 5). Funds are authorised to meet particular societal needs through approved programs. In the case of geoscience programs, broad socioeconomic objectives are set, whilst the precise technical requirements are determined between the delivery agency and the client. Program performance and outcomes are measured through evaluation. The challenge for a delivery agency operating in this environment is to maintain alignment of the program with the objectives and to ensure that the clients' needs are being met in a way consistent with the government's authorisation.

Different sorts of users of geoscience information need information in different forms. Characteristics of the different fields using geoscience information are:

Policy requirements:

- Technical advice needs to be expressed in a readily understood form based on demonstrated competence and comprehensive knowledge.
- Timeframes are often short: ability to respond quickly depends on ready access to all relevant existing information and the capacity to analyse it.
- Geoscience input — ability to market its relevance and to communicate to non-technical stakeholders is essential.

Resource and environmental management and risk mitigation:

- Technical information and data need to be focused on management outcomes underpinned by a sound theoretical base, demonstrated competence and all relevant data.
- Timeframes are variable, but generally allow time to undertake relevant original work.
- Relevance of geoscience input is not always understood by managers; standing with technical peers is essential; strong ability to communicate and translate information into pragmatic applications for stakeholders with differing technical knowledge (e.g. community and catchment groups) is needed.

Petroleum and mineral exploration:

- Concepts and basic data must be focused on prospectivity and underpinned by a strong theoretical base, acknowledged competence and relevant data. The petroleum industry requirements of geological surveys tend to be more concept-based and acknowledging of innovation. The mineral industry requirements have tended to be more data based and less interested in concepts, although this appears to be slowly changing.
- Timeframes are variable, but delivery needs to be in phase with the exploration cycle.
- Ability to communicate with technical peers is essential with a tight focus on exploration implications.

The increasing emphasis on clients and their information requirements, and the pace of change in the operating environment mean that geological surveys now need a different *modus operandi*. Projects need to be targeted to be sufficient for the requirement, but have the capacity to meet needs in new and different ways and to create new products for the client. In targeted projects, the shorter time requirements of clients are to be met through multidisciplinary teams in which tools, specialists and facilities are used to answer specific questions. In other words, the approach must be strategic, adding value to existing knowledge with optimum and synergistic use of technology and skills. The approach will be less on obtaining comprehensive understanding of the geology of an area, but rather on how existing data and information can be adapted and supplemented by new work in a targeted way to meet particular requirements. Organisations need to identify and nurture these skills and the technology important to their future capability, and to systematically withdraw from the skills and technology that are not central to their business mission. Where skills do not exist in-house, they need to be provided through partnership arrangements.

These requirements are similar to the characteristics of many service provider organisations, but a unique aspect of the activities of geological surveys is the need to manage geoscience information at a national scale. A consistent message from the analysis of societal and client needs is that management of information is central to the future. 'Modern geological surveys are in the geoscience information business' (Price 1992 p. 99). The technology is now available to exploit the data collected before the development of modern information technology and to capitalise quickly and effectively on new information as it is collected and assimilated. This will require the development of complete databases of all relevant data with ready access and facility to exchange information according to national standards.

Of equal importance is the requirement to create a software environment of decision support systems based on geoscience principles to make effective use of the large amount of data and to facilitate the provision of geoscience input to meeting client needs. It will be equally important that clients have access to the software environment in an interactive mode and that we can provide the associated technological support if the information resource is to be truly available to meet their needs. Herein lies the future core of geological survey work for which geological surveys are uniquely placed in terms of the long-term resource commitments and knowledge required to meet this challenge.

Conclusions

The need for geoscience information from geological surveys has dramatically expanded beyond the traditional role associated with national development of the minerals and petroleum industries. New clients for this information are the stakeholders in ecologically sustainable development. Their information needs relate to policy and management outcomes rather than to technical reports and papers, although a strong demonstrated technical and theoretical base is a prerequisite to gaining acceptance by these clients. At the same time the traditional client base in the petroleum and mineral industries has been changing. In both cases the result is a need for stronger client interaction, marketing and identification with clients' aspirations if a geological survey is to stay at the forefront of the geoscience information business.

Such an agency needs to be flexible and responsive to a rapidly changing environment. Data acquisition, which has been the traditional strength of a geological survey, will diminish in relative importance as data become more readily available from disparate sources in digital form. To assist in this process surveys must develop new skills in geoscience information management, and possess the technology and key skills of geological survey work. If this can be achieved, then a national geological survey will prosper in an increasingly competitive world.

Acknowledgments

This paper draws on background papers developed in AGSO, in particular those by T Loutit, L. Jaques, C. Simpson, D. Denham and C. Pigram, whose assistance is greatly appreciated.

References

Australian Bureau of Agricultural and Resource Economics (ABARE), 1995. Australian Commodities, 2(3), 279–416.

Bernknopf, R.L., Brookshire, D.S., Soller, D.R., McKee, J., Sutter, J.F., Mati, J.C. & Campbell, R.H., 1993. Societal value of geological maps, U.S. Geological Survey Circular IIII, 53 pp.

Cook, P.J., 1995. The role of the geological surveys in the 21st century. Episodes, 17, 106–110.

Eaton, G.P., 1995. Evolution of the global job market in the earth sciences. Address to the American Geophysical Union, Baltimore M.D. June 1 1995. http://www.usgs.gov/whats-new/global-jobs.html. Modification date 6-13-95 (MP).

Edgvane, K.D. & Burne, R.V., 1995. Coastal/marine ecosystem biogeography: implications for biodiversity studies and ecosystem management. Australia Marine Sciences Association Annual Scientific Conference, University Technological Sciences Sydney, July 3D6 1995, Program & Abstracts, 59.

Herriman, N., 1989. Successful ore discovery: the need for goals, good science and geological maps. BMR Research Symposium. Geoscience Mapping Towards the 21st Century. Bureau of Mineral Resources, Canberra.

Industry Commission, 1995. Research and development. Australian Government Publishing Service, Canberra, 893 pp.

Koslow, T. & Exon, N., 1995. Seamount discoveries prompt calls for exploration and conservation. Australian Fisheries, 54(2), 10–13.

Laffan, M., 1995. Assessment and classification of site productivity and land suitability for eucalypt plantations in Tasmania. ACLEP (Australian Collaborative Land Evaluation Program) Newsletter, June, 12–16.

Large, R., 1992. Mineral Symposium, ABARE National Agricultural and Resources Outlook Conference, Canberra.

Loutit, T., 1995. Onshore Sedimentary & Petroleum Geology. AGSO94 Yearbook of the Australian Geological Survey Organisation, 4–5.

Lowder, G., 1994. Discovery 2000 — a beginning. Minfo New South Wales Mining and Exploration Quarterly, 45, 1.

McGrath, S.P. & Loveland, P.J., 1992. The soil geochemical atlas of England and Wales. Blackie Academic and Professional, London and New York, 101 pp.

National Aeronautics and Space Administration, 1986. Earth System Science, Washington DC, 48 pp.

Ocean Outlook, 1994. Ocean outlook: a blueprint for the oceans. A report from the Congress 16–17 November 1994 and a Scientific Program Proposed by the Steering Committee.

Offshore Strategy. Promoting petroleum exploration offshore Australia, 1990. Australian Government Publishing Service, Canberra, 18 pp.

Pearce, D., Barbier, E. & Markandya, 1990. Sustainable development, economics and environment in the Third World. Earthscan Publications Ltd., London, 217 pp.

Price, R.A., 1992. National geological surveys: their present and future roles. Episodes, 15(2), 98–100.

Resource Assessment Commission, 1993. Coastal Zone Inquiry: Final Report. Australian Government Publishing Service, Canberra, 664 pp.

Shoalwater Bay Military Training Area Resource Assessment, 1993. Australian Bureau of Agricultural and Resource Economics, Australian Geological Survey Organisation, Bureau of Resource Sciences, Canberra, 290 pp.

Thwaites, R., 1995. An opportunity for quantitative land resource assessment in Queensland forestry. ACLEP (Australian Collaborative Land Evaluation Program) Newsletter, June, 2–5.

Waring, T., 1997. Australia and the global supply of minerals and energy — an overview. Proceedings National Agricultural and Resources Outlook Conference, Canberra 4–6 February 1997, Minerals and Energy, Volume 3. ABARE, Canberra, 3–20.

Williams, N., 1996, AGSO/BMR — 50 years of changing needs — what now? Abstracts, AGSO Jubilee Symposium Canberra 21 February, 1996. Australian Geological Survey Organisation, Record 1996/1, 19.

Williams, N. & Huleatt, M., 1996. The importance of regional geological mapping in minerals exploration. Proceedings National Agricultural and Resources Outlook Conference, Canberra 6–8 February 1996, Minerals and Energy, Volume 3. Australian Bureau of Agricultural and Resource Economics Canberra, 51–59.

Woodall, R., 1984. Success in mineral exploration: a matter of confidence. Geoscience Canada, 11(1), 41–46.

Wyborn, L.A.I. & Needham, R.S., 1990. Environmental geochemistry and the "sickness country" concept. Submission by the Bureau of Mineral Resources, Geology & Geophysics to the Resource Assessment Commission. Kakadu Conservation Zone Inquiry (unpublished).

AGSO Journal of Australian Geology & Geophysics, 17(1), 13–25

The sustainability of mineral use

R.W.R. Rutland[1]

The sustainability of mineral use, like the definition of sustainable consumption patterns, is a function both of the availability of resources and of the environmental impacts of resource use. The environmental impacts are a major factor in modifying consumption patterns and therefore in moderating demand.

In spite of a number of factors tending to moderate demand, global demand is likely to continue to increase for the foreseeable future, mainly as a result of the continuing increase in global population which is unlikely to be stabilised before the end of the 21st century.

Discussion of the relationship between population, resource use and the environment, using the concepts of the support square and per capita resource use, illustrates the unsustainability of present demand trends and of the consumption patterns that cause them. It is therefore important that strategies to maintain supply be linked to effective strategies to move to sustainable consumption patterns.

The capacity to meet the demand for minerals is also being reduced, as the global population increases, by concerns about environmental impacts and by competing land uses. It is therefore important that minerals issues should be part of the integrated approach to land management promoted in Agenda 21.

Sustainability is a long-term concept involving inter-generational equity, and the time scales are beyond those of the reasonably foreseeable future. It is not possible to see beyond a definable 'horizon of sustainability' and the 'precautionary principle' is therefore important for global management strategies. The paper identifies a number of steps at the global level which can assist in the management of resources and of environmental impacts during the period of transition to more sustainable consumption patterns. These relate to improvements in knowledge (1) for assessing the impacts of mineral use in the industrial system as an essential basis for improving efficiency and determining optimum consumption patterns; (2) for monitoring the chemical health of the global land surface; and (3) for assessing resource potential in ways that can be integrated with other land use information, and that can be used to push back the horizon of sustainability.

Introduction

The spectacular technological and economic development and consequent population growth of the last two centuries has been made possible by the expansion of the use of all kinds of natural resources, and, most fundamentally, by the use of petroleum and mineral resources. These resources are the foundation of our energy, manufacturing, communication and construction industries (Fig. 1), and of modern agricultural industry with its dependence on mechanisation, fertilisers and pesticides. Consumption of these resources continues to increase.

There has been concern about the depletion and possible exhaustion of these resources since the beginning of this century, a concern which reached its most forceful expression in the 'Limits to Growth' report for the Club of Rome (Meadows et al. 1972). While its estimates of impending scarcity were wide of the mark, the report was a landmark in the debate on sustainability.

However, in recent decades, concern that the availability of resources would set limits to national or global economic growth has been largely displaced by the more immediate concern that limits are being set by the environmental impacts of the increasing human population and its increasing consumption of resources. Land degradation and pollution of air and water are perhaps the most pressing problems but the issue which has captured the greatest attention from governments (and economists) has been that of global warming arising from greenhouse gases, and from CO_2 emissions in particular. The pressure to move away from fossil fuels towards renewable (or nuclear) energy sources is largely due to the evidence of environmental impacts rather than to recognition either of the clearly finite limits of petroleum supply, or of the desirability of conserving this valuable chemical or material commodity for purposes other than its calorific value.

These environmental issues are therefore having increasing impact on consumption patterns and on the likely future demand for resources. Moreover, the increasing concerns about the environmental impacts of mineral exploitation, extraction and use are also leading to economic or socio-political constraints on the capacity to meet demand. That capacity is also being reduced more generally by competing land uses as the global population increases.

Thus there is a close nexus between minerals and the environment, and minerals issues should be central to the planning for sustainable development, and specifically for the assessment of sustainable consumption patterns.

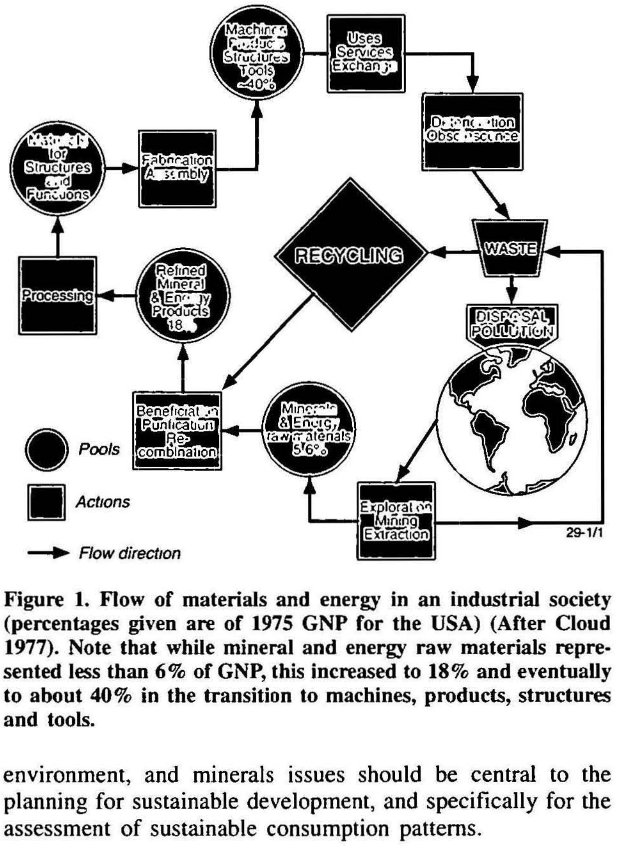


Figure 1. Flow of materials and energy in an industrial society (percentages given are of 1975 GNP for the USA) (After Cloud 1977). Note that while mineral and energy raw materials represented less than 6% of GNP, this increased to 18% and eventually to about 40% in the transition to machines, products, structures and tools.

Sustainability

The growth of world population and production, combined with unsustainable consumption patterns, places increasingly severe stress on the life-supporting capacities of our planet.

Agenda 21, Chap. 5, Demographic dynamics and sustainability. Report of the United Nations Conference on Environment and Development, Rio de Janiero, 3–14 June 1992. Vol. 1, Resolutions adopted by the Conference. (United Nations Publication Sales no. E.93.I.8 and corrigenda), resolution 1, annex II.

[1] Research School of Earth Sciences, Australian National University, Canberra ACT 0200, Australia

The concept of a sustainable society was encapsulated in the Brundtland Report *Our common future* (World Commission on Environment and Development [WCED] 1987) as one that 'meets the needs of the present without compromising the ability of future generations to meet their own needs'.

The report recognises that 'the concept of sustainable development does imply limits — not absolute limits but limits imposed by the present state of technology and social organisation on environmental resources, and by the ability of the biosphere to absorb the effects of human activities'. On the depletion of non-living resources, the report suggests that 'the rate of depletion should take into account the criticality of that resource, the availability of technologies for minimising depletion and the likelihood of substitutes being available. Thus land should not be degraded beyond reasonable recovery. With minerals and fossil fuels, the rate of depletion and the emphasis on recycling and economy of use should be calibrated to ensure that the resource does not run out before acceptable substitutes are available. Sustainable development requires that the rate of depletion of non-renewable resources should foreclose as few future options as possible' *(underlining added).*

Arising from the 1992 United Nations Conference on Environment and Development (UNCED), Agenda 21 was developed to address the pressing problems of today and to prepare the world for the challenges of the next century. The Commission on Sustainable Development (CSD) was established to promote the implementation of Agenda 21.

Agenda 21 (Chapter 4, para. 10 (e)) also called for identification of 'balanced patterns of consumption worldwide which the earth can support in the long term'. The concept of **sustainable consumption patterns** carries with it the concept of sustainable production patterns, and is usually considered in terms of the environmental impacts which provide the short-term constraint. It is recognised in Chapter 4, for example, that 'the major cause of the deterioration of the global environment is the unsustainable pattern of consumption and production, particularly in industrialised countries'.

The concept of sustainable consumption patterns must embrace the varying **levels** of resource use in different countries and the overall global level of resource use. A definition of sustainable development must therefore take into account both of the key minerals issues — the capacity of the environment to absorb the effects of resource use and the sustainability of supply of essentially non-renewable resources. It is necessary that the supply be sustainable for as long as required, and in the longer term, this may become the critical constraint on consumption patterns.

These key minerals issues are linked by the overriding issue of global population growth. Some aspects of the relationship between population, resource use and the environment are discussed below, before further discussion of environmental impacts and the availability of resources.

Population growth and increasing demand

Global population will almost certainly reach 8 billion over the next 30 to 40 years and is unlikely to stabilise much below 12 billion before the end of the 21st century (see, for example, Arizpe et al. 1992). This will place enormous additional burdens on the natural environment. In addition to the goal of an environmentally stable future, embodied in the principles noted above, the United Nations has the parallel goal of improving the living standards in less developed countries. To achieve this, the demand for materials and energy would increase, both from the rapidly increasing population and from the aspirations for improved living standards.

The Brundtland Report (WCED 1987), considered that global economic expansion by a factor of five to ten would be required in order to meet the demand for improved living standards for an increasing population. For this WCED apparently believed that such growth could be achieved largely by more efficient use of materials and energy and by improved technology to reduce environmental impacts (Arizpe et al. 1992 p. 69).

In this connection, there has been an attempt to distinguish between **economic growth**, which involves increased inputs of energy and materials, and **economic development**, which can take place through increased efficiency, without increased consumption of material capital, (e.g. Daly & Cobb 1989). Arizpe et al. (1992 p. 69), take the view that 'WCED is too optimistic — that a factor of 5–10 increase cannot come from development alone and if it comes mainly from growth it will be devastatingly unsustainable'.

There has been significant decoupling of economic expansion from **metallic minerals** consumption in developed countries in recent decades. The intensity of use (kg/million $GNP) of aluminium, for example, has declined since 1975 (Fig. 2A; Wellmer & Kürsten 1992 fig. 12). The substitution by non-metallic materials, combined with efficiencies in the use of the metals, partly accounts for the lower intensity of use of metals in industrialised countries. A further factor is the great increase in value-adding in finished products. But although the relative importance of metals has declined in industrial economies, there has not been significant decline in the total quantities of metals used (Fig. 2B), i.e. there has **not** been an overall dematerialisation).

Globally, consumption continues to increase. While demand

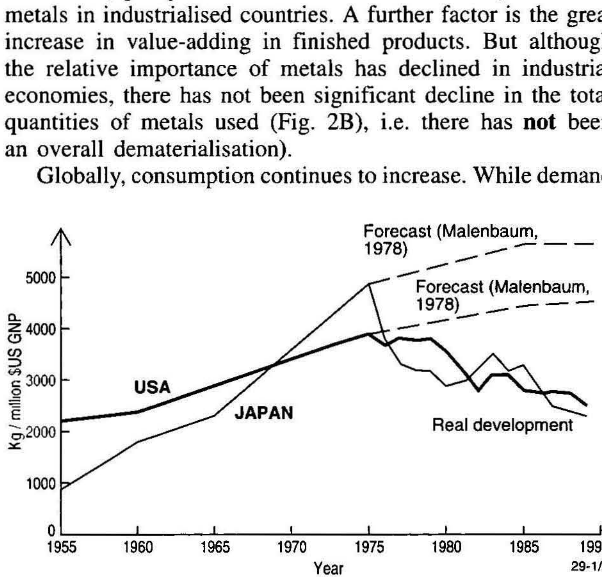


Figure 2A. Comparison of intensity of use of aluminium in Japan and USA, showing a change in trend in the mid-1970s.

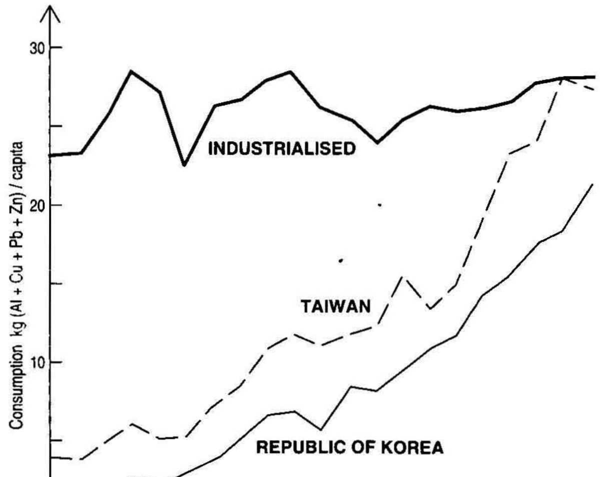


Figure 2B. Aggregated base metal consumption showing how consumption in Taiwan and Korea has grown to match the relatively constant level in industrialised countries.

for some minerals has stabilised, and is even declining in some developed countries, this will be more than offset by increasing demand in the developing countries and particularly those of South East Asia, where population is also growing rapidly. For example, the per capita consumption in the newly industrialising countries of Korea and Taiwan has grown rapidly over the last thirty years to levels similar to those of industrialised countries (Fig. 2B). Recent analyses also show that between 1950 and 1990 the population in underdeveloped regions grew from 68% to 77% of the global population, while the share of consumption of various metals grew from between 1% and 5% in 1950 to between 12% and 25% in 1990 (Wellmer & Kürsten 1992).

For similar reasons, it seems likely that demand for the **fuel minerals** will continue to increase for several decades (e.g. Bookout 1989, Masters et al. 1991). In this case, however, the demand for energy has continued to increase in the developed as well as the developing countries. Estimates on a 'business as usual' or 'as-is' basis (Ogawa 1991, quoted by Foster 1993), indicate that world primary energy consumption will approximately double between 1975 and 2000 and could double again in each of the following 25 years for an estimated population of 9.8 billion in 2050. Foster (1994) comments that he does not pretend that the forecast will come to pass but adds that uncertainty about the outcome cannot mask the trend.

The increased materials and energy use indicated above will inevitably be accompanied by increased environmental impacts.

Population, resource use and the environment can also be clearly linked through the concepts of per capita use of resources, and per capita use of space. The **total impact** of the human population is the product of population and per capita resource use (Daly 1977) and it can be controlled by controlling either or both of these factors.

The support square

The per capita use of space is given by the total available area divided by the total population and has been called the support square, which has been described as 'the scrap of land that must supply all the resources that an individual uses throughout a life, and that must fulfil the same purpose for others who follow. Somehow that space must also consume most of the solid wastes left over' (Skinner 1989). This human-focused concept actually overstates the area available per capita in so far as the needs to preserve areas of natural environment and biodiversity must also be recognised: the same total area has to support most other land-based species as well as the human population.

Nevertheless, the diminishing size of the support square gives a very graphic illustration of the impact of the increasing human population (Fig. 3). The average global support square towards the end of the next century is likely to be about 100 m square, about the same as the local support square for Europe today. But Europe offers a particularly favourable environment for human habitation. Moreover, Europe, like other developed countries in North America and Asia, is not wholly dependent on the local support square: it obtains a substantial part of its resources from other less densely populated regions. To a considerable extent, therefore, environmental impacts associated with the production of both renewable and non-renewable resources are also borne elsewhere. As the global population increases, with its attendant environmental impacts, pressures on land use will also increase, as will pressures on the natural environment and on biodiversity. It will become increasingly difficult to guarantee external supplies of both renewable and non-renewable resources.

In the case of metallic and fuel resources, the global distribution is very uneven and there are very large trade flows. It will be increasingly necessary to recognise mineral

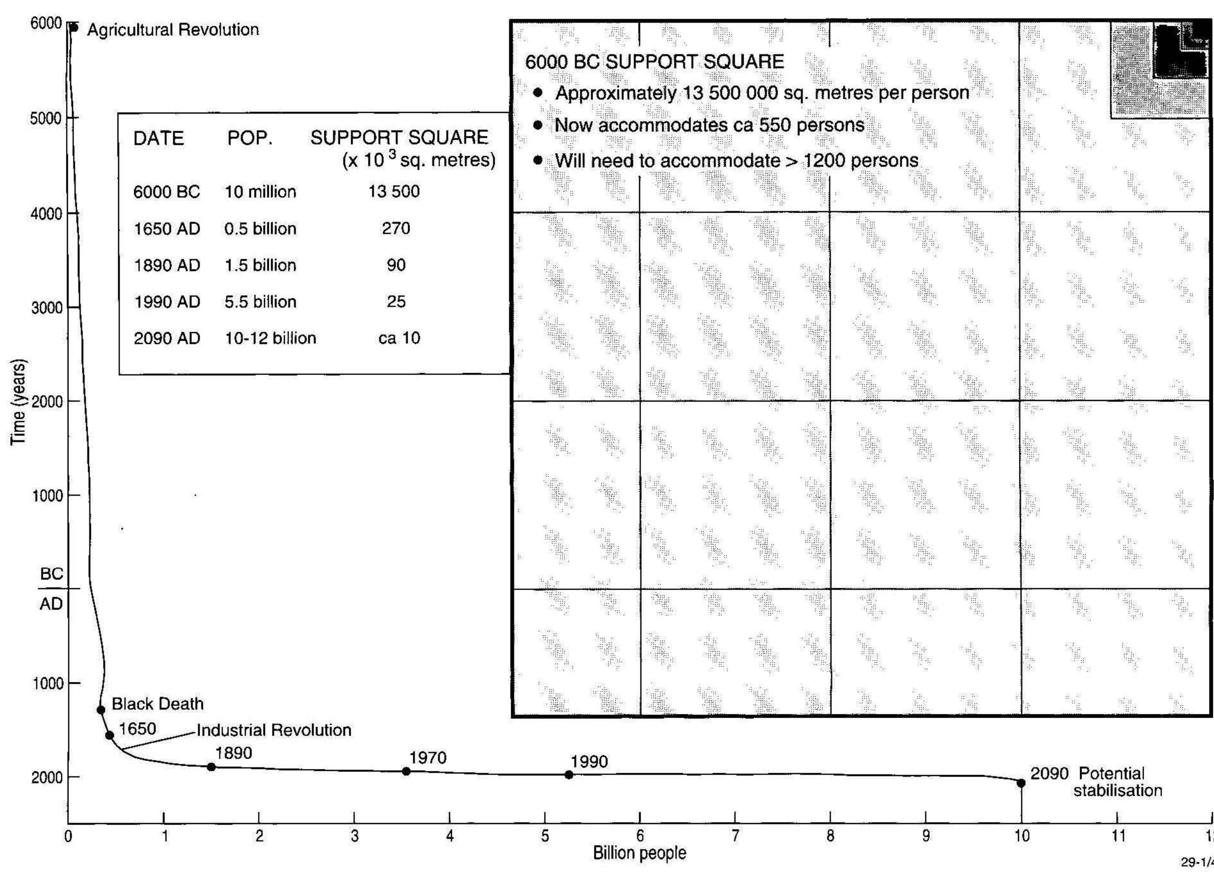


Figure 3. Growth in global population, and resulting shrinkage of the global support square.

supply as a global problem, requiring global cooperation and management. The competition for land use is likely to close additional areas to exploration and development, as it has already done for parts of Europe. It is therefore especially important that issues of mineral supply be considered as part of the proposed **integrated approach to the planning and management of land resources** under Agenda 21, both at the national and international levels (see *Metallic minerals*, below).

Per capita resource use

The average global per capita consumption of all minerals has been estimated at close to 10 tonnes a year (Skinner 1989). The **total impact** for the global population involves the displacement of about 50 billion tonnes of minerals per annum, a figure substantially greater than the amount of material moved by natural processes. A large proportion of this material consists of industrial minerals which are re-located from quarries to the sites of growing cities and to transport networks.

In the major industrial countries, consumption is much higher than the global average. In Germany, for example, it has been estimated that the average individual, in a lifetime of 70 years, consumes about 772 tonnes of construction materials, nearly 54 tonnes of other industrial minerals, about 363 tonnes of fuel minerals and about 43 tonnes of metals (mainly steel) (Fig. 4). Allowing for the quantities of ore and overburden that are involved in producing the final products, it is likely that some 1600 tonnes of rock are consumed by each individual, or well over 20 tonnes a year.

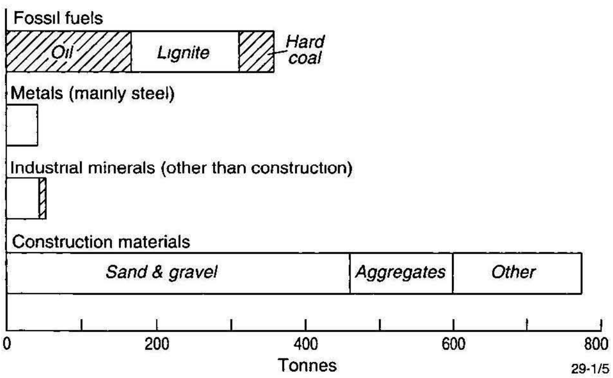


Figure 4. Material consumed in a lifetime of 70 years in Germany. About 17.5% of the total is imported (indicated by shaded areas) — mainly fossil fuels and metallic ores (after Wellmer 1994, personal communication).

The volume of rock involved over the average lifetime is more than 500 cubic metres, which corresponds to an area more than 7 metres square excavated to a depth of 10 metres — approximately 0.5% of the 100 metre support square.

If this kind of consumption were maintained for a population of 10–12 billion, the **total impact** would be more than quadrupled to over 200 billion tonnes, or approaching one hundred cubic kilometres of rock, each year. It is hardly possible to argue that such consumption rates are sustainable, either in terms of environmental impacts or availability of resources.

This discussion serves to illustrate the unsustainability of present demand trends and the consumption patterns that cause them. Clearly, every effort must be made to decouple economic expansion as far as possible from increased use of materials and energy. The fundamental needs of sustainable development are to minimise the primary inputs of materials and energy and to minimise the environmental impacts of these inputs. This will involve acceleration of present trends towards dematerialisation, combined with recycling and substitution. However, as indicated above, the process of adjustment will be counteracted by the increasing global population and the demand for improved living standards. Targets for sustainable consumption patterns will need to be set in the light of the best possible knowledge of the impact of resource use (see **Environmental impacts of minerals use**, below), and of the availability of resources (see **Availability of resources**, below).

Environmental impacts of minerals use

The present relationships between atmosphere, hydrosphere, lithosphere and biosphere are the result of evolution throughout Earth history. The interactions are complex but rates of change resulting from natural processes are relatively slow on human time-scales and the natural environment is in a state of dynamic quasi-equilibrium. Soils, in particular, form part of an oxidised zone resulting from interaction between bed-rock, air, water, plants and animals.

The environmental impacts of minerals extraction and use arise from the disturbance of this natural balance of Earth processes. In the case of phosphate, for example, phosphate which has been sequestered by natural exogenic processes over hundreds of millions of years is being returned to the land surface on a time scale of a few hundreds of years. Similarly, in the case of fuel minerals the current rate of consumption is more than a million times the natural rate of accumulation.

Metallic ore deposits, in contrast, are largely formed by endogenic processes and are unusual concentrations of elements which normally have very low concentrations in soils and water. The ores are largely derived from below the oxidised zone which is broadly in equilibrium with air and ground water. Both natural weathering and mineral processing therefore involve oxidation of these ore minerals and the release of various pollutants, including sulphur dioxide and toxic trace elements.

Fuel minerals

The environmental impacts of the combustion of fuel minerals are well known and have become a global policy issue (see e.g. Steering Committee of the Climate Change Study 1995). The emissions of carbon dioxide and other greenhouse gases continue to increase. If the forecasts of increased consumption of energy given above were realised (Foster 1993), carbon dioxide emissions could increase by a factor of 5.5 between 1975 and 2050. This emphasises what dramatic changes would be required to stabilise the emissions at 1990 levels by the year 2000, as proposed by the Framework Convention on Climate Change.

On the basis of climate models and current knowledge of the carbon cycle, it has been estimated that, depending on the success of the measures taken, carbon dioxide concentrations in the atmosphere will reach between two and three times pre-industrial levels for a population of 11.3 billion by the year 2100 (Inter-Governmental Panel on Climate Change 1994). It has also been estimated that the enhanced greenhouse effect will cause a rise in the average surface temperature of the Earth during the next century of between 1.5° and 4.5°C, and a rise in sea level of between 30 and 110 cm. The latter will add to the effects of coastal subsidence in many areas of very high population density (including mega-cities) in the coastal zone.

The analyses indicate that stabilisation at twice the present level of CO_2 (or less) will require an eventual and sustained reduction of emissions to substantially below present levels, with obvious implications for the substitution of coal and petroleum. However, recent analyses suggest that an **immediate** reduction of emissions is not essential for the achievement of such long-term targets. With orderly planning, they might be achieved by more cost-effective mitigation strategies, which allow adherence to the 'business-as-usual' scenario for up to three decades — but temperature and sea-level rise would initially be more rapid (Wigley et al. 1996).

In any case, it will be necessary to adapt to the consequences of significant climate change. Fortunately, not all these consequences will be negative (e.g. Petit-Maire 1995, for the Sahara) and, in the context of population growth, it might be argued that climate change is of lesser importance than land degradation, water quality and availability, other forms of pollution, and the preservation of biodiversity.

Clearly, although there are many uncertainties in the scientific assessment of climate change, much information is available for decisions in this area, and the precautionary principle requires that steps be taken to mitigate the environmental impacts. Agenda 21 (Chapter 4, para. 24) noted that (in general) significant changes in consumption and production patterns seem unlikely to occur in the near future 'without the stimulus of prices and market signals that make clear to producers and consumers the environmental costs of the consumption of energy, materials and natural resources, and the generation of wastes'. With respect to energy, a panel of the US Academy of Sciences in 1991 concluded (Tickell 1994) that 'on the basis of the principle that the polluter should pay, pricing of energy production and use should reflect the full cost of the associated environmental problems.' For the longer term the panel envisaged 'that a mix of renewable energy resources, together with nuclear power, would gradually assume a larger role in the new price structure. The transition to them raises more political and economic than technical problems'. The timing and nature of the changes in the energy mix will have major implications for the mineral industry as a whole.

Metallic minerals

One of the major impacts of the use of the metallic minerals arises from the energy used in their production. Gases released in the production process, notably sulphur dioxide, have also caused environmental problems, such as acid rain. A number of the metals are toxic, and can cause unacceptable pollution (as with lead in petrol).

Thus, the ideal industrial ecosystem (Fig. 5) would have minimal input of primary materials. Inputs can be reduced by more efficient processing throughout the cycle, by reduction of waste and by recycling. For example, in 1994 in the USA, reclaimed metals and mineral materials accounted for about one quarter of the total mineral raw materials used (USBM 1995). For specific metals, such as lead and copper, recycled material already accounts for well over half the total consumption in some industrialised countries.

Environmental impacts can be further reduced by improved treatment of waste and, where necessary and possible, by substitution. For example, there may be the opportunity to use waste materials from the production of high value commodities to substitute for primary raw materials in the

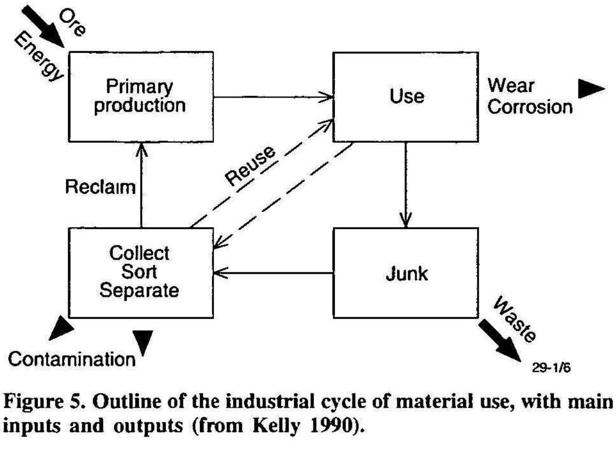


Figure 5. Outline of the industrial cycle of material use, with main inputs and outputs (from Kelly 1990).

production of low value commodities. Fly ash and rea-gypsum and, potentially, sulphur from the production of electricity based on coal can be used as substitutes for primary raw materials for cement production, gypsum and other sources of sulphur.

New materials and composites are increasingly being used instead of metals (e.g. Kelly 1990). Some of these substitutes, such as ceramic materials, inorganic glasses and optical fibre, are derived from relatively common rock-forming minerals, while others, notably plastics, depend on the supply of fuel minerals. Relatively few are from renewable materials. In general, therefore, substitution involves replacing one non-renewable resource with another. It does not contribute substantially to dematerialisation, although it may reduce environmental impacts.

One reason for substituting other materials for metals for some purposes is that less energy is required, for example in the production of paper and plastic products, than for equivalent metallic products. However, the differences are not large enough to overcome other factors, such as particular properties or the ability to be processed for particular purposes. There have also been great improvements in the energy efficiency of primary production and forming of metals, and further improvements are possible.

Some attempts have been made to assess the environmental impact of extraction and processing involved in the primary production of metals. For example, a pilot study of mass balances (inputs and outputs) in the production of nickel has been undertaken in Germany (Wellmer, BGR Germany, pers. comm. 1994). The study took into account the different flows involved in the processing of laterite ores of various types and of sulphide ores. Information from such studies can be used in assessing total environmental impacts of alternative source materials.

Attempts are also being made to assess the environmental impacts of the production of various goods by focusing on material inputs, and taking into account all phases of product life cycles (Hinterberger et al. 1994). The underlying thesis is that knowledge of the total material inputs involved would allow the assessment of the relative merits of different materials and products and assist in an overall dematerialisation strategy. It is therefore proposed to classify products according to **'material input per service unit'.** Such studies will be valuable in promoting an understanding of environmental impacts. Clearly, general reductions in material inputs would also lead to reductions in energy usage, in waste, and in any toxic chemical flows.

It seems likely, however, that the main approach to environmental impacts related to metallic minerals will continue to be through addressing the problems of particular identified adverse outputs from the industrial ecosystem. Agenda 21, for example, deals specifically with the environmentally sound management of toxic chemicals in Chapter 9, of hazardous wastes in Chapter 20 and of solid wastes and sewage-related issues in Chapter 21. Nevertheless, these strategies will be more effective if it is recognised that these outputs are specifically related to material inputs which may themselves be subject to modification. Remedial measures may therefore be possible at all stages in the mineral cycle.

It is therefore suggested that greater efforts should be made at the international level to coordinate, assess and disseminate technological knowledge of appropriate strategies for improving the efficiency of the industrial ecosystem and reducing environmental impacts. Such efforts could be coordinated by the United Nations with the cooperation of non-government organisations such as the International Council on Metals and the Environment.

The assessment of pollution, at all scales, requires that we monitor the health of the surface of the solid Earth in the same way as is being undertaken for the oceans and atmosphere. The natural concentration of trace elements reflects the variability of the geology and knowledge of the natural variation can be critical in assessing human impacts (e.g. Wyborn et al. 1996).

The International Geochemical Mapping Project of the International Geological Correlation Program (Darnley et al. 1995, p. 15) has addressed the need for a coherent, systematic, worldwide, multi-element geochemical database and has determined the basic requirements and likely costs. It points out that such a database is pertinent to administrative and legal issues and that it 'contains information directly relevant to economic and environmental decisions involving mineral exploration, extraction and processing; manufacturing industries; agriculture; forestry; many aspects of human and animal health; waste disposal; and land use planning'. It has established that available data are substantially incomplete and internally inconsistent. Evidently, the data required could be obtained by enlisting the cooperation of national geological surveys. The necessary central coordination could be carried out by an appropriate United Nations agency.

Industrial minerals

In the context of land-use planning, it is clearly important to take particular note of the demand for **industrial minerals** which, as noted above, form the dominant component of total material usage. Debate concerning the depletion of mineral resources has been mainly concerned with the metallic and fuel minerals, with the implicit assumption that supplies of industrial minerals are inexhaustible. However, because of the enormous quantities involved, and because they are not readily recycled, the supply of industrial minerals raises particular problems of environmental impacts.

It might be supposed that once the main infrastructure of industrialised countries has been established, the needs for **construction materials** (for replacement and maintenance) would be significantly reduced, thus contributing to the process of dematerialisation (Phillips 1987). Apparently, however, this stage has not yet been reached in Europe. Although population there has stabilised, the annual consumption of construction materials continues to increase and there is widespread concern about the environmental impacts of quarrying and transport (Mineral Resources and Sustainable Development 1994). To meet the demand, there has also been an increase in the amount of sand and gravel derived from shallow offshore areas, and coastal superquarries have also been developed. It has been suggested, in this connection, that sustainable development of the coastal zone may require imposition 'of a littoral or "thalassic" tax which, like a carbon tax, takes a global view of the "polluter pays" principle' (Cook 1995).

Thus the rate of consumption of construction materials and the environmental impacts are clearly important issues for the promotion of sustainable construction industry activities (Agenda 21, Chapter 7G).

Amongst industrial minerals, phosphate has a special importance because of its essential contribution to the productivity of the agricultural industry (International Strategic Minerals Inventory 1984). Phosphate production increased roughly sixfold between 1950 and 1980 to around 150 million tonnes a year (roughly 30 kg per capita globally and near 50 kg per capita in some countries). It has fallen recently because of the near-collapse of output from the former USSR, but is likely to continue to rise in the future to meet the needs of the growing global population. Reserves are very large (Northolt et al. 1989, USBM 1995) but clearly finite. However, as with petroleum, the principal concern is with the environmental impacts of phosphate use as a result of the greatly increased levels of phosphate, especially in inland waters. But there is no substitute and it is difficult to control and reduce consumption.

Availability of resources

General outlook

The supply of resources is essentially a response to demand which has largely been regulated by price. Over recent decades, because of the success of the mineral industry in meeting demand, there has been plentiful and low-cost supply, which stimulates consumption and therefore demand. This situation was temporarily changed by the oil price shocks of the 1970s. In general, as discussed above, levels of demand do change through time, not only as a result of changes in cost, but also because of substitution, recycling, technological advances or environmental concerns At present, the changes in demand are being driven largely by the environmental concerns. It is necessary to consider, however, whether resulting consumption patterns are also sustainable in terms of availability of resources.

The issues involved in assessing the sustainability of mineral supply have been comprehensively addressed in the scientific literature, but they have received little explicit attention in Agenda 21 or in the more general debate on sustainability. There has been a tendency to take too pessimistic a view, using published figures of ore reserves, or too optimistic a view on the basis that mineral resources are essentially infinite and that solutions will be found to the technological problems when scarcities of conventional mineral deposits emerge.

For the very long term it is hardly possible to predict how far technological advances, or specific scarcities, will lead to reduced demand (dematerialisation) or successful substitution, especially of energy, by renewable resources. Eventually, the development of non-polluting energy sources may largely solve the problems of resource supply by allowing extraction of minerals from sources which cannot at present be exploited economically and without unacceptable environmental impacts; but the timescale of any such development is very uncertain.

Because of the potential changes in the level of demand (for the reasons noted above) and potential changes in the nature of the supply (e.g. from lower grade sources as the result of technological change), it is not practicable to assess the overall 'lifetime' of the resources of a particular commodity. It is possible, for example, that the demand for a particular commodity will disappear, in which case it would cease to be a commodity or resource, and its 'lifetime' would be infinite.

Instead it is appropriate to consider a 'horizon of sustainability'. This defines how far we can look ahead to assured supplies of particular commodities making particular assumptions about the nature of demand. This horizon of sustainability can be extended further into the future as appropriate knowledge of resources is developed. On the timescales of sustainable development it is necessary to consider not only those resources which have already been identified but also the scope for discovering new deposits. These 'identified' and 'undiscovered' resources are discussed below.

The horizon of sustainability: identified and undiscovered resources

Identified resources. Most existing national and international assessment programs are limited to the assessment of 'identified resources' and especially of 'demonstrated economic resources' (DER), that part of total resources which has been identified by exploration and drilling, and which can be extracted economically under current conditions (Fig. 9.)

At present, therefore, knowledge of future availability of resources (and of whether production and consumption patterns are sustainable) is based essentially on the assessments of **identified resources,** which are not readily related to prospective mineral provinces or to longer term mineral potential. It is clear, however, from these assessments, that **the supply of mineral and petroleum resources over timescales of a**

few decades is well assured. The trend towards internationalisation of major resource companies, allied with increasingly effective exploration methods, has permitted the ready maintenance of the world's stock of economic identified resources. Technological advances in mining methods and mineral processing (e.g. for gold and copper), and strong competition, have also resulted in stable or declining deflated commodity prices (Fig. 6; see also Wellmer & Kürsten 1992).

The current stock of DER can be related to the changing demand for (and therefore production of) a particular commodity by the resource-production ratio (R/P ratio). This ratio gives the number of years that supply could be maintained from current stocks at the current level of demand. Time series of production, DER and R/P ratio illustrate the availability of mineral resources through time and give an indication of the effects of past changes in demand, and of major social and economic impacts. They also demonstrate that the stock of DER is not a fixed stock, subject only to depletion. It is continually being renewed, either by discovery of new economic resources or transfer from the large pool of known but sub-economic resources, as a result of technological advances or of price rises induced by scarcity.

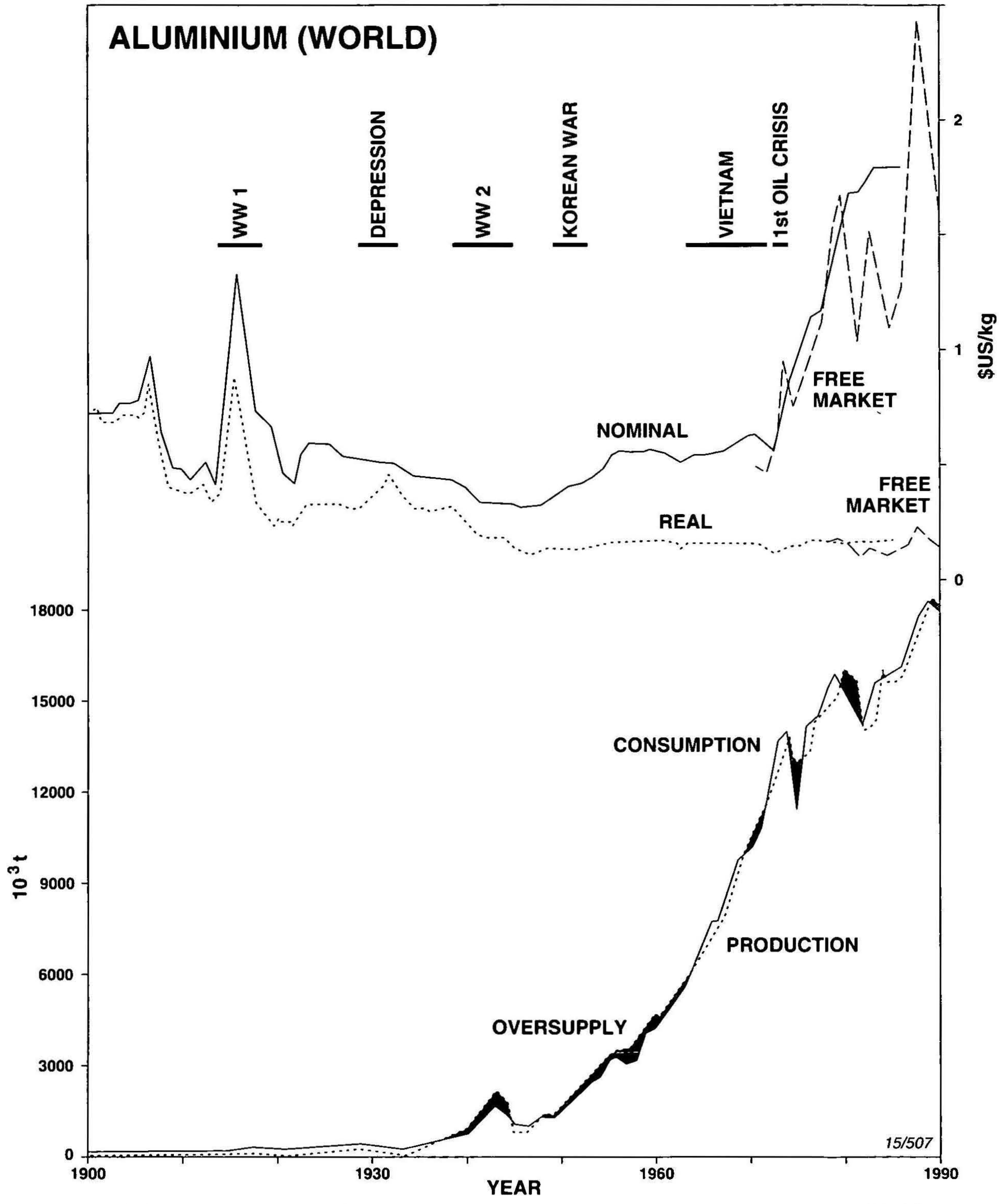


Figure 6. Aluminium as an example of declining real prices while production and consumption have grown rapidly (after Wellmer & Kürsten 1992).

In general, the annual production of mineral commodities has increased greatly and fairly steadily this century, but DER have also grown so that the R/P ratios have been maintained. In the case of bauxite, DER have been maintained but the R/P ratio has declined because of greatly increased annual production.

The large tonnage commodities such as coal, iron ore, bauxite and phosphate have large R/P ratios (hundreds of years). However, these are near-surface deposits and the capacity to continue to renew the stock of DER is in some doubt. Mining of these commodities also has the largest immediate (if transitory) environmental impact.

For most of the metallic minerals, R/P ratios are much smaller (some tens of years), but again it has been possible to maintain these ratios. This reflects the capacity of the minerals industry to take a relatively long-term view of future demand and of the factors likely to affect it, and to make appropriate investment in exploration and development. The time lag between such investment and the establishment of new DER is typically ten years or more. R/P ratios therefore provide a clear horizon of sustainability, usually of around thirty or forty years.

The information on identified resources does not, however, provide an assurance of supply over the longer timescales of sustainable development — to potential stabilisation of the global population at the end of the next century, or to the potential steady state development further in the future.

The evident continuing success in maintaining R/P ratios must be set against the relatively short period since the industrial revolution over which the resources have been exploited, and the exponential growth in demand (Fig. 7). These are essentially non-renewable resources and the economic resources have been severely depleted during the 20th century at an ever increasing rate. On the timescales of long-term sustainability, the situation for metallic ore deposits is not fundamentally different from that for petroleum (e.g. von Engelhardt et al. 1976, Cloud 1977; illustrated in Fig. 8). The trend towards utilisation of lower grade ores is already well established. In the absence of dramatic changes in consumption patterns, there would, at some time in the future, be real scarcity of resources **of the kinds currently mined.** It is clearly desirable to have as much advance warning as possible of potential **mineral shocks.**

As indicated above, technological solutions may well become available, but during the period of increasing global population and increasing demand for mineral resources, prudent management, according to the precautionary principle,

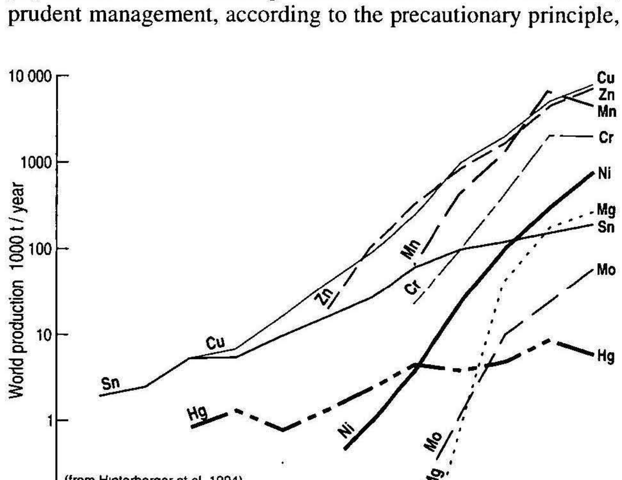


Figure 7. Exponential growth of metals production since the industrial revolution. Note the logarithmic scale for production (after Hinterberger et al. 1994).

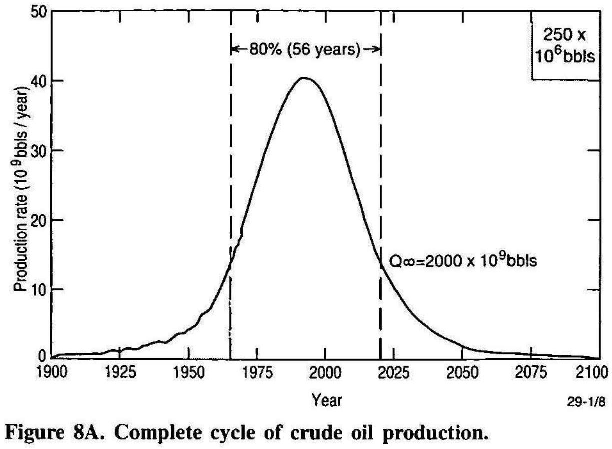


Figure 8A. Complete cycle of crude oil production.

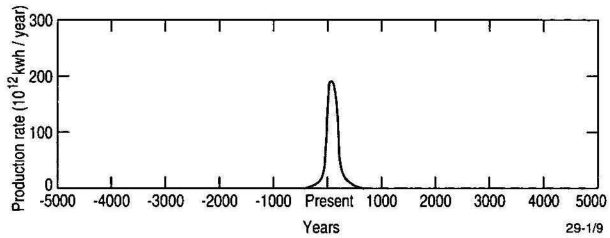


Figure 8B. The epoch of fossil fuel exploitation viewed in the perspective of 10 000 years of human existence, past and future. (Both from Engelhardt et al. 1976, after Hubbert 1974).

clearly requires further knowledge of the sustainability of supply, and of sustainable consumption patterns, beyond the present horizon of sustainability. The concern is not one of 'running out of resources', but of avoiding potential problems, and trying to ensure the optimum and efficient use of available resources with minimum environmental impact.

At present, we have very limited knowledge of the global potential for discovering new deposits, and this deficiency needs to be addressed. There are also increasing pressures on land use, which may make it increasingly difficult to explore for, and develop, the available resources. These issues are discussed in the following sections.

Undiscovered resources. Current information for petroleum (oil and natural gas) is quite comprehensive. As well as reasonably reliable figures for reserves, there are also good estimates of the quantity of undiscovered resources (or potential resources), especially for crude oil. Early estimates of the potential were too low and were soon overtaken by actual production figures. However, as the processes of generation, migration and entrapment of petroleum have become better understood, it has been possible to assess potential within reasonably narrow limits. A recent estimate (Masters et al. 1991) of ultimate resources (cumulative production plus reserves plus mean undiscovered resources) of crude oil gives a figure of 2079 billion barrels (in a range between 1800 and 2480 at the 90% confidence level), which is very similar to various estimates made 20 or 30 years earlier (see also Fuller 1993).

In contrast, there are no reliable global estimates for the undiscovered resources of metallic minerals. There are many different deposit types, the processes of generation of metallic mineral deposits are very complex and less well known than for petroleum, and methods for estimating undiscovered resources are much less reliable. Most methods are of a very general character and do not permit information about undiscovered resources to be linked in integrated systems of land use and management.

This is unfortunate, since it tends to foster the notion that these resources are unlimited, a notion based on the fact that the total amount of most metals and other mineral commodities in the Earth's crust is indeed very high. However, in most rocks, the metallic elements occur in very low concentrations, and not in discrete minerals, so that their extraction is not feasible. Metallic mineral deposits, whether currently economic or sub-economic, result from particular conjunctions of processes in the Earth's crust which have led to the deposition of unusual concentrations of metals as ore minerals in particular locations.

Mineral deposits of the kinds currently regarded as economic or sub-economic might be quite limited by comparison with the resources already identified. Their distribution is highly heterogeneous at various scales both areally and in depth. Different kinds of deposit are characteristic of different geological environments so that their occurrence is limited to particular mineral provinces. Moreover, many deposit types are formed close to the Earth's surface so that the opportunities for discoveries below the uppermost few kilometres are relatively limited.

Many of the more prospective terrains of the world have already been extensively explored by modern methods so that the more readily detected deposits have already been discovered. However, exploration techniques are continually being improved in the search for concealed deposits. Other terrains have only recently been opened to the international exploration industry and there will undoubtedly be major discoveries over the next few decades. It is likely, therefore, that the horizon of sustainability can be pushed back into the second half of the 21st century. Nevertheless, in the present state of knowledge, it is prudent to recognise that there may be real problems of meeting demand over the next century. These problems may be increased by political factors and, more generally, by restrictions on the availability of land for exploration (see below).

Strategic factors

The uneven distribution of mineral deposits worldwide means that the sources of supply of some commodities, such as platinum, chromium, vanadium and manganese, are highly restricted geographically. In recent decades, for example, South Africa and the former Soviet Union have accounted for over 80% of world mine production of platinum group metals. Such commodities will continue to attract high exploration interest in order to establish other sources of supply.

The dependency on supplies of minerals from outside sources has led some countries to the identification of 'strategic minerals', an imprecise term which embraces the concepts of criticality and vulnerability. At the national level, the criticality of a mineral depends on its contribution to the national economy and general physical well-being, and critical minerals may in some cases be vulnerable to interruptions of supply. A report by the Office of Technology Assessment (US Congress 1985) noted that

> Only three nations (South Africa, Zaire and the USSR) account for over half of the world's production of chromium, cobalt, manganese and platinum group metals. These metals are essential in the production of high temperature alloys, steel and stainless steel, industrial and automotive catalysts, electronics and other applications that are critical to the US economy and the national defense.

A fuller knowledge of potential sources of supply is therefore likely to be of increasing interest to many countries.

A number of countries have participated in producing an International Strategic Minerals Inventory (ISMI, recently renamed International Studies of Minerals Issues), and this has produced valuable information on the identified resources of many commodities (mainly metals, but also important non-metallic industrial minerals, such as phosphate and graphite). ISMI has not attempted to make estimates of undiscovered resources (Fig. 9).

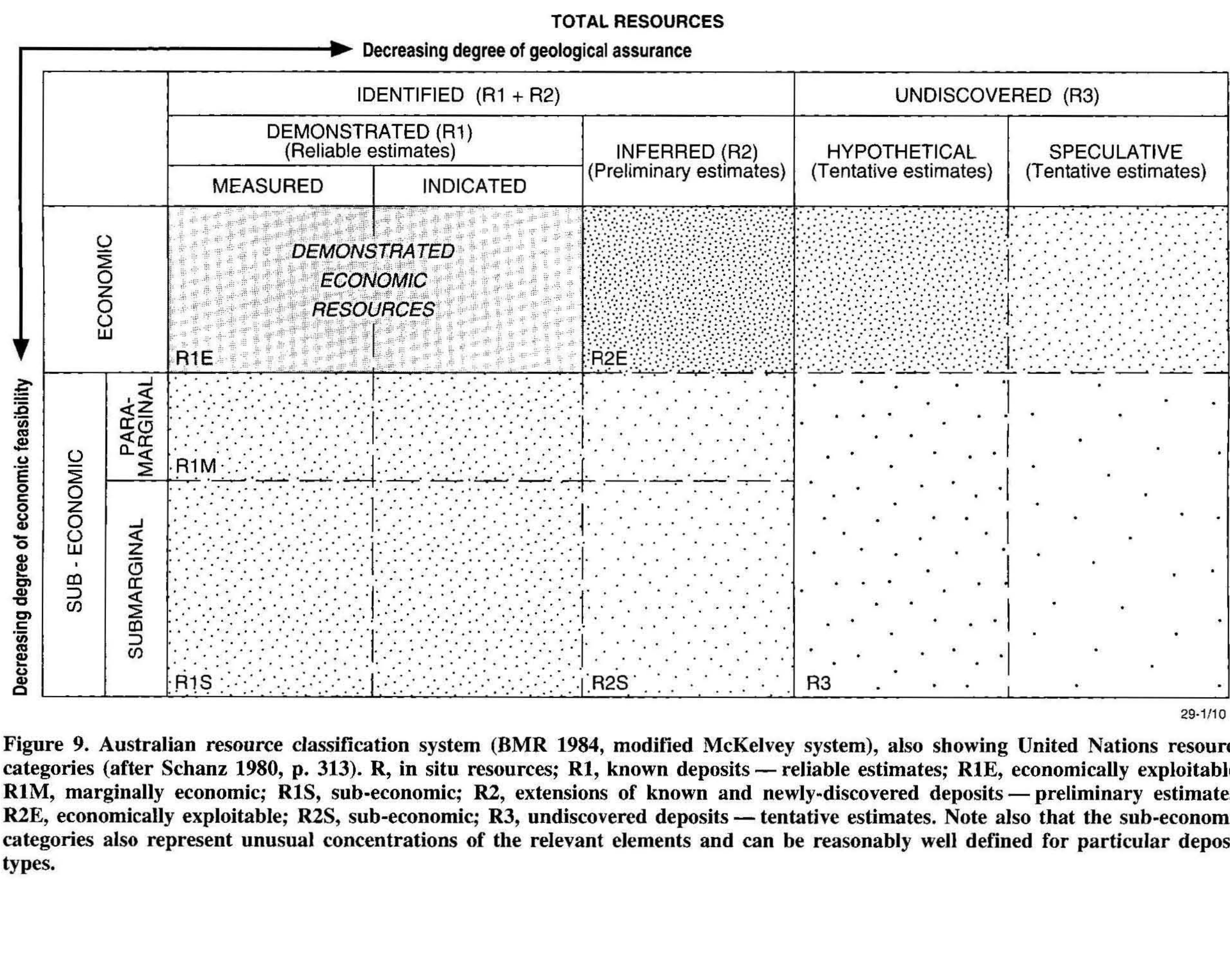


Figure 9. Australian resource classification system (BMR 1984, modified McKelvey system), also showing United Nations resource categories (after Schanz 1980, p. 313). R, in situ resources; R1, known deposits — reliable estimates; R1E, economically exploitable; R1M, marginally economic; R1S, sub-economic; R2, extensions of known and newly-discovered deposits — preliminary estimates; R2E, economically exploitable; R2S, sub-economic; R3, undiscovered deposits — tentative estimates. Note also that the sub-economic categories also represent unusual concentrations of the relevant elements and can be reasonably well defined for particular deposit types.

Availability of land for exploration

Population pressures worldwide and the attendant environmental impacts are causing increasingly severe competition for land use, and there has also been a reaction against mining in some countries. The need to meet the global demand for mineral resources from the most efficient sources worldwide is not readily appreciated either by local communities, whose lifestyles may be affected by major mining projects, or by national conservation movements. This is especially the case if the demand is perceived to be the result of extravagant or wasteful consumption patterns with undesirable environmental impacts. It is therefore important that strategies to maintain supply be linked to effective strategies to move to sustainable consumption patterns.

Population pressure and environmental impacts can militate against the vigorous exploration programs which would need to be pursued in the most prospective areas worldwide if essential mineral supply is to be maintained in the short term. It is important, therefore, for governments to recognise that only a relatively small proportion of the continental areas is highly prospective for each of the various metallic mineral deposit types. These areas need to be identified, and their mineral potential taken into account in determining the needs for mineral exploration, in a global context and using an integrated approach to land use planning. This will not be feasible unless a comprehensive information base is developed on mineral resource potential, which can be integrated with other land use information.

This is recognised in general terms in Agenda 21, Chapter 10 (Integrated approach to planning and management of land resources) viz:

> Integration should take place at two levels, considering, on the one hand, all environmental, social and economic factors (including for example impacts of the various economic and social sectors on the environment and natural resources) and on the other all environmental and resource components together (i.e. air, water, biota, land geological and natural resources).

Assessment of mineral resource potential

Areas which are prospective for particular mineral deposit types have been called **permissive tracts** and their identification is the first step in the assessment of mineral potential and of unidentified resources. Such permissive tracts can be identified on the basis of geoscientific mapping programs carried out by national geological surveys. In Australia, for example, mapping under the National Geoscience Mapping Accord would allow the identification of permissive tracts within the main mineralised provinces (Fig. 10). In conjunction with information on mineral occurrences, such mapping permits **qualitative** estimates of prospectivity and resource potential. It provides the basis for assessment and investment in exploration by mining companies. Mineral maps and/or metallogenic maps can be produced as by-products of geological surveys (e.g. Emberger 1993).

It is also desirable to assess not only the most likely

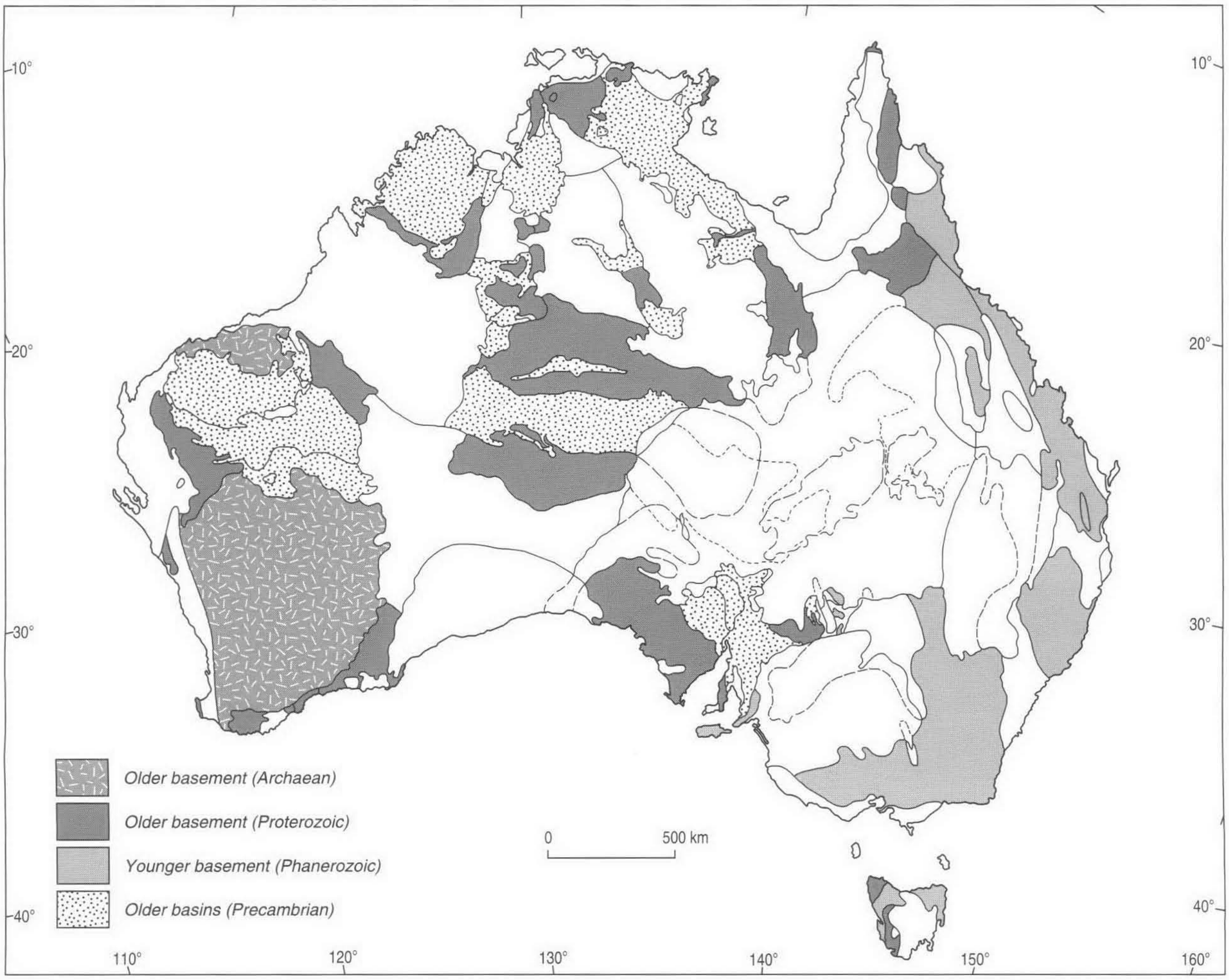


Figure 10. Main basement provinces of Australia and associated Precambrian basins — the principal areas prospective for metallic mineral deposits.

geographic sources, but also the quantities of undiscovered deposits in both the hypothetical and speculative categories (Fig. 9). Hypothetical resources are those 'which may reasonably be expected to exist in a known mining district or mineral province under known geological conditions' while speculative resources are those 'which may occur either in known types of deposits in a favourable geological setting where no discoveries have previously been made, or in yet unknown types of deposits which remain to be recognised' (BMR 1984). Both these categories contain deposits which are likely to be economic or marginally economic under current criteria.

Several approaches have been made to the **quantitative** assessment of undiscovered mineral resources (e.g. Dorian & Zwartendyk 1984). Most notably, a three-part method of quantitative assessment has been applied by the US Geological Survey since 1975. 'Its original purpose was to provide quantitative resource information in a form consistent with an economic analysis so that mineral resource values could be compared with other competing uses of land' (Singer 1993). These assessments are most reliable where the mineral deposit geology is already well-known — usually for relatively small geological provinces.

More recently, there has been a proposal to make a **national** three-part assessment, providing 'a consistent, usable minimum level of current mineral-resource information together with estimates of total undiscovered mineral endowment' for the entire United States (McCammon & Briskey 1992 p. 259). It was suggested that such an assessment 'is essential for ensuring that all domestic mineral resources will be considered in planning the optimum use of the Nation's public lands and for securing long-term mineral supplies from national and international sources'.

As a first step in this program, a two-year preliminary quantitative national assessment based on existing national data was also proposed. This would produce 'maps showing the outlines of tracts that are permissive for the types of deposits concerned' (McCammon & Briskey 1992 p. 61). Such an assessment would be of great value for planners in the United States, and similar assessments would be of even greater value in developing countries with substantial resource potential. Moreover, the value of all such national assessments would be greatly enhanced if they could be examined in the context of global potential and global resource needs. However, whatever the fate of the proposed US program, it is unrealistic at the present time to propose a similar **quantitative** assessment worldwide. In most countries, the level of geological knowledge is inadequate for assessments by this three-part method and there are impediments to the acquisition of such knowledge, both in terms of expertise and financial resources (cf. Harris et al. 1993).

A more realistic goal globally would be to produce maps delineating 'permissive tracts' worldwide using internationally agreed criteria. This would involve only the first steps in the proposed US preliminary assessment, viz

- compile existing data
- apply limited mineral-deposit models
- construct maps of delineated permissive tracts.

Such maps would provide the basis for iterative assessments of undiscovered resources as data became available. Although the assessments of individual tracts would generally be, at best, semi-quantitative, the global picture so provided would allow much more realistic assessments of sustainability beyond the horizon currently provided by identified resources.

More immediately, such maps would help the consideration of minerals issues within an integrated approach to land-use planning. They would allow the needs for mineral exploration and development to be assessed in relation to other land-use needs. Moreover, since perceived mineral resource potential has been identified as the most important single criterion for the international mining industry in assessing the investment environment for exploration, a global program identifying permissive tracts worldwide would aid the efficient and socially harmonious operation of the industry. For example, it would help local and national populations to appreciate the wider global interest in keeping the principal permissive tracts of the world open for exploration and development as far as possible if the global endowment of mineral resources is to be effectively managed and used.

In this context, it is important to distinguish between the scope of mineral exploration and the scope of mineral developments. It must be recognised that, although it is necessary to explore over large areas, this can largely be accomplished by non-intrusive techniques, such as aeromagnetic surveys, and exploration is not generally incompatible with other land uses. Mineral development after successful exploration will continue to affect only relatively small areas. If current best practice in integrating environmental and development concerns is implemented, the short-term environmental impacts of mining can be minimised within acceptable limits, and the long-term impacts can be negligible. It also needs to be emphasised that, under appropriate environmental guidelines, the needs for exploration and development are not incompatible with other forms of land use, including agriculture and national parks.

The Committee on Natural Resources of the United Nations Economic and Social Commission has therefore concluded that the United Nations 'could make a major contribution to the long term management and sustainable development of mineral resources through developing a global knowledge base, at appropriate scales, of the potential for mineral resource exploration and development' (CNR 1994). The committee recognised that the program should be based on GIS technology, which would allow integration with other land management information.

The Committee on Natural Resources was aware that much of the information required is already being collected in many countries, and that a number of existing international organisations (both government and non-government) could help develop such global knowledge (see also Harrison 1993). The World Bank has also recognised the importance of such information for developing countries. It is believed, therefore, that a global knowledge base could be developed at relatively low cost by building on existing efforts of many institutions at the national and regional levels. The UN needs to define the global mission and provide the necessary coordinating mechanisms, including the development of globally consistent approaches to the definition of permissive tracts, and the assessment of resource potential.

Countries with well established geological surveys and mineral industries, and those with existing international minerals programs, can play leading roles in developing such a program at the regional level. Others, such as the countries of the former Soviet Union, need substantial assistance to ensure that information gathered in the past on a confidential basis is not irrevocably lost, but contributes to the global knowledge base.

Such global knowledge is clearly essential if sustainable consumption patterns are to be developed to take into account resource availability as well as environmental impacts.

Conclusion

The non-renewable nature of mineral (including fuel mineral) resources raises special problems in applying concepts of sustainable development.

Present global trends are towards increasing consumption and therefore increasing environmental impacts, in spite of significant efforts in many countries. This is largely because of global population growth and the requirement for improved standards of living in developing countries.

Growth in per capita consumption to levels currently enjoyed by the developed countries for a future global population of 10–12 billion is clearly not sustainable. If the desired global economic expansion by a factor of 5 or 10 is to be achieved in a sustainable way, then it must be decoupled as far as possible from increased input of materials and energy. 'We must devise models for a steady-state society, in which population size is in broad balance with the availability of resources' (Tickell 1991). In these circumstances, it is imperative that minerals issues be given prominence in the implementation of Agenda 21.

As noted above (**Availability of resources**), in the long term, it is hardly possible to predict how far technological advances, or specific scarcities, will lead to reduced demand (dematerialisation) or successful substitution, especially of energy, by renewable resources.

It is therefore impractical to apply the criterion of sustainability proposed by the World Bank economist Herman Daly, i.e. the rate of use of a non-renewable resource should be no greater than the rate at which a renewable resource, used sustainably, can be substituted for it (Daly 1990). Experience suggests that substitution will not occur on a major scale until a clear need has been established, so that changes in consumption patterns will be relatively abrupt.

The issue of sustainable consumption patterns needs to be considered both in terms of the capacity of the environment to absorb the impacts of resource use and the capacity to sustain the supply. The former is more readily assessed and the problems are relatively well-known, but impacts such as those of phosphate use in agriculture and of combustion of fuel minerals have proved difficult to control. In these cases the transition to more sustainable patterns faces severe political and economic problems.

In the case of metallic minerals, it is suggested that the UN could make a significant contribution in coordinating, assessing and disseminating technological knowledge of appropriate strategies for improving the efficiency of the industrial ecosystem and reducing environmental impacts (*The support square*, and *Per capita resource use*, above).

There are also significant gaps in knowledge needed for assessing pollution. A program to monitor the chemical health of the global land surface needs to be implemented. This will allow anthropogenic impacts to be assessed in the context of natural variation and will also assist in the delineation of permissive tracts.

In relation to supply, it is possible to determine the viability of consumption patterns out to a horizon of sustainability. This is only a few decades for demonstrated economic resources, but can be extended by assessment of undiscovered resources. The sustainability of supply can therefore be defined on a rolling basis in relation to this extended horizon of sustainability, which can also take into account changes in demand as a result of improved efficiency of use, recycling and substitution.

Given present trends it seems unlikely that the global consumption of resources can be stabilised, and eventually reduced, for some decades. It is likely, therefore, that there will be increasing difficulty in containing environmental impacts and in meeting the demand for minerals from the finite sources of supply currently regarded as economic or sub-economic. To help manage this situation during the period of transition to more sustainable consumption patterns, it is recommended that:

1. A global knowledge of the potential for mineral exploration and development (especially the identification of 'permissive tracts') should be developed, and integrated with other land information so that land-use planning can properly take into account national and global needs for mineral exploration and development.
2. Based on this knowledge, global estimates of undiscovered resources should be made, so that the horizon of sustainability can be extended as far as possible and maximum warning obtained of potential mineral shocks.

These steps are relatively non-controversial and low cost. Much of the information required is already being collected at national level. It would be appropriate for the United Nations to coordinate the collection of information to provide a global framework for policy formulation at both international and national levels. National geological surveys can make unique contributions both in relation to environmental impacts and in relation to resource availability.

Acknowledgments

This paper was prepared in the course of working on a strategy paper for the United Nations Committee on National Resources. Discussions with members of that committee are gratefully acknowledged. The paper has also benefited from comments by a number of individuals in Australian government and industry.

Helpful comments by the reviewers of the paper, Herman Daly and Don Singer, are also acknowledged with appreciation.

References

Arizpe, L., Costanza, R. & Lutz, W., 1992. Population and natural resource use. In: Dooge, J.C.I. et al. (editors), An agenda of science for environment and development into the twenty-first century. Cambridge University Press.

Bookout, J.F., 1989. Two centuries of fossil fuel energy. In: World Natural Resources Colloquium. 28th International Geological Congress, Washington DC.

BMR, 1984. BMR refines its mineral resources classification system. Australian Mineral Industries Quarterly, 36(3), 73–82.

Cloud, P., 1977. Entropy materials and posterity. Geologische Rundschau 66(3), 678–696.

CNR, 1994. United Nations Committee on Natural Resources, Report on the Second Session (22 February–4 March 1994). Economic and Social Council; Official Records, 1994; Supplement no. 6 (E/1994/26; E/C.7/1994/13).

Cook, P.J., 1995. Societal trends and their impact on the coastal zone and adjacent seas. Proceedings of the International Conference 'Coastal Change 95', Bordomer–IOC, Bordeaux, 876–891.

Daly, H.E., 1977. Steady state economics. W.H. Freeman, San Francisco, 185 pp.

Daly, H.E., 1990. Towards some operational principles of sustainable development. Ecological Economics, 2, 1–6.

Daly, H.E. & Cobb, J.B.Jnr., 1989. For the common good: redirecting the economy toward community, the environment, and a sustainable future. Beacon, Boston.

Darnley, A.G. et al., 1995. A global geochemical database for environmental and resource management: recommendations for international geochemical mapping; Final Report of IGCP Project 259. Earth Sciences 19, UNESCO Publishing.

Dorian, J.P. & Zwartendyk, J., 1984. Resource assessment methodologies and applications. Materials in Society, 8(4), 659–679.

Emberger, A . 1993. Geological mapping and mineral maps. In: McDivitt, J.F. (editor), International mineral development source book. Forum for International Mineral Development, Colorado School of Mines.

Foster, R.J., 1993. Energy, economics, environment and politics. OPEC Review, Autumn 1993, 293–303.

Fuller, J., 1993. The oil industry today. The British Association Lectures. The Geological Society Publishing House, Bath, U.K.

Harris, D.P. et al., 1993. Evaluation of the United States

Geological Survey's three step assessment methodology. a research report to the USGS under contract 1434-92-C-30041. USGS Open File Report 93-258-B.

Harrison, J.M., 1993. International organisations assisting mineral development. In: McDivitt, J.F. (editor), International mineral development source book. Forum for International Mineral Development, Colorado School of Mines.

Hinterberger, F., Kranendonk, S., Welfens, M.J. & Schmidt-Bleek, F., 1994. Increasing resource productivity through eco-efficient services. Wuppertal Papers no. 13, May 1994. Wuppertal Institute.

Inter-governmental Panel on Climate Change, 1994. Radiative forcing of climate change. The 1994 report of the Scientific Assessment Group of IPPC: summary for policy makers. WMO & UNEP, Geneva.

International Strategic Minerals Inventory, 1984. Summary report: phosphate. US Geological Survey Circular, 930-C.

Kelly, A., 1990. The future of metals. Minerals Industry International, 996, 5–15.

McCammon, R.A. & Briskey, J.A.Jr., 1992. A proposed national mineral resource assessment. Nonrenewable Resources 1(4), 259–266.

Masters, C.D., Root, D.H., & Attanasi, E.D., 1991. Resource constraints in petroleum production potential. Science 253, 146–152.

Meadows, D.H., Meadows, D.L., Randers, J. & Behrens, W.W. III, 1972. The limits to growth: a report for the Club of Rome's project on the predicament of mankind. Universe Books, New York/Earth Island, London.

Meadows, D.H., Meadows, D.L. & Randers, J., 1992. Beyond the limits: global collapse or a sustainable future. Earthscan Publications Ltd, London.

Mineral Resources and Sustainable Development: a workshop, 1994. British Geological Survey, Technical Report WF/94/12.

Northolt, A.J.G., Sheldon, R.P. & Davidson, D.F. (editors), 1989. Phosphate deposits of the world, vol. 2: Phosphate rock resources. Cambridge University Press, 600 pp.

Petit-Maire, N., 1995. Plus 1°C? It happened before. Science International. ICSU Newsletter, 59, 13–18.

Phillips, W.G.B., 1987. Factors affecting the long-term availability of bulk minerals for the construction industry. In: McLaren, D.J. & Skinner, B.J. (editors.), Resources and world development (Report of Dahlem Workshop). John Wiley and Sons, 327–345.

Schanz, J.J., 1980. The United Nations endeavour to standardise mineral resource classification. Natural Resources Forum, 4(3), 307–313.

Singer, D.A., 1993. Basic concepts in three-part quantitative assessments of undiscovered mineral resources. Nonrenewable Resources 2(1), 69–81.

Skinner, B.J., 1989. Resources in the 21st century: can supplies meet needs? In: World Natural Resources Colloquium. 28th International Geological Congress, Washington D.C.

Steering Committee of the Climate Change Study, 1995. Steering Committee of the Climate Change Study. Australian Academy of Technological Sciences and Engineering, February 1995.

Tickell, Sir Crispin, 1991. What we must do to save the planet. New Scientist, 7 September, 2.

Tickell, Sir Crispin, 1994. The future and its consequences. The British Association Lectures 1993, 20–24.

UNCED, 1992. Report of the United Nations Conference on Environment and Development, United Nations, New York, 1993.

USBM, 1995. Mineral commodity summaries 1995. United States Department of the Interior, Bureau of Mines.

US Congress, 1985. Strategic materials: technologies to reduce US import vulnerability. Office of Technology Assessment, OTA-ITE-248, Washington, DC.

von Engelhardt, W., Goguel, J., Hubbert, M.K., Prentice, J.E., Price, R.A. & Trumpy, R., 1976. Earth resources, time and man — a geoscience perspective. Environmental Geology, 1, 193–206.

WCED, 1987. Our common future. World Commission on Environment and Development. Oxford University Press.

Wellmer, F-W. & Kürsten, M., 1992. International perspective on mineral resources. Episodes, September, 182–194.

Wigley, T.M.L., Richels, R. & Edmonds, J.A. 1996. Economic and environmental choices in the stabilisation of atmospheric CO_2 concentrations. Nature, 379, 240–243.

Wyborn, L.A.I., Jagodzinski, E.A., Jacobsen, G. & Needham, R.S., 1996. Environmental geochemistry in the southern Kakadu region and the 'sickness country' concept: the need for baseline geochemical studies. Geological Society of Australia, Abstracts no. 41, 483.

AGSO Journal of Australian Geology & Geophysics, 17(1), 27–31

The origin of the Earth

Stuart Ross Taylor[1]

It is not possible to consider the formation of the Earth in isolation without reference to the formation of the rest of the solar system. A brief account is given of the current scientific consensus on that topic, explaining the origin of an inner solar system rocky planet depleted in most of the gaseous and icy components of the original solar nebula. Volatile element depletion occurred at a very early stage in the nebula, and was probably responsible for the formation of Jupiter before that of the inner planets. The Earth formed subsequently from accumulation of a hierarchy of planetesimals. Evidence of these remains in the ancient cratered surfaces and the obliquities (tilts) of most planets. Earth melting occurred during this process, as well as from the giant Moon-forming impact. The strange density and chemistry of the Moon are consistent with an origin from the mantle of the impactor. Core-mantle separation on the Earth was coeval with accretion. Some speculations are given on the origin of the hydrosphere.

The relation of the Earth to the solar system

Although the Earth is unique, it is not possible to discuss the origin of the Earth separately from that of the other 'terrestrial' planets, or from that of the entire solar system. The presence of its unique satellite, the Moon, must also be accounted for. The planets are usually divided into three major groups: the small terrestrial planets (Mercury, Venus, Earth, Mars), the gas giants Jupiter and Saturn, and the smaller ice giants Uranus and Neptune. Pluto is only called a planet by courtesy. It is one of the larger icy bodies from the Kuiper Belt, is only 20% of the mass of the Moon, and is similar to Neptune's satellite, Triton.

The inner solar system with its small rocky planets is distinct from the region of the giant planets that dominate the outer reaches of the system. A basic reason for this division was the depletion of volatile elements and loss of gaseous elements in the inner nebula. This was probably associated with violent solar activity in the earliest stages of nebular evolution. Jupiter and Saturn formed while the gas was still present and at a significantly earlier stage than the inner planets. The terrestrial planets accreted later from left-over planetesimals after the gaseous components of the nebula had been dissipated.

The solar nebula

The planets all rotate around the sun anticlockwise when viewed from above and all lie close to the Earth–Sun or ecliptic plane. This pattern is due to the formation of the Sun and planets from a spinning disk of dust and gas, the solar nebula, a concept that was formulated by the French scientist Pierre-Simon, Marquis de Laplace (1749–1827), about 200 years ago.

The solar system began to form about 4570 million years ago from the solar nebula. This is the age of the refractory inclusions in some meteorites, which appear to be the oldest objects that formed in the nebula. The disk of dust and gas had become separated as a fragment from a larger molecular cloud in a spiral arm of the galaxy about 12–15 billion years after the Big Bang.

Following its separation as a fragment of a molecular cloud, the primitive solar nebula consisted mainly (98%) of gas (H and He) with about 2% of heavier elements, which are divided into 'ices' (water, ammonia, methane, about 1.5%) and 'rock' (about 0.5%) (Levy & Lunine 1993). We are well informed about the composition, for the non-gaseous elements, of the primordial solar nebula. This is because of the close correspondence between the abundances of the non-gaseous chemical elements in the solar photosphere, and in the CI chondritic meteorites.

[1] Department of Nuclear Physics, Research School of Physical Sciences, Australian National University, Canberra, ACT 0200, Australia

Volatile element depletion in the inner nebula and the formation of Jupiter

The inner solar system is depleted not only in gas, but also in elements that are volatile below about 1200K (Taylor 1992). This depletion is well illustrated by the abundance of potassium (a moderately volatile element) compared with that of uranium (a refractory element). Both these elements are gamma-ray emitters, which means that geochemical measurements can be made for Venus, Mars and the Moon as well as the Earth (Fig. 1). The Venusian data come from gamma-ray measurements made by the Russian Venera and Vega landers, the Martian data from the meteorites from that planet that have landed on Earth.

Potassium and uranium are distinctly different in chemical properties, ionic radius and valency. However, both elements are concentrated in residual melts during crystallisation of

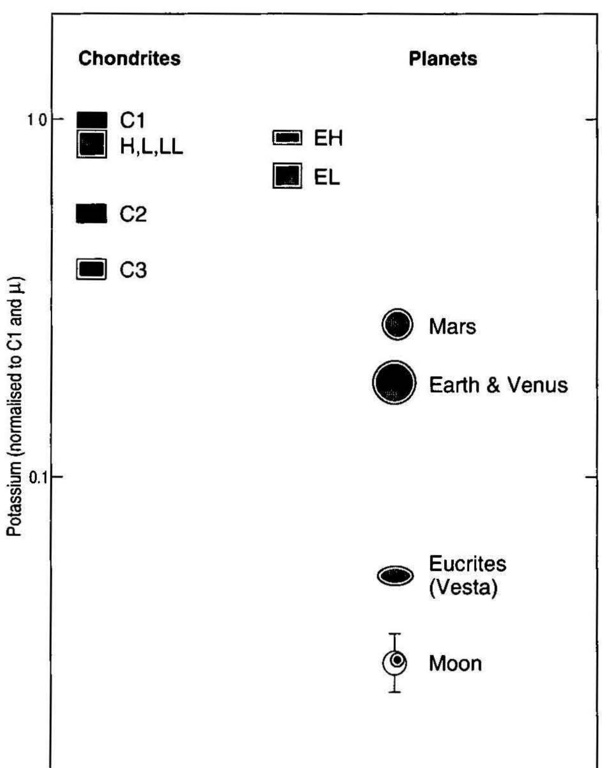


Figure 1. Depletion of volatile elements, here represented by potassium, relative to refractory elements (e.g. uranium) is widespread in the inner solar system. Most classes of meteorites have higher K/U ratios than the planets. The Moon is more highly depleted in K and other volatile elements than the terrestrial planets (after Humayun & Clayton, 1995).

basaltic silicate melts, since they are both excluded from the common rock-forming minerals in basalts (i.e. they are 'incompatible' elements). Thus they tend to preserve their bulk planetary ratios during planetary differentiation. It has occasionally been suggested that potassium could behave as a metal at high pressures and so be buried in the metallic planetary cores. However, although potassium is depleted on Mars, the pressure at the centre of Mars (r=3390 km; 400 kb) is insufficient to allow potassium to enter a Martian core. In addition to potassium, many other volatile elements are also depleted in the Earth relative to primitive nebular abundances. Most of these elements have chemical properties which make it unlikely that they would enter into metallic phases.

Perhaps potassium, which is a moderately volatile element, could have been boiled off during a high temperature stage of planetary accretion. However, it turns out that elements of the atomic weight of potassium cannot be lost from the terrestrial planets once these bodies have reached their present size. The K/U ratio should vary with planetary size, if they were boiled off in some manner during accretion. This is not observed. Neither are there any signs of isotopic fractionation that would occur in such a process (Humayun & Clayton 1995).

The Rb/Sr isotopic systematics show that the Earth is depleted in volatile rubidium relative to refractory strontium. Rubidium has closely similar properties to potassium, so that it is unlikely that either element is present in the mantle in its primordial solar nebular abundance. It is a clear conclusion that, in common with the other volatile elements, rubidium was depleted in the precursor material from which the Earth accreted.

The time of volatile depletion in the inner nebula is given by the ages of the meteorites. The Pb-U and Rb-Sr ages give the time of separation and depletion of volatile lead and rubidium relative to refractory uranium and strontium from the primordial solar nebula values. This depletion in volatile elements occurred before 4566 m.y. The most likely explanation is that when the growing sun reached critical mass and thermonuclear burning began, violent T Tauri and FU Orionis activity swept out the gas and uncondensed elements from the inner nebula. Metre-sized planetesimals survived and these subsequently accreted into the terrestrial planets. Water condensed as ice in the nebula at 160K at what is termed a 'snow line' at 4–5 A.U. (Astronomical Unit, the mean distance between the Earth and the Sun, 1.496 10^8 km) (Taylor 1992; Levy & Lunine 1993).

A massive core of about 15 Earth masses was able to form within about a million years due to this pile-up of water ice at 5 A.U. This core was sufficiently massive to trap hydrogen and helium by gravitational attraction and so Jupiter grew rapidly. The early formation of Jupiter had profound consequences. It depleted the asteroid belt and pumped up the orbital inclinations and eccentricities of the remaining asteroids so that they were unable to collect themselves into a planet The region where Mars would later accumulate was starved. Mars is only 1/3000 as massive as Jupiter. Within the inner nebula, only bodies large enough (metre size) to survive the early intense heating episodes from the early Sun were left.

The accretion of the Earth from planetesimals

Earlier views that the Earth and the inner planets accreted from fine dust have been discarded. Such a process of planetary formation would lead either to planets of uniform or smoothly varying composition. Obliquities would be expected to be zero and rotation rates likewise uniform and possibly very low or zero.

The former presence of a hierarchy of planetesimals is indicated by several lines of evidence. Direct evidence for the previous existence of bodies up to 100 km in diameter comes from the observation that ancient surfaces on planets and satellites are saturated with craters. From Mercury, close to the Sun, out to the satellites of Uranus, a massive bombardment struck planets and satellites. The lunar surface is the classic example. Craters, from micron-sized pits due to impact of tiny grains on lunar samples, up to giant ringed basins over 1000 km in diameter, are present (Fig. 2). The planetesimals ranged in size from a few metres up to Mars-sized objects. The asteroids are our best current examples (Fig. 3) (Binzel et al. 1990).

Figure 2. Mare Orientale, 900 km in diameter, is a type example of a multi-ring basin formed by the impact of a planetesimal or asteroid perhaps 100–200 km in diameter. The concentric rings of mountains were formed within a few minutes, 3800 million years ago (courtesy NASA Lunar Orbiter IV 187M).

Were there larger intermediate-sized bodies (Moon–Mercury–Mars size) in the hierarchy of objects which accreted to form the terrestrial planets? The major piece of evidence for the presence in the early solar nebula of very large objects (of lunar, Mars and Earth-sized masses) comes from the obliquity or inclination of the planets to their axis of rotation. A body the size of the Earth crashing into the planet would be needed to tip Uranus through 90°. Although smaller collisions are required to account for the tilt of the other planets, objects at least as large as Mars (1/10 Earth mass) must have been responsible, since the impacts of a multitude of smaller (Phobos-size) bodies will average out.

How many of these very large objects were there? Computer simulations of the accretion process in the inner solar system show that about 100 moon-size bodies, 10 Mercury-size and 3 to 5 Mars-size bodies would have formed the final population of planetesimals existing just before the final sweep-up into

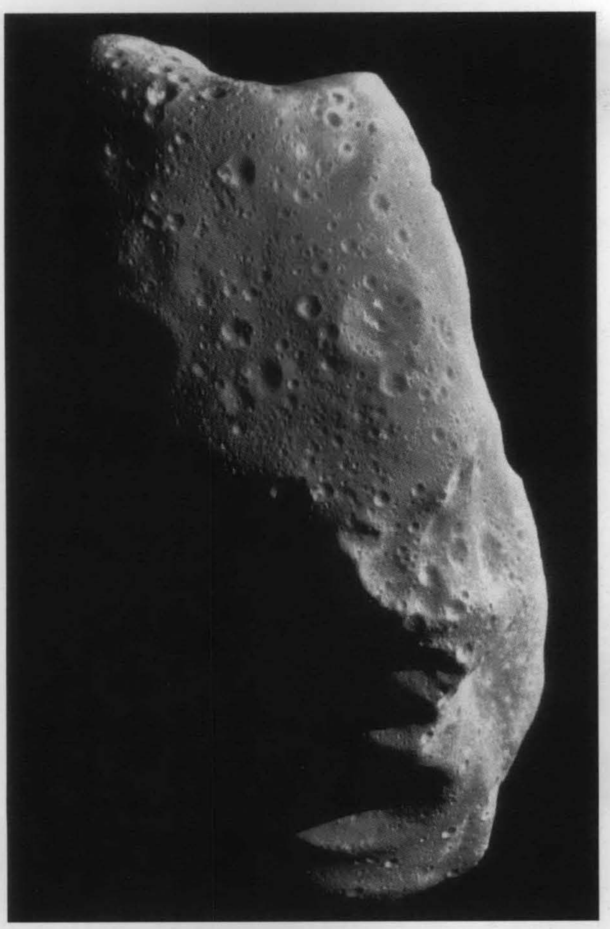


Figure 3. An analogue for a small planetesimal. The S-class asteroid 243 Ida, $56 \times 24 \times 21$ km. Its density of 2.6 ± 0.5 g/cm^3 is consistent with a bulk chondritic composition and a porous structure.

Venus and the Earth. Mars, only 1/10 of the mass of Earth, and Mercury, only about 1/20 Earth mass, are survivors of this population.

A large impact is probably responsible for the strange facts that Mercury has such a high density and a small rocky mantle. The current view of Mercury is that a body about 1/6 of its mass struck Mercury at a late stage in the accretion of the planet. The collision disrupted the planet: most of the mantle silicate was lost to space, but the iron core clumped together with a smaller silicate mantle. The origin of the Moon also involves a collision of an object larger than Mars with the Earth, resulting this time in the production of a low density satellite. From this discussion, it is apparent that there is ample evidence for the existence of large precursor bodies or planetesimals in the early solar system.

The Earth and the other inner planets formed through the accretion of those planetesimals which were left in the inner nebula after the gas was swept away. At an early stage in solar nebular evolution, the dust in the rotating disk of the solar nebula began to clump together, beginning with grains and proceeding through metre-sized lumps to objects of kilometre size, finally reaching dimensions of hundreds to thousands of kilometres during the final stage before planetary accretion. The larger planetesimals were probably melted early in solar system history and differentiated into metallic cores and silicate mantles. Based on evidence from meteorites, even some relatively small planetesimals underwent internal differentiation into metallic cores and silicate mantles quite early in their history. The larger planetesimals almost certainly had already gone through at least one intraplanetary melting episode, with core formation occurring before they were accreted by the inner planets. Such bodies may have been broken up by collisions and reaccreted in differing proportions of metal and silicate fractions, so that much diversity of composition among the accreting bodies can be expected (e.g. Newsom & Jones 1990). Possibly, some undifferentiated planetesimals were added late in the accretionary sequence. Such a 'late veneer' could account for the excess siderophile elements in the upper mantle, as well as adding water.

The terrestrial planets collected somewhat differing populations of planetesimals and differ in composition among themselves. The absence of a planet in the asteroid belt, in which over 4000 small bodies have been labelled, is due to the influence of massive Jupiter, which swept up, or ejected many of the bodies. As noted earlier, the total mass of the many small objects in the belt is less than 5% of the mass of the Moon. The small size of Mars is due to a similar cause: starvation caused by massive Jupiter, which formed earlier and depleted the neighbourhood.

Over 120 asteroids (the Apollos) perturbed from the asteroid belt are in Earth-crossing or Earth-approaching orbits. Once in these orbits, they have lifetimes of only about 100 m.y. Somewhere between 100 and 1000 tons of meteoritic material, mainly as dust, falls on the Earth each day. Every few million years an asteroid large enough to form a 20 km diameter crater hits the Earth. The extinction of 70% of species, including all the dinosaurs at the end of the Cretaceous Period, 65 million years ago, was most probably due to the impact of an asteroid some 10 km in diameter. The evidence for the collision includes a worldwide 'spike' in iridium (rare in the Earth's crust, but more abundant in meteorites), quartz grains shocked by pressures of hundreds of kilobars, and soot from massive fires, at the exact Cretaceous–Tertiary boundary. No internal single geological process can account for these facts.

Although the inner planets are chondritic in a broad sense, it does not appear possible to construct the Earth and the other terrestrial planets out of the building blocks supplied by the currently sampled population of meteorites. There are many differences in detail between the present population of meteorites and the compositions of the terrestrial planets. These include volatile/refractory element ratios such as K/U, oxygen isotopes, noble gases, and density. These rule out any of the known meteorite classes as potential candidates for the source material for the inner planets. If they are providing us with an adequate sample of the inner asteroid belt, then there were substantial differences between that area and the zone sunwards of about 2 A.U. The most significant difference appears to have been a generally greater depletion of the volatile elements in that region in which the terrestrial planets accumulated (Wasson 1985; Kerridge & Matthews 1988). As always, there are exceptions to this tidy scheme. Thus the eucrites derived from the asteroid Vesta are also depleted in volatile elements.

The asteroid belt was depleted by the early formation of Jupiter, so that very little material was left in that location. This makes the belt a poor quarry from which to build the terrestrial planets. The current view is that the Earth and the other terrestrial planets accreted from a hierarchy of planetesimals of varying sizes, a process taking perhaps 50 to 100 million years. The noble gases (He, Ne, Ar, Kr, and Xe) and hydrogen are strongly depleted in the Earth relative to solar abundances. By the time the Earth, Venus, Mars and Mercury accreted, the gaseous components of the nebula were long gone. The hydrogen and helium gas in the nebula is swept away on time scales of 3–10 million years. The cause of this early volatile loss in the inner portions of the solar nebula appears to be connected with intense solar activity (T Tauri and FU Orionis stages) around the time that the early Sun joined the main sequence stage of stellar evolution.

Because of the differences in planetary compositions, there appears to have been little lateral mixing in the nebula. The terrestrial planets either accumulated from rather narrow (perhaps <0.5 A.U.) concentric zones in the solar nebula or from different populations of planetesimals. The present asteroid belt has a zoned structure, ranging from apparently differentiated objects in the inner belt, to apparently primitive ones which dominate in the outer reaches. Although the zones in the asteroid belt have been broadened through time by collisions, they may be an analogue for the original structure of the nebula.

Formation of the Moon: consequences for the Earth

The Earth's Moon is a unique satellite; the satellites of the other outer planets are mainly rock–ice mixtures, formed by accretion around their parent planets, or by subsequent capture. None of the other terrestrial planets except Mars possesses moons, but Phobos and Deimos, the tiny Martian moons, are probably captured asteroids. The lunar orbit is neither in the equatorial plane of the Earth nor in the plane of the ecliptic, but is inclined at 5.1° to the latter. The Moon has a high mass relative to that of its primary planet, compared with the satellites of the giant planets. The bulk density of the Moon (3.34 g/cm^3) is much less than that of the Earth (5.54 g/cm^3) or of the other inner planets. It is attributable to a low metallic iron content. The angular momentum of the Earth–Moon pair is anomalously high compared with that of the other inner planets. Some event or process spun up the system, although it is not rotating rapidly enough for classical fission to occur.

The Moon has an unusual composition by either cosmic or terrestrial standards. It is strongly depleted in volatile elements (eg. K, Pb, Bi) by a factor of about 50 compared to the Earth, or 200 relative to primordial solar nebula abundances, and is enriched in refractory elements (e.g. Ca, Al, Ti, U) by about a factor of 1.5 compared with the Earth. This has been confirmed recently by the data from the Clementine Mission (Lucey et al. 1995). The bulk lunar composition contains about 50% more FeO than current estimates of the terrestrial mantle. The Moon is bone dry, no indigenous H_2O having been detected at ppb levels (Taylor 1982).

No previous theories of lunar origin (capture, fission or double planet) survived the encounter with the Apollo sample data. A giant collision is now thought to be the most likely explanation for the origin of the Moon. The theory proposes that during the final stages of accretion of the terrestrial planets, and when the Earth was close to its present size, it suffered a grazing impact, at about 5 km/sec, with an object about 0.14 Earth masses (i.e. over 30% larger than Mars). Both this body and the Earth are assumed to have differentiated at that stage into a metallic core and silicate mantle. The collision disrupted the impacting body, much of which went into orbit about the Earth. Gravitational torques due to the asymmetrical shape of the Earth after the impact were sufficient to accelerate material into orbit. The core of the impacting body accreted to the Earth within about 4 hours. Most of the metal core ended up in the Earth, with the metal penetrating the mantle and ending up wrapped about the Earth's core (Taylor 1987).

The material out in lunar orbit either immediately coalesced to form a totally molten Moon, or broke up into several moonlets that subsequently accreted to form a partly molten Moon. At least half the Moon was molten shortly after accretion, with the feldspathic highland crust crystallising from this 'magma ocean'. The impact event was sufficiently energetic to vaporise much of the material which subsequently recondensed to make up the Moon. This explains such unique geochemical features as the bone dry nature of the Moon, the extreme depletion of very volatile elements and the enrichment of refractory elements in the Moon.

What were the implications for the Earth of the single impact origin of the Moon? The event probably triggered or enhanced complete mantle melting. However, the accretion of the Earth from large planetesimals, rather than fine dust, virtually guaranteed planetary melting, since the heat dumped in by such events cannot readily be lost.

Core, mantle, hydrosphere and atmosphere

The current scenario is that separation of the metallic core from the silicate mantle appears to have occurred promptly following the accretion of the Earth. Possibly, core formation was effectively coincident with accretion. No early sialic crust formed. Since the terrestrial planets accreted from planetesimals which were already mostly differentiated into metallic and silicate phases, little further reaction between metal and silicate may have occurred once these bodies accreted to the Earth. Possibly, the iron cores in the planetesimals were still molten

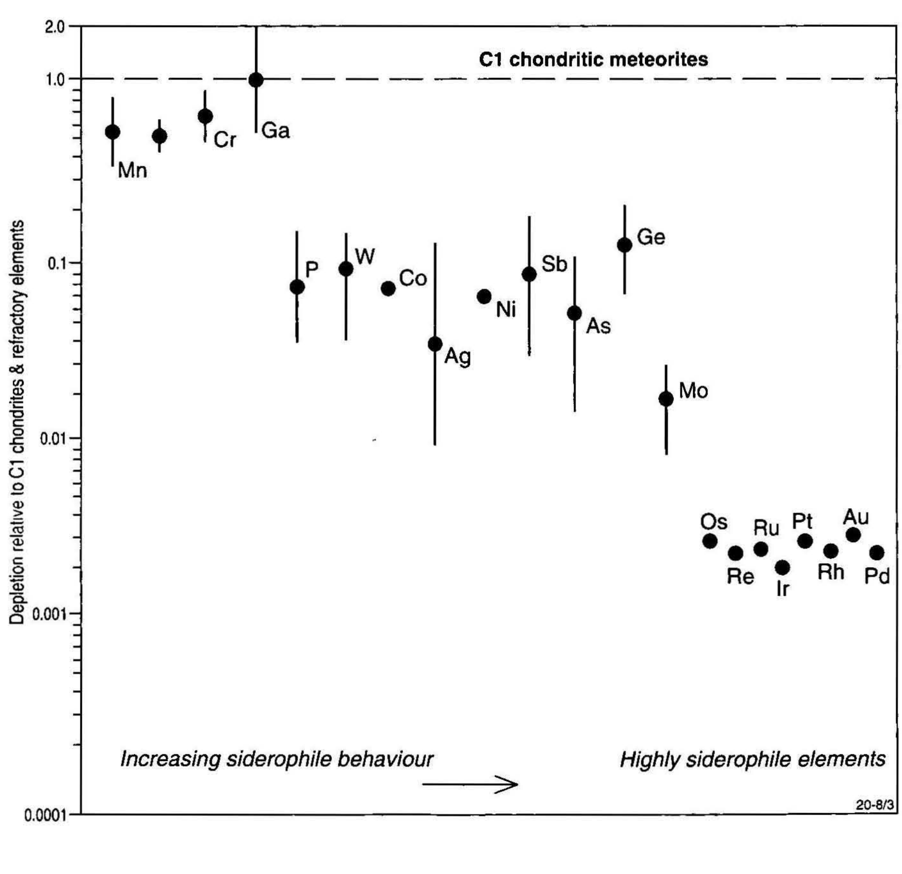


Figure 4. The abundance of siderophile elements in the upper mantle of the Earth, and in the Moon, normalised relative to primitive CI concentrations. The relatively high levels of Ni and Co, and the uniform 'chondritic-type' abundance pattern of the highly siderophile elements, indicate a lack of equilibration between the silicate mantle and the metallic iron core (courtesy of H.E. Newsom).

when they accreted to the Earth. In this model, core formation is effectively instantaneous and coeval with accretion, rather than occurring over hundreds of millions of years. However, melting of large planetesimals and their accretion to the Earth is likely to occur over a period of 50 to 100 million years, and separation of metal may have been delayed for such a period. This would be consistent with recent data from the decay of ^{182}Hf to ^{182}W that suggests such a time interval (Lee et al. 1995).

In addition, Re, Au, Ni, Co and the platinum group elements (Ru, Rh, Pd, Os, Ir, Pt) are highly abundant in the upper mantle of the Earth (Fig. 4). Thus the present upper mantle did not achieve equilibrium with the core. The distribution of platinum group elements appears to be rather uniform, and they are present in approximately primordial ratios, although depleted relative to iron. Their abundances are determined from samples only from the top 100 to 200 km of the mantle, so the composition of the lower mantle is essentially unknown.

A common explanation for the overabundance of platinum group elements in the upper mantle is that they result from late accretion of planetesimals rich in platinum group elements. A similar scenario of late accretion of volatile-rich planetesimals is commonly invoked to account for the volatile (e.g. H_2O) inventory of the Earth. A cometary influx could be an equally viable source. Another possibility is that the present abundance of highly siderophile elements in the Earth's mantle is due to addition of a small portion of the core of the lunar-forming impactor; the rest of the impactor's core accreted to the Earth's core without significant interaction with the Earth's mantle.

The metallic Fe/Ni outer core of the Earth contains about 10% of a light element. If high-pressure core–mantle equilibrium was not attained in the early Earth, then it seems unlikely that oxygen entered the core, since this requires megabar pressures, as is the case for potassium. Sulphur then becomes a possible candidate for the light element in the Earth's core. However, sulphur is a volatile element and the terrestrial budget may be inadequate. Other possible candidates are silicon and carbon.

Large collisions in the final stages of accretion are likely to have removed any primitive atmosphere which might have formed. The Earth accreted long after the hydrogen and helium, the principal components of the primordial solar nebula, had been dissipated and there is no sign of any primitive atmosphere. The present atmosphere and hydrosphere of the Earth appear to be entirely secondary in origin, formed by degassing from the interior or by late accretion from comets and asteroids from beyond Mars.

The isotopic composition of the noble gases helium, argon and xenon have provided crucial evidence that there was a sudden early degassing or outgassing event. Early extensive degassing is indicated by the noble gas data, which indicate that the mean atmospheric age is greater than 4 billion years. Thus most of the primitive volatiles were degassed from the mantle in the first half billion years after accretion, before there was significant addition to the mantle of ^{40}Ar from the decay of radiogenic ^{40}K. The xenon data indicate that up to 80% of the degassing occurred within about 50 million years following accretion. This early rapid degassing would be consistent with a molten mantle, that resulted both from the accretion of large planetesimals and from the formation of the Moon by a massive collision with the Earth.

Water was about twice as abundant as 'rock' in the primordial solar nebula. Since water either was never condensed or was lost along with the other volatiles in the early solar heating event, little water was available in the zone from which the Earth and the other inner planets were formed. Thus the source of water now in the Earth is an interesting problem. However, the total water content of the Earth is probably less than 500 ppm, an amount so small in comparison with the abundance of water in the early nebula that it could be ignored to a first approximation, except for our total dependence on it. Some water, perhaps present in hydrated minerals in already formed planetesimals, might have survived the early intense heating which drove the volatiles out of the inner solar system. Most probably came as a late-accreting veneer from beyond Mars since the minerals in the early nebula appear, from the meteorite evidence, to have been anhydrous. The late veneer that supplied the excess platinum group elements to the upper mantle is one possible source.

Water-ice is expected to be a stable phase in the nebula only at temperatures below 160K at nebula pressures. This means that water ice will occur only beyond 4–5 A.U. from the sun, in the outer reaches of the asteroid belt. This is consistent with the observation that icy satellites are restricted to the region of the giant planets. Carbonaceous chondrites, probably typical of asteroid compositions beyond about 3 A.U. contain up to 20% water by weight. Thus most of the terrestrial water was possibly derived from planetesimals or comets from beyond Mars, perhaps late in the accretional history of the Earth. If comets comprised 10% of the bodies responsible for the bombardment between 4.4 and 3.8 billion years they could supply the appropriate amount of water for the terrestrial oceans. Such a model is not without problems, since comets impact at high velocity, and so may remove earlier atmospheres and hydrospheres.

References

Binzel, R.P., Gehrels, T. & Matthews, M.S., 1990. Asteroids II. University of Arizona Press, Tucson.

Humayun, M. & Clayton, R.N., 1995. Potassium isotope chemistry: genetic implications of volatile element depletion. Geochimica et Cosmochimica Acta, 59, 2131–2148.

Kerridge, J.F. & Matthews, M.S., 1988. Meteorites and the early solar system. University of Arizona Press, Tucson, 1269 pp.

Lee, D.-C. & Halliday, A.N., 1995. ^{182}Hf-^{182}W chronometry, terrestrial core formation and early solar system evolution. EOS, 76, F338 (abstract).

Levy, E.H. & Lunine, J.I. 1993. Protostars and planets III. University of Arizona Press, 1293 pp.

Lucey, P.G., Taylor, G.J. & Malaret, E., 1995. Abundance and distribution of iron on the Moon. Science, 268, 1150–1153.

Newsom, H.E. & Jones, J.H., 1990. Origin of the Earth. Oxford University Press, New York, 378 pp.

Taylor, S.R., 1982. Planetary science: a lunar perspective. Lunar and Planetary Institute, Houston, Texas, 481 pp.

Taylor, S.R., 1987. The origin of the Moon. American Scientist 75, 469–477.

Taylor, S.R., 1992. Solar system evolution: a new perspective. Cambridge University Press, 307 pp.

Wasson, J.T., 1985. Meteorites: their record of solar-system history, Freeman, NY, 267 pp.

AGSO Journal of Australian Geology & Geophysics, 17(1), 33–48

Earth's evolution and mineral resources, with particular emphasis on volcanic-hosted massive sulphide deposits and banded iron formations

Michael Solomon[1] & Shen-su Sun[2]

Oh, never let us never, never doubt
What nobody is sure about
The Microbe (Hilaire Belloc)

From 3.5 Ga the overall uniformity of the composition and character of sedimentary and mantle- and crustal-derived igneous rocks, and the persistence of ore types related to convergent continental margins (e.g. volcanic-hosted massive sulphide, porphyry Cu-Mo, Sn-W and lode Au), indicates continuity of tectonic processes broadly similar to the present. Secular variations in abundance of these ores reflect tectonic cycles having periods of perhaps several hundred million years, but other ores reflect long-term changes in tectonic style and the composition of the ocean and atmosphere. Thus the development of extensive continental margins and depositional basins from about 3.0 Ga heralded development of giant Au-U conglomerate and banded iron formation deposits, with at least the former suppressed in the Palaeoproterozoic as a result of increased atmospheric P_{O_2}. Aggregation of large continents and the development of extensive basins allowed formation of giant Cu-Co, Pb-Zn and U-platinum group elements-Au ores from about 2.0 Ga. There is no firm evidence that mantle heterogeneity has contributed to ore distribution, nor that Archaean crust or mantle was anomalously enriched in Au or platinum group elements.

Mineralogical and isotopic data from volcanic-hosted massive sulphide deposits support the hypothesis that there was abundant sulphate in deep oceans from 3.5 Ga, and the lack of Pb and barite in Late Archaean ores may be related to the mafic-rich composition of the local shallow crust, steep thermal gradients, anoxic, sulphate basin waters or the compositions of particular magma types. Both the ocean margin (Holland 1973) and the hydrothermal plume/gravity current models for the origin of banded iron formations are broadly compatible with their composition, secular distribution and timing with respect to glaciation, ocean anoxia and tectonic activity. In the gravity current model the possible impact of the hydrothermal activity on climate warrants further investigation.

Introduction

The origins of many of Earth's mineral resources are closely linked to the physical and chemical evolution of the planet and the life it supports, and particularly its atmosphere and hydrosphere. Many mineral deposits are sensitive signatures of particular tectonic and magmatic environments so that there are periods in Earth's history that favour specific ore types. In this paper we focus on those deposits that by their distribution patterns and composition contribute to the debate on the evolution of the Earth, and particularly volcanic-hosted massive sulphide deposits and Fe and Mn ores related to banded iron formations. Recent reviews of mineral deposit distribution include Meyer (1985, 1988), Hutchinson (1992), Barley & Groves (1992), Lambert et al. (1992) and Kirkham & Roscoe (1993).

Our time scale to the beginning of the Proterozoic is as follows: Hadean: 4.5 to 3.9 billion years (Ga); Early Archaean: 3.9 to 3.5 Ga; Middle Archaean: 3.5 to 3.0 Ga; Late Archaean: 3.0 to 2.5 Ga.

The evolving Earth

The Hadean

The possibility that there are 'enriched' mantle sources favourable for some styles of mineralisation raises questions on secular and spatial variations in abundance of the ore-forming elements in the mantle and crust, and particularly the possibility of mantle heterogeneity generated during Earth's accretionary stage (see discussion by Taylor, and Taylor & McLennan in this issue on the accretional history of Earth and evolution and composition of the continental crust). Core formation was probably contemporaneous with the accretion of planetesimals at ~4.5 Ga as a result of temperature increase and melting caused by accretional energy and heat generated by short-lived nuclides (Stevenson 1983). During core formation highly siderophile elements, such as platinum group elements and Au, were effectively concentrated into the core. After the completion of core formation, addition of a small amount (<1 mass %) of oxidised accretional 'veneer' introduced the currently observed highly siderophile elements into Earth's mantle (Kimura et al. 1974, McDonough & Sun 1995, Taylor this issue), and it is possible that this process led to heterogeneous distributions of platinum group elements and Au in the mantle. In addition, core-mantle interaction and rising of their boundary layer may have introduced some platinum group elements and Au, a process especially important in Hadean time (McDonough, Harvard University, pers. comm. 1994), and Tredoux et al. (1989) and McDonald et al. (1995) speculated that such enrichment may have occurred in the Archaean through underplating of mantle plume-derived and subducted lithosphere.

During Hadean time a circum-global magma ocean may have existed in the upper mantle and its evolution and final solidification could have resulted in mineralogical and chemical layering of the mantle, i.e. large scale mantle heterogeneity. However, this would probably have been counterbalanced by vigorous mantle convection. Although debate continues, available geophysical and geochemical observations favour a 'pyrolite' type mantle without major differences in composition between upper and lower mantle (McDonough & Sun 1995).

The Archaean

Judging from dates close to 4 Ga on gneisses in Canada, China and the USA, and of detrital zircons aged 4.1–4.2 Ga from Western Australia, the crust, atmosphere and hydrosphere may be several hundred million years older than the 3.87 Ga rocks of the Isua Supercrustal Belt. The overall uniformity of both chemical composition and range of types of sedimentary and igneous rocks from 3.87 Ga points strongly to the continuity of tectonic processes broadly similar to those of Recent Earth (Lowe 1992a, Lowe & Ernst 1992, Windley 1992, 1993). A survey of the abundance ratios between Ti (a lithophile element) and Pd (a chalcophile element) in S-undersaturated, high temperature, mantle-derived melts, including komatiites and picrites ranging in age from 3.4 Ga to ~90 million years (Ma) and fertile mantle peridotites, shows near constant values of $2\text{–}3.10^5$ (Sun et al. 1991, McDonough & Sun 1995; Fig. 1).

[1] Centre for Ore Deposit and Exploration Studies, University of Tasmania, Hobart, Tasmania 7005, Australia
[2] Australian Geological Survey Organisation, GPO Box 378, Canberra, ACT 2601, Australia

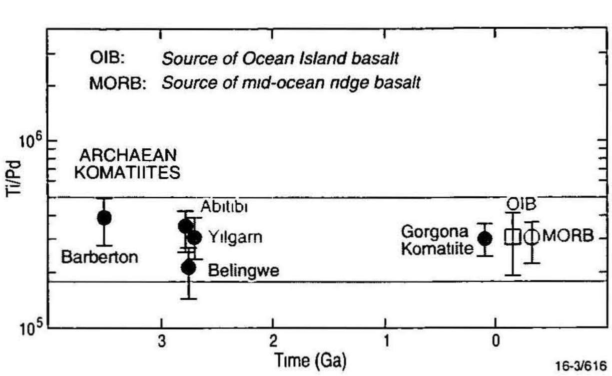


Figure 1. Ti/Pd values in the convecting mantle over time, based on study of komatiites from 3.45 to 0.06 Ga and ultramafic mantle xenoliths, ophiolites and alpine peridotites (Sun et al. 1991, McDonough & Sun 1995, McDonough pers. comm. 1995).

This indicates a lack of obvious secular variation in the abundance of Pd (and other platinum group elements) in the convecting mantle, and the same can be inferred for gold. Similarly, Jochum et al. (1993) found that Sn/Sm appears to be constant in mantle-derived rocks for the last 3.4 Ga at a value ~0.32. Nevertheless, secondary enrichment of Pd, Pt and Au can occur in komatiites, boninites and some shoshonites and plume-related picrites and tholeiites due to S-undersaturation and second-stage melting (Sun et al. 1991).

Continuity of crustal processes is also indicated by the similarity and persistence of certain ore types from the Early Archaean to the present, particularly volcanic-hosted massive sulphide (from 3.5 Ga), porphyry copper (from 3.3 Ga), and Sn-W and syndeformational lode gold deposits (from about 2.7 Ga), all typical of Phanerozoic convergent continental margins before, during and after orogenesis (Hutchinson 1981, Sawkins 1990a, Barley & Groves 1992; Fig. 2). Archaean mesothermal lode Au provinces are no richer in that metal than Phanerozoic examples (see below).

Earth's surface rind in the earliest Archaean probably consisted of thin platelets of simatic and lesser sialic material undergoing movement and subduction under the influence of shallow convection, involving relatively rapid flow due to the hotter mantle (Lowe & Ernst 1992). Deep penetration of the mantle by subducted material may have been inhibited by shallow melting in earlier Archaean times and the few crustal blocks probably grew by rapid accretion, with little interference from neighbours, to reach diameters >1000 km (Lowe & Ernst 1992, McCulloch 1993). Periodically, large volumes of crust, both ultramafic–mafic and felsic, could have formed, been recycled back into the mantle and reprocessed in the crustal environment through lithosphere subduction, meteorite impact to about 3.7 Ga (Maher & Stevenson 1988), and crustal remelting. Stein & Hofmann (1995) proposed periods of rapid crustal growth as a result of mantle overturn and major orogenesis when two-layer mantle convection gave way to single-cell, whole-mantle convection, during which magmas generated by buoyant plumes underplated crust or formed oceanic plateaus, and subduction slabs penetrated deep into the mantle. The chemical composition of the continental crust cannot be maintained by long-term depletion of the isolated upper mantle, requiring episodic or continuous replenishment from the lower mantle. Recent numerical modelling by Davies (1995) indicates that for a two-layered convecting mantle the 'cold finger' effect of lithosphere subduction would effectively buffer the temperature in the Archaean upper mantle to only ~100°C hotter than the present day upper mantle (with a potential temperature of ~1300°C and an adiabatic gradient of ~0.3°C/km) while the potential temperature of the lower mantle could have been about three hundred degrees hotter.

Mantle plumes originating at the thermal boundary layers, the core/lower mantle and/or the upper/lower mantle, could have played a major role in generation of the voluminous Archaean continental flood basalts and oceanic plateaux from the plume heads (Campbell & Griffiths 1990), as well as introducing large layered mafic-ultramafic intrusions and kimberlites from the deep mantle and causing magma underplating, crustal melting, and growth of refractory, and thus lighter, lithospheric mantle and continental lithosphere. Survival of diamonds formed in the Archaean would be favoured in the cold, thick and stable lithospheric mantle formed in this way, eventually being brought to the surface by kimberlites generated by later melting of the mantle plume at great depths.

Des Marais (1994a) proposed that much of the earlier Archaean crustal material was probably submerged, restricting the influence of weathering on the compositions of the oceans and atmosphere. However, the report of an unconformity at 3.5–3.46 Ga in the Pilbara Block, with a possible related palaeosol, indicates that orogenesis, exposure and weathering were occurring at that time (Buick et al. 1995). Domal intrusion of intermediate to silicic plutons from about 3.4 Ga (e.g. the Pilbara Block, Bickle et al. 1993) or earlier, and regional metamorphism and anatexis (the high grade gneiss terrains), caused stabilisation and thickening of crustal platelets with increasing recycling in the crustal environment as evidenced by the growing Eu anomaly (Taylor & McLennan 1995).

The Late Archaean saw the merging of terrains along convergent margins, accretion of volcanic arcs in greenstone belts and the development of cratonic and craton margin sequences. Extensive platforms were established with major rifts and passive margins generated in extensional settings, and foreland basins in convergent settings (Lowe & Ernst 1992, Grotzinger & Ingersoll 1992). For example, Witt (1995) identified Late Archaean tholeiitic and high-Mg sills in the Kalgoorlie Terrane in Western Australia that are consistent with formation in the marine continental margin basins proposed by Barley et al. (1989) for this terrane. The identification of a 2.7 Ga suture extending into the upper mantle to a depth of 65 km in the Abitibi Belt implies collisional events and deep subduction similar to that of recent times (Calvert et al. 1995). Major Ni-Cu, Au and base metal mineralisation in the Yilgarn and Superior provinces was associated with orogenesis between 2.8 and 2.6 Ga (Barley et al. 1995). One of the earliest large basins (covering some 26 000 km^2), the foreland-type Witwatersrand Basin dated at between 3.1 and 2.7 Ga, contained sediments with detrital uraninite and gold, probably derived from hydrothermally altered ≤ 3.05 Ga U- and Au-rich granites (Robb & Meyer 1990), though the amount of Au therein must have been exceptionally high (Kirkham & Roscoe 1993).

The Late Archaean was a period of rapid growth of continental crust, particularly near 2.7 Ga, and by the end of the Archaean more than half of the Earth's continental crust may have been formed (e.g. Taylor & McLennan, this issue). Gradual changes occurred in the overall composition of the upper crust, owing to a reduction of melting of the subducted oceanic crust when the mantle cooled, an increase in melting and reprocessing of the pre-existing crust, and cordilleran as well as island arc-type, calcalkaline magmatism becoming more common. Thus average Palaeoproterozoic upper continental crust was enriched in large ion lithophile elements (e.g. K, Rb, Pb, U) compared to average upper Archaean crust (Taylor & McLennan this issue).

The evolution of life and the early oceans

The earliest organic compounds on Earth may have formed by photochemical reactions in the atmosphere and on the surface, or as a result of the introduction of material by extraterrestrial bodies during impact (see Kasting 1993). Another possibility of increasing interest is an origin using the chemical energy developed as a result of mixing hot

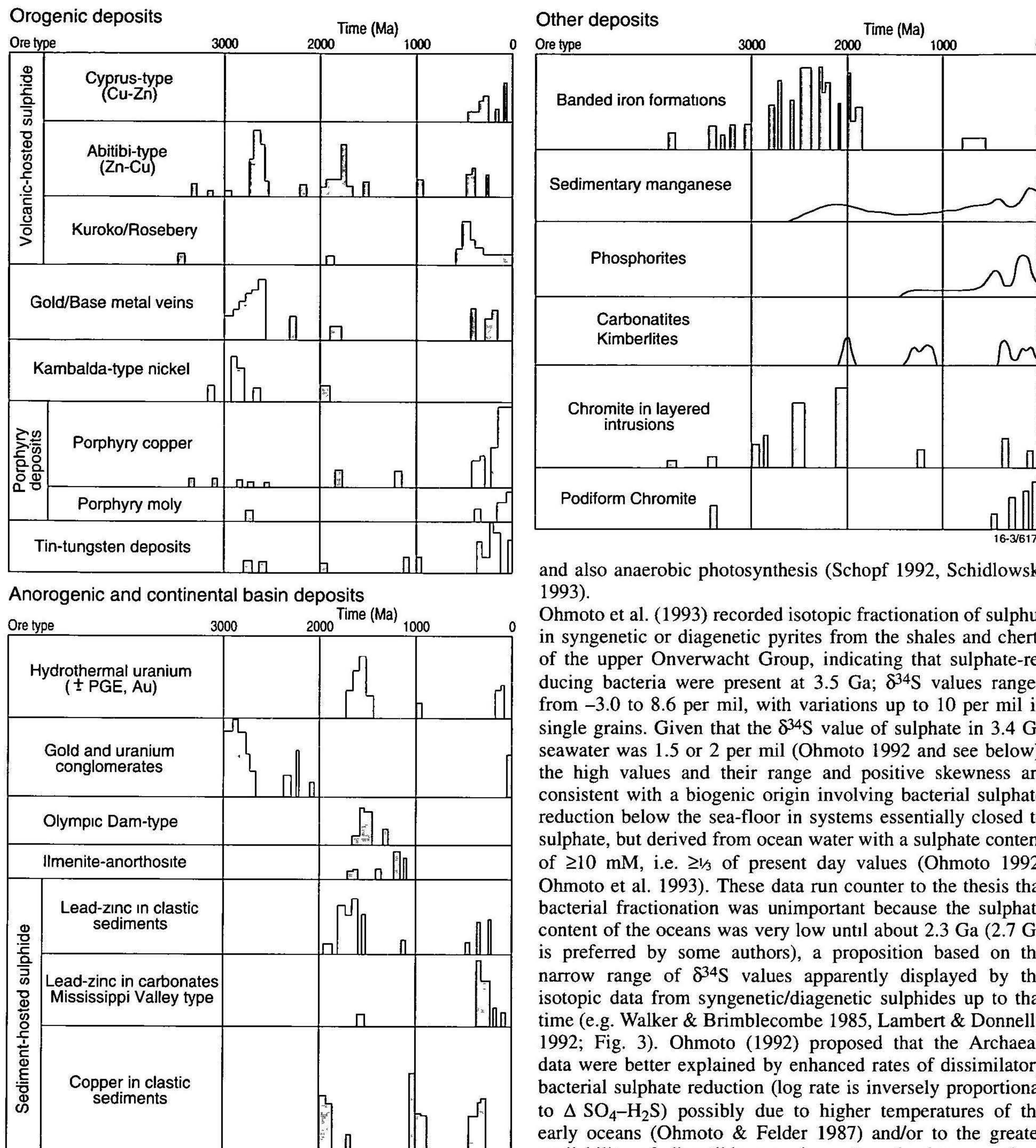


Figure 2. The distribution in time of some important types of mineral deposit, from Meyer (1985, 1988), Lambert et al. (1992), Klein & Beukes (1992) and Barley & Groves (1992). The vertical bars for each ore type represent the approximate proportions of the total tonnage from 3.8 Ga.

seafloor vent waters with cold ocean waters (Russell et al. 1993, Shock et al. 1995). The oldest fossils known appear to be the prokaryotic stromatolites and cellularly preserved filamentous and colonial microfossils in the carbonaceous cherts of the Warrawoona Group in Western Australia, probably ≤3.47 Ga (Schopf & Packer 1987, Thorpe et al. 1992, M.E. Barley pers. comm. 1995), and in cherts of the Onverwacht and Fig Tree groups of the Barberton Greenstone Belt, dated at about 3.45 Ga and 3.45–3.23 Ga, respectively (Byerly et al. 1986, de Ronde & de Wit 1994). The microfossils indicate the presence of autotrophic and heterotrophic organisms, which probably included cyanobacteria capable of oxygen-producing and also anaerobic photosynthesis (Schopf 1992, Schidlowski 1993).

Ohmoto et al. (1993) recorded isotopic fractionation of sulphur in syngenetic or diagenetic pyrites from the shales and cherts of the upper Onverwacht Group, indicating that sulphate-reducing bacteria were present at 3.5 Ga; $\delta^{34}S$ values ranged from –3.0 to 8.6 per mil, with variations up to 10 per mil in single grains. Given that the $\delta^{34}S$ value of sulphate in 3.4 Ga seawater was 1.5 or 2 per mil (Ohmoto 1992 and see below), the high values and their range and positive skewness are consistent with a biogenic origin involving bacterial sulphate reduction below the sea-floor in systems essentially closed to sulphate, but derived from ocean water with a sulphate content of ≥10 mM, i.e. ≥⅓ of present day values (Ohmoto 1992; Ohmoto et al. 1993). These data run counter to the thesis that bacterial fractionation was unimportant because the sulphate content of the oceans was very low until about 2.3 Ga (2.7 Ga is preferred by some authors), a proposition based on the narrow range of $\delta^{34}S$ values apparently displayed by the isotopic data from syngenetic/diagenetic sulphides up to that time (e.g. Walker & Brimblecombe 1985, Lambert & Donnelly 1992; Fig. 3). Ohmoto (1992) proposed that the Archaean data were better explained by enhanced rates of dissimilatory bacterial sulphate reduction (log rate is inversely proportional to Δ SO_4–H_2S) possibly due to higher temperatures of the early oceans (Ohmoto & Felder 1987) and/or to the greater availability of digestible organic matter, the latter partly a function of higher atmospheric P_{CO_2}. Trudinger (1992) suggested that if the phylogeny of archaeobacteria is correctly understood then bacterial sulphate reduction probably predated 3.45 Ga, and he raised the possibility of the existence of other bacterial sulphate reducers in the past than the groups presently known, groups having different fractionation characteristics that may have evolved to sulphate reduction via sulphur and then sulphite reduction.

Abundant sulphate in the Early/Middle Archaean oceans is also indicated by the barite lenses at North Pole in the Warrawoona Group in Western Australia and in the Fig Tree Group in Africa. These barites are believed to be replacements of evaporative gypsum (Perry et al. 1971, Lambert et al. 1978, Buick & Dunlop 1990, Ohmoto 1992), and the associated sedimentary rocks in Western Australia indicate extensive areas of shallow, nearshore probably (but not proven) marine environments (Buick & Dunlop 1990). Other barites of broadly similar age in the Onverwacht Group in South Africa and in the Sargur Group of the Dhawar Craton in India are thought to be of hydrothermal or detrital origin (Reimer 1990, Deb et al. 1991). Ohmoto (1992) has shown from sulphur-isotope distributions that Archean barites are unlikely to have been derived by oxidation of hydrothermal H_2S, but it is possible that some of the barites (and coeval sulphides) are primary precipitates derived from low temperature seawater cycling (e.g. Vearncombe 1995). The sulphur isotope compositions of all the barites mentioned range from 2 to 21 per mil (mostly 3 to 6 per mil, Strauss 1993), suggesting derivation from ocean sulphate having a δ^{34} value of 1.5–2.0 per mil (Ohmoto 1992). Further evidence of abundant sulphate in the Early/ Middle Archaean oceans of the Pilbara region includes the barite in 3.46 and 3.26 Ga volcanic-hosted massive sulphide deposits that probably formed in water ≥1.5 km deep (see *Volcanic-hosted massive sulphide deposits*, below). The scarcity of barite in Late Archaean volcanic-hosted massive sulphide deposits may reflect a low ocean sulphate content but other possible explanations are discussed below.

Kasting (1993) tentatively concluded that Archaean atmospheric P_{O_2} was $<10^{-3}$ atm, i.e. was essentially anoxic (Fig. 4), but Towe (1991) argued from his estimates of organic carbon and iron preserved in early sediments that biological aerobic respiration must have been the major sink for oxygen, and that such a level of activity required free oxygen in the atmosphere and oceans. The high ocean sulphate content indicated by the evidence discussed above also implies free oxygen in the atmosphere (Ohmoto 1992), as does the (limited) Archaean palaeosol data (Holland 1992a). Hutchinson's (1992) proposal that Archaean banded iron formations have more reduced assemblages than their Proterozoic counterparts is compatible with a relatively low P_{O_2} in the Archaean.

Grotzinger & Kasting (1993) suggested that the abundance of massive abiotic aragonite carbonate sediments in the Archaean and Palaeoproterozoic indicates that the early oceans were markedly more saturated in $CaCO_3$ than at present, and that this might account for the paucity of evaporitic sulphates during that period (by removing Ca^{2+} during evaporation before gypsum saturation). Their alternative explanation, that the sulphate content of seawater was low, seems unlikely in view of the evidence referred to above. Using the P_{CO_2} modelled for Earth's early atmosphere (higher than present, Kasting 1993) and other constraints, Holland & Kasting (1992) and Grotzinger & Kasting (1993) modelled possible pH values for the early oceans one or two units lower than the present day (pH ~8). Grotzinger & Kasting (1993) concluded that the formation of the soda ocean of Kempe & Degens (1985) was unlikely, a setback for economic geologists wishing to solve metal transport problems with highly alkaline fluids. Relatively reduced, acid early ocean waters may have allowed solution

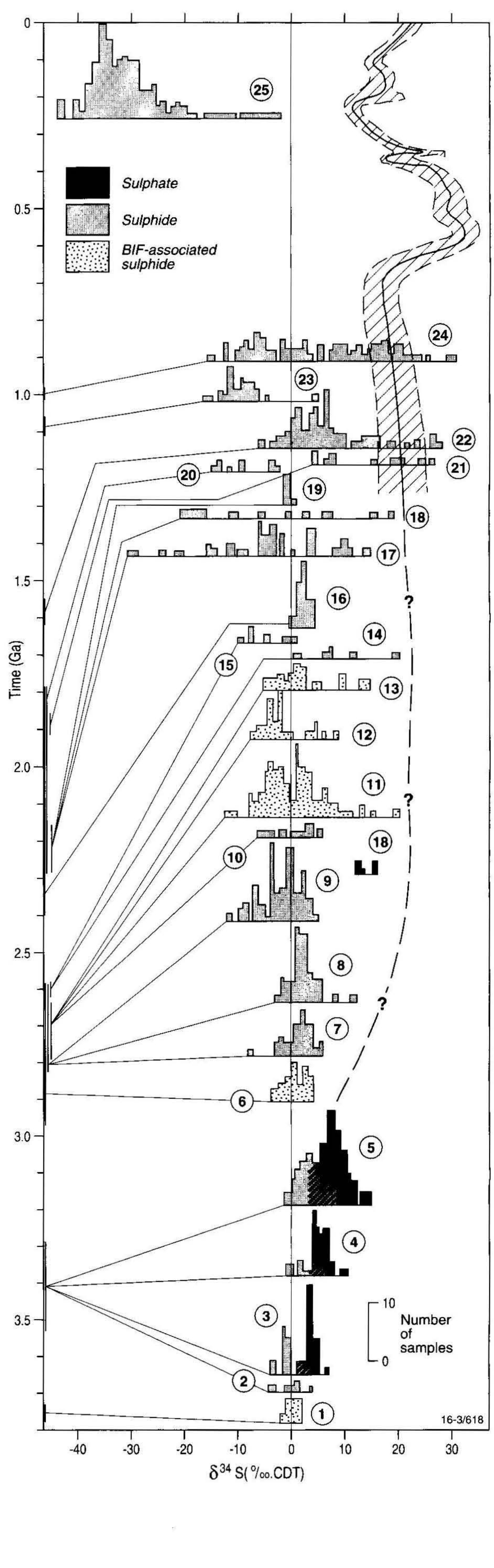


Figure 3. The distribution of sulphur isotopic values from sulphates and (mostly sedimentary) sulphides with time, from Schidlowski (1986). Hayes et al. (1992) and Strauss (1993) have plotted more data for the Proterozoic and discussed the global conservation problems of the positive weighting of the δ^{34}S data. The Neoproterozoic-Phanerozoic sulphate curve is for seawater sulphate. 1, BIF, Isua Belt, Greenland. 2, Onverwacht Group, South Africa. 3, Warrawoona Megasequence, Pilbara, Western Australia. 4, Fig Tree Group, South Africa. 5, Iengra Series, Siberia. 6, BIF, Zimbabwe. 7, Shale, Yilgarn, Western Australia. 8, Deer Lake greenstone belt, Minnesota, USA. 9, Birch-Uchi greenstone belt, Canada. 10, Fortescue Group, Western Australia. 11–13, Michipicoten, Woman River and Lumby–Finlayson BIFs, Canada. 14, Steeprock Lake series, Canada. 15, Ventersdorp Group, South Africa. 16, Cahill Formation, Pine Creek, Australia. 17, Transvaal Supergroup, South Africa. 18 (two plots), Lorraine and Gordon Lake formations, Canada. 19, Frood Series, Sudbury, Canada. 20, Shale, Outukumpu, Finland. 21, Onwatin Slate, Sudbury, Canada. 22, McArthur Basin, Australia. 23, Adirondack sediments, Canada. 24, Nonesuch Shale, Canada. 25, Kupfershiefer, Europe.

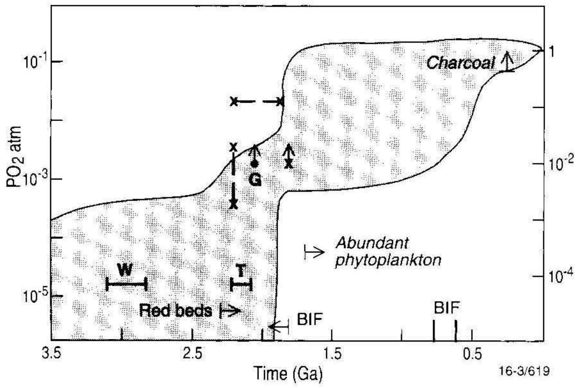


Figure 4. The variation in atmospheric P_{O_2} with time according to Kasting (1993). Crosses indicate tentative estimates and ranges of P_{O_2} from palaeosol studies (Holland et al. 1989, Holland & Beukes 1990, Holland 1992a). G: the requirements for *Grypania spiralis* from the Michicopoten Iron Formation (Han & Runnegar 1992, Runnegar 1993). W: the approximate possible age range of the uraninite-pyrite-bearing Witwatersrand Basin (Robb & Meyer 1990). T: the age of the hematite-magnetite-bearing Tarkwaian conglomerates (Krupp et al. 1994). The time of redbed formation is from Lowe (1992b). The cessation of significant banded iron formation coincides approximately with the sharp rise in P_{O_2} from about 2.0 to 1.8 Ga, but banded iron formations reappear in the period 0.8–0.6 Ga. The very low P_{O_2} values for the Archaean atmosphere seem unlikely in view of the evidence of substantial sulphate in Archaean oceans.

of more Fe^{2+} and Mn^{2+} than at present. High Mg/Ca ratios (a corollary of extensive carbonate precipitation) and high Fe contents of the oceans is indicated by the abundance of sedimentary ferroan (and manganiferous) dolomite and ankerite in Early and Middle Archaean sequences, and of ferroan dolomite and siderite in regional alteration assemblages (Veizer et al. 1989a, b). The Mg-rich chlorite commonly present in the altered footwalls of Archaean volcanic-hosted massive sulphide deposits (Franklin et al. 1981, Meyer 1988) may be due to relatively high Mg^{2+} contents in seawater-derived ore solutions compared to those of the Phanerozoic. Increase in oceanic pH may have been a factor (additional to increased P_{O_2}) in shutting off giant-sized banded iron formation from about 1.8 Ga.

Despite intensive research, particularly in Australia, it seems that biological activity has been essential to metal fixation in only a few types of ore deposit, for example, pyritic shales (e.g. Brukunga in South Australia, Seccombe et al. 1985), sediment-hosted Pb-Zn deposits in Ireland (e.g. Anderson et al. 1989) and possibly some Cu-Co ores (e.g. the Kupferschiefer; Fig. 3). Proposals that the stratiform Pb-Zn deposits of the Phanerozoic Selwyn Basin in Canada contain largely biogenic sulphide have been seriously undermined by Ohmoto et al. (1990). Sulphide microbialites have been reported from the Silvermines Pb-Zn deposit in Ireland (Russell 1996). A complex biota was probably associated with other seafloor ore-forming systems (e.g. volcanic-hosted massive sulphide deposits and Au-U conglomerates), but does not appear to have been crucial to ore formation, though hydrothermal processes may have been vital to the development of the earliest organisms (Russell 1996).

The Proterozoic-Phanerozoic

Collisional and obducted material and sediments of the craton margin are more common in Palaeoproterozoic than Archaean sequences, and long-lived platforms (passive margins?) of Palaeoproterozoic age provided the environment for a surge in BIF and Mn sedimentation (Fig. 2), perhaps during periods of continental fragmentation and dispersal (see *Banded iron formations and manganese deposits*, below). During the late Palaeoproterozoic large continental masses supported extensive basins in which the earliest sedex Pb-Zn-Ag deposits were formed, e.g. in northern Australia from about 1.7 Ga (e.g. Mount Isa), and in Korea and north China, possibly from ~1.9 Ga (e.g. Jiande). Hairpin bends of the palaeomagnetic apparent polar wandering path of north Australia for the mid-Proterozoic, probably related to major changes in relative plate motion, coincided with basin mineralisation (Idnurm et al. 1994, Loutit et al. 1994). Other giant basin deposits include the Cu-Co ores of Udokan at about 2.0 Ga and the Zambia/Zaire deposits at about 1.0 Ga, and the earliest important hydrothermal U-Au-platinum group element assemblages were formed in extensive, fractured cratonic basins in Australia and Canada from about 1.7 Ga (e.g. Jabiluka, Rabbit Lake). The period from about 2.0 Ga through the Mesoproterozoic saw abundant felsic magmatism in and on the continents, and some oxidised, alkaline, S-poor granitic magmas yielded large tonnages of hydrothermal Fe (e.g. Kiruna at 1.9 Ga, Olympic Dam at 1.6 Ga). The scarcity of greenstones (Condie 1994) and lack of volcanic-hosted massive sulphide and porphyry-related ores between about 1.7 and 1.4 Ga may be related to development of a supercontinent during this period (Hoffman 1989). According to Veizer et al. (1989c) and Lowe (1992a), there was steady-state equilibrium between addition and consumption of crustal material from about 1.8–1.5 Ga, with less new crust added and craton rifting and associated sedimentation prominent.

There is no longer any support for the model proposed by Etheridge et al. (1987) that envisaged Palaeoproterozoic orogeny and crustal growth (e.g. in the Pine Creek Inlier) as a result of mantle upwelling and underplating beneath large sialic masses rather than plate tectonic processes (Windley 1992, 1993, Myers 1993, Solomon & Groves 1994). Nevertheless the origin of the extensive felsic magmatic activity that characterised northern Australia and other continents in the late Palaeoproterozoic has been much debated (Hoffman 1989).

The lack of significant redbeds before about 2.2 Ga, the presence of large deposits of detrital uraninite and pyrite and the removal of Fe and other components from palaeosols before that time suggest there may have been low atmospheric P_{O_2}/P_{CO_2} during the Archaean and into the earliest Proterozoic (Lowe 1992b, Holland 1992a, Kirkham & Roscoe 1993, Krupp et al. 1994; Fig. 4). A significant rise or rises between about 2.0 and 2.3 Ga (Kirkham & Roscoe 1993 suggest between 2.3 and 2.4 Ga) is indicated by the appearance of redbeds and iron oxide-bearing conglomerates, and also by retention of Fe^{III} in palaeosols. Holland (1992a) tentatively suggested a value of P_{O_2} of about 4×10^{-3}–3×10^{-4} atm (2×10^{-2}–1.5×10^{-3} PAL) for the pre-(?) 2.2 Ga Heckpoort palaeosol, while Mossman & Farrow (1992) obtained values of 2×10^{-2} and $<10^{-2}$ PAL for the pre-2.2 Ga pre-Huronian Supergroup profiles of the Elliot Lake palaeosol (Kirkham & Roscoe 1993). Holland & Beukes (1990) thought that the Kuruman banded iron formation profile, with weathering between 2.2 and 1.9 Ga, probably formed at P_{O_2} levels of 1.5×10^{-1} PAL. The presence of *Grypania spiralis* in 2.1 Ga iron formations from Michigan indicates a P_{O_2} level of $\geq10^{-2}$ PAL (Han & Runnegar 1992), probably in both ocean and atmosphere (Runnegar 1993), and formation of the 2.0 Ga Udokan Cu-Co ores involved reduction of Cu-bearing oxidised solutions. Highly oxidised shallow groundwaters (of meteoric origin?) were probably vital to the generation of the 1.6 Ga hematite-Cu-U-REE ores at Olympic Dam and the 1.5–1.7 Ga Pine Creek U-platinum group element-Au deposits (Solomon & Groves 1994, Haynes et al. 1995). P_{O_2} must have reached at least 6–10% of the present value by the close of the Proterozoic to have supported the newly diversifying biota (Holland 1992b).

The causes of the Palaeoproterozoic increase in atmospheric P_{O_2} may have been biological, e.g. increased photosynthesis

(Holland 1992b), or increased burial rate caused by evolutionary changes to the biota (Logan et al. 1995), or the result of increased and faster burial of reduced organic C from about 2.9 Ga, mostly during periods of continental breakup and development of extensive sedimentary basins on new continental margins (Des Marais et al. 1992, Des Marais 1994b). The latter authors based their opinions on the $\delta^{13}C$ values of organic C and survival of C-bearing sediments, both at a maximum in the periods 2.3–2.0 Ga and 1.1–0.7 Ga. As reviewed by Kirkham & Roscoe (1993) these are also periods of glaciation (related to lowering of P_{CO_2}) and development of large provinces of banded iron formations. With respect to the latter, it is noted that Fe is an essential component of the modern photosynthetic production of organic compounds, and that addition of Fe to ocean surface water dramatically stimulates the phytoplankton growth rate, lowers ocean P_{CO_2} and increases the concentration of dimethyl sulphide, a potential precursor of atmospheric sulphate particles, raising the possibility of lowering ocean and atmospheric temperature (see the Ironex experiments, Martin et al. 1994, Kumar et al. 1995, Frost 1996 and related papers). Is it possible that the substantial and protracted Palaeoproterozoic and Neoproterozoic oceanic inputs of Fe envisaged for banded iron formations (see *Banded iron formations and manganese deposits*) had significant impacts on the fauna and climate, possibly lowering temperature to the extent that glaciation occurred?

Cyclic changes in atmospheric P_{O_2}, such as modelled for the late Palaeozoic by Graham et al. (1995), were probably superimposed on the overall increase through the Palaeoproterozoic. The cause and effect relationship between atmospheric P_{O_2}/P_{CO_2} and biological activity is complex (e.g. Kasting 1993, Graham et al. 1995), and other factors such as a decline in release of reduced gases (possibly related to mantle oxidation) may also be important (Kasting et al. 1993).

Similar evidence of relatively sulphate-rich oceans in the Early and Middle Archaean (see above) is also seen in Late Archaean and Palaeoproterozoic rocks. For example, the sulphur of apparently stratiform barite lenses from the area of the 2.7 Ga Hemlo deposit in Canada ($\delta^{34}S$ of about 2–12 per mil; Cameron & Hattori 1985) may be a product of mixing seawater sulphate partially reduced during seawater cycling with ambient seawater sulphate, indicating abundant ocean sulphate and a Late Archaean isotopic composition close to Early Archaean values (Ohmoto 1996). Anhydrite is present in the Archaean Geco and the Palaeoproterozoic Anderson Lake deposits, both of volcanic-hosted massive sulphide type (i.e. involving seawater cycling) though highly metamorphosed (Franklin et al. 1981). Stratiform barite ($\delta^{34}S$ = 17.1–21.2 per mil) and sulphides (6.2–8.5 per mil) from the 1.9 Ga Aravalli Supergroup in India (Deb et al. 1991), and the 1.8 Ga Åsen barite ($\delta^{34}S$ = 1.5–15.0 per mil) in the Skellefte district of Sweden (from Strauss 1993), may be of similar origin.

Vein barite at the Eastern Creek deposit (18.4–24.7 per mil) in the 1.6–1.7 Ga McArthur Basin in Australia was probably deposited from basin water ultimately of marine origin (Muir et al. 1985), as were barite grains disseminated in McArthur Basin sediments (19.9–31.5 per mil, Walker et al. 1983). There is abundant evidence of precursor gypsum in the McArthur Basin (Jackson et al. 1987), and bedded anhydrites are common from about 1.4 Ga (Strauss 1993). These late Palaeoproterozoic results indicate oceanic sulphate $\delta^{34}S$ values substantially higher than those of the Archaean by about 1.9 Ga. In part at least, the lack of evaporitic sulphates in the Palaeoproterozoic might reflect the increasing likelihood of dissolution of anhydrite and gypsum with increasing age (but see earlier discussion of changes in ocean composition).

Superimposed on the overall crustal evolution from at least 2.9 Ga are second-order cycles of several hundred million years, involving continental assembly, culmination and breakup. Greenhouse conditions and oceanic anoxia are associated with new ocean basins and enhanced ocean ridge exhalative activity following breakup, and glaciation is attendant on continental aggregation (e.g. Veevers 1990, Barley et al. 1995), so there is a strong link between the tectonic cycles and climatic changes. The distribution of ore deposit types in time reflects the various stages. For example, major phosphate and manganese ores (e.g. the Cambrian Georgina phosphates and the Cretaceous Groote Eylandt Mn ores in Australia) follow sea level changes resulting from continental breakup, and peaks in abundance of volcanic-hosted massive sulphide ores worldwide in the Ordovician–Silurian and Mesozoic–Tertiary reflect volcanism related to enhanced oceanic growth and subduction following continental fragmentation (Rona 1988, Barley & Groves 1992). These post-breakup periods were also characterised by development of extensive shale basins, some of which were anoxic (see below), deposition of turbidites, and by low $\delta^{13}C$ in carbonates and high $\delta^{34}S$ in evaporites (Holser et al. 1988, Titley 1991).

Palaeoproterozoic phosphorites are known from the Baltic Platform, but phosphate deposits are most abundant worldwide in the Neoproterozoic–Cambrian period (Fig. 2), probably reflecting the surge in diversification of eukaryotes at that time and the appearance of shelly faunas (Cook et al. 1990). Another surge in biological diversification occurred in the Ordovician, tying in with formation of anoxic basins, and enhanced volcanic arc and mid-ocean ridge activity.

A review of mineralisation through time

Most major mineral deposit types have characteristic first-order age/frequency distributions (Fig. 2), reflecting evolutionary trends in the compositions of the mantle, crust and hydrosphere, changes in tectonic style and the increasing significance of recycling with time (Veizer et al. 1989c, Lambert et al. 1992). The influence of second-order variations, of the order of 100 my, has been noted above. Here we discuss a few of the important ore types.

Magmatic platinum group element, nickel, copper and chromite deposits derived from mafic and ultramafic magmas

Deposits of this group occur in orogenic belts and in large cratonic intrusions from about 3.0 Ga. Layered mafic/ultramafic complexes with important platinum group element deposits, such as Stillwater (2.7 Ga) and the Bushveld Complex (2.0 Ga), are probably related to mantle plumes, and the Ni-Cu-platinum group element deposits of the Norilsk region in Siberia are hosted by 0.25 Ga flood basalts of mantle plume origin (Naldrett et al. 1995). Mixing of S-undersaturated magma and sulphide-bearing magma or crustal S was commonly critical to Ni-Cu and platinum group element ore formation; for example, under certain conditions ground erosion and assimilation of sulphide-bearing sediments by Ni-rich, S-undersaturated komatiitic lavas produced massive Ni-Cu lenses at the base of the flows (Groves et al. 1986), particularly in the Late Archaean of the Yilgarn Block and the Superior Province.

Magmatic porphyry copper-gold and copper-molybdenum deposits, and tin-tungsten and tungsten-molybdenum deposits

Porphyry-related Cu-Au ores generally formed in subduction-related island arcs with little or no continental crust, and Cu-Mo ores typically in convergent continental margins with thick crust (Titley & Beane 1981). Because of their shallow level of emplacement and terrestrial setting their frequencies decline rapidly, particularly for the island arc types, of which the oldest are probably the Late Ordovician Goonumbla and Copper Hill in New South Wales. The Cu-Mo types, however, extend into the Middle Archaean (e.g. 3.32 Ga at Gobbos in the Pilbara Block, ME Barley, University of WA, pers. comm. 1995). Their presence supports the uniformity of related tectonic processes throughout the greater part of Earth's history. Most of the surviving examples relate to the formation of subduction-related magmatic arcs following the breakup of Pangaea.

Orogenic Sn-W and W-Mo deposits are also essentially of magmatic origin. They are related to mostly post-orogenic emplacement of highly fractionated felsic melts (tin granites) on the craton side of magmatic arcs in continental convergent margins during subduction (Sawkins 1990a). Emplacement levels range from subvolcanic to pegmatitic, the former surviving erosion like the Cu-Mo ores, but the latter, and also intermediate-depth vein and greisen types, being relatively common in Late Archaean rocks (e.g. Blockley 1980). Anorogenic Sn-W deposits, like those of northern Nigeria (0.6-0.7 Ga), Rondonia (2.0 Ga) and in the Bushveld complex (e.g. Zaiplaats, 2.05 Ga), are found in stable cratons and have a better survival record. Sawkins (1976, 1990a) tied these magmatic events to melting attendant on mantle hotspot activity (mantle plumes?) related to continental fragmentation.

The Sm/Nd signatures in two Australian tin granites of orogenic type suggest that melting in the sialic crust was related to emplacement of mantle-derived melts (Mackenzie et al. 1988, Sun & Higgins 1996). Orogenic Sn-W provinces lie within (mostly) extensive linear granitic magmatic belts, but are commonly circular or oval in outline with sharp boundaries — perhaps the provinces also reflect the size and position of large mantle-derived melts emplaced in the deep sialic crust.

Mesothermal lode gold deposits

Gold-bearing quartz veins approximately contemporaneous with collision-induced deformation are found in greenstone belts and turbidite-rich greywacke-sandstone-mudstone sequences of the continental margin from about 2.7 Ga to 0.1 Ga, with major peaks at about 2.6–2.7, 0.38 and <0.2 Ga. The low salinity, CO_2-bearing fluids were probably derived from metamorphic dewatering at temperatures of more than 350°C and travelled via deep fractures and terrane boundaries. While post-Archaean examples originated mostly near the brittle-ductile transition, Archaean examples formed over a greater range of pressure and temperature. Syn- and post-ore intrusion of granitoids is characteristic (e.g. the Victorian and Pine Creek goldfields in Australia). The Late Archaean deposits of the Yilgarn Block and the Abitibi belt are of similar age and style, mostly forming in the period 2.6–2.7 Ga as part of a chain of events that culminated in the formation of one or more large continents (Myers 1995). The overall Au production/km^2 of such famous fields as the Yilgarn (~2.7 Ga) and Victoria (~0.38 Ga) in Australia is similar, suggesting that secular changes in the Au content of upper crustal rocks are not significant to ore formation. Titley (1991) noted that lode Au deposits were particularly common in early Palaeozoic and late Mesozoic turbidites and suggested that the Au was sourced in the host sediments during oceanic anoxia that followed continental fragmentation. However, it seems more likely that the physical make-up of the sandstone-shale sequences (favouring formation of, for instance, saddle reefs) and the reducing nature of the shales may be the more significant factors in gold deposition (Solomon & Groves 1994).

Sediment-hosted stratiform copper-cobalt and lead-zinc deposits

In contrast to the three groups outlined above, the commonly giant-sized sediment-hosted Cu ± Ag ± Co and Pb-Zn-Ag deposits make their first appearances at about 2.0 Ga (Gustafson & Williams 1981, Kirkham 1989). These deposits formed in large, cratonic or epicratonic, mostly mudstone-carbonate-sandstone sedimentary basins from probable high salinity basinal fluids during basin formation. Essential requirements to ore formation probably included the development of thick continental crust over regions large enough to support extensive shallow platforms at least locally undergoing evaporation to provide fluid salinity (e.g. northern Australia at 1.7–1.6 Ga, the Zechstein Sea at 0.28 Ga) and, in the case of most Pb-Zn ores, abundant biota to allow fluid reduction by organic matter. Broken Hill-type Pb-Zn deposits are associated with felsic volcanic rocks and may have a different genesis (Beeson 1990). Many Pb-Zn deposits other than the Broken Hill type appear to have formed in reduced marine environments (e.g. in mid-Proterozoic northern Australia) and from reduced, deep basinal fluids, but the Cu-Co ores were probably derived by reduction of shallow-circulating, oxidised, Cu-rich fluids (Gustafson & Williams 1981). High heat producing granites have been suggested as heat sources for circulating Pb-Zn fluids in northern Australia (Solomon & Heinrich 1992). The argument that these giant Pb-Zn ore deposits did not form until the crust had been enriched in Pb and Zn by late Palaeoproterozoic felsic magmatism (Sawkins 1989, Barley & Groves 1992, Lambert et al. 1992) is not supported by analytical data for earlier Palaeoproterozoic and Archaean sedimentary and igneous rocks (e.g. Bickle et al. 1993, Solomon & Groves 1994), but it is true that Palaeoproterozoic upper crust is richer in K, Rb, U and Pb than Archaean upper crust.

Sawkins (1990a) suggested these deposits formed during advanced rifting of basins, and drew attention to the concentration of sediment-hosted Cu ores in the early Neoproterozoic, a probable time of continental breakup. Hinman et al. (1994) demonstrated complex transpressional faulting during formation of the 1.64 Ga (R.W. Page unpublished data) HYC Pb-Zn-Ag deposit in northern Australia, also a time of rapid changes of plate motion, interplate tectonism and basin-wide hydrothermal activity (Idnurm et al. 1994).

Volcanic-hosted massive sulphide deposits and ocean compositions

Volcanic-hosted massive sulphide deposits, present intermittently from 3.5 Ga to today, may be classified as Cu-Zn ophiolite-related, Zn-Cu ± Pb and Zn-Pb-Cu (e.g. Franklin 1986). They form today mostly in back-arc basins of various types and at mid-ocean ridges, testifying to the persistence and consistency of plate tectonic and associated hydrothermal processes from 3.5 Ga. The 3.46 and 3.26 Ga Zn-Pb-Cu deposits from the eastern Pilbara Craton (Barley 1992, 1993; Vearncombe et al. 1995), probably the oldest base metal ores on Earth, contain barite, confirming abundant sulphate in at least some of the Early and Middle Archaean oceans, as suggested independently by Ohmoto (1992). They are broadly similar in form, textures and mineral and metal content to the Tertiary Kuroko deposits in Japan and/or some modern seafloor deposits of back-arc basins. This implies that they formed under conditions that inhibited boiling of the ore solutions (water depth was ≥1.8 km for the Japanese deposits; Pisutha-Arnond & Ohmoto 1983). Barite in the Big Stubby deposit has $\delta^{34}S$ values of 11–13 per mil (Lambert et al. 1978), which may be interpreted, following Ohmoto et al. (1983), as mixtures of hydrothermal sulphate (partially reduced oceanic sulphate, possibly via anhydrite) and ambient oceanic sulphate having a value of about 1.5 per mil (Ohmoto 1992). The sulphide-sulphur (about -4.3 to -0.9 per mil) probably represents mixtures of reduced sulphur derived from oceanic sulphate, and of sulphide-sulphur of direct magmatic origin or dissolved from mostly volcanic rocks having $\delta^{34}S$ values near zero per mil.

Zn-Cu volcanic-hosted massive sulphide deposits without Pb and mostly without sulphate are common in the 3.1–2.7 Ga greenstone belts of the Superior province of Canada, and occur in other cratons. Local host rocks are almost invariably felsic volcanics. In the Yilgarn Block deposits of similar age and also without sulphate have minor Pb, for example, Zn/Pb ratios of 13.4 and 14.6 at Teutonic Bore and Scuddles, respectively (Barley 1992), compared to 3.0 for Kuroko deposits (Tanimura et al. 1983) and 3.0–4.1 for 3.5 Ga deposits in the Pilbara Block (Barley 1992). Sawkins (1990b) and Barley (1992) argued that volcanic-hosted massive sulphide ores are of magmatic origin, the metal content of ore solutions being related to the composition of the coeval magma, and Barley (1992) suggested that lack of barite is related to ore formation in anoxic, sulphate-free, ocean water. Similar sulphate-free basin water has been proposed for the source of the solutions responsible for the Early Proterozoic volcanic-hosted massive sulphide deposits of Arizona (Eastoe et al. 1990), and also the Late Palaeozoic, pyrrhotite-bearing, barite-free Zn-Cu ± Pb deposits of the West Shasta district, California (Eastoe & Gustin 1996). One of the difficulties of this model is that if, as proposed, the ores are deposited from solutions that are buoyant in seawater, they will not form the sulphate chimneys that appear to be vital to the process of trapping sulphides and building the mounds (e.g. Goldfarb et al. 1983).

There are other possible explanations for both lack of Pb and absence of barite. Ohmoto (1996) suggested these features resulted from a greater degree of reworking of the massive sulphide ore (= leaching of barite and galena) at the high temperature stage due to the higher mass flux caused by an elevated temperature gradient in the Archaean compared to that in subsequent periods (see Ohmoto et al. 1983). Coeval and older Pb-bearing deposits in the Pilbara suggest steep temperature gradients were confined to certain terranes. An alternative explanation, discussed by Franklin et al. (1981), is that most of the metallic components of the Cu-Zn ores, whether Archaean or younger, were derived by dissolution in underlying terrains dominated by mafic rocks which, containing more reductants, such as Fe^{II}, and low Pb contents, cause total reduction of oceanic sulphate during hydrothermal circulation and low Pb contents in the ore solutions. Modern mid-ocean ridge fluids sourced in basalts have these characteristics (Von Damm et al. 1985), and Franklin et al. (1981) noted that Canadian Archaean greenstone belts, which contain mostly Cu-Zn, low Pb deposits, have >90% of mafic rocks by volume.

Eastoe & Gustin (1996) postulated that few volcanic-hosted massive sulphide deposits on the sea-floor would survive in oxidised seawater and that preservation ideally requires anoxic conditions. They pointed out that Ordovician rocks contain more deposits than any other period, an anomaly enhanced when the preservation potential is taken into account, and this was a period of widespread oceanic anoxia and enhanced volcanic arc activity following continental breakup (Veevers 1990). However, there are other factors to consider, as noted by Eastoe & Gustin (1996), including rapid burial and the observation that some volcanic-hosted massive sulphide deposits differ from those discussed so far in being mostly much larger in area and tonnage, rather sheet-like and rich in Zn and mostly rich in Pb. They appear to have formed from reduced, saline solutions that reversed buoyancy on exhalation and precipitated much of the total metal content of the solutions in basins, the resulting deposits being shielded from oxic seawater by the spent ore fluids. High salinities postulated for this group (Solomon 1981, Green et al. 1981) have recently been found in fluid inclusions at Hellyer, Tasmania (Khin Zaw et al. 1996, Solomon & Khin Zaw in press), and a partial or wholly magmatic origin for the Hellyer fluids is likely, this being the only reasonable source of chloride.

Banded iron formations and manganese deposits

Banded iron formations are potentially important as indicators of ancient atmospheric and oceanic conditions (e.g. Towe 1991, Holland 1992b, Kasting 1993, Morris 1993). They are chemical sediments found in rocks from 3.8 Ga (Isua) to about 0.6 Ga (Rapitan and others) but there is a marked peak in occurrence between about 2.75 and 1.9 Ga, and there are none between about 1.8 and 0.8 Ga and post-0.6 Ga (Klein & Beukes 1992; Fig. 2). Until about 2.6 Ga banded iron formation occurrences tend to be smaller (<50 m thick and strike lengths <10 km; Guilbert & Park 1986) than those of Palaeoproterozoic age, although several Late Archaean banded iron formations extend along strike for more than 30 km, for example at Koolyanobbing in Western Australia (BHP Staff 1975) and in the Chitradurga Schist Belt in India (Rao & Naqvi 1995). About 90% of known iron formations are Palaeoproterozoic in age, and occur in five provinces (from a review by Kirkham & Roscoe 1993); they reach maximum size in the ~2.5 Ga Hamersley Basin, which originally probably covered some 150 000 km^2 and contained some 3×10^{17} kg Fe (Trendall 1975, Morris & Trendall 1988). The Archaean banded iron formations are commonly associated with mafic volcanic rocks, but evidence of volcanic activity is generally more limited or absent in Proterozoic types, except in the Hamersley province, where Barley et al. (1997) have shown that some 30 000 km^3 of bimodal igneous rocks were more or less coeval at 2470 ± 4 to 24491 ± 3 Ma with the development of the Hamersley iron formations.

The Dales Gorge member of the Hamersley banded iron formations, perhaps the most intensively studied banded iron formation unit (Trendall & Blockley 1970, and review in Morris 1993), consists of 17 banded iron formation macrobands (chert-iron oxide, 2–15 m thick) that alternate with 16 thinner ferruginous shale macrobands (chert-carbonate-silicate with shale bands). Many of the macrobands can be traced over the whole basin (Trendall 1975), and Ewers & Morris (1981) found some had near-constant composition over distances of more than 100 km. Mesobands of chert (5–15 mm thick) and Fe-rich material occur in macrobands, and the chert mesobands commonly display microlaminae (0.2–1.5 mm thick, Morris 1993). The probable primary minerals in the Hamersley banded iron formation were microcrystalline chert and hydrated iron oxide, carbonate (mostly siderite) and greenalite/stilpnomelane or some silicate precursor (Ayres 1972, Morris 1993); haematite was probably generated during early diagenesis. The similar (and probably contemporaneous) banded iron formations of the Kuruman Iron Formation of the western Transvaal Basin are thought to have formed in water 100–700 m deep, below wave base and the photic zone, during marine transgression over a shallow-water carbonate platform or shelf that may have been part of a back-arc basin (Klein & Beukes 1989, Beukes et al. 1990).

The genesis of banded iron formations

Most recent workers, following Gross (1983 and earlier papers) and Morris & Horwitz (1983), have favoured a hydrothermal source for a large part of the Fe and silica of banded iron formations, for example, Towe (1991), Klein & Beukes (1992) and Morris (1993), with deposition in continental margin basins or on platforms. The Fe-Si-rich waters may have arrived by (a) upwelling related to winds or currents of reduced ocean water, Fe^{III} depositing as a result of mixing with shallow, oxidised waters (Holland 1973, 1984; Beukes et al. 1990; Fig. 5A), or (b) convective upwelling initiated by ocean ridge activity (Morris 1993), or (c) buoyancy forces generated at ocean ridge vents, with Fe and silica being deposited from gravity currents driven towards the coastal sites (Solomon & Groves 1994, Isley 1995; Fig. 5B).

In the *upwelling models* the waters contain ≥1 ppm Fe and up to 120 ppm SiO_2 (Holland 1984, Dove & Rimstidt 1992) and come from a markedly stratified ocean with large, deep, fluid reservoirs carrying the reduced Fe and connected to open oceans. With each upwelling pulse or phase the Fe or Fe-rich water must be resupplied to the reduced ocean, ultimately by ridge vent activity. The REE compositions and the marked positive Eu anomaly in many banded iron formations (normalised to North American Shale), particularly the high-Fe Dales Gorge material of Morris (1993), cannot be explained by continental weathering and indicate significant input from reduced ocean-floor vent fluids that have reacted with basalt at high temperatures (e.g. Klein & Beukes 1989, Danielson et al. 1992). Alibert & McCulloch (1993) estimated $50 \pm 10\%$ of Nd was supplied to the Dales Gorge-Joffre banded iron formation units from hydrothermal sources.

To supply the 3×10^{17} kg Fe of the Hamersley Basin, and allowing say 50% Fe from terrestrial input and 1 ppm Fe in the fluid, requires 1.5×10^{23} kg of reduced, possibly slightly acid, Fe-bearing ocean water. At Hollands (1984) annual rate of upwelling of 3×10^{16} kg water, the Hamersley banded iron formation deposits could be laid down in about 5×10^{6} years, well within the 20×10^{6} years allowed by the dates of Barley et al. (1997). Modern annual upwelling rates are substantially higher along several passive continental margins (e.g. Smith 1992), and the Fe supply could also have been faster, owing to a more acid, reduced and warmer ocean (see earlier discussion). The *convection* model of Morris (1993) differs only slightly from the above and has the problem that vigorous convection would return fluids to the deep ocean rather than direct them to the continental margin, and would tend to mix the oxygenated and anoxic waters. In the modern oceans, mid-ocean ridge plume dispersal is affected by ocean circulation, not the reverse (Edmond et al. 1982).

Morris (1993) has discussed in some detail the precipitation of Fe and SiO_2 following mixing of oxidised and reduced waters, highlighting the difficulties in accounting for more or less coeval but physically distinguished chert and Fe deposition. His models for the varve-like bands involve photochemical oxidation of Fe^{2+} in the surface zone, storm mixing of reduced and oxidised waters to resupply the shallow water, and evaporation to deposit carbonate and silica, implying seasonal

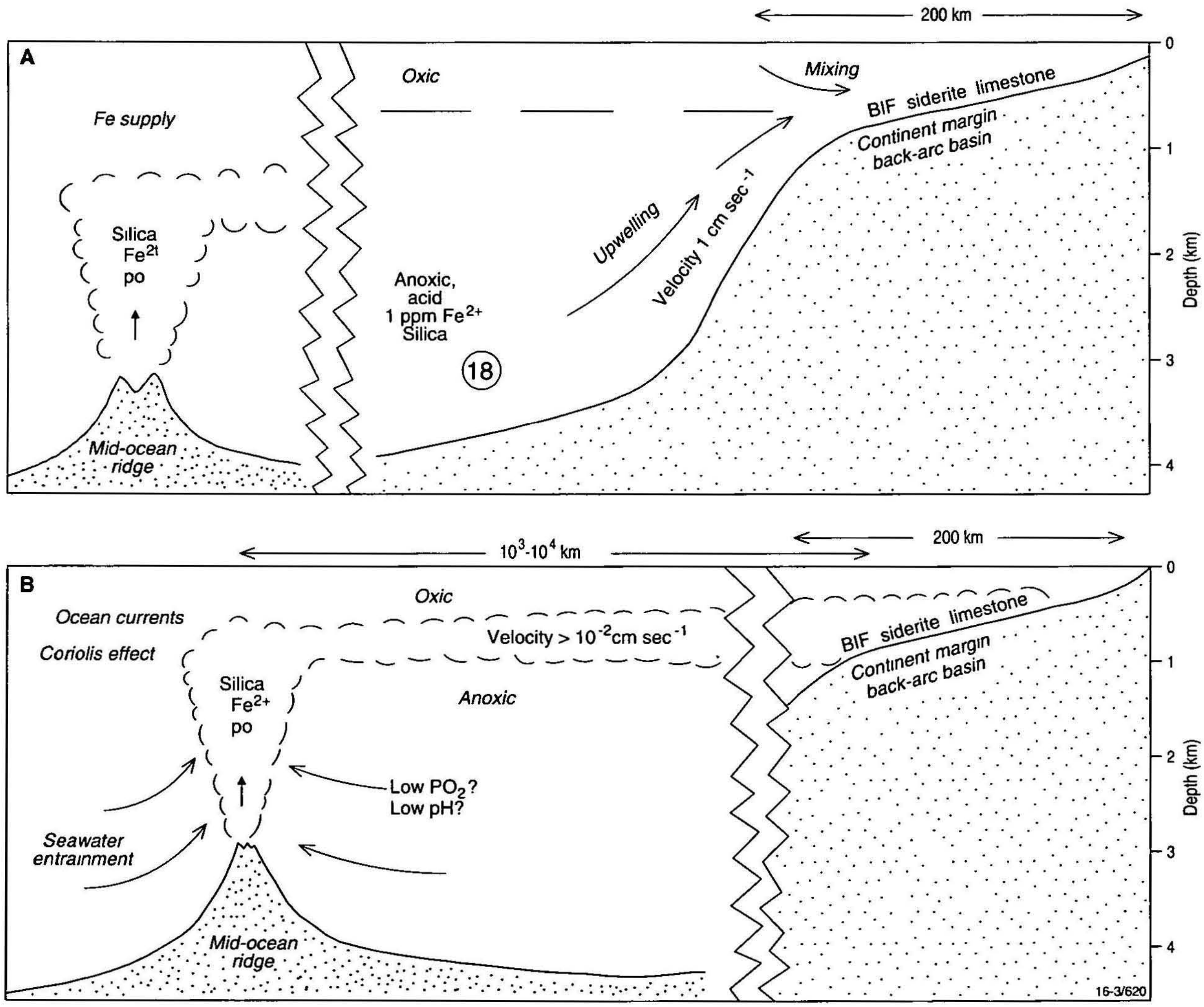


Figure 5. Sketches to illustrate the major competing models for supplying Fe to giant BIF basins in the Palaeoproterozoic.
A. The upwelling model of Holland (1973, 1984), with alongshore currents creating upwelling of deep, reduced, slightly acid(?) ocean waters into shallow, oxidised ocean water where oxidation of Fe^{2+} leads to precipitation of Fe hydroxide gels. Fe is re-supplied by exhalative ocean ridge activity greater than the present. The banded iron formation–siderite–limestone transition is from Beukes et al. (1990). Possible mechanisms of silica and Fe(hydroxide?) deposition are discussed by Morris (1993).

B. The hydrothermal plume/gravity current model of Solomon & Groves (1994) and Isley (1995), with pyrrhotite (and possibly Fe^{2+}) transported vertically and then laterally to basins on the continental margin, with the mass flux and fluid temperature at the sea-floor considerably greater than the present. If oxidation of sulphide and reduced Fe is effected it is likely to occur in the plume (cf. Isley 1995). The oxidised products are carried to the continental margin in the gravity current in which mixing with ocean water is minimal and the velocity is more or less maintained. Stratification of the ocean with respect to P_{O_2} is not an essential ingredient of this model (see text for discussion).

controls. Mn is at low levels in most, but not all, Palaeoproterozoic banded iron formations and probably remains in solution as reduced species (Holland 1984, Nelsen et al. 1987). For those banded iron formations with significant Mn, a higher P_{O_2} is indicated or perhaps a longer transport time, Mn oxidation being notably sluggish (Roy 1992).

For the *hydrothermal plume model* to supply the Fe of the Hamersley Basin requires ridge activity much more vigorous than that observed today, involving relatively high geothermal gradients, and/or higher magmatic temperatures, and/or larger magma bodies. These factors could create larger and hotter sub-seafloor convection systems and thereby higher vent temperatures, higher buoyancy flux, including higher mass flux (see Cathles 1981, 1983) and, most critically, higher Fe contents in the fluids. The recognition of a major basalt-rhyolite province coeval with Hamersley banded iron formations supports the idea of an increased Fe input to the oceans at that time (Barley et al. 1997).

The present-day global flux of Fe^{2+} at mid-ocean ridges is about 2×10^9 kg a year and the fluids at 350°C carry about 100 ppm Fe (composition from Von Damm et al. 1985; fluid flux from Wolery & Sleep 1988). Fluids 50–75°C hotter might carry 1000 ppm Fe and 1000 ppm SiO_2 (Seyfried et al. 1991). Doubling the modern mass flux and focusing the *global* Fe output to the Hamersley Basin yields 4×10^{10} kg Fe a year, requiring nearly 5×10^6 years to deposit the Hamersley banded iron formations, assuming 50% from terrestrial sources, just matching the upwelling model. The hydrothermal Fe supply could have been further augmented if Fe > H_2S in early Precambrian vent fluids (Walker & Brimblecombe 1985, Wolery & Sleep 1988, Holland & Kasting 1992), limiting the proportion of Fe^{2+} trapped as sulphides in the chimneys and mounds (see below), a condition that might arise partly because of the high temperature, partly because of a lower oceanic sulphate content (and hence lower contribution to the vent fluids via sulphate reduction), and also possibly because the source mafic rocks were sulphide-undersaturated. Total oxidation of modern hydrothermal reductants is possible at a P_{O_2} of 10^{-3} PAL (Holland 1992b) so even the enhanced outputs envisaged here would only require an oxygen pressure near 10^{-2} PAL, even with abundant Mn deposition. Increased terrestrial input of Fe to offshore basins compared to the present is also likely, because during glaciations (see *Proterozoic-Phanerozoic*) there may be an increased supply of aeolian dust, as recorded in Antarctic ice formed during the last maximum glaciation (Kumar et al. 1995).

In modern mid-ocean ridge vents the escaping Fe^{2+} (i.e. the fraction not trapped in basal mounds) is precipitated as sulphide (mostly pyrrhotite) almost immediately on mixing with ambient seawater (Fe and H_2S are approximately balanced, Edmond et al. 1982). It is then carried upwards by the plume, only a few per cent being lost through fallout from the plume flanks (Converse et al. 1984). Rising a few hundred metres in the plume the sulphide oxidises to various hydroxides, for example:

$$FeS + 6H_2O \rightarrow FeOOH + 4.5H_2 + 2H^+ + SO_4^{2-}$$

probably forming submicron size particles like those found by Nelsen et al. (1987) over the Mid-Atlantic Ridge. Excess Fe^{2+} in plumes could form Fe hydroxides directly, for example:

$$Fe^{2+} + 2H_2O \rightarrow FeOOH + 2H^+ + 0.5H_2.$$

If the vent fluids entered anoxic, sulphate-free water, thus inhibiting chimney and hence mound formation (Goldfarb et al. 1983), then almost the entire Fe content of the fluid might be vented. Silica would also have precipitated on cooling, perhaps as silica gel, albeit more slowly, due to kinetic effects (Rimstidt & Barnes 1980). The reduced Fe must eventually be oxidised, requiring at least an oxygenated ocean layer, but a stratified ocean is not required as in the upwelling model. Isley (1995) calculated a supply of 2.5×10^{10} kg Fe^{2+} a year to the depositional site by first removing H_2S and various oxidants from the plume fluids, and allowing oxidation in the gravity current near the continental margin. However, pyrrhotite deposition would be immediate upon quenching even in anoxic conditions, and particulate sulphides would be transported (with any excess Fe^{2+}) until oxidation occurred on entering oxygenated water. Spreading laterally at depth on becoming neutrally buoyant on a pycnocline, or at the surface, the horizontal velocity of the gravity current would be maintained almost indefinitely in a calm ocean (provided the mass flux was maintained at source) because mixing with ambient water is minimal, but in modern oceans the plumes are bent by currents and Coriolis forces, thereby focusing the lateral spread of the current (e.g. Edmond et al. 1982). The particulate-rich fluids may have spread over distances of the order of 10^3–10^4 km from the source, like modern vent fluids (as evidenced by their ^3He and Mn contents, Lupton et al. 1980, Klinkhammer & Hudson 1986). Deposition presumably occurred when current velocities fell below particle settling velocities, but there is no obvious explanation in this model for the finer scale banding of the Transvaal and Hamersley banded iron formations.

The lack or scarcity of silica in Phanerozoic ironstones can be explained in both plume and upwelling models by the emergence of silica-secreting organisms at the beginning of the Phanerozoic, assuming there was sufficient time during Fe transport in the gravity current model for the biota to consume most of the silica. The sustained surge in Fe and other nutrients into the ocean waters entrained in the plume probably accelerated both consumption and overall biomass (e.g. Frost 1996 and related papers).

Tectonics

During the Late Archaean/Palaeoproterozoic period, continental masses developed of sufficient size to support long-lived and extensive basins. Wide continental margins originated during continental breakup, providing for the first time the extensive, stable, clastic-starved basins in which banded iron formations accumulated (Gross 1983, Schissel & Aro 1992, Alibert & McCulloch 1993). The combination of large marginal basins with abundant ocean ridge activity and appropriate chemical conditions in the oceans could have set up ideal conditions for formation of the giant Proterozoic Fe deposits. The marine transgressions recognised by Klein & Beukes (1989) in the Transvaal Supergroup would be more likely products of increased ocean spreading. Buhn et al. (1992) proposed that the Neoproterozoic banded iron formation and Mn deposits at Otjosondu in Namibia are related to opening of the Khomas Sea.

On the southern margin of the Pilbara Block the Fortescue Group basalt lavas and dykes (which include some komatiites), aged 2.77–2.69 Ga (Blake & Barley 1992), may be expressions of mantle plume activity, the protracted thermal decay and volcanism continuing during subsidence of the Hamersley Basin (Barley et al. 1997), and a similar story may apply to the Ventersdorp lavas of southern Africa and the Transvaal Basin.

Isley (1995) suggested that mid-ocean ridge crests were probably at shallower depths than today, but unless the water was ≥ 3 km deep, fluids such as those discussed above would have boiled below the rock surface, a process likely to have deposited Fe and Cu sulphides, significantly lowering the Fe content of the fluids (see Drummond & Ohmoto 1985).

Sedimentary manganese deposits

Manganese deposits are as old as 3.0 Ga, but are most common in the Mesozoic and Cainozoic (Roy 1992; Fig. 2). Large deposits in sedimentary rocks may have been deposited from upwelling oxygenated water at the margins of more reduced basins in which Mn remained in solution, in many cases following marine transgression accompanied by development of anoxic basins (Roy 1992). The manganese of those deposits accompanying banded iron formations, which include the largest (the 2.2 Ga Kalahari deposit), and even non-banded iron formation-related occurrences, may be ultimately sourced in ocean ridge vents (Fe/Mn ~3 at 21°N, Edmond et al. 1982).

Review of banded iron formation

The upwelling-stratified ocean model accounts fairly well for the distribution of banded iron formations in time, with increased atmospheric P_{O_2} (for which there is independent evidence) inhibiting the development of large anoxic ocean basins from about 1.8 Ga, perhaps combined with increased oceanic pH (Figs 2, 4, 5). However, the fundamental controls over basin formation and size were probably tectonic, with extensive platforms and sufficiently large anoxic basins not developing until the earliest Proterozoic. The Neoproterozoic banded iron formations can also be related to an important period of continental rifting and oceanic basin development. Oscillations in oceanic P_{O_2} related to tectonic cycles of a few hundred million years were probably of minor significance. Attendant surges in biological activity and burial of organic residues might have triggered changes in atmospheric P_{O_2} that led to the Proterozoic glaciations, of which the major were more or less coeval with banded iron formations.

The plume/gravity current model is quite clearly related to tectonic processes, but perhaps more directly than previously realised if the dramatic increase in Fe and nutrient content due to ocean ridge and other volcanic vent activity can drive increased biological activity and export production to anoxic basins to a level sufficient to cause glaciation. The giant banded iron formation provinces are not synchronous worldwide and do not involve entire oceans so there is doubt as to the significance of the hydrothermal and related processes to climate.

Summary

Several features relating to both ore formation and Earth's evolution have emerged in this review.

- The secular stability of the chemical composition of igneous and sedimentary rocks points to the operation of similar tectonic processes to those of the present, and this is supported by the occurrence from 3.5 Ga of mineral deposits having essentially present-day appearance and composition (e.g. volcanic-hosted massive sulphide, porphyry Cu, Sn-W, mesothermal lode Au, etc).
- Secular changes in other ore types (other than due to preservation) follow the long-term thermal and chemical evolution of the Earth, for example:
  (i) the restriction of komatiite-related Ni-Cu ores to the early Precambrian, perhaps because of the cooling mantle, and
  (ii) the lack of deposits such as the stratiform, sediment-hosted Cu-Co, Zn-Pb and U-PGE types before the increase of atmospheric P_{O_2} during the Palaeoproterozoic.
- Mantle heterogeneity does not appear to be important in controlling ore distribution in time or place, nor is there evidence in Au and platinum group element occurrences that these elements were enriched in the Archaean crust or mantle.
- Mantle plumes released from the core/mantle boundary at various stages through Earth's history were probably the direct or indirect (e.g. secondary melting) cause of formation of several ore types (e.g. large platinum group elements, Cr), and large bodies of mantle-derived melts emplaced in the continental crust may be responsible for the development of orogenic as well as anorogenic Sn-W provinces.
- Long-term changes in tectonic style have profoundly affected variation in mineralisation. For example, extensive passive margins and large intracratonic basins developed during the Late Archaean and Palaeoproterozoic as sialic crustal masses enlarged and began to interfere with their neighbours, changes that allowed formation of the giant Palaeoproterozoic banded iron formations and Late Archaean U-Au conglomerates, and, from 2.0 Ga, sediment-hosted Pb-Zn and Cu-Co ores and hydrothermal U deposits.
- Tectonic cycles of a few hundred million years, involving compilation, fracturing and breakup of continental masses, account for much of the second-order (a few hundred m.y.) variation in secular occurrence of, for example, volcanic-hosted massive sulphide, syndeformational lode Au, porphyry Cu and Sn-W ores, and also sedimentary Mn and phosphate deposits.
- There was probably substantial sulphate in at least some deep and shallow oceans from at least 3.5 Ga, a finding based on the occurrences of evaporitic sulphate, the sulphur isotope studies of Ohmoto (1992) and Ohmoto et al. (1993), and the presence of volcanic-hosted massive sulphide deposits in the eastern Pilbara Block; sulphates then appear intermittently in mineral deposits until about 1.4 Ga, from which time bedded evaporitic sulphates are common.
- The $\delta^{34}S$ value of oceanic sulphate was about 1.0–1.5 per mil from 3.5 Ga into the Palaeoproterozoic, after which it rose rapidly (and the sulphate content probably also increased), reaching Phanerozoic values by about 1.7 Ga.
- Atmospheric P_{O_2} probably increased over the period 2.3–2.0 Ga, based on palaeosol and palaeontological evidence, together with the appearance of redbeds and disappearance of significant detrital uraninite and pyrite. Both the upwelling current and hydrothermal plume models for the origin of banded iron formation are viable, but we raise the possibility that surges in banded iron formation related to intense hydrothermal vent activity might possibly have contributed to atmospheric composition rather than being controlled by it.

Acknowledgments

We are grateful to Mike Barley, Ron Berry, Clive Burrett, Gary Davidson, Chris Eastoe, Dave Huston, Subhash Jaireth, Ian Lambert, Trevor McDougall, Dick Morris, Mike Russell, Don Sangster and Roger Skirrow for suggesting improvements to the manuscript.

References

Albers, J.P. & Bain, J.H.C., 1985. Regional setting and new information on some critical geological features of the West Shasta district, California. Economic Geology, 80, 2072–2091.

Alibert, C. & McCulloch, M.T., 1993. Rare earth element and neodymium isotopic compositions of the banded iron-formations and associated shales from Hamersley, Western Australia. Geochimica et Cosmochimica Acta, 57, 187–204.

Anderson, I.K., Andrew, C.J., Ashton, J.H., Boyce, A.J., Caulfield, J.B.D., Fallick, A.E. & Russell, M.J., 1989. Preliminary sulphur isotope data of diagenetic and vein sulphides in the Lower Palaeozoic strata of Ireland and southern Scotland: implications for Zn + Pb + Ba mineralisation. Journal of the Geological Society of London, 146, 715–720.

Ayres, D.E., 1972. Genesis of iron-bearing minerals in banded iron formation mesobands in the Dale Gorge member, Hamersley Group, Western Australia. Economic Geology, 67, 1214–1233.

Barley, M.E., 1992. A review of Archean volcanic-hosted massive sulfide and sulfate mineralization in Western Australia. Economic Geology, 87, 855–872.

Barley, M.E., 1993. Volcanic, sedimentary and tectonostratigraphic environments of the 3.46 Ga Warrawoona Megasequence: a review. Precambrian Research, 60, 47–67.

Barley, M.E. & Groves, D.I., 1992. Supercontinent cycles and the distribution of metal deposits through time. Geology, 20, 291–302.

Barley, M.E., Eisenlohr, B., Groves, D.I., Perring, C.S. & Vearncombe, J.R., 1989. Late Archaean convergent margin tectonics and gold mineralization: a new look at the Norseman-Wiluna belt, Western Australia. Geology, 17, 826–829.

Barley, M.E., Groves, D.I., Krapez, B. & Kerrich, R., 1995. The Late Archean: a major anomaly in global metallogeny. Abstracts Tectonics and Metallogeny of Early/Mid Precambrian Orogenic Belts, Montreal, 1995.

Barley, M.E., Pickard, A.L. & Sylvester, P.J., 1997. Emplacement of a large igneous province as a possible cause of banded iron formation 2.45 billion years ago. Nature, 385, 55–58.

Beeson, R., 1990. Broken Hill-type lead-zinc deposits — an overview of their occurrence and geological setting. Transactions of the Institution of Mining and Metallurgy, 99, B163–B175.

Beukes, N.J., Klein, C., Kaufman, A.J. & Hayes, J.M., 1990. Carbonate petrography, kerogen distribution, and carbon and oxygen isotope variations in an Early Proterozoic transition from limestone to iron-formation deposition, Transvaal Supergroup, South Africa. Economic Geology, 85, 663–690.

BHP Staff, 1975. Koolyanobbing iron ore deposits, W. A. Australasian Institute of Mining and Metallurgy, Monograph 5 (1), 940–942.

Bickle, M.J., Bettanay, L.F., Chapman, H.J., Groves, D.I., McNaughton, Campbell, I. H & de Laeter, J.R,. 1993. Origin of the 3500–3300 Ma calc-alkaline rocks in the Pilbara Archaean: isotopic and geochemical constraints from the Shaw Batholith. Precambrian Research, 60, 117–149.

Blake, T.S. & Barley, M.E., 1992. Tectonic evolution of the Late Archaean to Early Proterozoic Mount Bruce Megasequence Set, Western Australia. Tectonics, 11, 1415–1425.

Blockley, J.G., 1980. The tin deposits of Western Australia. Geological Survey of Western Australia, Mineral Resources Bulletin 184.

Buhn, B., Stanistreet, I.G. & Okrusch, M., 1992. Late Proterozoic outer shelf manganese and iron deposits at Otjosondu (Namibia) related to the Damaran oceanic opening. Economic Geology, 87, 1393–1411.

Buick, R. & Dunlop, J.S.R., 1990. Evaporitic sediments of Early Archaean age from the Warrawoona Group, North Pole, Western Australia. Sedimentology, 37, 247–277.

Buick, R., Thornett, J.R., McNaughton, N.J., Smith, J.B., Barley, M.E. & Savage, M., 1995. Record of emergent continental crust ~3.5 billion years ago in the Pilbara craton of Australia. Nature, 375, 574–577.

Byerly, G.R., Lowe, D.R. & Walsh, M.M., 1986. Stromatolites from the 3,300–3,500-Myr-old Swaziland Supergroup, Barberton Mountain Land, South Africa. Nature, 319, 489–491.

Calvert, A.J., Sawyer, E.W., Davis, W.J. & Ludden, J.N., 1995. Archaean subduction inferred from images of a mantle suture in the Superior Province. Nature, 375, 670–674.

Cameron, E.M., 1983. Evidence from early Proterozoic anhydrite for sulphur isotopic partitioning in Precambrian oceans. Nature, 304, 54–56.

Cameron, E.M. & Hattori, K., 1985. The Hemlo gold deposit: a geochemical and isotopic study. Geochimica et Cosmochimica Acta, 49, 2041–2050.

Campbell, I.H. & Griffiths, R.W., 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth and Planetary Science Letters, 99, 79–93.

Cathles, L.M., 1981. Fluid flow and genesis of hydrothermal ore deposits. In: Skinner, B.J. (editor) Economic Geology Anniversary Volume. Economic Geology Publishing Company, El Paso, Texas, 424–457.

Cathles, L.M., 1983. An analysis of the hydrothermal system responsible for massive sulfide deposition in the Hokuroku Basin of Japan. Economic Geology Monograph 5, 439–487.

Condie, K.C., 1994. Greenstones through time. In: Condie, K.C. (editor), Archean crustal evolution. Developments in Precambrian Geology 11, Elsevier, Amsterdam, 85–120.

Converse, D.R., Holland, H.D. & Edmond, J.M., 1984. Flow rates in the axial hot springs of the East Pacific Rise (21°N): implications for the heat budget and the formation of massive sulphide deposits. Earth and Planetary Science Letters, 69, 159–175.

Cook, P.J., Shergold, J.H., Burnett, W.C. & Riggs, S.R., 1990. Phosphorite research, a historical overview. In: Northolt, A.J.G. & Jarvis, I. (editors), Phosphorite research and development. Geological Society of London, 1–22.

Danielson, A., Moller, P. & Dulski, P., 1992. The europium anomalies in banded iron formations and the thermal history of the oceanic crust. Chemical Geology, 97, 89–100.

Davies, G.F., 1995. Punctuated tectonic evolution of the Earth. Earth and Planetary Science Letters, 136, 363–379.

Deb, M., Hoefs, J. & Baumann, A., 1991. Isotopic composition of two Precambrian stratiform barite deposits from the Indian Shield. Geochimica et Cosmochimica Acta, 55, 303–308.

de Ronde, C.E.J. & de Wit, M.J., 1994. Tectonic history of the Barberton greenstone belt, South Africa: 490 million years of Archean crustal evolution. Tectonics, 13, 983–1005.

Des Marais, D.J., 1994a. The Archean atmosphere: its composition and fate. In: Condie, K.C. (editor), Archean crustal evolution. Developments in Precambrian Geology 11, Elsevier, Amsterdam, 505–523.

Des Marais, D.J., 1994b. Tectonic control of the crustal organic carbon reservoir during the Precambrian. Chemical Geology, 114, 303–314.

Des Marais, D.J., Strauss, H., Summons, R.E. & Hayes, J.M., 1992. Carbon isotope evidence for the stepwise oxidation of the Proterozoic environment. Nature, 359, 605–609.

Dove, P.M. & Rimstidt, J.D., 1992. Silica-water interaction. Reviews in Mineralogy 29, 259–308.

Drummond, S.E. & Ohmoto, H., 1985. Chemical evolution and mineral deposition in boiling hydrothermal systems. Economic Geology, 80, 126–147.

Eastoe, C.J. & Gustin, M.M., 1996. Volcanogenic massive sulfide deposits and anoxia in the Phanerozoic oceans. Ore Geology Reviews, 10, 179–197.

Eastoe, C.J., Gustin, M.S., Hurlbut, D.F. & Orr, R.L., 1990. Sulfur isotopes in Early Proterozoic massive sulfide deposits: new data from Arizona and implications for ocean chemistry. Precambrian Research, 46, 353–

364.
Edmond, J.M., von Damm, K.L., McDuff, R.E. & Measures, C.I., 1982. Chemistry of hot springs on the East Pacific Rise and their effluent dispersal Nature, 297, 187–191.
Etheridge, M.A., Rutland, R.W.R. & Wyborn, L.A.I., 1987. Orogenesis and tectonic processes in the Early to Middle Proterozoic of northern Australia. In: Kroner, A. (editor), Proterozoic lithospheric evolution. American Geophysical Union Geodynamics Series 17, 131–147.
Ewers, W.E. & Morris, R.C., 1981. Studies of the Dales Gorge member of the Brockman Iron Formation, Western Australia. Economic Geology, 76, 1929–1953.
Franklin, J.M., 1986. Volcanic-associated massive sulfide deposits an update. In: Andrew, C.J., Crowe, R.W.A., Finlay, S., Penell, W.M. & Pyne, J.F. (editors), Geology and genesis of mineral deposits of Ireland. Irish Association for Economic Geology, 49–69.
Franklin, J.M., Lydon, J.W. & Sangster, D.F., 1981. Volcanic-associated massive sulfide deposits. In: Skinner, B.J. (editor), Economic Geology Anniversary Volume. Economic Geology Publishing Company, El Paso, Texas, 485–627.
Frost, B.W., 1996. Phytoplankton bloom on iron rations. Nature, 383, 475–476.
Goldfarb, M.S., Converse, D.R., Holland, H.D. & Edmond, J.M., 1983. The genesis of hot spring deposits on the East Pacific Rise, 21°N. Economic Geology Monograph 5, 184–197.
Goodfellow, W.D., 1987. Anoxic stratified oceans a source of sulphur in sediment-hosted stratiform Zn-Pb deposits (Selwyn Basin, Yukon, Canada). Chemical Geology, 65, 359–382.
Graham, J.B., Dudley, R., Aguilar, N.M. & Gans, C., 1995. Implications of the late Palaeozoic oxygen pulse for physiology and evolution. Nature, 375, 117–120.
Green, G.R., Solomon, M. & Walshe, J L, 1981. The formation of the volcanic-hosted massive sulfide ore deposit at Rosebery, Tasmania. Economic Geology, 76, 304–338.
Gross, G.A., 1983. Tectonic systems and the deposition of iron-formation. Precambrian Research, 20, 171_{187}.
Grotzinger, J.P. & Ingersoll, R.V., 1992. Proterozoic sedimentary basins. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 47–50.
Grotzinger, J.P. & Kasting, J.F., 1993. New constraints on Precambrian ocean composition. Journal of Geology, 101, 235–243.
Groves, D.I., Korkiakoski, E.A., McNaughton, N.J., Lesher, C.M. & Cowden, C.A., 1986. Thermal erosion by komatiites at Kambalda and genesis of nickel ores. Nature, 319, 136–139.
Guilbert, J.M. & Park, C.F.Jr, 1986. The geology of ore deposits. W H Freeman and Co., New York.
Gustafson, L.B. & Williams, N., 1981. Sediment-hosted stratiform deposits of copper, lead and zinc. In: Skinner, B.J. (editor), Economic Geology 75th Anniversary Volume. Economic Geology Publishing Company, El Paso, 139–178.
Han, T.M. & Runnegar, B., 1992. Megascopic eukaryotic algae from the 2.1 billion-year-old Negaumee Iron-formation, Michigan. Science, 257, 232–235.
Hayes, J.M., Lambert, I.B. & Strauss, H., 1992. The sulfur-isotopic record. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 129–132.
Haynes, D.W., Cross, K.C., Bills, R.T. & Reed, M.H., 1995. Olympic Dam ore genesis: a fluid-mixing model. Economic Geology, 90, 281-307.
Hinman, M., Wall, V.J. & Heinrich, C.A., 1994. The interplay between sedimentation, deformation and hydrothermal activity at the McArthur Pb-Zn (-Cu) deposit. Geological Society of Australia Abstracts no. 37, 176–177.
Hoffman, P.F., 1989. Speculations on Laurentia's first gigayear (2.0 to 1.0 Ga). Geology, 17, 135–138.
Holland, H.D., 1973. The oceans: a possible source of iron in iron-formation. Economic Geology, 68, 1169–1172.
Holland, H.D., 1984. The chemical evolution of the atmosphere and oceans. Princeton University Press, Princeton, New Jersey, 351 pp.
Holland, H.D., 1992a. Distribution and paleoenvironmental interpretation of Proterozoic paleosols. In: Schopf, J.W. & C Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 153–155.
Holland, H D, 1992b. Chemistry and evolution of the Proterozoic ocean. In: Schopf, J.W. & C Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 169–172.
Holland, H.D. & Beukes, N.J., 1990. A paleoweathering profile from Griqualand West, South Africa: evidence for a dramatic rise in atmospheric oxygen between 2.2 and 1.9 BYBP. American Journal of Science, 290-A, 1–34.
Holland, H.D. & Kasting, J.F., 1992. The environment of the Archean Earth. In: Schopf, J.W. & C Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 21–24.
Holser, W.T., Schidlowski, M., Mackenzie, F.T. & Maynard, J.B., 1988. Geochemical cycles of carbon and sulfur in chemical cycles. In: Gregor, C.B., Garrels, R.M., Mackenzie, F.T. & Maynard, J.B. (editors), The evolution of the Earth. John Wiley and Sons, New York, 105–173.
Hutchinson, R.W., 1981. Mineral deposits as guides to supracrustal evolution. In: O'Connel, R.J. & Fyfe, W.S. (editors), Evolution of the Earth. American Geophysical Union, Geodynamics Series 5, 120–132.
Hutchinson, R.W., 1992. Mineral deposits and metallogeny: indicators of Earths evolution. In: Schidlowski, M., Golubic, S., Kimberley, M.M., McKirdy, D.M. & Trudinger, P.A. (editors), Early organic evolution: implications for mineral and energy resources. Springer Verlag, Berlin, 521-545.
Idnurm, M., Giddings, J.W., Plumb, K.A. & Wyborn, L.A.I., 1994. Exploration implications of palaeomagnetic overprint chronology in the McArthur Basin. Australasian Institute of Mining and Metallurgy Annual Conference 1994, 153–156.
Isley, A.E., 1995. Hydrothermal plumes and the delivery of iron to banded iron formation. Journal of Geology, 103, 169–185.
Jackson, M.J., Muir, M.D. & Plumb, K.A., 1987. Geology of the southern McArthur Basin, Northern Territory. Bureau of Mineral Resources, Australia, Bulletin 220.
Jochum, K.P., Hofmann, A.W. & Seufert, H.M., 1993. Tin in mantle-derived rocks: constraints on Earth evolution. Geochimica et Cosmochimica Acta, 57, 3585–3595.
Kasting, J.F., 1993. Earth's early atmosphere. Science, 259, 920–926.
Kasting, J.F., Eggler, D.H. & Raeburn, S.P., 1993. Mantle redox evolution and the oxidation state of the Archean atmosphere. Journal of Geology, 101, 245–257.
Kempe, A. & Degens, E.T., 1985 An early soda ocean? Chemical Geology, 53, 95–108.
Khin Zaw, Gemmell, J.B., Large, R.R., Mernagh, T.P. & Ryan, C.G., 1996. Evolution and source of ore fluids in the stringer system, Hellyer VHMS deposit, Tasmania, Australia: evidence from fluid inclusion microthermometry and geochemistry. Ore Geology Reviews, 10,

251–278.
Kimura, K., Lewis, R.S. & Anders, E., 1974. Distribution of gold and rhenium between nickel-iron and silicate melts: implications for the abundances of siderophile elements on the Earth and Moon. Geochimica et Cosmochimica Acta, 38, 683–701.
Kirkham, R.V., 1989. Distribution, settings, and genesis of sediment-hosted stratiform copper deposits. Geological Association of Canada Special Paper 36, 3–38.
Kirkham, R.V. & Roscoe, S.M., 1993. Atmospheric evolution and ore deposit formation. Resource Geology Special Issue, 15, 1–17.
Kirschivink, J.L., 1992. Late Proterozoic low-latitude global glaciation: the snowball Earth. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 51–52.
Klein, C. & Beukes, N.J., 1989. Geochemistry and sedimentology of a facies transition from limestone to iron-formation deposition in the Early Proterozoic Transvaal Supergroup, South Africa. Economic Geology, 84, 1733–1774.
Klein. C. & Beukes, N.J., 1992. Time distribution, stratigraphy, and sedimentologic setting, and geochemistry of Precambrian iron-formations In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 139–146.
Klinkhammer, G. & Hudson, A., 1986. Dispersal patterns for hydrothermal plumes in the South Pacific using manganese as a tracer. Earth and Planetary Science Letters, 79, 241–249.
Krupp, R., Oberthur, T. & Hirdes, W., 1994. The early Precambrian atmosphere and hydrosphere: thermodynamic constraints from mineral deposits. Economic Geology, 89, 1581–1598.
Kumar, N., Anderson, R.F., Mortlock, R.A., Froelich, P.N., Kubik, P., Dittrich-Hannen, B. & Suter, M., 1995. Increased biological productivity and export production in the glacial Southern Ocean. Nature, 378, 675–680.
Lambert, I.B. & Donnelly, T.H., 1992. Global oxidation and a supercontinent in the Proterozoic: evidence from stable isotopic trends. In: Schidlowski, M., Golubic, S., Kimberley, M.M., McKirdy, D.M. & Trudinger, P.A. (editors), Early organic evolution: implications for mineral and energy resources. Springer Verlag, Berlin, 408–413.
Lambert, I.B., Donnelly, T.H., Dunlop, J.S.R. & Groves, D.I., 1978. Stable isotopic compositions of early Archaean sulphate deposits of probable evaporitic and volcanogenic origins. Nature, 276, 808–811.
Lambert, I.B., Beukes, N.J., Klein, C. & Veizer, J., 1992. Proterozoic mineral deposits through time In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 59–62.
Large, R.R., Doyle, M., Raymond, O., Cooke, D., Jones, A. & Heasman, L., 1996. Evaluation of the role of Cambrian granites in the genesis of world class VHMS deposits in Tasmania. Ore Geology Reviews.
Logan, G.A., Hayes, J.M., Hieshima, G.B. & Summons, R.E., 1995. Terminal Proterozoic reorganization of biogeochemical cycles. Nature, 376, 53–56.
Loutit, T.S., Wyborn, L.A.I., Hinman, M.C. & Idnurm, M., 1994. Palaeomagnetic, tectonic, magmatic and mineralization events in the Proterozoic of northern Australia. Australasian Institute of Mining and Metallurgy Annual Conference 1994, 123–128.
Lowe, D R, 1992a. Major events in the geological development of the Precambrian Earth. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 67–75.
Lowe, D R, 1992b. Other geological indicators. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 157–158.
Lowe, D.R. & Ernst, W.G., 1992. The Archean geologic record. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 13–19.
Lupton, J.E., Klinkhammer, G.P., Normark, W.R., Haymon, R., Macdonald, K.C., Weiss, R.F. & Craig, H., 1980. Helium-3 and manganese at the 21°N East Pacific Rise hydrothermal site. Earth and Planetary Science Letters, 50, 115–127.
Mackenzie, D.E., Black, L.P. & Sun, S.S., 1988. Origin of alkali-feldspar granites: an example from the Poimena Granite, northeastern Tasmania, Australia. Geochimica et Cosmochimica Acta, 52, 2507–2524.
Maher, K.A. & Stevenson, D.J., 1988. Impact frustration of the origin of life. Nature, 331, 612–614.
Martin, J.H., Coale, K.H. et al., 1994. Testing the iron hypothesis in ecosystems of the equatorial Pacific Ocean. Nature, 371, 123–129.
McCulloch, M.T., 1993. The role of subducted slabs in an evolving earth. Earth and Planetary Science Letters, 115, 89–100.
McDonald, I., de Wit, M.J., Smith, C.B., Bizzi, L.A. & Viljoen, K.S., 1995. The geochemistry of the platinum-group elements in Brazilian and southern African kimberlites. Geochimica et Cosmochimica Acta, 59, 2883–2903.
McDonough, W.F. & Sun, S.S., 1995. The composition of the Earth. Chemical Geology, 120, 223–253.
Meyer, C., 1985. Ore metals through geologic history. Science, 227, 1421–1428.
Meyer, C., 1988. Ore deposits as guides to geologic history of the Earth. Annual Reviews of the Earth Sciences, 16, 147–171.
Morris, R.C., 1993. Genetic modelling for banded iron-formation of the Hamersley Group, Pilbara Craton, Western Australia. Precambrian Research, 60, 243–286.
Morris, R.C. & Horwitz, R.C., 1983. The origin of the iron-formation-rich Hamersley Group of Western Australia deposition on a platform. Precambrian Research, 21, 273–297.
Morris, R.C. & Trendall, A.F., 1988. Discussion — A model for the deposition of the microbanded Precambrian iron formations. American Journal of Science, 288, 664–673.
Mossman, D.J. & Farrow, C.E.G., 1992. Paleosol and ore-forming processes in the Elliot Lake district of Canada. In: Schidlowski, M., Golubic, S., Kimberley, M.M., McKirdy, D.M. & Trudinger, P.A. (editors), Early organic evolution: implications for mineral and energy resources. Springer Verlag, Berlin, 69–75.
Muir, M.D., Donnelly, T.H., Wilkins, R.W.T. & Armstrong, K.J., 1985. Stable isotope, petrological, and fluid inclusion studies of minor mineral deposits from the McArthur Basin: implications for the genesis of some sediment-hosted base metal mineralisation from the Northern Territory. Australian Journal of Earth Sciences, 32, 239–260.
Myers, J.S., 1993. Precambrian history of the West Australian craton and adjacent orogens. Annual Review of Earth and Planetary Science, 21, 453–485.
Myers, J.S., 1995. The generation and assembly of an Archaean supercontinent: evidence from the Yilgarn craton, Western Australia. In: Coward, M.P. & Ries, A.C. (editors), Early Precambrian processes. Geological Society of London Special Publication 95, 143–154.
Naldrett, A.J., Fedorenko, V.A., Lightfoot, P.C., Kunilov, V.I., Gorbachev, N.S., Doherty, W. & Johan, Z., 1995. Ni-Cu-PGE deposits of Norilsk region, Siberia: their formation in conduits for flood basalt volcanism. Trans-

actions of the Institution of Mining and Metallurgy, 104, B18–B36.
Nelsen, T.A., Klinkhammer, G.P., Trefry, J.M. & Trocine, R.P., 1987. Real-time observation of dispersed hydrothermal plumes using nephelometry: examples from the Mid-Atlantic Ridge. Earth and Planetary Science Letters, 81, 245–252.
Ohmoto, H., 1992. Biogeochemistry of sulfur and the mechanisms of sulfide-sulfate mineralization in Archean oceans In: Schidlowski, M., Golubic, S., Kimberley, M.M., McKirdy, D.M. & Trudinger, P.A. (editors), Early organic evolution: implications for mineral and energy resources. Springer Verlag, Berlin, 378–397.
Ohmoto, H., 1996. Formation of volcanogenic massive sulfide deposits: the Kuroko perspective. Ore Geology Reviews, 10, 135–177.
Ohmoto, H. & Felder, R.P., 1987. Bacterial activity in the warmer, sulphate-bearing, Archaean oceans. Nature, 328, 244–246.
Ohmoto, H., Drummond, S.E., Eldridge, C.S., Pisutha-Arnond, V. & Lenagh, T.C., 1983. Chemical processes of ore formation. Economic Geology Monograph 5, 570–604.
Ohmoto, H., Kaiser, C.J. & Geer, K.A., 1990. Systematics of sulphur isotopes in recent marine sediments and ancient sediment-hosted basemetal deposits. In: Herbert, H.K. & Ho, S.E. (editors), Stable isotopes and fluid processes in mineralization. University of Western Australia, Geology Department and University Extension Publication 23, 70–120.
Ohmoto, H., Kakegawa, T. & Lowe, D.R., 1993. 3.4-billion-year-old biogenic pyrites from Barberton, South Africa: sulfur isotope evidence. Science, 262, 555–557.
Perry, E.C., Monster, J. & Reimer, T., 1971. Sulfur isotopes in Swaziland System barites and the evolution of the Earths atmosphere. Science, 171, 1015–1016.
Pisutha-Arnond, V. & Ohmoto, H., 1983. Thermal history, and chemical and isotopic compositions of the ore-forming fluids responsible for the Kuroko massive sulfide deposits in the Hokuroko district of Japan. Economic Geology Monograph 5, 523–558.
Rao, T.G. & Naqvi, S.M., 1995. Geochemistry, depositional environment and tectonic setting of the BIFs of the Late Archaean Chitradurga Schist Belt, India. Chemical Geology, 121, 217–243.
Reimer, T.O., 1990. Archaean baryte deposits of southern Africa. Journal of the Geological Society of India, 35, 131–150.
Rickard, D.T., Zweifel, H. & Donnelly, T.H., 1979. Sulfur isotope systematics in the Esen pyrite-barite deposits, Skellefte district, Sweden. Economic Geology, 74, 1060–1068.
Rimstidt, J.D. & Barnes, H.L., 1980. The kinetics of silica-water reactions. Geochimica et Cosmochimica Acta, 44, 1683–1699.
Robb, L.J. & Meyer, F.M., 1990. The nature of the Witwatersrand hinterland: conjectures on the source area problem. Economic Geology, 85, 511–536.
Rona, P.A., 1988. Hydrothermal mineralization at oceanic ridges. Canadian Mineralogist, 26, 431–465.
Roy, S., 1992. Environments and processes of manganese deposition. Economic Geology, 87, 1218–1217.
Runnegar, B., 1993. Algae and oxygen in Earths ancient atmosphere. Science, 259, 835.
Russell, M.J., 1996. The generation at hot springs of sedimentary ore deposits, microbialites and life. Ore Geology Reviews, 10, 199–214.
Russell, M.J., Daniel, R.M. & Hall, A.J., 1993. On the emergence of life via catalytic iron-sulphide membranes. Terra Nova, 5, 343–347.
Sawkins, F.J., 1976. Metal deposits related to intracontinental hotspot and rifting environments. Journal of Geology, 84, 653–671.
Sawkins, F.J., 1989. Anorogenic felsic magmatism, rift sedimentation and giant Proterozoic Pb-Zn deposits. Geology, 17, 657–60.
Sawkins, F.J., 1990a. Metal deposits in relation to tectonics. Second edition. Minerals and Rocks 17. Springer Verlag, Berlin, 461 pp.
Sawkins, F.J., 1990b. Integrated tectonic-genetic model for volcanic-hosted massive sulfide deposits. Geology, 18, 1061–1064.
Schidlowski, M., 1986. Evolution of the early sulphur cycle. In: Geochemistry of the Earth surface and processes of mineral formation, Proceedings of International Meeting, Granada, Spain, 1986.
Schidlowski, M., 1993. The initiation of biological processes on Earth: summary of empirical evidence. In: Engel, M.H. & Macko, S.A. (editors), Organic geochemistry. Plenum Press, New York, 639–655.
Schissel, D. & Aro, P., 1992. The major Early Proterozoic sedimentary iron and manganese deposits and their tectonic setting. Economic Geology, 87, 1367–1374.
Schopf, J.W., 1992. Paleobiology of the Archean. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere. Cambridge University Press, Cambridge, 25–39.
Schopf, J.W. & Packer, B.M., 1987. Early Archaean (3.3-billion to 3.5-billion-year-old) microfossils from the Warrawoona Group, Australia. Science, 237, 70–73.
Seccombe, P.K., Spry, P.G., Both, R.A., Jones, M.T. & Schiller, J.C., 1985. Base metal mineralization in the Kanmantoo Group, South Australia: a regional sulfur isotope study. Economic Geology, 80, 1824–1841.
Seyfried, W.E.Jr, Ding, K. & Berndt, M.E., 1991. Phase equilibria constraints on the chemistry of hot spring fluids at mid-ocean ridges. Geochimica et Cosmochimica Acta, 55, 3559–3580.
Shock, E.L., McCollom, T. & Schulte, M.D., 1995. Geochemical constraints on chemolithoautotrophic reactions in hydrothermal systems. Origins of Life and Evolution of the Biosphere, 25, 141–159.
Smith, R.L., 1992. Coastal upwelling in the modern ocean. In: Summerhayes, C.P., Prill, W.L. & Emeis, K.C. (editors), Upwelling systems: evolution since the early Miocene. Geological Society of London, Special Publication, 9–28.
Solomon, M., 1981. An introduction to the geology and metallic mineral resources of Tasmania. Economic Geology, 76, 194–208.
Solomon, M. & Groves, D.I., 1994. The geology and origin of Australia's mineral deposits. Geology and Geophysics Monograph 24. Oxford University Press, Oxford, 951 pp.
Solomon, M & Heinrich, C.A., 1992. Are high heat producing granites essential to the origin of giant lead-zinc deposits at McArthur River and Mount Isa, Australia? Exploration and Mining Geology, 1, 85–91.
Solomon, M. & Khin Zaw, in press. The formation on the sea floor of the Hellyer volcanogenic massive sulfide deposit. Economic Geology.
Stein, M. & Hofmann, A.W., 1995. Mantle plumes and episodic crustal growth. Nature, 372, 63–68.
Stevenson, D.J., 1983. The nature of the Earth prior to the oldest known rock record: the Hadean Earth. In: Schopf, J.W. (editor), Earth's earliest biosphere: its origin and evolution. Princeton University Press, Princeton, New Jersey, 32–40.
Strauss, H., 1993. The sulfur isotope record of Precambrian sulfates: new data and a critical evaluation of the existing record. Precambrian Research, 63, 225–246.

Sun, S-s. & Higgins, N.C., 1996. Neodymium and strontium isotope study of the Blue Tier Batholith, NE Tasmania and its bearing on the origin of tin-bearing alkali feldspar granites. Ore Geology Reviews, 10, 339–365.

Sun, S-s., Wallace, D.A., Hoatson, D.M., Glikson, A.Y. & Keays, R.R., 1991. Use of geochemistry as a guide to platinum group element potential of mafic-ultramafic rocks: examples from the west Pilbara Block and Halls Creek mobile zone, Western Australia. Precambrian Research, 50, 1–35.

Tanimura, S., Date, J., Takahashi, T. & Ohmoto, H., 1983. Part II. Stratigraphy and structure of the Hokuroku district. Economic Geology Monograph 5, 24–39.

Taylor, S.R. & McLennan, S.M., 1995. The geochemical evolution of the continental crust. Reviews of Geophysics, 33, 241–265.

Thorpe, R.I., Hickman, A.H., Davis, D.W., Martensen, J.K. & Trendall, A.F., 1992. Conventional U-Pb geochronology of Archaean felsic units in the Marble Bar region, Pilbara craton, Western Australia. Precambrian Research, 56, 169–189.

Titley, S.R., 1991. Phanerozoic ocean cycles and sediment-hosted gold ores. Geology, 19, 645–648.

Titley, S.R. & Beane, R.E., 1981. Porphyry copper deposits. Part I. Geologic settings, petrology and tectogenesis. In: Skinner, B.J. (editor), Economic Geology Anniversary Volume. Economic Geology Publishing Company, El Paso, Texas, 214–235.

Towe, K.M., 1991. Aerobic carbon cycling and cerium oxidation: significance for Archean oxygen levels and banded iron formation. Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section), 97, 113–123.

Tredoux, M., de Wit, M.J., Hart, R.J., Armstrong, R.A., Lindsay, N.M. & Sellschop, J.P., 1989. Platinum group elements in a 3.5 Ga nickel-iron occurrence: possible evidence of a deep mantle origin. Journal of Geophysical Research, 94, 795–813.

Trendall, A.F., 1975. Hamersley Basin. Geological Survey of Western Australia Memoir 2, 119–143.

Trendall, A.F. & Blockley, J.G., 1970. The iron formations of the Precambrian Hamersley Group, Western Australia. Geological Survey of Western Australia, Bulletin 119.

Trudinger, P.A., 1992. Bacterial sulfate reduction: current status and possible origin. In: Schidlowski, M., Golubic, S., Kimberley, M.M., McKirdy, D.M. & Trudinger, P.A. (editors), Early organic evolution: implications for mineral and energy resources. Springer Verlag, Berlin, 368–377.

Vearncombe, S., 1995. Volcanogenic massive sulphide-sulphate mineralisation at Strelley, Pilbara Craton, Western Australia. PhD thesis, University of Western Australia, Perth, 153 pp.

Vearncombe, S., Barley, M., Groves, D.I., McNaughton, N.J., Mikucki, E.J. & Vearncombe, J.R., 1995. 3.26 Ga black smoker-type mineralization in the Strelley Belt, Pilbara Craton, Western Australia. Journal of the Geological Society of London, 152, 587–590.

Veevers, J.J., 1990. Tectonic-climatic supercycle in the billion-year plate tectonic eon: Permian Pangean icehouse alternates with Cretaceous dispersed-continents greenhouse. Sedimentary Geology, 68, 1–16.

Veizer, J., Hoefs, J., Ridler, R.H., Jensen, L.S. & Lowe, D.R. 1989a. Geochemistry of Precambrian carbonates I. Archean hydrothermal systems. Geochimica et Cosmochimica Acta, 53, 845–857.

Veizer, J., Hoefs, J., Lowe, D.R. & Thurston, P.C., 1989b. Geochemistry of Precambrian carbonates II. Archean greenstone belts and Archean sea water. Geochimica et Cosmochimica Acta, 53, 859–871.

Veizer, J., Laznicka, P. & Jansen, S.L., 1989c. Mineralization through geologic time: recycling perspective. American Journal of Science, 289, 484–524.

Von Damm, K.L., Edmond, J.M., Measures, C.I., Grant, B.C., Trull, T., Walden, B. & Weiss, R., 1985. Chemistry of submarine hydrothermal solutions at 21°N East Pacific Rise. Geochimica et Cosmochimica Acta, 49, 2197–2220.

Walker, J.C.G. & Brimblecombe, P., 1985. Iron and sulfur in the pre-biologic ocean. Precambrian Research, 28, 205–222.

Walker, R.N., Gulson, B. & Smith, J.W., 1983. The Coxco deposit. A Proterozoic Mississippi Valley-type deposit in the McArthur River district, Northern Territory, Australia. Economic Geology, 87, 214–249.

Windley, B, 1992. Proterozoic collisional and accretionary orogens. In: Condie, K.C. (editor), Proterozoic crustal evolution. Developments in Precambrian Geology, 10, 419–446. Elsevier, Amsterdam.

Windley, B.F., 1993. Uniformitarianism today: plate tectonics is the key to the past. Journal of the Geological Society of London, 150, 7–19.

Witt, W.K., 1995. Tholeiitic and high-Mg mafic/ultramafic sills in the Eastern Goldfields Province, Western Australia: implications for regional settings. Australian Journal of Earth Sciences, 42, 407–422.

Wolery, T.J. & Sleep, N.H., 1988. Interactions of geochemical cycles with the mantle. In: Gregor, C.B., Garrels, R.M., Mackenzie, F.T. & Maynard, J.B. (editors), Chemical cycles in the evolution of the Earth. John Wiley and Sons, New York, 77–103.

AGSO Journal of Australian Geology & Geophysics. 17(1). 49–54

The mantle beneath Australia

B.L.N. Kennett[1]

The configuration of earthquakes around Australia enables these natural events to be used as probes into the seismic structure of the upper mantle. For the region below northern Australia, the combination of short-period and broad-band observations has enabled the construction of radial velocity profiles for P and S velocities and attenuation. There is a range of evidence for lateral variations in seismic velocity in the lithosphere and the upper mantle transition zone. The lateral variations in structure can be investigated directly by a number of techniques, including the analysis of travel-time residuals and waveform inversion, especially for surface waves. Line profiles have revealed strong contrasts in P velocities across the shield edge in southeastern Australia. Broader scale P wave tomography has extended the definition of three-dimensional structure, especially in the northern part of the continent. Surface wave studies have begun to reveal the three-dimensional variations in shear wave structure beneath the continent by exploiting the records from portable broad-band stations. These results show that some of the contrasts in surface geology between Precambrian and Phanerozoic outcrop are reflected in depth; some structures extend to depths of 200 km or more, e.g. the Mt Isa Block and the New England Block.

Introduction

The surficial geology of the Australian continent is composed of an assemblage of crustal blocks that can be broadly grouped into the Precambrian western and central cratons and the Phanerozoic eastern province. Structural differences between the Precambrian shield and eastern Australia are inferred from surface wave dispersion (cf. Muirhead & Drummond 1991; Denham 1991), and teleseismic travel-time residuals (Drummond et al. 1989), whose origin is due to structures which certainly extend below 100 km depth.

This paper reports on a variety of studies which have been undertaken to try to constrain seismic structure in the lithosphere, asthenosphere and the transition zone beneath. Most of the studies exploit the extensive seismic activity in the belt which runs through Indonesia, New Guinea and its offshore islands, Vanuatu, Fiji and the Tonga–Kermadec zone. Many of the seismic events enable investigation of the upper mantle structure from their refracted arrival recorded at arrays of portable stations in northern Australia (Fig. 1).

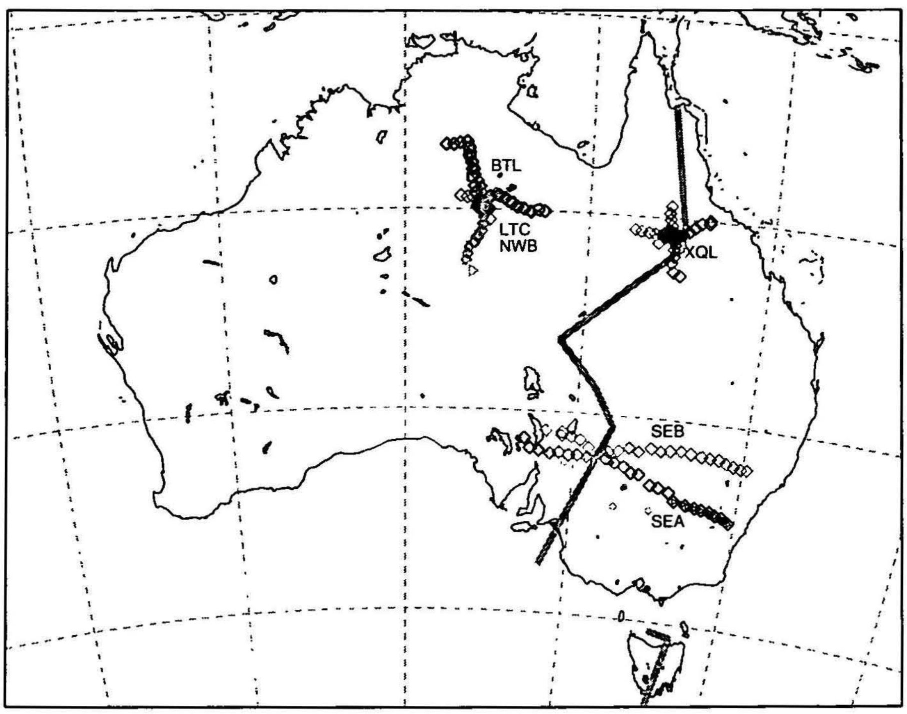


Figure 1. Configuration of short-period seismic array experiments 1985–1992 in relation to the inferred position of the major structural boundary between Precambrian western and central Australia and the Phanerozoic east.

P and S velocity profiles

A number of experiments have been carried out in northern Australia aimed at delineating the major features of the mantle's velocity profile, notably the work of Hales et al. (1980), who used records of Indonesian earthquakes at various of depths from a number of stations in a travel-time analysis. The resulting model is rather complex, with many small discontinuities and low-velocity zones, which may reflect the mapping of three-dimensional structure into a one-dimensional profile. A subsequent reinterpretation by Leven (1985), using comparisons between observed and synthetic seismograms, led to a somewhat simplified structure, but retained a prominent velocity contrast near 210 km depth.

Muirhead & Drummond (1991) have provided an excellent summary of the results of various experiments, using explosive sources, which have provided constraints on P wave velocity structure in several parts of Australia down to 200 km. Control on lithospheric S velocities is somewhat weaker, even though surface wave dispersion (Denham 1991) provides additional information.

The natural sources to the north of Australia are too far away for refracted wave arrivals (for either P or S) to easily constrain shallower structure and so this is inferred from earlier studies or by using local refraction results. For example, Bowman & Kennett (1991) used the aftershocks of the 1988 Tennant Creek earthquakes to investigate regional S wave propagation in central Australia and were also able to infer the velocity profile in the crust and uppermost mantle. Bowman & Kennett (1993) made further use of these aftershocks to develop a set of travel times for P and S waves travelling in the shield structures of western Australia, and to find a compatible velocity model for the lithosphere.

Short-period studies

The Research School of Earth Sciences carried out a sequence of experiments in 1985–1987, using short-period vertical seismometers to record the natural seismicity in the Indone-

[1] Research School of Earth Sciences, Australian National University, Canberra ACT 0200, Australia

sia/New Guinea region. The LTC experiment in 1985 and the NWB experiment in 1986 were based around the Warramunga seismic array near Tennant Creek. The 1987 XQL experiment in northern Queensland linked into the seismic station at Charters Towers.

Dey et al. (1993) summarised the results from the LTC and NWB experiments, which show significant variation in P wave velocity structure in the upper mantle between paths for events along the Flores arc, studied by Bowman & Kennett (1990), and the paths to events in New Guinea. The shallow structure has to be inferred, but the P velocity structure is well constrained from above the base of the lithosphere, near 210 km, down to below the 410 km discontinuity. The interpretation confirms the need for a velocity contrast near 210 km depth (see Fig. 3). The event distribution in the NWB and LTC experiments gives rather limited control on the nature of the 660 km discontinuity.

The analysis to determine the velocity profiles was based on composite record sections, using many events recorded at 20–30 portable recorders equipped with vertical component sensors with a dominant period near 1 Hz. Stacking was used to reinforce the coherent arrivals, with all arrivals within 10 km epicentral distance being combined into a single trace. The influence of variable source time functions was minimised by stacking the envelope of the seismograms. This approach works well for P waves, and the branches associated with the main upper mantle discontinuities can be clearly seen in the record sections. However, the corresponding sections for S waves show a clear arrival associated with the lithosphere, which cannot easily be traced beyond 2000 km, and no branches associated with greater depth.

The XQL experiment in Queensland allowed the investigation of deeper structure, and Cummins et al. (1992) showed that in short-period data there is no obvious signature of the postulated discontinuity near 520 km. Consequently, any P velocity transition will have to be spread over at least 25 km, which would still produce an influence on long period records.

Broad-band studies

A broad-band sensor has been operated at the Warramunga array (WRA) in northern Australia since late in 1988, and over a period of years it has been possible to build up record sections covering the range of interest for the upper mantle by using events in the Indonesia/New Guinea earthquake belt. The records from the permanent station have been augmented by portable broad-band stations deployed up to 300 km from WRA. The surface conditions in this region are such that good results can be obtained for SV waves on radial component records after rotation to the great circle path. The high surface velocities lead to very little contamination by converted P waves; which is a considerable improvement over previous S wave studies of the upper mantle, which have been restricted to SH waves.

Figure 2 shows a composite record section covering the P and S wave components returned from the upper mantle. This section has been constructed from unfiltered radial components associated with many events, and clearly displays the benefit of broad-band recording. The onset of S waves shows high-frequency behaviour (greater than 1 Hz) out to 2000 km, but beyond this distance the S wave arrivals have a significantly lower frequency (0.2 Hz at 3000 km) and this is also seen for later arrivals at shorter distance. The loss of higher frequencies is less pronounced for P waves. The travel-time curves for the upper mantle model illustrated in Figure 3 are superimposed on the record section to aid recognition of the phase onsets.

The change in frequency content for the S waves returned from greater depth has been analysed by Gudmundsson et al. (1994) to determine the attenuation structure with depth under northern Australia. They used the slope of the spectral ratio between P and S wave arrivals on the same record to constrain the differential attenuation between P and S. This differential information can be interpreted with a knowledge of the velocity structure and requires strong attenuation of S waves in the asthenosphere between 210 and 410 km. In a parallel analysis, Kennett et al. (1994) used the composite record sections, together with the earlier information from the short period studies, to build velocity profiles for P and S. These velocity models have been refined by comparison of observed and synthetic seismograms, including the influence of attenuation (Fig. 3). An advantage of this study is that both P and S velocity profiles are determined for the same events and the P/S velocity ratio can be well determined, which is particularly useful for studies of mantle composition. The depth variation

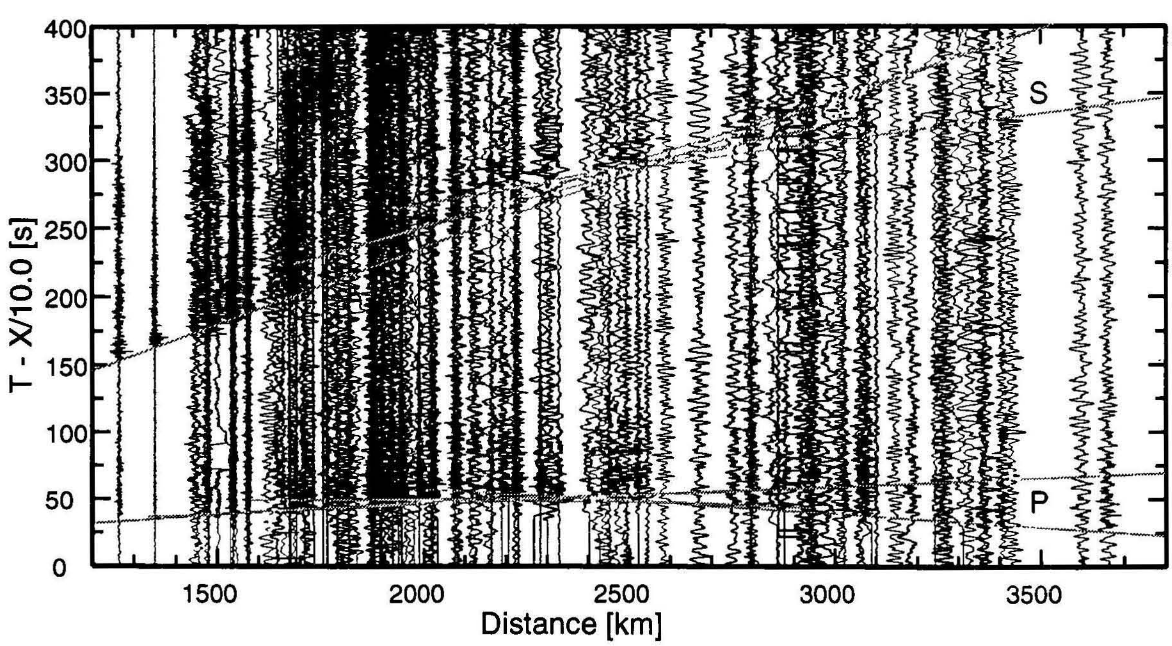


Figure 2. A composite record section of unfiltered broad-band seismograms, recorded at the Warramunga array, for paths beneath northern Australia. The section is constructed from the radial component for each path and covers the P and SV waves returned from the upper mantle. The travel-time curves for the velocity model illustrated in Figure 3 are superimposed on the section.

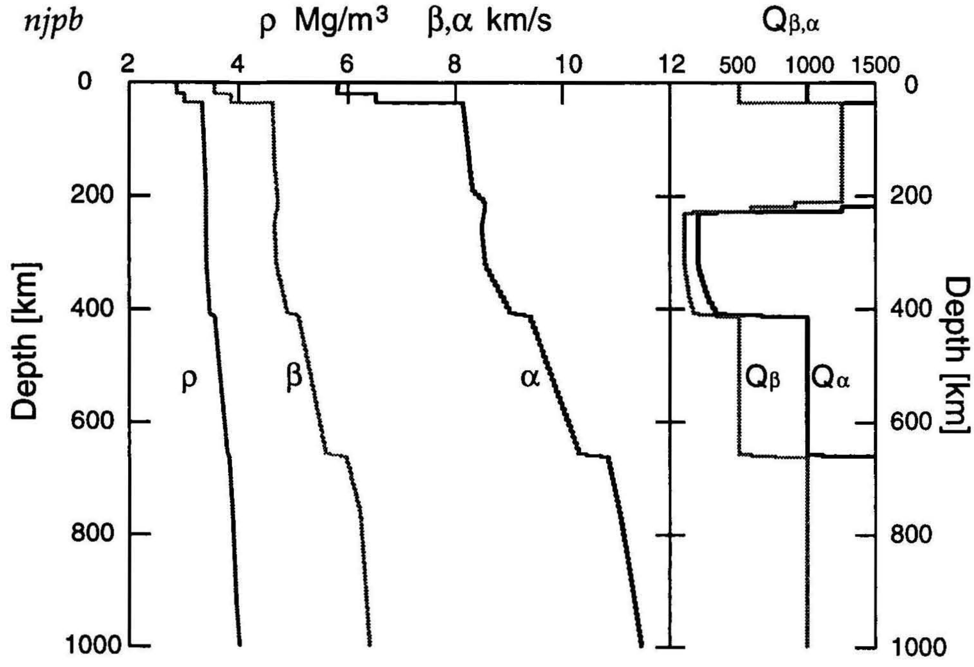


Figure 3. P and S velocity and attenuation structure for the upper mantle beneath northern Australia determined from a combination of short-period and broad-band observations.

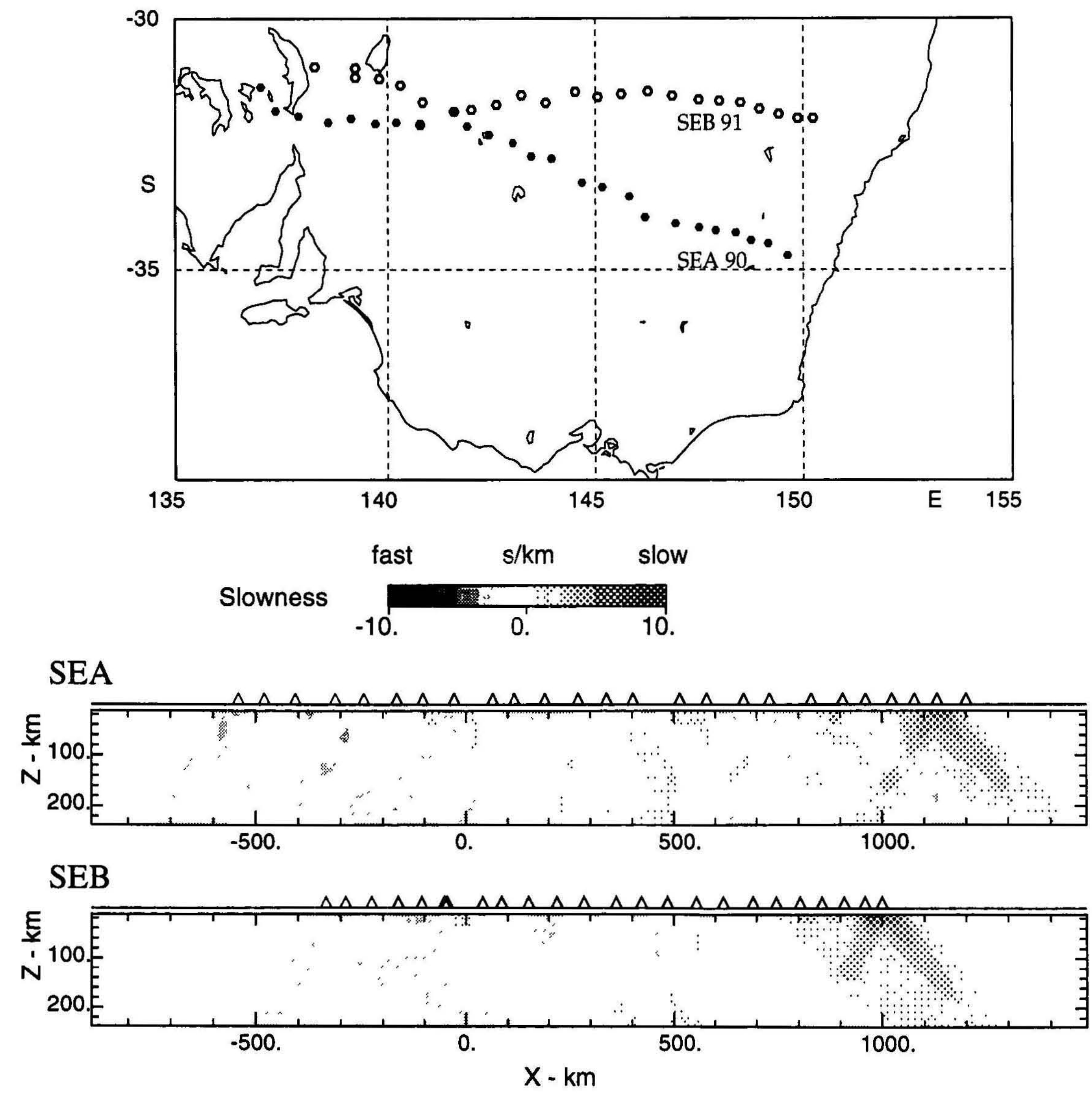


Figure 4. P velocity variations across the shield edge in southeastern Australia determined from the inversion of teleseismic delay times along the SEA and SEB profiles.

of the P/S velocity ratio is in good general agreement with the results for the shield areas of north America obtained by combining the P velocity profile of LeFevre & Helmberger (1989) with the S wave structure of Grand & Helmberger (1984).

Comparison of the radial and tangential components of the S wavefield indicates that for the arrivals from the upper mantle discontinuities there is a systematically earlier arrival on the SH component of more than a second. These indications of seismic anisotropy for the refracted S wave arrivals have been analysed by Tong et al. (1994), who determined the direction of fast propagation and time shift between the S components by correlation analysis. The nature of the observed anisotropy is such that it cannot be explained by structure local to the receivers. A level of anisotropy of the order of 1% in both the lithosphere and the asthenosphere beneath would explain the data quite well.

Three-dimensional structure

The favourable position of the Australian continent relative to world seismicity can be exploited in a number of ways to obtain information on the three-dimensional seismic structure in the mantle.

Body wave studies

One class of experiment that can provide useful information on seismic structure is to determine the systematic patterns of travel-time residuals across an array of portable stations, and then to invert for the structure along the line. In 1990 and 1991, two arrays, 1500 km long, of short-period recorders were deployed to cross the contact between Phanerozoic and Precambrian rocks in southeastern Australia (Fig. 1). The profiles had two stations in common, near Broken Hill, so that the residual patterns could be tied together. In each case, the residual patterns for teleseismic P waves show a systematic trend to early arrivals to the west on the older Precambrian outcrop with a time differential of 0.6–1.0 s between the ends of the profile, depending on the azimuth of the source. These delay patterns confirm the results of the preliminary survey by Cleary et al. (1972), using the nuclear explosion Longshot in the Aleutians. The interpretation of the residuals requires a substantial contrast in seismic properties in the upper part of the mantle, extending to at least 200 km depth, very close to the edge of the Precambrian outcrop (Vahau & Kennett 1995). The two-dimensional structures inferred along the two profiles are illustrated in Figure 4; there is significant structure beneath the Murray Basin, which may be related to the assemblage of the Lachlan Fold Belt.

A further source of information on velocity structure under northern Australia comes from P wave tomography studies, using the sources in the Indonesian and New Guinea region (Widiyantoro & van der Hilst 1996). This work uses the travel-time residuals for P and pP for teleseismic paths, as well as paths to Australian stations (both permanent and portable) in a tomographic inversion for structure in a zone covering the southern Philippines, Malaysia, Indonesia, Papua New Guinea and northern Australia. In order to minimise the influence of structure external to the region, the inversion for P wave velocity follows the approach introduced by Inoue et al. (1990), in which a detailed grid is used for the region of interest and a coarser grid for the region outside. Figure 5 shows a cross section through the P wave model for the depth interval 160–200 km, which shows very strong contrasts between the high velocities in the lithosphere under northern Australia and the lower velocities behind the subduction zone in Indonesia.

Surface wave studies

The configuration of the seismicity around the Australian continent is very favourable for tomographic methods using surface wave trains. The Research School of Earth Sciences has completed a major field-based program to install some 60 portable broad-band stations across Australia over three and a half years (van der Hilst et al. 1994). The project is using a set of up to 12 portable instruments, which occupy sites for 5 months at a time before being moved to a new location, so that a full continental array can eventually be synthesised. The mobility of the arrays has led to the project being named Skippy.

The deployments commenced in May 1993 with 8 stations

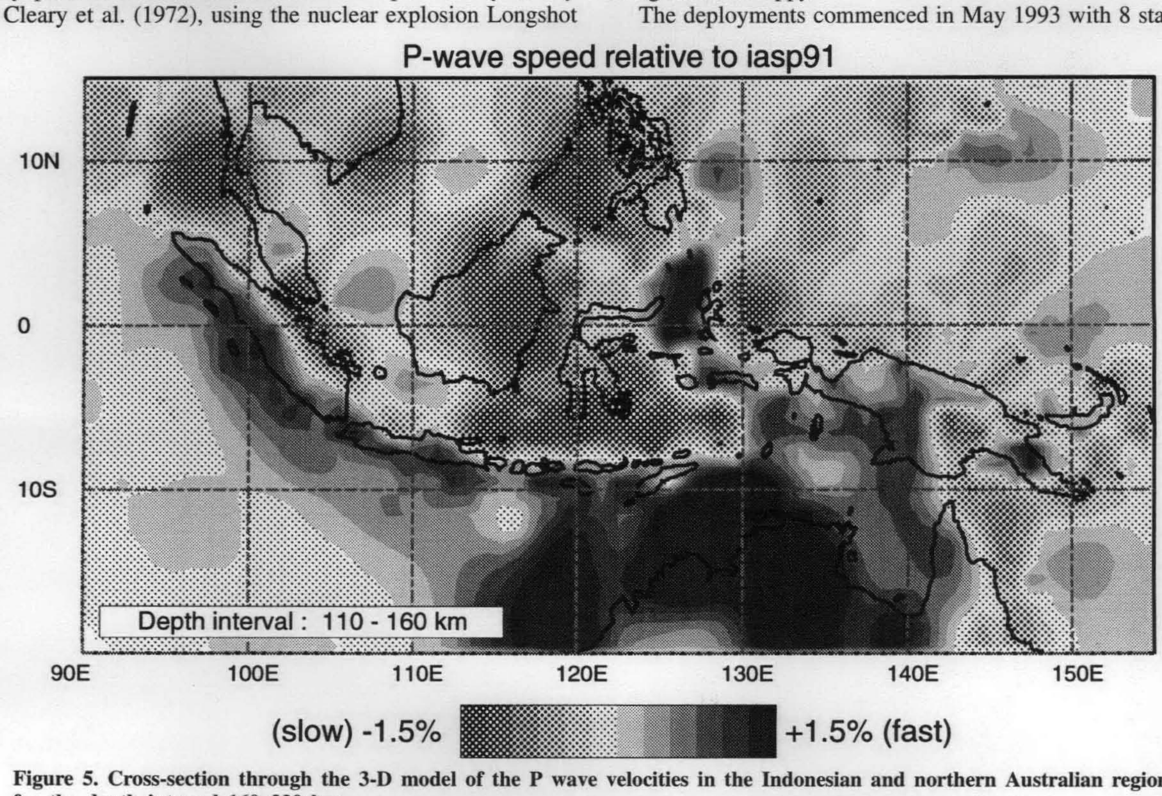


Figure 5. Cross-section through the 3-D model of the P wave velocities in the Indonesian and northern Australian region for the depth interval 160–220 km

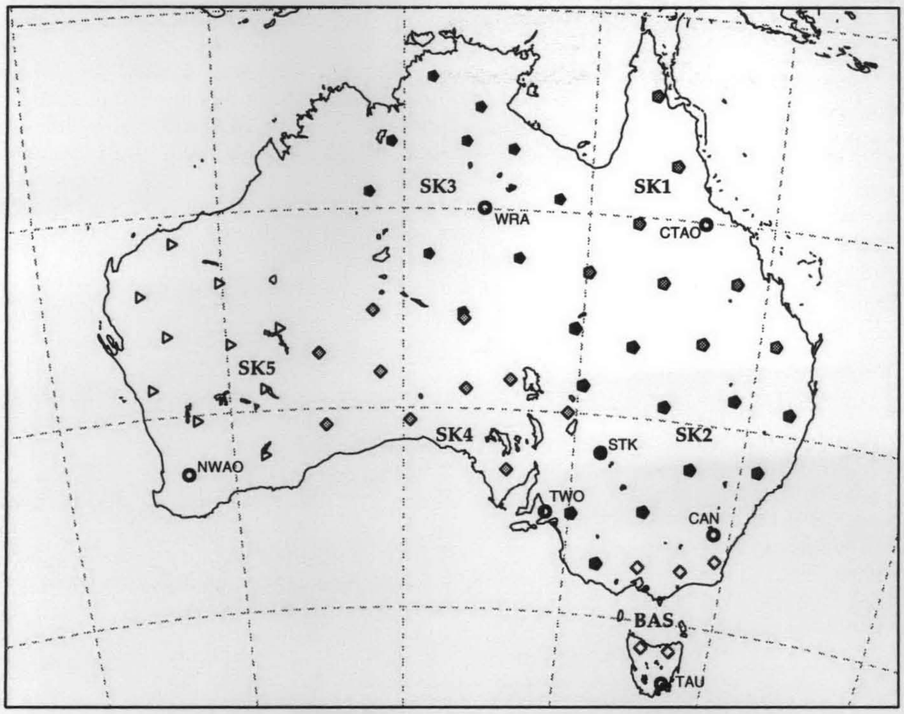


Figure 6. Configuration of broad band surveys. SK1, May–Oct 1993; SK2, Nov 1994–Mar 1995; SK3, May–Oct 1994; BAS, Nov 1994–Feb 1995; SK4, Mar–Aug 1995; SK5, planned Sept 1995–Mar 1996. Permanent stations with high fidelity recording are indicated by a double circle and station name.

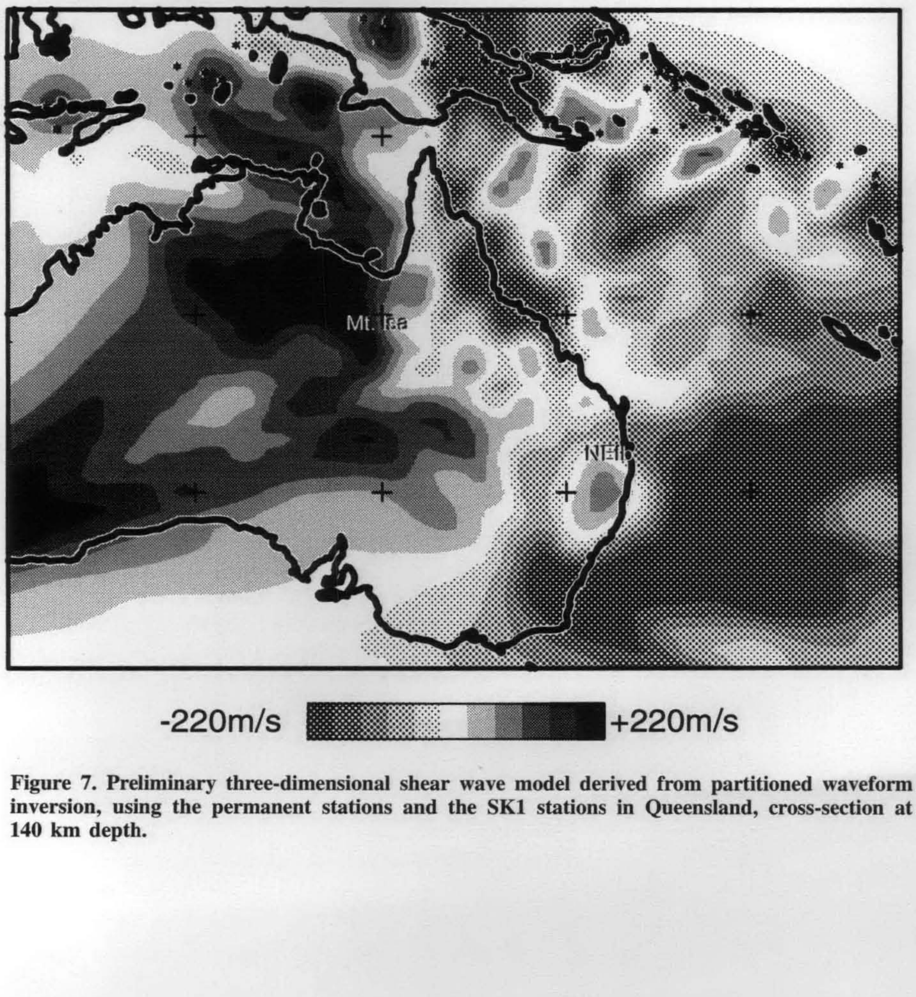


Figure 7. Preliminary three-dimensional shear wave model derived from partitioned waveform inversion, using the permanent stations and the SK1 stations in Queensland, cross-section at 140 km depth.

in Queensland and by the end of August 1995 had covered all the continent except western Australia (Fig. 6). The data set is supplemented by records of suitable events from the permanent broad-band stations. The broad-band records provide a wide range of data, but the primary object is the delineation of lithospheric and mantle structure, using waveform inversion for the shear wave and surface wave portion of the seismogram. The analysis is based on the partitioned waveform inversion technique introduced by Nolet (1990). A nonlinear optimisation is used to find a stratified model which gives the best fit to the observed seismograms, which should represent the average structure along the great circle between source and receiver. The assemblage of path averages is then used in a linear inversion to recover the three-dimensional shear wave structure.

The results of the inversion process for the data from the stations in Queensland (van der Hilst et al. 1995) are presented in Figure 7 for a depth of 140 km. At this level in the lithosphere there are significant contrasts in shear wave velocity. The relatively low shear wave speeds along the Queensland coast may well be associated with Quaternary volcanism. The outline of the higher wave speeds correlates quite well with the surface expression of the Tasman line separating Precambrian and Phanerozoic Australia; though we can note the rather high velocities beneath the Mt Isa Block, which extend to substantial depth (at least 300 km). Another interesting feature is the high shear wave speeds associated with the New England Block. These intriguing results have been derived from the analysis of data from just the first part of the Skippy project. More detail on lithospheric and mantle structure beneath the Australasian region has been revealed as data from the later deployments have been incorporated into the inversion for three-dimensional structure (Zielhuis & van der Hilst 1996, van der Hilst et al. in press).

Conclusion

Over the last decade, there has been a significant increase in knowledge of the P and S wave velocities in the mantle, particularly beneath northern Australia, based on the use of both short-period and broad-band seismic information. The current generation of three-dimensional studies based on the use of portable broad-band seismometers has the potential to dramatically increase the level of understanding of structure in the lithosphere and the underlying mantle of the Australian region. In particular, it should be possible to relate the surficial contrasts between eastern and western Australia to the nature of the structures in the mantle beneath.

Acknowledgments

The study of mantle structure beneath Australia has involved many members of the Seismology Group at the Research School of Earth Sciences, both in the field and in subsequent analysis. I would like to thank Doug Christie, John Grant, Armando Arcidiaco, Tony Percival, Gus Angus and Jan Hulse for their efforts in the field in often trying and uncomfortable conditions. Roger Bowman, Phil Cummins, Oli Gudmundsson, Cheng Tong, Rob van der Hilst, Alet Zielhuis and Sri Widiyantoro have all been involved with different aspects of the structural analysis.

References

Bowman, J.R. & Kennett, B.L.N., 1990. An investigation of the upper mantle beneath northwestern Australia using a hybrid seismic array. Geophysical Journal International, 101, 411–424.

Bowman, J.R. & Kennett, B.L.N., 1991. Propagation of Lg waves in the North Australian craton: influence of crustal velocity gradients. Bulletin of the Seismological Society of America, 81, 592–610.

Bowman, J.R. & Kennett, B.L.N., 1993. The velocity structure of the Australian shield from seismic travel times. Bulletin of the Seismological Society of America, 83, 25–37.

Cleary, J.R., Simpson, D.W. & Muirhead, K.J., 1972. Variations in Australian upper mantle structure from observations of the Cannikin explosion. Nature, 236, 111–112.

Cummins, P.R., Kennett, B.L.N., Bowman, J.R. & Bostock, M.G., 1992. The 520 km discontinuity? Bulletin of the Seismological Society of America, 83, 323–336.

Denham D., 1991. Shear wave crustal models for the Australian continent. In: Drummond, B.J. (editor), The Australian lithosphere. Geological Society of Australia, Special Publication 17, 59–66.

Dey, S.C., Kennett, B.L.N., Bowman, J.R. & Goody, A., 1993. Variations in upper mantle structure under northern Australia. Geophysical Journal International, 114, 304–310.

Drummond, B.J., Muirhead, K.J., Wellman, P. & Wright, C., 1989. A teleseismic travel-time residual map of the Australian continent. BMR Journal of Australian Geology & Geophysics, 11, 101–105

Finlayson, D.M., 1982. Geophysical differences in the lithosphere between Phanerozoic and Precambrian Australia. Tectonophysics, 84, 287–312.

Grand, S. & Helmberger, D.V. 1984. Upper mantle shear structure of North America. Geophysical Journal of the Royal Astronomical Society, 76, 399–438.

Gudmundsson, O., Kennett, B.L.N. & Goody, A., 1994. Broadband observations of upper mantle seismic phases in northern Australia and the attenuation structure in the upper mantle. Physics of the Earth and Planetary Interiors, 84, 207–226.

Hales, A.L., Muirhead, K.J. & Rynn, J.W., 1980. A compressional velocity distribution for the upper mantle. Tectonophysics, 63, 309–348.

Inoue, H., Fukao, Y., Tanabe, K. & Ogata, Y., 1990. Whole mantle P-wave mantle tomography. Physics of the Earth and Planetary Interiors, 59, 294–328.

Kennett, B.L.N., Gudmundsson, O. & Tong, C., 1994. The upper-mantle S and P velocity structure beneath northern Australia from broad-band observations. Physics of the Earth and Planetary Interiors, 86, 85–98.

LeFevre, L.V. & Helmberger, D.V., 1989. Upper mantle P velocity structure of the Australian shield. Journal of Geophysical Research, 94, 17749–17765.

Leven, J.H., 1985. The application of synthetic seismograms in the interpretation of the upper mantle P-wave velocity structure in northern Australia. Physics of the Earth and Planetary Interiors, 38, 9–27.

Muirhead, K.J. & Drummond, B.J., 1991. The seismic structure of the lithosphere under Australia and its implications for continental plate tectonics. In: Drummond, B.J. (editor), The Australian lithosphere. Geological Society of Australia, Special Publication 17, 23–40.

Nolet, G., 1990. Partitioned waveform inversion and two-dimensional structure under the Network of Autonomously Recording Seismographs. Journal of Geophysical Research, 95, 8499–8512.

Tong, C., Gudmundsson, O. & Kennett, B.L.N., 1994. Shear wave splitting in refracted waves returned from the upper mantle transition zone beneath northern Australia. Journal of Geophysical Research, 99, 15783–15797.

Vahau, A. & Kennett, B.L.N., 1995. Delay-time tomography across the shield edge in southeastern Australia. Annual Report for 1994, Research School of Earth Sciences, Australian National University, 25–26.

van der Hilst, R., Kennett, B., Christie, D. & Grant, J., 1994. Project Skippy explores the mantle and lithosphere beneath Australia. EOS, Transactions American Geophysical Union, 75, 177, 180, 181.

van der Hilst, R.D., Zielhuis, A., Weekes, J. & Kennett, B.L.N., 1995. The Skippy project. Annual Report for 1994, Research School of Earth Sciences, Australian National University, 23–25.

van der Hilst, R.D., Kennett, B.L.N. & Shibutani, T., in press. Lithospheric and mantle structure beneath Australia. In: The Australian lithosphere. American Geophysical Union, Geodynamics Monograph Series.

Widiyantoro, S. & van der Hilst, R.D., 1996. Structure and evolution of lithospheric slab beneath the Sunda Arc, Indonesia. Science, 271, 1566–1570.

Zielhuis, A. & van der Hilst, R.D., 1996. Upper-mantle shear velocity beneath eastern Australia from inversion of waveforms from Skippy portable arrays. Geophysical Journal International, 127, 1–16.

AGSO Journal of Australian Geology & Geophysics, 17(1), 55–62

The origin and evolution of the Earth's continental crust

Stuart Ross Taylor[1] & Scott M. McLennan[2]

The present upper crustal composition of the Earth is attributed largely to intracrustal differentiation, resulting in the production of K-rich granites. The crust grows episodically and it is concluded that at least 60 per cent of it was emplaced by the late Archaean (ca 2.7 Ae). Archaean tonalites and trondhjemites resulted from slab melting of young hot oceanic crust. In contrast, most subduction-related rocks, now the main contributors to crustal growth, are derived from the mantle wedge above subduction zones. The contrast between the processes responsible for Archaean and post-Archaean crustal growth is attributed to faster subduction of younger (hotter) oceanic crust in the Archaean (ultimately due to higher heat flow) compared with subduction of older cooler oceanic crust in more recent times. The terrestrial continental crust appears to be unique compared to crusts on other planets and satellites in the solar system, ultimately, a consequence of the presence of water on Earth.

Introduction

The continental crust constitutes only 0.40 per cent of the mass of the Earth. Although it might seem so small that it could be ignored to a first approximation, the crust contains over 30 per cent of the bulk Earth budget for several of the most incompatible elements, such as Cs, Rb, K, U, Th and La. It is thus a major geochemical reservoir, particularly since the crust is not easily recycled back into the mantle. For these reasons, its composition is a major constraint on all geochemical models of bulk Earth composition and evolution.

41.2 per cent of the surface area of the Earth, or 2.10×10^8 km^2, is occupied by continental crust, of which 71.3 per cent, or 1.50×10^8 km^2, lies above sea level. There are four submerged microcontinents and ten major continental blocks (Cogley 1984). The average elevation of the continents above the mean sea floor (oceanic crust) is about 5 km. The elevation of the area above the 200 m isobath (i.e. the shelf/slope break) is 690 m. The mean elevation of the continental crust above present sea level is 125 m.

Crustal thickness varies between 10 and 80 km, correlating with the size of the continental block and the age of the last tectonic event. The average thickness is 41 km (Christensen & Mooney 1995). The volume of the crust is about 8.3×10^9 km^3 — this includes the submerged continental masses and sediment on the ocean floor derived from the continents, and has an error of about ± 10 per cent, because it depends on the variations in crustal thickness. The base of the crust is defined as the Mohorovicic Discontinuity (Moho). At this boundary, compressional wave velocities (Vp) increase from about 7 to about 8 km.s^{-1}. The Moho may be absent locally and is often not sharp. Thus, the crust is usually defined as material with seismic shear wave velocity Vs<4.3 km.s^{-1} or Vp<7.8 km.s^{-1}. Because of the possibility of underplating by basic or ultramafic material and probable interlayering of high-velocity mantle material with lower velocity crustal material, the base of the crust is likely to be very complicated in detail.

A mid-crustal boundary, the Conrad Discontinuity, is occasionally observed at a depth of about 10–20 km. It is often absent or poorly constrained by seismic data, especially in shield areas. Sometimes it is gradational over several kilometres. Curiously, super-deep drill holes have failed to identify many of the discontinuities that were based on interpretations of the geophysical evidence.

Estimates of crustal density range from 2.7 to 2.9 gm.cm^{-3} increasing with depth. If an average density of 2.8 gm.cm^{-3} is adopted, a crustal mass of 2.34×10^{25} gm (± 7%) is obtained. On this basis, the continental crust forms 0.40 per cent of the mass of the whole Earth and 0.61 per cent of the mass of the crust and mantle.

[1] Department of Nuclear Physics, Research School of Physical Sciences, Australian National University, Canberra, ACT 0200, Australia

[2] Department of Earth and Space Sciences, State University of New York, Stony Brook NY 11794-2100, USA

Age of the crust

The Sm–Nd isotope system is the most robust approach to dating the crust because it is least prone to resetting during metamorphism. Fractionation of Sm from Nd takes place at the time of mantle melting. Thus the depleted mantle Nd model age (T_{DM}) of crustal igneous rocks reflects their age of extraction from the mantle. However, processes such as intracrustal melting, metamorphic resetting, and assimilation of older material may affect Sm/Nd ratios, and estimates of average crustal age based on the Nd isotope system may still represent a minimum age (McLennan 1988).The average age of continental crust is about 2.0–2.3 Ae (DePaolo et al. 1991; McCulloch & Bennett 1993). If about 60 per cent of the crust was in place by 2.6 Ae, a mean age of 2.4 Ae for the continental crust is obtained.

Composition of the post-Archaean crust

The continental crust is particularly heterogeneous, as a glance at a geological map illustrates. Compositional changes may occur on a scale of metres, and it might be thought difficult, if not impossible, to calculate an average composition. However, the processes of erosion and sedimentation have carried out an efficient sampling of the upper crust and this information is contained in sedimentary rock sequences. Some elements, of which the rare earth elements (REE) are an excellent example, are transferred quantitatively during erosion and sedimentation from parent rocks into clastic sediments because they are not readily partitioned into fluids during weathering or diagenesis.

The REE abundance patterns in post-Archaean clastic sedimentary rocks show extreme uniformity on a global scale. Thus REE patterns for composite shale samples from Europe (ES) and North America (NASC) are similar to those for the post-Archaean Australian average shale (PAAS). These patterns are distinguished by relatively flat heavy-REE abundances — about 10 times chondritic, by light-REE enrichment and a quite uniform depletion in Eu (Eu/Eu* = 0.65). This uniformity extends both within and between continents. It is thus interpreted to represent the REE abundances in the upper continental crust exposed to weathering (Fig. 1).

The concentrations of other insoluble elements, such as Th and Sc, are also, like the REE, a measure of upper crustal abundances. Sc, although trivalent and a member of the same group (III) of the periodic table as the REE, is a much smaller ion, and is concentrated in basic rocks, entering early crystallising pyroxenes. In contrast, Th is typically concentrated in granitic rocks. The Th/Sc ratio in sedimentary rocks thus forms an index of the relative proportions of granitic and basic rocks.

By using elemental ratios that either are constant across a wide range of igneous compositions (e.g. K/U) or vary systematically with bulk composition (e.g. K/Rb), it is possible to extrapolate to obtain the upper crustal abundances of a number of other elements. In this manner the abundance of

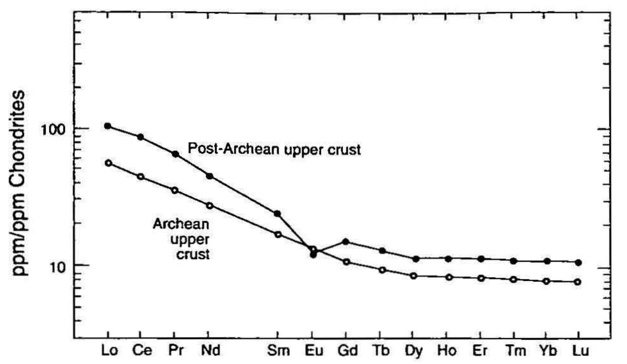


Figure 1. Chondrite-normalised REE diagram, showing estimates of post-Archaean and Archaean upper crust, from Taylor & McLennan (1985).

Rb can be obtained from K/Rb (250); Sr from Rb/Sr (0.3), while U can be obtained from the upper crustal Th/U ratios (3.8) and K from K/U ratios (10,000). The normative mineralogy of the upper crust, based on its major element composition is given in table 4 of Taylor & McLennan (1995). The composition of the post-Archaean *upper* crust is well established, with several estimates converging on a composition close to that of granodiorite (Taylor & McLennan 1995, table 3). Some recent studies employing such approaches include those of Taylor & McLennan (1985), Wedepohl (1991) and Ronov et al. (1992). There is less agreement about the composition of the *bulk* crust (see later)

Origin of post-Archaean upper crust

The upper crustal composition cannot be representative of the whole crust, because of element balance calculations, and heat flow/heat production data. Strong evidence for an intracrustal origin of the upper crust is provided by the Eu-depletion in post-Archaean sedimentary rocks. Eu anomalies rarely occur in igneous rocks derived from the mantle. No primitive mantle-derived volcanic rock shows a relative depletion in Eu. The depletion of Eu that characterises chondrite-normalised REE patterns in clastic sedimentary rocks is not due to surficial processes of oxidation or reduction. This element is present as the trivalent ion in sediments. Under the reducing conditions typical of magmas, much of the Eu is divalent. Thus the depletion in Eu is the signature of an earlier igneous event. The Eu-depleted K-rich granites and granodiorites that now dominate the upper crust were formed by intracrustal melting. The depletion of Eu observed in the upper crust is due to the retention of Eu in residual plagioclase in the lower crust. Plagioclase is only stable to a pressure of 10 kbars (a depth of 40 km on the Earth). This sink for Eu is thus consistent with the experimental studies of intracrustal melting for granite origin (e.g. Wyllie 1983). Two sources of heat are available to initiate intracrustal melting. One is the heat generated by radioactive decay of K, U and Th. The second source is underplating of the crust by basaltic magmas and mantle plumes. This is less easily evaluated, but is needed, since the crustal radioactive sources are probably inadequate.

Information about how long this process has been operating can be obtained from the sedimentary record. Loess from widely scattered localities across the globe has uniform REE patterns. This indicates that during the period represented by the source regions of the loess, the processes producing the upper crust did not change Loess is derived from source rocks that extend back almost 2 b.y. Loess samples (Taylor et al.1983) from China, Europe, New Zealand and North America have Nd depleted mantle model ages (T_{DM}) extending back to 1700 m.y. Such sedimentary REE patterns also exist for sedimentary rocks with Nd model ages older than about 2 b.y. (e.g. McLennan & Hemming 1992). These data indicate that the composition of the upper continental crust was uniform and produced by similar processes back to at least 2 b.y. The processes producing the upper crustal composition that is being sampled during the formation of sedimentary rocks have, accordingly, been the same since well back into the Proterozoic.

Lower crust

Geophysical data show the very diverse nature of the lower crust. It appears to be at least as heterogeneous as the upper crust, and is likely to be very complex in detail, an example of which appears to be the Ivrea Zone in Northern Italy (Voshage et al. 1990). Owing to the inaccessibility of the lower crust and the absence of some averaging technique, such as provided for the upper crust by sediments, it is much more difficult to arrive at a representative composition for it than for the upper crust. Current understanding of the petrogenesis of most granitic rocks and the ubiquitous presence of negative Eu anomalies in sedimentary rocks indicate that intracrustal partial melting must be a fundamental process governing the composition and chemical structure of the lower continental crust. Xenoliths and granulite facies rocks, both providing enigmatic information, are available as potential samples (Rudnick & Presper 1990).

Xenoliths are frequently found in volcanic pipes and flows and record P–T conditions indicating derivation from the lower crust. They are commonly much more basic in composition than the granulite facies regions, and frequently show a relative enrichment in Eu. The positive Eu anomaly is mostly related to the accumulation of cumulate phases rather than to simple residues from partial melts (e.g. Rudnick & Taylor 1987; Rudnick 1992b).

Granulite facies regions are possible samples of the lower crust and they commonly possess positive Eu anomalies; however, these are typically found in the more acidic compositions rather than in mafic material that could represent residues after partial melting. Many such terranes appear, on compositional grounds, to be upper crust that has been buried in Himalayan-type collisions. Possibly, most regional granulites formed in mid-crust regions, so they are not a good model on which to base lower crustal compositions (e.g. Mezger 1992).

In summary, the lower crust appears to be essentially the basic residue left after extraction of the granodioritic upper crust together with additions from underplating by basaltic magmas. Measurements of Poisson's ratio (V_p/V_s; Zandt & Ammon 1995) provide strong support for the basic nature of the lower crust.

The Archaean–Proterozoic boundary

The Archaean–Proterozoic transition marks a major change, both in the volume of crust and intracrustal differentiation. The crustal processes responsible for these changes took place during the late Archaean and are recorded in early Proterozoic sedimentary rocks. The presence of large masses of unsubductable continental crust changed the tectonic regime from the multi-plate Archaean crust and produced the present linear (e.g. South America) or arcuate subduction zones (e.g. western Pacific arcs).

McLennan et al. (1979) found, in the early Proterozoic (2.5–2.2 b.y.) Huronian sedimentary succession of Canada, a progressive change from REE patterns without Eu depletions at the base to typical post-Archaean patterns at the top of the sequence. Similar changes in sedimentary REE patterns have been noted in early Proterozoic successions of the Pine Creek Geosyncline and Hamersley Basin, Australia, and the Slave Province of Canada (see review in Taylor & McLennan 1985). This change reflects an episodic change in upper crustal composition and is related to a massive emplacement of K-rich granitic rocks, depleted in Eu, in the upper crust toward the close of the Archaean. This process ('cratonisation') produces massive intracrustal melting to produce granites, transfers heat-producing elements to the upper crust and generally 'stabilises' the crust. It was non-synchronous over the globe and extended over several hundred million years.

Many geological events correlate with a major change at the Archaean–Proterozoic boundary. The proliferation of banded iron-formations can be related in part to the development of stable shelves during the late Archaean/early Proterozoic. The dramatic increase in ^{87}Sr in marine carbonates from that period (e.g. Veizer 1983) is, similarly, due to an upper crustal enrichment in ^{87}Rb in the K-rich granites that came to dominate the upper crust (coupled with a reduction of the mantle flux of Sr from submarine volcanics, resulting from a cooling Earth). The widespread occurrence of uranium deposits in basal Proterozoic sediments is due to the enrichment of the upper crust in incompatible elements, owing to intracrustal melting in the late Archaean. The first supercontinent probably formed at this time (e.g. Hoffman 1992).

The Archaean crust

The composition of the Archaean upper crust as revealed in the sedimentary record stands in marked contrast to that of the post-Archaean crust (Taylor & McLennan 1995, table 3). A significant difference is shown by the REE patterns in the Archaean sedimentary rocks, which, relative to those of the post-Archaean crust, typically show no Eu anomalies and a lower enrichment in the LREE (Fig. 1). These differences in REE patterns between Archaean and post-Archaean clastic sediments have been documented in many studies (see summary in Taylor & McLennan 1985). They form a crucial observation for models of the evolution of the continental crust.

In detail, there is a great variation in REE patterns in Archaean sediments. This stands in contrast to the very uniform post-Archaean sedimentary REE patterns. Both very steep and flat patterns are locally abundant. The flat patterns are derived from basaltic precursors to first-cycle sediments. The steep patterns occur in first-cycle sediments derived directly from tonalites, trondhjemites and granodiorites (TTG igneous suites). Both REE patterns are derived from the Archaean 'bimodal igneous suite', which largely dominates the Archaean upper crust. The most common patterns bear a superficial resemblance to the REE patterns of island-arc volcanic rocks, such as andesites. This similarity is the result of derivation largely from a mixture of the ubiquitous bimodal suite of felsic igneous rocks (tonalites, trondhjemites, granodiorites or the 'TTG suite' and their volcanic equivalents) and basaltic rocks, which are dominant in many Archaean terrains.

Some workers (e.g. Gibbs et al. 1986; Gao & Wedepohl 1995) have argued that the difference between the REE patterns of Archaean and post-Archaean sediments is without age significance, claiming that it is a consequence of differing tectonic settings This means that post-Archaean greywackes should be identical to those of Archaean age. Except for those of fore-arc basins of oceanic island arcs, in all post-Archaean tectonic environments, the younger sediments display the crucial signature of Eu depletion. Archaean greywackes are clearly petrographically distinct from their modern counterparts in being plagioclase-rich (tonalitic sources), but having relatively few andesitic rock fragments (McLennan 1984). They appear to have formed in tectonic settings such as back-arc, continental arc, trailing edge and foreland basins. The evidence that Archaean turbidites are derived from the 'bimodal basaltic–TTG suite' rather than from 'island-arc andesites' negates the proposed analogy between conditions in the Archaean and modern arc environments.

Small areas of Archaean crust are preserved in high-grade metamorphic terrains. Such terrains exist in Greenland, India, Montana–Wyoming, Canadian Shield, Western Gneiss Terrain, Australia and the Limpopo Belt, South Africa. These regions form a tectonic environment distinct from that of the greenstone belts. The REE patterns in these sediments form two groups. Commonly, they are highly metamorphosed equivalents of Archaean greenstone belt sediments (e.g. Kapuskasing; Taylor et al. 1986) The second group of sedimentary rocks (Limpopo, Western Gneiss Terrain, Taylor et al. 1986) contains samples with REE patterns indistinguishable from typical post-Archaean patterns. These sediments must have been deposited on a stable shelf environment, most probably occurring on small stable mini-cratons.

Both groups of REE patterns occur in close proximity in meta-sediments in Archaean high-grade terrains in India, Greenland and the Limpopo Belt, demonstrating that the sediments were derived from highly localised sources, as is the general case for Archaean sediments. Small areas of stable granitic crust (mini-cratons) in such regions must exist in close association with greenstone belts, but the granites are not contributing significant amounts of erosional debris to the greenstone belt sedimentary basins. The high-grade belts are preferentially preserved, as the greenstone belt environments undergo destruction by erosion and recycling (Veizer & Jansen 1985).

Most of the crust as sampled by the greenstone belt terrains was derived from areas where the bimodal suite of basalt and TTG dominated the land area being eroded to supply the sediments. Despite the close association of greenstone belts with 'granitic' terrains (Bickle et al. 1994), the limited extent of the Archaean granitic terrains with 'post-Archaean' REE signatures is indicated by the absence of such signatures in the greenstone belt sediments. Wind-borne dust should have been widespread in the Archaean, owing to the absence of vegetation. Any granitic terrain should have contributed the characteristic signature of Eu depletion to the greenstone belt sediments. The absence of this signature in the majority of Archaean sediments means that such high-grade terrains represented less than 10 per cent of the exposed Archaean crust (see discussion in Taylor et al. 1986). The zircon data of Stephenson & Patchett (1990) are also consistent with this view of the very limited extent of granitic cratons in the Archaean.

An alternative view has been proposed by Gao & Wedepohl (1995) and Bowring & Housh (1995), who claim that granites with Eu anomalies were formerly widespread in the Archaean, although now missing, owing to erosion. The evidence discussed above makes the former existence of widespread granitic crust in the Archaean unlikely, a view also consistent with the field evidence from Archaean cratons. Basically, what you see is what was there (Trendall 1996).

In modern environments such as arcs or Pacific deep-sea environments, where volcanic provenances dominate, there is ubiquitous evidence for upper crustal material with negative Eu anomalies (e.g. Ben Othmann et al. 1989). In contrast to the situation in post-Archaean time, the isolation of small Archaean cratons enabled the survival of distinct suites of sedimentary rocks with much variation in REE patterns.

Origin of Archaean crust

The Archaean crust probably consisted of many small fast-spreading plates (Pollack 1986). The tonalite–trondhjemite suite was produced by rapid subduction of warm basaltic crust (e.g. Martin 1993, Drummond & Defant 1990).The steep REE patterns of the Archaean TTG suites indicate that garnet was in the residue during partial melting. An origin by melting at mantle depths for the TTG suite is thus indicated, because garnet is only stable in mafic–ultramafic systems at depths below about 40 km. In the Southern Andes, where rapid subduction of young hot oceanic crust occurs today, the slab reaches melting temperatures before complete dehydration occurs, and 'Archaean' tonalites are produced (e.g. Martin 1986).

The Archaean crust thus formed as a mixture of piled-up basalt–komatiite and tonalite–trondhjemite intrusions and extrusives. Sedimentary data suggest that in the upper crust the ratio of basalt to TTG was about equal. Probably this ratio was typical of the Archaean bulk crust, which unlike the present crust, was not vertically zoned in this model. Only minor intracrustal melting occurred in the Archaean. Areas of the crust that had undergone such melting, generating upper crustal negative Eu anomalies, formed cratonic regions of limited extent. Perhaps they were only slightly larger than the present extent of early Archaean terranes in West Greenland–Labrador and the Minnesota River Valley. The limited extent of PAAS-type REE patterns in the Archaean sedimentary record shows that the cratonic regions were very localised. In contrast, the post-Archaean crust is highly stratified into upper and lower crust, owing to extensive intracrustal partial melting.

Bulk crustal composition

There is less of a consensus about both bulk crustal composition and mechanisms of crust formation, since arriving at the average composition of the bulk rather than the exposed crust is complicated. The most important constraint on models of bulk crustal composition relies on the interpretation of continental heat flow data.

Sedimentary rock data provide information only on that portion of the crust exposed to weathering and erosion, but the upper crust is not representative of the entire 41 km thickness of the continental crust. Mass balance calculations show that a 41 km thick crust with K, Th and U abundances equal to that of the present upper continental crust would require about 80–90 per cent of the entire Earth's complement of these elements to be present in the continental crust. Heat flow data show that the upper crust (about 10 km thick) is strongly enriched in the heat-producing elements (K, U, Th).

The present average heat flow from continental areas is about 50 $mW.m^{-2}$ compared to 100 $mW.m^{-2}$ for the oceans and 84 $mW.m^{-2}$ for the whole Earth. However, it is difficult to assign precise values to the contributions from crust and underlying mantle (e.g. Nyblade & Pollack 1993).

Even when the effects of a tectonic heat contribution are removed for younger crust, there is a well-established difference between the heat flow in Archaean and later Precambrian terrains (e.g. Nyblade & Pollack 1993). There appears to be a steep offset in the data at the Archaean–Proterozoic boundary. Because erosional levels are not significantly deeper in Archaean terrains (Watson 1976), the lower heat flow is not due to deeper erosion of the Archaean crust removing a surficial hot layer or an upper granitic layer, as envisaged by Gao & Wedepohl (1995). The difference in heat flow is attributed by Nyblade & Pollack (1993) to an increase in the thickness of the subcrustal lithosphere under Archaean cratons, which lowers surface heat flow by deflecting mantle heat flow around the cratons. We calculate the bulk Archaean crustal values for the heat-producing elements at 1.0% K, 3.8 ppm Th and 1.0 ppm U (McLennan & Taylor 1996).

Most of the crust was generated in the Archaean, with lesser amounts from island-arc volcanism being added later to make up the present crust. The overall crustal bulk composition in our models was calculated from a 60/40 mixture of Archaean bimodal and post-Archaean andesitic compositions (Taylor & McLennan 1985). These result in the following concentrations for the heat-producing elements in the bulk continental crust: 1.1% K, 4.2 ppm Th and 1.1 ppm U, which give the crustal component of the heat flow of 29 $mW.m^2$, or slightly over half of the total heat flow measured in the continental crust. Thus, despite a significant difference in the *upper* crustal composition between Archaean and post-Archaean time, there is probably little difference in bulk composition.

In contrast, other recent estimates of bulk crustal composition exceed the heat flow constraint. Thus, the models of Christensen & Mooney (1995) and Wedepohl (1995) predict a greater heat flow than is observed from the continental crust, *even assuming no contribution to the heat flux from the mantle.* The model of Rudnick & Fountain (1995) is only viable if the background mantle contribution to the measured heat flux is less than 10 $mW.m^2$, which we consider too low. In contrast, models that propose a basaltic crustal composition (e.g. Abbott & Mooney 1995) produce too little heat in the crust to account for the observed values (see McLennan & Taylor 1996 for an extended discussion).

Origin of the continental crust

There is a basic distinction between the igneous activity that contributed to the formation of the continental crust in the Archaean and post-Archaean epochs (Figs 2, 3). There were probably many more plates in the Archaean, owing to higher heat flow (e.g. Pollack 1986) resulting in the rapid recycling of young hot oceanic lithosphere. Such basaltic crust reaches melting temperatures before dehydration has occurred. Under these conditions partial melting occurs, leaving a hornblende–garnet residue. The resulting product is the TTG suite with a steep REE pattern and no Eu anomaly.

Generation of the high-Mg TTG rocks (tonalites, trondhjemites and granodiorites — the so-called sanukitoid suite; Shirey & Hanson 1984) appears to require low-pressure/high-temperature conditions. Such conditions could have been achieved in the Archaean by mantle melting associated with dehydration and/or partial melting of subducted young and hot oceanic lithosphere (Martin 1993).

The Archaean crust is formed from mixtures of the two dominant ('bimodal') igneous lithologies — Na-rich igneous rocks such as tonalites, trondhjemites and granodiorites or the TTG suite (and their volcanic equivalents), and basalts. REE patterns with Eu depletion similar to PAAS are rarely observed in the sedimentary record. They are restricted to cratonic sediments preserved in high-grade metamorphic terrains. These are interpreted as being derived from mini-cratons (Taylor et al. 1986) that were forerunners of the major development of cratons in the late Archaean.

A massive increase in the growth rate of the continental crust, well documented by Nd-isotope evidence, occurred over an extended period of 3.2–2.6 Ae, but differing for individual cratonic regions (e.g. Galer & Golstein 1991; McCulloch & Bennett 1994). Massive *intracrustal* melting of late Archaean crust produced an upper crust dominated by K-rich granodiorites and granites. This change is reflected in the REE patterns observed in the clastic sediments. These typically display significant depletion in Eu. At this time, the upper crust assumed its present composition, as Archaean-type REE patterns are swamped.

Such intracrustal melting often occurs within 50–100 million years of the derivation of new crust from the mantle (e.g. Moorbath 1978). Mantle plumes arriving beneath the crust are considered to be a prime cause of crustal melting. Possibly, the lithospheric keel beneath older cratons deflects mantle heat flow or mantle plumes towards younger marginal areas.

The number of plates became fewer as global heat flow diminished in the Late Archaean, and modern-style plate tectonics became the dominant tectonic theme. Oceanic crust was both older and colder by the time it reached the subduction zone. Older oceanic crust returns to the deep mantle without being remelted, and fluids from dehydration of the slab rise into the overlying mantle wedge, where they induce melting. This results in the production of the present subduction zone calc-alkaline suite and addition of this material to the crust.

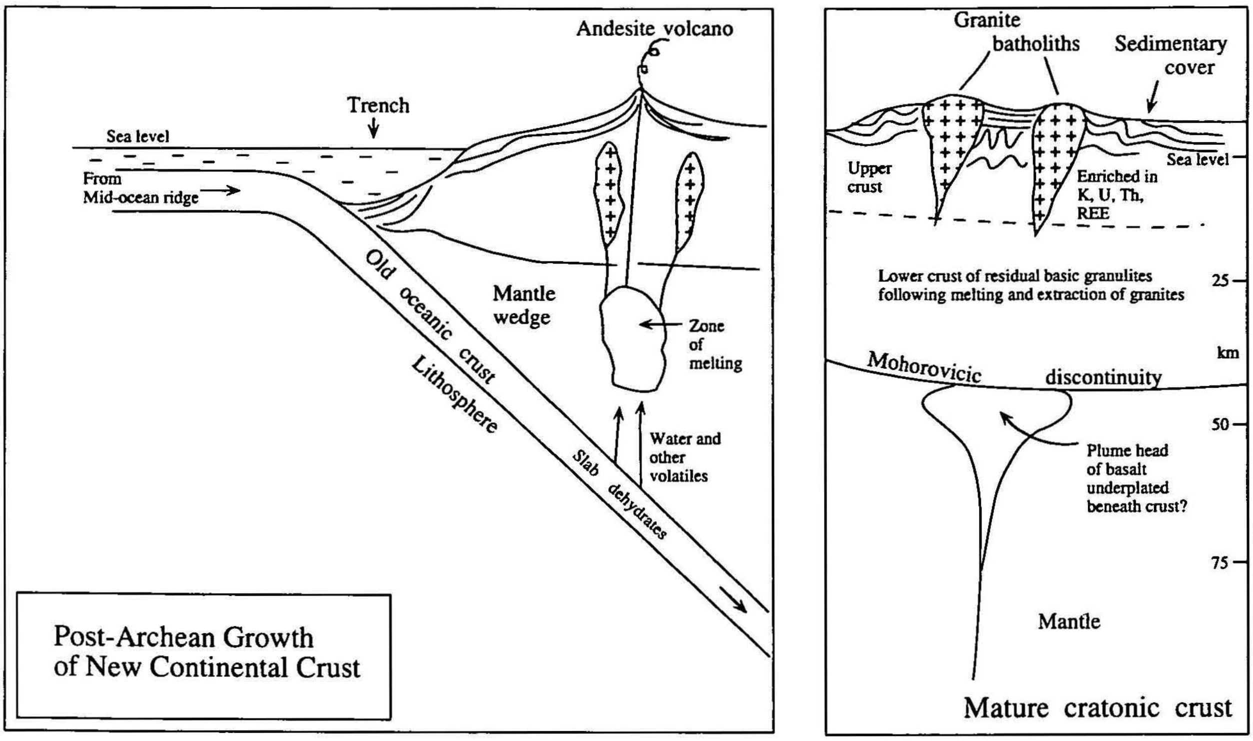


Figure 2. Schematic models of post-Archaean crustal development at subduction zones.

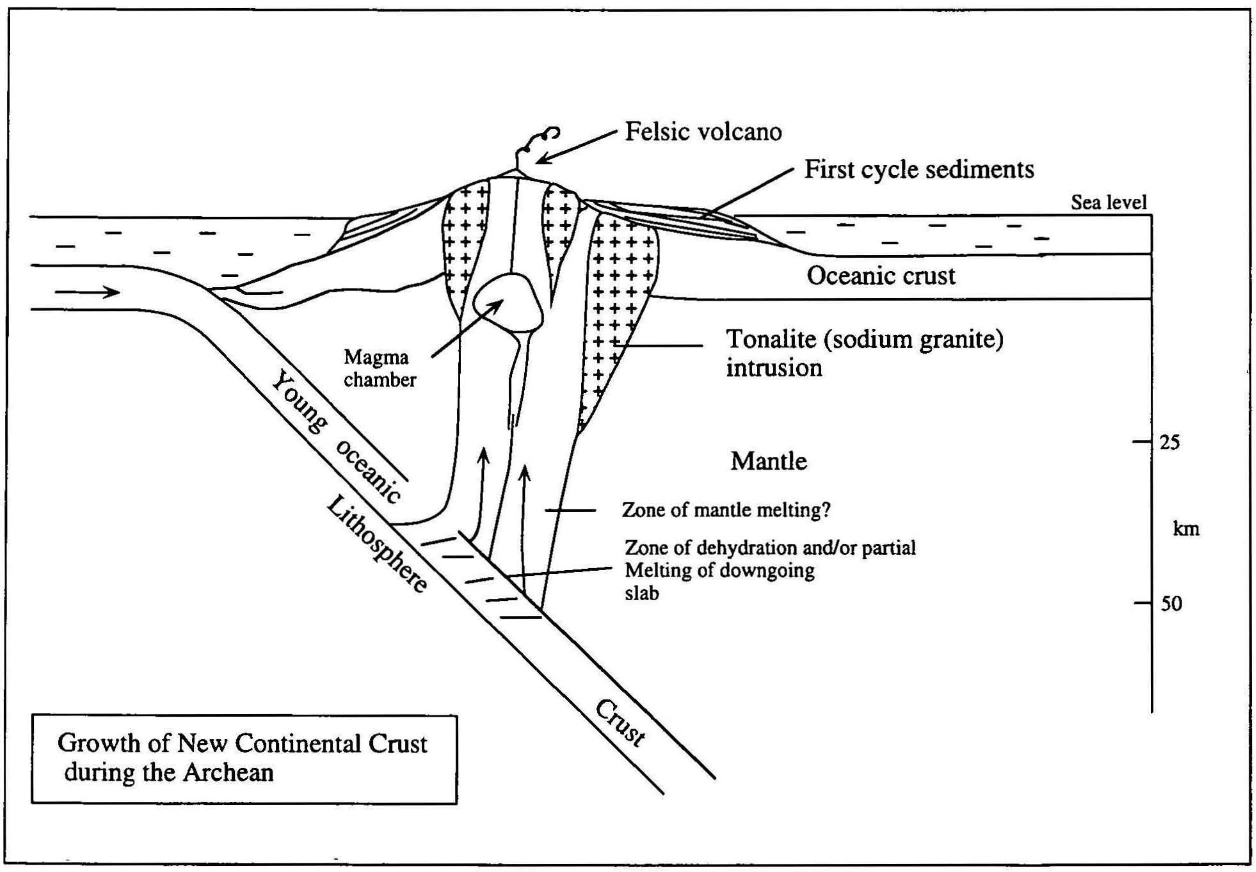


Figure 3. During the Archaean, subducting oceanic crust was younger and hotter on average and the pressure–temperature conditions of dehydration/melting of the subducting slab and overlying mantle differed considerably.

Episodic growth of the crust

One hypothesis proposes that the present mass of the crust formed very early in the Earth's history and has been recycled through the mantle. New additions are balanced by losses, resulting in a steady-state system (Armstrong 1991). The second proposes that the crust has grown throughout geological time in major episodic pulses (e.g. Moorbath 1978; Taylor & McLennan 1981, 1985). The REE abundances enable some testing of these models. They show a major change in the sedimentary record around the Archaean–Proterozoic transition (ca 2.5 Ae), which has been correlated with an earlier major pulse of crustal growth during the late Archaean.

Evidence from ^{10}Be and Pb isotopes shows that sediments have been recycled into the mantle (see review by McLennan 1988). The evidence from ^{10}Be and other geological, geochemical and isotopic constraints limits the amount of subducted sediments to a few per cent. Those models that propose massive recycling of the crust through the mantle encounter various difficulties. Data from long-lived radiogenic isotopes for mantle-derived rocks provide no independent constraints on this problem (Armstrong 1990). During sedimentary recycling processes, the sedimentary mass is largely cannibalistic, with little new material being added from the mantle (McLennan 1988). Thus, the mass of sediment available for subduction is $<1.6\times10^{15}$g.yr^{-1} (about 0.5 km^3.yr^{-1} of crust). This amount does not provide sufficient material to support a steady-state crustal mass.

For the past two billion years, sea level has been within about one kilometre of the present level. This constitutes the so-called freeboard constraint and is evidence that the volume of continents relative to oceans has been roughly constant over this period (e.g. Kasting & Holm 1992). The sedimentary record in the Archaean is much too fragmentary to make freeboard a restriction on crustal volume. The freeboard constraint appears valid at least back into the Early Proterozoic, but provides no constraints of crustal volume before the Proterozoic. Most models accommodate the constraint by proposing that most crustal growth had occurred by the late Archaean, with lesser additions since the Early Proterozoic.

Two observations inform us that early 'granitic' crusts were very limited in extent. There was no land vegetation in the Archaean (Cloud 1988). Hence, there must have been wide exposures of bare rock, with the consequent formation of dust and its transport by wind. Present-day mid-ocean sediments, derived largely by wind transport, show the tell-tale signature of Eu depletion of the upper continental crust. This signature is missing from virtually all Archaean sediments except for those found in the local cratonic areas discussed above. Most of the exposed crustal rocks from which the Archaean sediments were derived were basalts and the TTG suite, rather than granites and granodiorites.

The second critical observation is that of Stephenson & Patchett (1990): they analysed zircons, mainly from quartzites, from the Canadian Shield and the Wyoming, North Atlantic (Labrador, Greenland, Scotland) and Kaapvaal (Southern Africa) cratons. Zircon, being a highly durable mineral, survives many cycles of weathering, erosion and deposition of sediments. Stephenson & Patchett (1990) found that the age of zircon populations, dated by the ^{176}Lu–^{176}Hf technique, in Early Archaean quartzites, is mostly the same as that of the terrain in which they are found, and that there is a scarcity or absence of significantly older zircons. If there was an early sialic crust, then a massive population of ancient zircons derived from it by erosion should have survived and been recycled into younger Archaean sediments. Nutman (personal communication 1994) using U–Pb dating by ion-microprobe, likewise has found no older zircon populations in 3.8 Ae metasediments from Isua, Greenland.

In summary, the growth of the continental crust has proceeded in an episodic fashion throughout geological time (e.g. Moorbath 1978; Taylor & McLennan 1981, 1985), with a major increase in growth rate in the late Archaean (Fig. 4). The present crust has continued to grow by island-arc volcanism and related magmatism, followed by episodes of intracrustal melting (e.g. eastern Australia). There has been no significant change in this process, detectable in the sedimentary REE record, since Archaean–Proterozoic boundary time.

Subcrustal lithosphere

Are there deep roots to the continents? If so, what is their relationship to the overlying crust? There is good evidence that fast seismic P-wave velocities extend to depths of 220 km. Lateral variations in S-wave velocities extend much deeper, to 400 km. Thus, there appear to be deep keels, which are relatively cold and seem to be refractory in composition under the continents (Boyd & Gurney 1986, Jordan 1988). The heat-flow data (Nyblade & Pollack 1993) are consistent with the lithosphere beneath Archaean cratons being much thicker

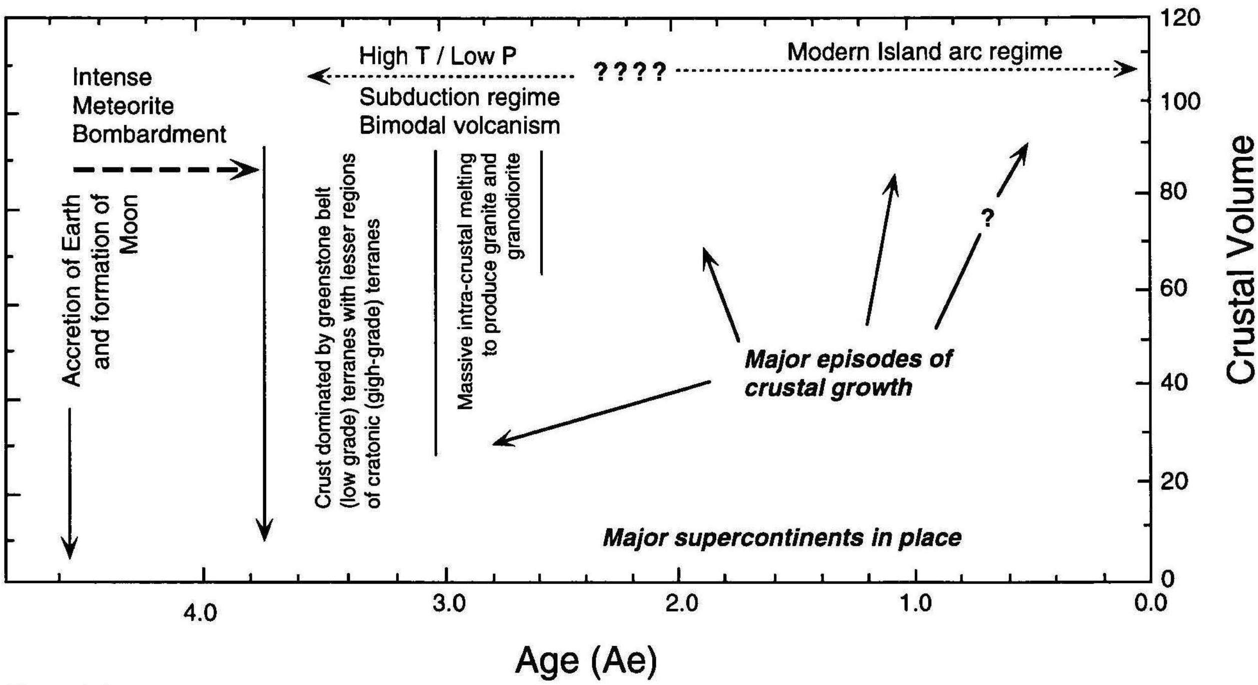


Figure 4. Schematic model of the growth and evolution of the continental crust. The actual values of crust present at any given time are not well constrained; however, a value of 50 per cent crust by about 2.5 Ae is a likely minimum value to satisfy freeboard constraints. Although major global episodes of crustal growth and differentiation are well documented during the late Archaean and at about 2.1–1.7 Ae, it is less clear if crustal growth has been episodic on a global scale during younger times.

than under younger cratons. Studies of inclusions in diamonds indicate that low temperatures have persisted beneath shields for up to 3000 m.y. (Pearson et al. 1995). Possibly, these deep roots represent an Mg-enriched zone of refractory, lower density material, a residue remaining from the extraction of an Fe-rich partial melt. Whether the roots are residual from the extraction of basaltic oceanic crust or are associated with the extraction of continental crust is an unresolved question. However, the presence of these ancient roots, apparently welded to the base of the crust, raises clear problems with those models that invoke delamination of basic lower crusts (e.g. Rudnick & Fountain 1995).

Early crusts

It is often thought that, by analogy with the Moon, the Earth formed an early anorthositic crust. Several reasons make this unlikely. Firstly, the Moon is richer in Ca and Al than the terrestrial mantle (Taylor 1987; Lucey et al. 1995), leading to the early appearance of plagioclase during crystallisation of the lunar magma ocean. Secondly, plagioclase is unstable at shallow depths (40 km) in the Earth, transforming to garnet and, thus, locking up Ca and Al in a dense phase. In contrast, plagioclase will be stable in the Moon to depths of several hundred kilometres.

Plagioclase will not float in a wet terrestrial basaltic magma. The oldest terrestrial anorthosites are not distinct from and closely resemble younger Archaean examples (John Myers, pers. comm.). They must share the same petrogenesis, which makes derivation from a primordial magma ocean of the older examples less likely. Finally, there is no sign of an ancient reservoir of Eu or of primitive $^{87}Sr/^{86}Sr$, which would have resided in an early Sr-rich Rb-poor anorthositic crust.

The conditions for the production of massive granitic crusts are probably unique to the Earth and require three or more stages of derivation from a primitive mantle composition. The Earth has transformed less than 0.5 per cent of its volume to continental crust of intermediate composition and less than 0.2 per cent of its volume into granitic continental crust (ie. upper continental crust) in over 4000 million years, so the process is inefficient. The highland feldspathic crust of the Moon, about 12 per cent of lunar volume, formed, in contrast, within a few million years, during crystallisation from a magma ocean.

No good evidence exists for the enduring geological myth of a primordial world-encircling crust of 'sial' or granite. Such models originated through false analogies with the production of a silicic residuum during crystallisation of basaltic magmas and conditions in an early molten Earth, and a lack of appreciation of the difficulties of producing granite. The absence of evidence for such a crust is discussed above . A primitive 'sialic' crust is refuted by, amongst other evidence, the absence of an old zircon population in younger sediments (Stephenson & Patchett 1990). The oldest preserved terrestrial rock is the 3.96 Ae Acasta Gneiss, in the Northwest Territories of Canada (Bowring et al. 1990).

The formation of crusts distinct from the bulk planetary composition is a common planetary phenomenon. However, crusts similar to the Earth's continental crust do not seem to have formed on the other terrestrial planets or on the icy satellites of the giant planets. On the inner planets, the crusts are basaltic, which are the typical partial melts derived from Fe-Mg silicate mantles. On the satellites of the giant planets, crusts involving water, methane or ammonia ices form by melting of ice-rock mixtures.

The significant feature about the Earth, in contrast to the other terrestrial planets, appears to be the presence of liquid water at the surface, coupled with plate tectonics and subduction, which enable recycling of subducted basaltic crust through the mantle. It is these processes that permit the slow production of continental crust (e.g. Campbell & Taylor 1983). The absence of subduction leads to the persistence of barren basaltic plains, such as we observe on other planetary bodies and the Moon.

References

Abbott, D. & Mooney, W., 1995. The structure and geochemical evolution of the continental crust: support for the oceanic plateau model of continental growth. US National Report, IUGG, Reviews of Geophysics, Supplement, 231–242

Armstrong, R.L., 1991. The persistent myth of crustal growth. Australian Journal of Earth Sciences, 38, 613–630.

Ben Othman, D., White, W.M. & Patchett, J., 1989. The geochemistry of marine sediments, island arc magma genesis, and crust-mantle recycling. Earth and Planetary Science Letters, 94, 1–21.

Bickle, M.J., Nisbet, E.G. & Martin, A., 1984. Archaean greenstone belts are not oceanic crust. Journal of Geology, 102, 121–138.

Bowring, S.A. & Housh, T., 1995. The Earth's early evolution. Science, 269, 1535–1540.

Bowring, S.A., Housh, T.B. & Isachsen, C.E., 1990. The Acasta Gneisses: remnants of the Earth's earliest crust. In: Newsom, H.E. & Jones, J.H. (editors), Origin of the Earth. Oxford University Press, New York, 319–343.

Boyd, F.R. & Gurney, J.J., 1986. Diamonds and the African lithosphere. Science, 232, 472–477.

Campbell, I.H. & Taylor, S.R., 1983. No water, no granites — no oceans, no continents. Geophysical Research Letters, 10, 1061–1064.

Christensen, N.I. & Mooney, W.D., 1995. Seismic velocity structure and composition of the continental crust: a global view. Journal of Geophysical Research, 100, 9761–9788.

Cloud, P., 1988. Oasis in space. W.W. Norton & Co., New York, 508 pp.

Cogley, J.G., 1984. Continental margins and the extent and number of continents. Review of Geophysics and Space Physics, 22, 101–122.

DePaolo, D.J., Linn, A.M. & Schubert, G., 1991. The continental crustal age distribution: methods of determining mantle separation ages from Sm-Nd isotopic data and application to the southwestern United States. Journal of Geophysical Research, 96, 2071–2088.

Drummond, M.S. & Defant, M.J., 1990. A model for trondhjemite–tonalite–dacite genesis and crustal growth via slab melting: Archaean to modern comparisons. Journal of Geophysical Research, 95, 21503–21521.

Galer, S.J.G. & Goldstein, S.L., 1991. Early mantle differentiation and its thermal consequences. Geochimica et Cosmochimica Acta, 55, 227–239.

Gao, S. & Wedepohl, K.H., 1995. The negative Eu anomaly in Archaean sedimentary rocks: implications for decomposition, age and importance of their granitic sources. Earth and Planetary Science Letters, 133, 81–94.

Gibbs, A.K., Montgomery, C.W., O'Day, P.A. & Erslev, E.A., 1986. The Archaean–Proterozoic transition: evidence from the geochemistry of metasedimentary rocks of Guyana and Montana. Geochimica et Cosmochimica Acta, 50, 2125–2141.

Hoffman, P.F., 1988. United plates of America, the birth of a craton. Annual Review of Earth and Planetary Sciences 16, 543–603.

Jordan, T.H., 1988. Structure and formation of the continental tectosphere. Journal of Petrology, 11–37.

Kasting, J.F. & Holm, N.G., 1992. What determines the volume of the oceans. Earth and Planetary Science Letters, 109, 507–515.

Lucey, P.G., Taylor, G.J. & Malaret, E., 1995. Abundance and distribution of iron on the Moon. Science, 268, 1150–1153.
Martin, H., 1986. Effect of steeper Archaean geothermal gradient on geochemistry of subduction-zone magmas. Geology, 14, 753–756.
Martin, H., 1993. The mechanisms of petrogenesis of the Archaean continental crust — comparison with modern processes. Lithos, 30, 373–388.
McCulloch, M.T. & Bennett, V.C., 1993. Evolution of the early Earth: constraints from ^{143}Nd–^{142}Nd isotopic systematics. Lithos, 30, 237–255.
McCulloch, M.T. & Bennett, V.C., 1994. Progressive growth of the Earth's continental crust and depleted mantle: geochemical constraints. Geochimica et Cosmochimica Acta, 58, 4717–4738.
McLennan, S.M., 1984. Petrological characteristics of Archaean graywackes. Journal of Sedimentary Petrology, 54, 889–898.
McLennan, S.M., 1988. Recycling of the continental crust. Pure and Applied Geophysics (PAGEOPH), 128, 683–724.
McLennan, S.M. & Hemming, S.R., 1992. Samarium/neodymnium elemental and isotopic systematics in sedimentary rocks. Geochimica et Cosmochimica Acta, 56, 887–898.
McLennan, S.M., Fryer, B.J. & Young, G.M., 1983. Rare earth elements in Huronian (Lower Proterozoic) sedimentary rocks: composition and evolution of the post-Kenoran upper crust. Geochimica et Cosmochimica Acta, 43, 375–388.
McLennan, S.M. & Taylor, S.R., 1996. Heat flow and the chemical composition of the continental crust. Journal of Geology, 104, 369-377.
Mezger, K., 1992. Temporal evolution of regional granulite terrains: implications for the formation of lowermost continental crust. In: Fountain, D.M., Arculus, R. & Kay, R.M. (editors), Continental lower crust. Elsevier, Amsterdam, 447–474.
Moorbath, S., 1978. Age and isotope evidence for the evolution of the continental crust. Philosophical Transactions of the Royal Society of London, A288, 401–413.
Nyblade, A.A. & Pollack, H.N., 1993. A global analysis of heat flow from Precambrian terrains: implications for the thermal structure of Archaean and Proterozoic lithosphere. Journal of Geophysical Research, 98, 12207–12218.
Pearson, D.G., Carlson, R.W., Shirey, S.B., Boyd, F.R. & Nixon, P.H., 1995. The stabilisation of Archaean lithospheric mantle: a Re-Os isotope study of peridotite xenoliths from the Kaapvaal craton. Earth and Planetary Science Letters, 13, 341–357.
Pollack, H.N., 1986. Cratonization and thermal evolution of the mantle. Earth.and Planetary Science Letters, 80, 175–182.
Ronov, A.B., Bredanova, N.V. & Migdisov, A.A., 1992. General trends in the evolution of the chemical composition of sedimentary and magmatic rocks of the continental Earth crust. Soviet Scientific Reviews, Section G, Geology Reviews, 1, 1–37.
Rudnick, R.L., 1992. Xenoliths — samples of the lower continental crust. In: Fountain, D.M., Arculus, R. & Kay, R.M. (editors), Continental lower crust. Elsevier, Amsterdam, 269–316.
Rudnick, R.L. & Fountain, D.M., 1995. Nature and composition of the continental crust: A lower crustal perspective. Reviews of Geophysics, 33, 267–309.
Rudnick, R.L. & Presper, T., 1990. Geochemistry of intermediate to high-pressure granulites In: Vielzeuf, D. & Vidal, Ph. (editors), Granulites and crustal evolution. Kluwer Academic Publishers, Dordrecht, 523–550.
Rudnick, R.L. & Taylor, S.R., 1987. The composition and petrogenesis of the lower crust: a xenolith study. Journal of Geophysical Research, 92, 13981–14005.
Shirey, S.B. & Hanson, G.N., 1984. Mantle-derived Archaean monzodiorites and trachyandesites. Nature, 310, 222–224,.
Stephenson, R.K. & Patchett, P.J., 1990. Implications for the evolution of the continental crust from Hf isotope systematics of Archaean detrital zircons. Geochimica et Cosmochimica Acta, 54, 1683–1697.
Taylor, S.R. & McLennan, S.M., 1981. The composition and evolution of the continental crust: rare earth element evidence from sedimentary rocks. Philosophical Transactions of the Royal Society of London, A301, 381–399.
Taylor, S.R. & McLennan, S.M., 1985. The continental crust: its composition and evolution. Blackwell, Oxford, 312 pp.
Taylor, S.R. & McLennan, S.M., 1995. The geochemical evolution of the continental crust. Reviews of Geophysics, 33, 241–265
Taylor, S.R., McLennan, S.M. & McCulloch, M.T., 1983. Geochemistry of loess, continental crustal composition and crustal model ages. Geochimica et Cosmochimica Acta, 47, 1897–1905.
Taylor, S.R., Rudnick, R.L., McLennan, S.M. & Eriksson, K.A., 1986. Rare earth element patterns in Archaean high-grade metasediments and their tectonic significance. Geochimica et Cosmochimica Acta, 50, 2267–2279.
Trendall, A.F., 1996. A tale of two cratons: speculations on the origin of continents. Journal of the Royal Society of Western Australia, 79, 141.
Veizer, J., 1983. Trace elements and isotopes in sedimentary carbonates. Reviews in Mineralogy 11, 265–299.
Veizer, J. & Jansen, S. L., 1985. Basement and sedimentary recycling—2: Time dimension to global tectonics. Journal of Geology, 93, 625–643.
Voshage, H., Hoffman, A. W., Mazzucchelli, M., Rivalenti, G., Sinigoi, S., Raczek, I. & Demarchi, G., 1990. Isotopic evidence from the Ivrea Zone for a hybrid lower crust formed by magmatic underplating. Nature, 347, 731–736.
Watson, J., 1976. Vertical movements in Proterozoic structural provinces. Philosophical Transactions of the Royal Society of London, A280, 629–640.
Wedepohl, K.H., 1991. Chemical composition and fractionation of the continental crust. Geologische Rundschau, 80, 207–223.
Wedepohl, K.H., 1995. The composition of the continental crust. Geochimica et Cosmochimica Acta, 59, 1217–1232
Wyllie, P.J., 1983. Experimental and thermal constraints on the deep seated parentage of some granitoid magmas in subduction zones. In: Atherton, M.P. & Gribble, C.D. (editors), Migmatites, melting and metamorphism. Shiva, Nantwich, 37–51.
Zandt, G. & Ammon, C.J., 1995. Continental crust composition constrained by measurements of crustal Poisson's ratio. Nature, 374, 152–154.

AGSO Journal of Australian Geology & Geophysics. 17(1). 63–72

Paleogeothermal gradients in Australia: key to 4-D lithosphere mapping

Suzanne Y. O'Reilly[1], William L. Griffin[1,2] & Oliver Gaul[1]

Empirical paleogeotherms constructed from geothermobarometry of deep-seated xenoliths and garnet ± chromite concentrates from basalts, lamproites and kimberlites around Australia reveal regions of different paleogeothermal signatures. Differences between the tectonically young eastern areas and the cratonic western part of the Australian continent correspond to those shown on a large scale by long-wavelength magnetic data, which integrate the total magnetic signature from the lithospheric column where temperatures are below the Curie Point. Surface heat-flow measurements may not reflect deeper geothermal gradients and model-dependent extrapolations to lower crust and mantle depths must be used cautiously. In eastern Australia, where xenolith data are available (coinciding with the basaltic provinces), there is a remarkably consistent geotherm, which is independent of the age of the basaltic volcanism. This inflected (advective) geotherm is higher than conventional ocean basin geotherms and reflects thermal perturbation associated with volcanic episodes. It records the thermal state at the time of the particular volcanic activity, and decays towards a conductive geotherm with a relaxation time of 40–50 m.y. Data from the eastern craton margin (in South Australia and western New South Wales) indicate significant changes in the thermal state through time, while Archaean and Proterozoic areas in Western Australia reflect typically low geotherms. Knowledge of a robust geotherm for a specific lithospheric column can be used to construct a realistic distribution of rock types with depth. This lithospheric column provides constraints for the geologically meaningful interpretation of geophysical data and for placing geochemical and mantle process information in a spatial context.

Introduction

The thermal flux of the Earth is the driving force for all tectonic processes. Thermal heterogeneities drive convective processes within the Earth (e.g. Stacey 1992) and are the cause of partial melting that leads to igneous activity. Such heat redistribution has driven the formation of the crust and determined the structure of the lithosphere. Most intraplate volcanism probably originates in the asthenosphere, below the lithosphere (for definitions of these terms as used here, see O'Reilly & Griffin 1996) and is the ultimate source of the Earth's atmosphere, oceans and most mineral resources. Determination of paleogeotherms for different portions of the Australian lithosphere through time, therefore, is important for an understanding of the evolution of different tectonic blocks and the distribution of mineral and energy resources.

Much research in crustal metamorphic petrology over the past 15 years has focused on determination of P–T–t (Pressure–Temperature–time) paths of different types of metamorphic terrane; these studies have detailed the effects and evolution of thermal perturbations at relatively shallow depths. P–T studies of crustal and mantle-derived xenoliths and xenocrystic minerals in volcanic rocks attempt to examine the thermal state of the Earth to much greater depth, and to relate variations in the geotherm (the distribution of temperature with depth) to large-scale tectonic processes.

These studies are constrained by an inherent uncertainty: we can only obtain samples of deep-seated xenoliths from volcanic regions, which have, of course, been affected by thermal perturbation. So it is impossible, in principle, to obtain information on completely undisturbed mantle or lower crust by this means. Nonetheless, the study of mantle and deep-crustal xenoliths brought to the surface by mantle-derived magmas makes an essential contribution to our knowledge of the interior of the Earth, and especially of the continental lithosphere.

The information provided by these samples is greatly enhanced where their depth of origin can be determined. Unfortunately, while a temperature estimate is possible for most xenolith types and some xenocrystic minerals, pressure can be directly estimated only for a small proportion of the total sample, especially in younger and hotter terranes. However, if the local geotherm at the time of eruption is known, the temperatures estimated for individual xenoliths or xenocrystic minerals can be referred to the geotherm to obtain an estimate of the depth from which each sample was derived. Thus, a very important application of xenolith-derived geotherms is their use in reconstructing lithosphere stratigraphy (in the sense of a distribution of rock types with depth) from the otherwise disjointed information contained in a suite of xenoliths or a heavy-mineral concentrate. When the information from these invaluable samples is placed in such a 'stratigraphic' context, it forms the basis for 4-D lithosphere mapping — the geological mapping of the lithosphere, in both space and time (O'Reilly & Griffin 1996).

In principle, the present-day geotherm can be derived from measurements of surface heat flow and thermal conductivity. While this is valuable information, its usefulness in lithosphere mapping is limited: the measurements themselves are subject to several types of perturbation and the extrapolation of surface measurements to depth is highly model-dependent (e.g. Sass & Lachenbruch 1979; Pollack & Chapman 1977; O'Reilly & Griffin 1985), requiring a model of the vertical distribution of K, U and Th and the mechanisms of heat transfer in a given section. The most important limitation, however, is that, like all geophysical data, heat-flow measurements provide information on the situation today, and this may not be relevant to times in the past, particularly when a given xenolith suite was erupted.

In this review we will discuss two approaches to determining paleogeotherms, using (1) xenoliths and (2) heavy-mineral concentrates, and illustrate their contributions to our understanding of the Australian lithosphere.

Geothermobarometry

The choice of geothermobarometers and a self-consistent calculation protocol are critical to producing a meaningful and constrained paleogeotherm. Calculations should be based on a suite of xenoliths representing a good vertical sample of a geographically (and temporally) restricted mantle volume. The suitability of the xenoliths should be critically evaluated, using petrographic and microstructural criteria and microprobe analysis of mineral composition as indicators of homogeneity and equilibrium. The geothermobarometers used must be well-calibrated and relevant for the P–T range and bulk composition of the xenoliths. Composite xenoliths allow evaluation of the comparability of different geothermobarometers for different bulk compositions, assuming mutual equilibration.

One significant problem for geothermobarometers based on Fe–Mg exchange lies in the inability of most microanalytical techniques to give Fe^{3+}/Fe^{2+} ratios, which are important in the calculation of site occupancy in the relevant minerals. The estimation from electron microprobe analyses of Fe^{3+} in

[1] Key Centre for Geochemical Evolution and Metallogeny of Continents (GEMOC), School of Earth Sciences, Macquarie University, Sydney, NSW 2109, Australia

[2] CSIRO Exploration and Mining, PO Box 136, North Ryde, NSW 2113, Australia

silicates has been shown by McGuire et al. (1989) to be an artefact controlled by analytical error and, therefore, it should not be attempted for Mg-rich assemblages. The effect of Fe^{3+} on T in such assemblages, therefore, is commonly ignored. An over-riding test of any geothermobarometry protocol is the compatibility of calculated P–T estimates with the stability fields of the observed mineral assemblages. A comprehensive analysis by Griffin et al. (1984) showed that the use of total Fe as Fe^{2+} is the preferred procedure for geothermobarometry calculations for Mg-rich garnet pyroxenites.

There is no set choice of specific geothermobarometers that can be applied universally; almost every suite of xenoliths has an individual complexity, and newer methods are not necessarily better than older ones. The principles and specific methodologies for some particular xenolith types are given in Griffin et al. (1984), Pearson & O'Reilly 1991, Pearson et al. (1991a, b, 1995), Kopylova et al. (1995) and Xu et al.(1996, in press). In areas of basaltic volcanism, the only mantle rock types that allow calculation of P and T are garnet websterites and rare garnet peridotites. So far, there is no useable geobarometer available for garnet-free peridotites: the Ca-in-olivine geobarometer of Kohler & Brey (1990) is extremely T sensitive and not applicable to spinel peridotites in regions with high geothermal gradients (O'Reilly et al. 1997).

Definitions of major lithosphere boundaries

The *crust–mantle boundary* (CMB) is defined here as the depth below which ultramafic rock types become dominant. It may or may not coincide with the *Moho*, which is defined as the seismic discontinuity where compressional wave velocities jump to about 8–8.2 km/sec (Griffin & O'Reilly 1987; O'Reilly & Griffin 1994).

The *lithosphere–asthenosphere boundary* (LAB) may be defined thermally (and rheologically) as the change from a conductive environment (lithosphere) to an adiabatic one (asthenosphere) and, therefore, should coincide with the top of the geophysically defined *low velocity zone* in young tectonic regions — where it may lie within a boundary layer (McKenzie & Bickle 1988). This definition implies that the asthenosphere is a region where mantle melting may be common and where there is continual mixing by convection. It is commonly considered to have a relatively fertile bulk composition from which basaltic melts (sensu lato) may be extracted. The lithosphere, by contrast, is usually considered to have been mechanically stable for a long time. Cratonic lithosphere is commonly geochemically depleted and relatively thick (150–200 km), lying within the tectosphere of Jordan (1988): the LAB here probably represents a zone of pronounced magma–wall rock interaction (Wyllie 1988; O'Reilly & Griffin 1996) and may coincide with the *Lehmann discontinuity* in older (cratonic) regions. Tectonically reworked and young lithosphere is less depleted — it also may have developed geochemical complexity through metasomatic events without convective homogenisation — and is commonly thinner (80–120 km) than cratonic lithosphere. In determining the LAB by garnet geochemistry and Garnet Geotherms (see below), we have adopted a combination of these criteria. The LAB, or base of the lithosphere, is taken as being at the T where the Garnet Geotherm no longer follows a simple conductive model. Above this level depleted garnets are abundant and below it they are rare.

Evolution of mantle composition

As well as differences in thickness between old and young lithosphere, there appears to be a systematic difference in the rock type mix and the chemical composition of Archaean, Proterozoic and Phanerozoic lithosphere (Griffin et al. 1996a, in press). Archaean mantle sections contain significant amounts of depleted garnet harzburgites, typically concentrated in zones 150–180 km deep, where the harzburgite/(lherzolite+harzburgite) ratio may be 10–60% (e.g. Griffin et al. 1996b). At shallower levels, depleted lherzolites make up more than 90% of the column, while more fertile lherzolites may occur at greater depths. In Proterozoic mantle sections, harzburgitic rocks are rare. Phanerozoic lithospheric mantle is characterised by more fertile compositions, abundant evidence of multiple metasomatic events and rare harzburgites. Boyd & Mertzman (1987) demonstrated that garnet peridotite xenoliths in kimberlites (representing old lithosphere) are extremely depleted in Ca+Al, have low clinopyroxene/garnet ratios, and have Mg/Si ratios too low to allow their derivation by extraction of basic or ultrabasic melts alone.

The trace-element composition of garnets from lithospheric mantle rock types also records secular changes in the composition of lithospheric mantle (Griffin, Ryan & O'Reilly, unpublished data; Griffin et al. 1996a, in press). Lherzolitic garnets from Archaean sections have high mean Zr/Y ($\geq$5) and low mean Y/Ga (<3), while Phanerozoic lherzolitic garnets have low mean Zr/Y ($\leq$1) and high mean Y/Ga ($\geq$4); garnets from Proterozoic sections have intermediate values. Comparison with xenolith data and numerical modelling based on partition coefficients indicate that most of these differences can be accounted for by an increase in average clinopyroxene/garnet ratio from Archaean through Proterozoic to Phanerozoic time. Examination of garnet lherzolites from areas with Phanerozoic tectonothermal ages shows that they have high clinopyroxene+garnet as well as high clinopyroxene/garnet, reflecting high average fertility (high (Ca+Al)), which is also shared by spinel lherzolites from many localities. These xenoliths, like 'oceanic' peridotites, have *mg* numbers vs Mg/Si ratios consistent with an origin as residues from the extraction of varying proportions of basaltic melts.

Xenolith-based geotherms

Eastern Australia

Few P–T data on Australian xenoliths were available before 1980. Ferguson et al. (1977) described a garnet–spinel lherzolite xenolith from a Tertiary alkali-basalt diatreme at Jugiong, NSW, and determined conditions of 1240°C, 22 kb. This one point could not define a geotherm, and Ferguson et al. projected a conductive model geotherm — higher than an oceanic basin model — through the point. Irving (1974) estimated conditions of 1050–1100°C, 15–17 kb for a garnet pyroxenite from a 194 Ma basaltic pipe at Delegate, NSW, by experimentally reproducing the mineral chemistry and modal composition of the natural sample.

The first xenolith-derived Australian geotherm for which the shape and position were determined by a significant number of individual samples was described by Griffin et al. (1984), who used a series of garnet websterites (garnet + clinopyroxene + orthopyroxene ± spinel) from the Pleistocene nepheline basanite maars at Lakes Bullenmerri and Gnotuk in western Victoria. The xenoliths proved to have homogeneous minerals, suggesting that they were in internal equilibrium at the time of their entrainment in the host magma, and their mineral assemblages allow the simultaneous determination of both P and T, yielding a P–T locus constrained to about ± 50°C over a significant depth range; this locus was interpreted as defining the geotherm (Fig. 1). Equally important, the bulk compositions of the pyroxenes allowed direct comparison with several experimental studies of the position of the pressure-sensitive boundary between the spinel websterite and garnet websterite phase assemblage fields. This allowed the independent testing of many different proposed geothermobarometers; only those which gave P–T conditions consistent with the phase assemblages of the individual samples were considered valid.

Finally, the Bullenmerri/Gnotuk suite contains several composite xenoliths, in which garnet-pyroxenite or spinel–pyroxenite veins cut through spinel lherzolite. On the assump-

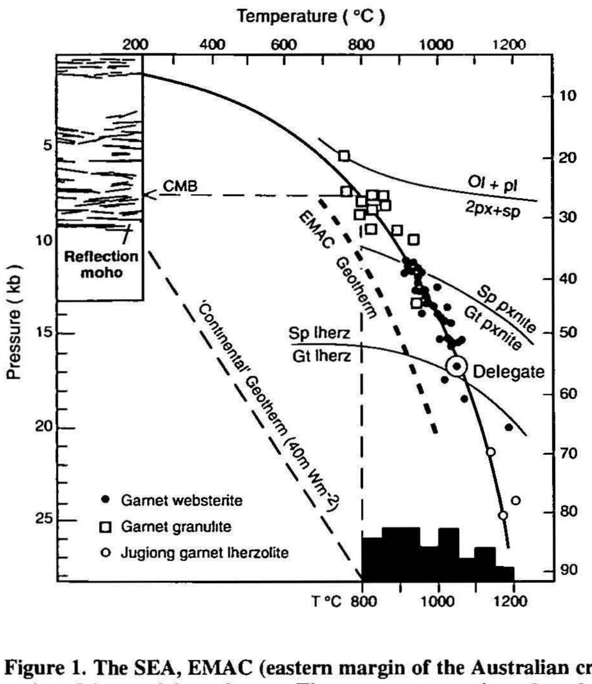


Figure 1. The SEA, EMAC (eastern margin of the Australian craton) and 'cratonic' geotherms. The garnet pyroxenites plotted as filled circles represent the original data from Bullenmerri and Gnotuk, as discussed in the text. Open squares are garnet granulites from O'Reilly et al. (1988). Open circles are P–T points for Jugiong garnet lherzolites (see text). The cartoon of the reflection seismic section is taken from Finlayson et al. (1993) with their location for the reflection Moho for western Victoria. The histogram along the lower axis is for 43 calculated temperatures for spinel lherzolites from the Bullenmerri locality. Graphical determination of the crust–mantle boundary depth by reference to a known geotherm is shown by the short-dashed line. Phase relation boundaries are as discussed in Griffin et al. (1984) and O'Reilly (1989). Note that the spinel–garnet lherzolite transition may occur over a zone up to about 10 km thick (Xu et al. in press; Ionov et al. 1993), but the line here represents the locus for bulk rock compositions with Mg number about 90 and for Cr contents typical of eastern Australian mantle lherzolites, using the experimental results of O'Neill (1981) and Nickel (1986).

tion that the two rock types were at the same T at the time of entrainment, these samples allowed evaluation of the many thermometers proposed for spinel lherzolite assemblages. Because the T estimates for the spinel peridotites were demonstrably consistent with the thermometers used to construct the geotherm, the depth of origin of individual spinel peridotite xenoliths could be determined with an accuracy of ± 3–4 kb by projection of these T estimates to this geotherm (Fig. 1).

The Bullenmerri/Gnotuk geotherm has recently been re-evaluated by Xu et al. (in press) to assess whether other geothermobarometers may be preferable to those originally used and to test the appropriate methodologies for rock types of different bulk composition and/or different conditions of equilibration to those found at Bullenmerri/Gnotuk. This is critical, because comparisons of equilibration conditions of different xenolith assemblages and the methodology of estimating P by reference to T of a known geotherm are valid only if the P–T estimation methods used give consistent and compatible results. However, this does not mean that only one method can be used. For example, in a xenolith suite containing Fe-rich pyroxenites, Mg-rich pyroxenites, spinel lherzolites and garnet lherzolites it will be necessary to use at least three different geothermobarometry calculation methods to enable valid T and P comparisons of rock types with different bulk composition (e.g. Xu et al. in press). In summary, the crucial criteria for compatibility as detailed above, are: (i) calculated P–T points must lie in the stability fields for the observed mineral assemblages in the xenoliths and (ii) geothermobarometry results for different rock types in composite xenoliths must be similar. The SEA geotherm, with the original calculation methods, was shown to be compatible with results using the Brey et al. (1990) Ca-in-orthopyroxene thermometer and Nickel-Green geobarometer (Nickel & Green 1985) for Mg-rich garnet pyroxenites and garnet lherzolites. However, this method was not valid for the Fe-rich (Mg-number of garnet less than 80) garnet pyroxenites, as most P–T points thus calculated plot in the spinel pyroxenite field (but on the same geotherm locus). The Ellis & Green (1979) thermometer (or later modifications) and Wood (1974) barometer give P–T values satisfying the compatibility criteria for such Fe-rich pyroxenites. In addition, the original Sachtleben & Seck (1981) temperatures for spinel lherzolites are compatible with Ca-in-orthopyroxene temperatures (Brey et al. 1990) and also with the Witt-Eickschen & Seck (1991) updates on the Sachtleben & Seck (1981) method.

The Bullenmerri/Gnotuk (B/G) paleogeotherm (Fig. 1) represents the thermal state of the lithosphere beneath this specific locality at the time of basalt eruption (~30,000 years ago), and was made possible by the unusual abundance here of garnet websterites from a large depth range. Small numbers of such xenoliths occur at other places along the 4000 km length of eastern Australia; P–T estimates from these xenoliths also plot along the B/G geotherm, even though they come from volcanic rocks covering a wide range of age and tectonic setting (O'Reilly & Griffin 1985; Griffin et al. 1987). Similarly, both the Jugiong garnet-spinel peridotites (Ferguson et al. 1977; authors' unpublished data) and the experimental reproduction of a Delegate garnet pyroxenite (Irving 1974; Fig. 1) are consistent with this geotherm (Fig. 1), and experimental data on two garnet websterites (Adam et al. 1992) agree within error. This xenolith-based paleogeotherm, therefore, is now referred to as the South-East Australia (SEA) geotherm, and it has a number of important implications.

(1) The shape of the B/G geotherm is much more strongly downward-concave than model geotherms based on conductive heat transport (Fig. 1), suggesting advective heat transport associated with upward movement of hot material. Cull et al. (1991) modelled the geotherm assuming that the heat source was basaltic magma, ponded at the crust–mantle boundary. This successfully reproduced the shape of the B/G geotherm, assuming an underplating rate of about 0.9 km/m.y. over a time span of 4–5 m.y. The volume of magma required for the underplating is about 10 times that observed in the surface eruptions, suggesting that such underplating has played a major role in building up the lower crust beneath eastern Australia.

Advective geotherms like the SEA one have been recognised in several other places around the world, including Spitsbergen (Amundsen et al. 1987) and the East Africa Rift Zone (Jones et al. 1983). All these, like eastern Australia, are areas of young intraplate volcanism, related to continental rifting. Such advective geotherms must decay towards conductive models, once volcanism ceases; Sass & Lachenbruch (1979) calculated a time constant of ~10 m.y. for this decay. Thus it would take only 40–50 m.y. for the SEA geotherm to fall back to a more normal continental conductive geotherm (Fig. 1). Because eastern Australian xenolith suites differing in age by up to 200 m.y. fall along this curve, the SEA geotherm is interpreted as reflecting the thermal state of the lithosphere at each locality, at the stage of thermal/tectonic development that resulted in the eruption of xenolith-bearing alkali basaltic volcanic rocks. The SEA geotherm will only coincide with the present-day geotherm (and areas of high heat flow) in areas like western Victoria and east-central and north Queensland, where volcanism extended into the Pleistocene.

(2) The SEA geotherm gives very high temperatures at relatively shallow depths, and steepens markedly with depth. Despite the obvious uncertainties in extrapolating to depth, the geotherm must intersect both the dry peridotite solidus and the 1280°C asthenospheric adiabat (McKenzie & Bickle 1988) at a depth of ~120±20 km. This depths coincides with the seismically determined depth to the top of the low velocity zone beneath south-eastern Australia (Muirhead & Drummond 1990). It is also similar to the depth of magma separation (~100 km) determined for Tertiary basalts from eastern Australia (O'Reilly & Zhang 1995) These estimates, and the lack of xenoliths from depths greater than 100 km in basalts from eastern Australia, suggest that (a) magmas segregate at and erupt from the low velocity zone or (b) rheological constraints make it difficult to sample the mantle within or below the low velocity zone, or both.

(3) As noted above, the SEA geotherm has allowed estimates of depth of origin for both the abundant spinel lherzolite xenoliths (representing mantle wall rocks) and the common mafic granulite xenoliths (representing lower crustal lithologies) in basaltic rocks along the length of eastern Australia (O'Reilly et al. 1990; O'Reilly & Griffin 1990). The distinction between dominantly mafic and felsic crustal rocks and dominantly ultramafic mantle rocks is fundamental in geology, and, commonly, has been equated with the Mohorovicic discontinuity (Moho) (Finlayson et al. 1993; O'Reilly & Griffin 1994), defined either by seismic refraction or seismic reflection studies (commonly the base of a zone of subhorizontal reflectors).

With xenolith T data and the reference geotherm, the depth at which ultramafic rocks become abundant can be determined, and this is taken as the crust–mantle boundary (CMB) (Fig. 1; Griffin et al. 1984; Griffin & O'Reilly 1987; O'Reilly & Griffin 1994, 1995). At the B/G locality and several other locations in eastern Australia the CMB occurs within a group of seismic reflectors and, commonly, several kilometres shallower than the Moho. The recognition that abundant spinel lherzolites occur at depths as shallow as 25–35 km wherever samples are available led to a major re-interpretation of seismic data from eastern Australia. Geophysical models based on a gradual increase in Vp with depth, and the lack of a typical Moho discontinuity, had led Finlayson et al. (1979) to suggest that the crust beneath eastern Australia was >50 km thick. The xenolith data lead instead to a model involving interlayering of mafic and ultramafic rocks near the CMB and a relatively thin crust (O'Reilly & Griffin 1985, 1996). Finlayson et al. (1993) subsequently confirmed this result for western Victoria, using high-resolution seismic reflection data.

Analyses of spinel–lherzolite xenolith suites from localities along the eastern seaboard have been used to map the regional variation in crustal thickness (Fig. 2). The thickest crust determined this way is in southeast Queensland and this is corroborated seismically by the results of the Eromanga traverse (Finlayson 1990). There is some correlation between depth to the CMB and topographic elevation, but, in detail, other factors, such as present-day heat flow, are important.

(4) The principle of thermal isostasy integrates topographic and physical (including density and thermal) characteristics of the lithosphere. Thermal data from xenoliths provide realistic parameters for assessing thermal isostasy for a given lithospheric column. Isostasy relates the elevation of a column of lithosphere to its mass; thermal isostasy further introduces the effect of the thermal state of the lithosphere on its mass distribution. A computer program ISOSTAT, developed by Paul Morgan (personal communication 1993), calculates the elevation and geotherm for a model lithospheric column by dividing the column into a number of layers (1–10). Each layer used needs thickness, density, heat production and thermal conductivity input; other input is surface temperature and heat flow for the locality.

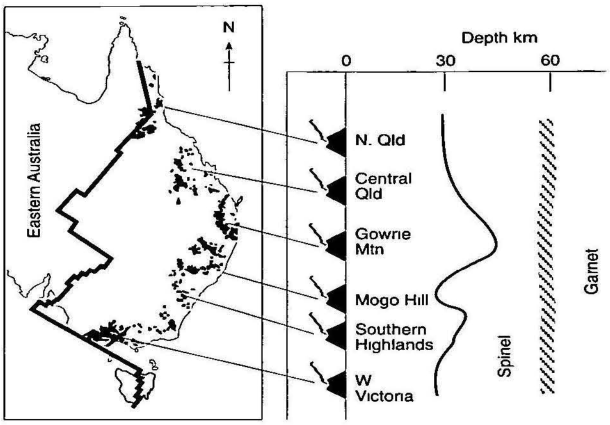


Figure 2. The upper part shows xenolith localities in eastern Australia in black: the heavy jagged line delineates the geophysically determined eastern limit of the Precambrian cratonic region of Australia and, hence, western limit of the Phanerozoic eastern Australian sequences (Pearson et al. 1991a). The lower section shows the variation in depth to Moho for six of the localities.

ISOSTAT determines the mass of the lithospheric column from the input parameters, and compares this with the mass of a purely asthenospheric column, approximated by a mid-ocean ridge (Lachenbruch & Morgan 1990). The amount of asthenospheric material which must be added to the lithospheric column to balance the masses determines the difference in elevation, and thus the elevation of the lithospheric column above sea level. By using crustal thickness from xenolith studies as one of the inputs for ISOSTAT, it is possible to check the appropriateness of the xenolith results.

An analysis of this kind was undertaken by Gaul (1993), using xenolith-derived crustal thicknesses from two places in eastern Australia, the Snowy Mountains (southeast NSW, elevation 1.2–1.7 km a.s.l.) and Mogo Hill (80 km north of Sydney, elevation 200 m a.s.l.). Crustal thicknesses determined were 30 km and 26 km, respectively; ISOSTAT returned elevations of 1.6 km and 0.0 km, in good agreement with the actual elevations. While the two localities have similar crustal thicknesses — that for the Snowy Mountains is also consistent with seismic modelling by Sambridge (personal communication, B. Kennett 1994, 1995) — they have quite different elevations. The Snowy Mountains have a heat flow of ~80 mW/m^2 whereas Mogo Hill's is ~50 mW/m^2 (Cull 1982, 1991); the differences are believed to reflect thermal decay since the end of volcanism: ~20–30 Ma in the Snowy Mountains, and 50–60 Ma at Mogo Hill. The ISOSTAT analysis shows that this difference in heat flow accounts for around 60% of the difference in elevation, while the 4 km difference in crustal thickness is responsible for only 40%. This shows the importance of including thermal parameters in isostasy modelling.

Common sense is required when applying this procedure. Firstly, the area must be in isostatic equilibrium or the results will be meaningless. Secondly, it must be remembered that xenoliths record a paleogeotherm for the area. Any uplift or change in crustal thickness since the eruption of the host rock will mean that the calculated elevation is no longer relevant. The use of thermal isostasy in this manner provides a means of checking crustal thicknesses determined from studies of mantle xenoliths, and makes it possible to use physical data to corroborate (or dispute) results from geochemical studies.

(5) The SEA geotherm also explains features of the large-scale magnetic patterns in Australia. Mantle peridotites are inherently non-magnetic, and the magnetic signature of any continental area is generated entirely by crustal rocks (Wasilewski 1987). The mafic granulites that appear to make up much of the lower crust in eastern Australia (Griffin & O'Reilly 1986; Rudnick et al. 1986) typically contain magnetite and have higher average magnetic susceptibility than the same lithologies in amphibolite facies. Where the temperature in the lower part of the crust exceeds ~575°C (the Curie point of magnetite), the overall magnetisation of the crust will be lower than in areas with cooler geotherms. On the SEA geotherm, the Curie point is reached at a depth of ~12 km, and deepest crustal granulites will be too hot to contribute to crustal magnetisation.

Mayhew & Johnson (1987) analysed MAGSAT data for Australia in terms of the average magnetisation of a 40 km slab, and found that the lowest values were recorded over western Victoria and east-central and northeast Queensland, both areas of Pleistocene volcanism. Other parts of the eastern seaboard have higher average magnetisation, consistent with the decay of the geotherm in these areas, following cessation of volcanism, bringing more of the lower crust to temperatures below the Curie point of magnetite. Local areas of significantly higher magnetisation correspond to gravity lows and are associated with highly magnetic granite bodies.

Eastern margin of the Australian craton (EMAC)

Pearson et al. (1991a, b, c) described a geotherm based on garnet granulite and garnet pyroxenite xenoliths from 260 Ma alkali basalt diatremes near White Cliffs, New South Wales and (probably post-Jurassic) kimberlites near Orroroo, South Australia. This geotherm lies at lower temperatures than the SEA geotherm (Fig. 1), but is still higher and more concave downward than most conductive models, implying advective heat transport. In contrast to the similar rock types from Bullenmerri/Gnotuk, many of the samples used for the EMAC geotherm show significant zoning of individual mineral grains. Pearson et al. interpreted this geotherm as reflecting cooling from a thermal event, leaving open the possibility that the P–T values were frozen in, rather than representing the ambient T at the time of eruption. In either case, the EMAC geotherm is significant because it suggests a different thermal regime on the edge of the craton at the time of basalt/kimberlite eruption.

The ultramafic rocks from the EMAC localities are too altered to allow meaningful T determinations, and the CMB has not been clearly defined in these areas. However, seismic data place the Moho at ~45 km — assuming that this is a maximum value for the CMB, many of the granulite xenoliths clearly are derived from mantle depths (up to 65 km). Chen et al. (1995) found that zircons from the granulite and pyroxenite xenoliths record four groups of U–Pb ages (1500–1600, ~800, ~600 and ~330 Ma), corresponding to tectonothermal events recognised in the crustal rocks from the same region. This result suggests that several magmatic underplating events may have occurred near the CMB in this region, without known eruptive counterparts.

East Kimberley area, Western Australia

No xenolith geotherms are available for the cratonic two-thirds of the Australian continent, simply because no xenolith suites have been found that will provide P–T estimates. This reflects both the lack of young volcanic rocks and the lack of deep mines in any of the deeply weathered older ones. The Argyle diamond mine, in a Proterozoic lamproite body, has yielded a few lherzolite xenoliths, in which pseudomorphs of pyroxenes and spinel record the previous existence of garnet. Jaques et al. (1990) used broad-beam microprobe analysis and the olivine–garnet thermometer to reconstruct the garnet composition, and estimated P–T conditions of 1140–1290°C, 50–60 kb for 20 of the xenoliths. These values lie near the 40 mW/m^2 conductive model geotherm shown in Figure 1 and illustrate the marked difference in thermal structure between this cratonic area in Proterozoic time and the Tertiary volcanic areas of eastern Australia.

Garnet Geotherms

Ryan et al. (1996) have described a technique for deriving geotherm data from the concentrates of Cr-pyrope garnet and chromite that are routinely collected during diamond exploration. This technique was developed specifically to provide an estimate of the geotherm in areas where xenolith suites are not available, such as the cratonic part of Australia. For each grain of Cr-pyrope garnet, a temperature (T_{Ni}) is calculated from the Ni content, using the Ni thermometer of Griffin et al. (1989) as revised by Ryan et al. (1996). This is followed by calculation of the pressure (P_{Cr}) at which that grain would have been in equilibrium with chromite. For Cr-saturated grains, P_{Cr} is a realistic estimate of the pressure at which the grain was entrained in the host magma; for grains that did not coexist with chromite in the mantle, P_{Cr} will be a minimum estimate.

In practice, analysis of a garnet concentrate gives a cloud of points in a T_{Ni}–P_{Cr} plot (Fig. 3), and the Garnet Geotherm is defined as the envelope of maximum P_{Cr} at each T_{Ni}, on the assumption that the grains along this line originally coexisted with chromite. This assumption can be checked by measuring the temperature (T_{Zn}) of chromite grains from the same concentrate, using their Zn content (Griffin et al. 1996b; Ryan et al. 1996), and comparing the T_{Zn} spectrum of the concentrate with the geotherm. The Garnet Geotherm is

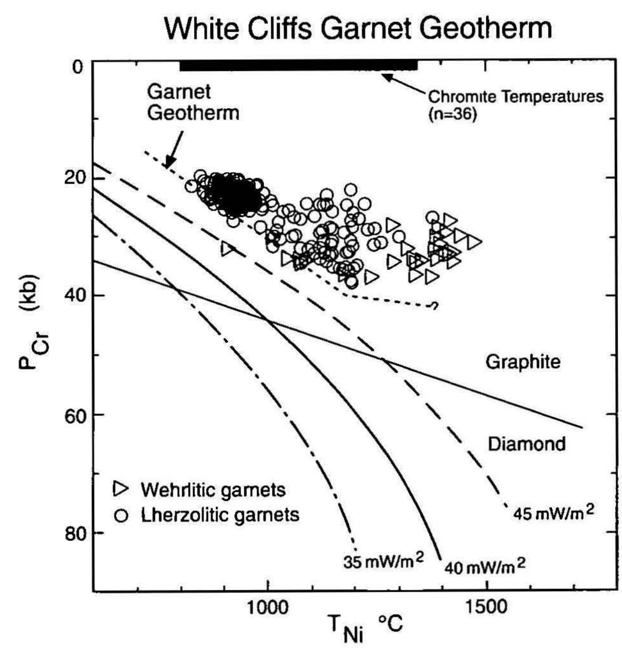


Figure 3. P_{Cr}–T_{Ni} **plot, showing data points for garnet concentrates from the White Cliffs basaltic diatremes. The Garnet Geotherm is the envelope of maximum** P_{Cr} **at each** T_{Ni}**; the 'kink' represents the deepest point at which the geotherm is believed to be conductive, and corresponds to a thermal perturbation at the base of the thermally defined lithosphere: the heavy lines at the top of the figure show the temperature range of chromites calculated by the** T_{Zn} **thermometer (see text for details). Model conductive geotherms (Pollack & Chapman 1977) corresponding to different surface heat flow (in mW/m**^{2}**) are shown for reference, together with the diamond–graphite equilibrium curve.**

considered reliable over the T range where both garnet and chromite were present in the mantle. This technique gives information directly comparable with that derived from xenolith thermobarometry, where the two techniques can be compared (Ryan et al. 1996).

These techniques also form the basis for placing the geochemical information content of heavy-mineral concentrates in a stratigraphic context, where it can be used to map the lithosphere (e.g. Griffin et al. 1996b). The transition from a depleted lithospheric signature to a magmatic 'asthenospheric' signature (enrichment in Zr,Y,Ti,Ga) in Cr-pyrope garnets has been used to define the base of the lithosphere (Fig. 4; Griffin & Ryan 1995; Ryan et al. 1996). As noted above, this 'chemical' definition of the base of the lithosphere corresponds in many areas to the 'thermal' definition based on the change from adiabatic to convective heat transport. The nature of the original host rock (lherzolitic or harzburgitic) can be determined from the chemical composition of garnets (and to some extent of chromites), so that the proportion of rock types at different stratigraphic levels can be determined. This has led to the recognition of different types of mantle, related to tectonothermal age (Griffin 1993). Some examples are given below, to illustrate the potential of the technique; relatively few localities have been studied so far.

White Cliffs, western New South Wales

The Garnet Geotherm for alkali-basaltic diatremes near White Cliffs is shown in Figure 3. It follows a conductive model that lies significantly above most 'cratonic' geotherms, corresponding to a surface heat flow of about 48 mW/m^2. This is consistent with the relatively high heat flow in the area today (Cull 1982, 1991; Mayhew & Johnson 1987) and may be typical of relatively unperturbed continental margin/mobile belt settings. The Garnet Geotherm lies at lower T than the projection to depth of the EMAC geotherm defined by granulite and pyroxenite xenoliths, some of which came from the White Cliffs pipes (Pearson et al. 1991b). This difference strongly suggests that the granulite P–T values represent a frozen-in intermediate stage in the thermal evolution of the lithosphere in this area, rather than the ambient conditions at the time of eruption. A similar conclusion was drawn for the southern margin of the Kaapvaal craton in South Africa, where T estimates for granulite xenoliths lie well above those estimated for the equivalent depths by upward projection of the geotherm defined by mantle xenoliths (Pearson et al. 1995).

The lithosphere–asthenosphere boundary (LAB) defined by the high-T limit of Y-depleted garnets lies near a depth of 130 km. This depth is comparable with a lithosphere thickness (defined by the depth to the LVZ) of 120 km for eastern Australia (Muirhead & Drummond 1990). The garnets have all been derived from lherzolitic rock types (except for a cluster of high-temperature wehrlitic garnets), and details of their trace-element chemistry are typical of garnets from mantle volumes with Late Proterozoic or younger tectonothermal ages (Griffin, Ryan & O'Reilly, unpublished data; Griffin et al. 1996a, in press).

Eastern margin of the Australian craton (EMAC), South Australia

The Adelaide Fold Belt of South Australia is intruded by several diamondiferous kimberlites of Jurassic age and a large number of barren kimberlites of unknown age. In addition, the presence of diamonds and abundant Cr-pyrope garnets in the basal Triassic sediments of the Springfield and Boolcunda Basins implies the presence of Permian or early Triassic diamondiferous kimberlites, probably under the basins. Analysis of garnet concentrates from these three sources shows that the Triassic–Jurassic diamondiferous kimberlites are characterised by Garnet Geotherms near the 40 mW/m^2 conductive model typical of many cratonic areas (Fig. 5). The LAB lay near 150–160 km when the diamondiferous kimberlites erupted and may have shallowed somewhat between Triassic and Jurassic time, in response to a thermal–magmatic event recorded in the metasomatic signatures of high-T garnets (Shee et al. 1993).

The barren kimberlites, on the other hand, reflect a significantly higher geotherm (near a 43 mW/m^2 model) and a shallower LAB, as well as a very prominent signature of melt-related metasomatism at depth (Fig. 5). This led Shee et al. (1993) to suggest that the barren kimberlites are younger

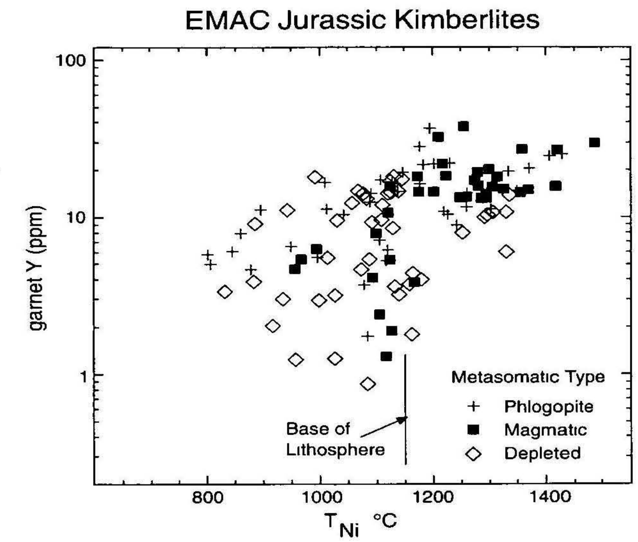


Figure 4. Y–T_{Ni} plot for garnets from Jurassic diamondiferous kimberlites along the eastern margin of the Australian Craton, showing the 'Y edge' taken as the base of the depleted lithosphere. Symbols correspond to different metasomatic signatures, identified by their Zr–Ti–Y relationships.

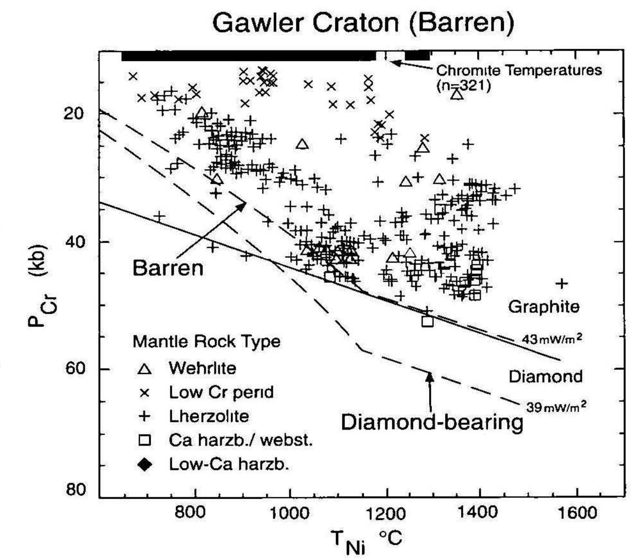


Figure 5. P_{Cr}–T_{Ni} plot, showing data points and Garnet Geotherms for garnet concentrates from the barren (post-Jurassic?) kimberlites of the Gawler Craton (eastern margin of the Australian craton). Points are coded by rock type, as determined by Ca–Cr relationships (abbreviations are: perid.–peridotite; webst.–websterite; harzb.–harzburgite). The Garnet Geotherm for these concentrates approximates a conductive model geotherm corresponding to a 43 mW/m^2 surface heat flow. The lower Garnet Geotherm for the diamond-bearing kimberlites from the same area also is shown; the 'kink' corresponds to the temperature of the 'Y edge' shown in Figure 4. The heavy lines at the top of the figure show the temperature range of chromites calculated by the T_{Zn} thermometer.

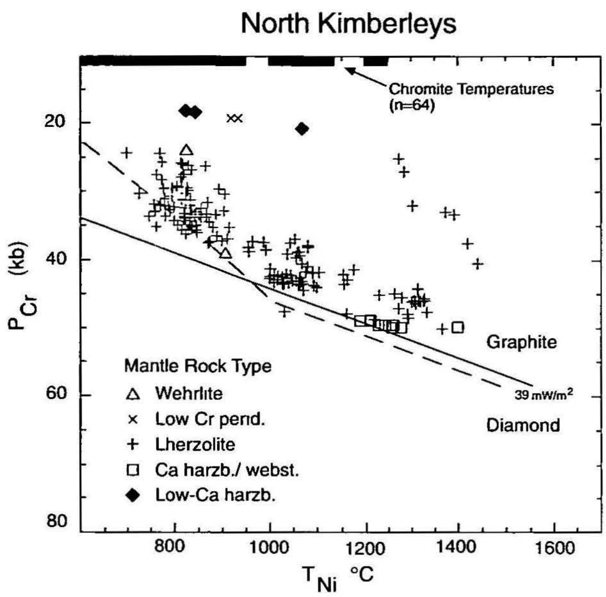


Figure 6. P_{Cr}–T_{Ni} **plot, showing data points for garnet concentrates from the kimberlites of the North Kimberley area of West Australia. Points are coded by rock type, as determined by Ca–Cr relationships (abbreviations as for Figure 5). The Garnet Geotherm for these concentrates approximates a conductive model geotherm corresponding to a 39 mW/m² surface heat flow. The heavy lines at the top of the figure show the temperature range of chromites calculated by the** T_{Zn} **thermometer.**

than the diamondiferous ones (a suggestion not yet tested by radiometric dating) and that the rise in the geotherm and the thinning of the lithosphere were related to Jurassic–Cretaceous magmatic activity near the base of the lithosphere.

Although this area lies near the edge of the Gawler Craton, the garnets are dominantly derived from lherzolitic rock types, with only minor harzburgitic material represented in the older kimberlites. The trace-element chemistry of the garnets is consistent with a Proterozoic tectonothermal age.

North Yilgarn, Western Australia

The Nabberu Basin, on the northern edge of the Archaean Yilgarn Block of Western Australia, is intruded by a number of ultramafic lamprophyres of Paleozoic age. Some of these are diamondiferous, while others, such as the Buljah Pool body (Hamilton & Rock 1990), are barren. The Garnet Geotherm lies near a 38 mW/m² conductive model and is strongly inflected at relatively shallow depth, reflecting heating near the base of the lithosphere; this perturbation in turn makes the LAB difficult to define with any precision.

Harzburgitic garnets are very abundant, in both diamondiferous and barren bodies, and the presence or absence of diamonds depends on the levels of mantle sampled during the ascent of individual bodies (Griffin & Ryan 1995). Although the area lies on the edge of the Late Proterozoic Capricorn Orogen, the mantle lithology is typical of areas with Archaean tectonothermal ages; this observation may suggest that the relatively young surface rocks overlie Archaean cratonic lithosphere and that reworking of the upper crust did not strongly affect the lithospheric mantle.

North and East Kimberley areas, Western Australia

The Kimberley Block of northwestern Western Australia is widely regarded as an Archaean craton, despite the lack of radiometric evidence to support this interpretation. The northern part of the block (the North Kimberley and East Kimberley districts) is intruded by numerous Late Proterozoic kimberlites. The Garnet Geotherm for these districts (Fig. 6) lies near a 40 mW/m² conductive model and is inflected at relatively shallow depth. The LAB appears to lie near 160 km, and diamondiferous kimberlites are those that have sampled between this depth and ~130 km, where the geotherm crosses into the diamond stability field (Fig. 6; Griffin & Ryan 1995). Harzburgitic garnets are very rare in this data set and only moderately subcalcic; however, more abundant harzburgitic garnets, including some significantly subcalcic ones, have been reported from other localities in the East Kimberley district (H. Lucas personal communication 1996) and from the Beta Creek area in the North Kimberley district (Striker Resources 1995). The overall chemistry of the lherzolitic garnets is intermediate between typical Archaean and Proterozoic samples (Griffin & Ryan, unpublished data). These data suggest that Archaean mantle has survived at least locally within the northern part of the Kimberley Block, although some may have been modified during Early–Middle Proterozoic time.

Conclusions, and the relevance of paleogeotherms

Australia can be divided into regions of different paleogeothermal signature, defined by using empirical geotherms. The

Figure 7. E–W section across the Lachlan Fold Belt, Murray Basin, Adelaide Zone to the Gawler Block, from Mayhew & Johnson (1987). From lower right: SEA geotherm, general deep stratigraphy, DM is a projection of the Dartmouth–Marulan seismic line (Finlayson et al. 1979), TL–Tasman line, SS–Stewart Shelf. Vertical sections with numbers represent seismic layering with P-wave velocities in km/sec. The thick line in the upper section shows the vertical integration of magnetisation (∫mdz) in thousands of amps calculated from MAGSAT data: filled circles are heat-flow data from near the section line: filled boxes enclose groups of heat-flow determinations (scale at right) within the Lachlan Fold Belt (right) and the eastern margin of the craton (left box). Full data sources are in references in Mayhew & Johnson (1987). Vp versus depth curve and seismic Moho interpretation for DM are from Finlayson et al. (1979); correlation with rock type is from O'Reilly & Griffin (1985).

geothermal regions correspond to those defined on a broad scale by long-wavelength magnetic data, which reflect the magnetic signature from the crustal (or crust/mantle) column at temperatures below the Curie Point for magnetite (Mayhew & Johnson 1987; O'Reilly & Griffin 1990). Surface heat-flow measurements may not reflect the geothermal gradients at depth and model-dependent extrapolation should be used cautiously. Mayhew & Johnson (1987) combined the magnetic, thermal, seismic and petrological information for the Australian lithosphere to give an integrated cross-section of the variation of thermal structure with depth from the eastern continental margin to the western edge of the Gawler craton in South Australia (Fig. 7).

Time is an important factor, as xenolith-derived geotherms give the P–T environment at the time of eruption. Older cratonic regions that have been undisturbed by subsequent tectonism may still reflect their ancient thermal structure, but young tectonically active regions (e.g. eastern Australia) are thermally perturbed and represent various stages of relaxation from the peak thermal regime towards a conductive geotherm. Differences in the thermal state significantly affect the seismic signature because (1) seismic waves travel more slowly in hot mantle than in stable cratonic mantle, and (2) mafic lithologies will convert from granulite to eclogite during cooling, giving high (mantle-type) seismic velocities to the lower crust (Griffin & O'Reilly 1987; O'Reilly et al. 1990; Austrheim 1991).

Paleogeotherms provide the fundamental data for defining mantle domains in terms of bulk composition, lithostratigraphy and thermal state, which are the basis for geological mapping of the lithosphere in space and time. Relating these mantle domains to a better understanding of tectonic evolution will help define the large-scale evolution of mantle processes with time, and their influence on the development of metallogenic provinces. The mantle provides the heat to drive element redistribution in the crust via magmatic and fluid activity; and mantle processes, including the formation and destruction of lithospheric mantle, are ultimately responsible for crustal generation and the formation of most types of ore deposits. The information derived from paleogeotherms therefore has important applications, not only to diamond exploration (e.g. Griffin & Ryan 1995), but also to understanding the localisation of some specific types of ore deposit, including komatiite-hosted Ni deposits, magmatic PGE deposits and epithermal, magma-associated ore systems.

Acknowledgments

We thank Norm Pearson, Chris Ryan and Tin Tin Win for their extensive contributions to the research summarised here, and Hans Lucas, Simon Shee, Chris Smith and Bruce Wyatt for discussions and samples. The manuscript was improved by thoughtful and constructive reviews from Dennis Hodge and Paul Morgan. The review of an anonymous referee also assisted in clarifying the manuscript by emphasising to us the necessity of explaining concepts simply and in detail, even if previously published. This work has been supported by funding from the Australian Research Council and the Macquarie University Research Grants scheme. This is publication number 21 from the National Key Centre for Geochemical Evolution and Metallogeny of Continents (GEMOC).

References

Adam, J., Green, T.H. & Day, R.A., 1992. An experimental study of two garnet pyroxenite xenoliths from the Bullenmerri Maar of western Victoria, Australia. Contributions to Mineralogy and Petrology, 111, 505–514.

Amundsen, H.E.F., Griffin, W.L. & O'Reilly, S.Y., 1987. The lower crust and upper mantle beneath northwestern Spitsbergen: evidence from xenoliths and geophysics. Tectonophysics, 139, 169–185.

Austrheim, H., 1991. Eclogite formation and dynamics of crustal roots under continental collision zones. Terra Nova, 3, 492–499.

Boyd, F.R. & Mertzman, S.A., 1987. Composition and structure of the Kaapvaal lithosphere, southern Africa. In: Mysen, B.O. (editor), Magmatic processes: physicochemical principles. The Geochemical Society, Special Publication 1, 13–24.

Brey, G.P. & Kohler, T., 1990. Geothermobarometry in four-phase lherzolites. II. New thermobarometers, and practical assessment of existing thermobarometers. Journal of Petrology, 31, 1353–1378.

Brey, G.P., Kohler, T. & Nickel, K.G., 1990. Geothermobarometry in four-phase lherzolites I. Experimental results from 10 to 60 kb. Journal of Petrology, 31, 1313–1352.

Chen, Y.D., O'Reilly, S.Y., Kinny, P.D. & Griffin, W.L., 1995. Dating lower crust and upper mantle events: an ion microprobe study of xenoliths from kimberlitic pipes, South Australia. Lithos, 32, 74–94.

Cull, J.P., 1982. An appraisal of Australian heat-flow data. BMR Journal of Australian Geology & Geophysics, 7, 11–21.

Cull, J.P., 1991. Heat flow and regional geophysics in Australia. In: Cermak, V. & Rybach, L. (editors), Terrestrial heat flow and the lithosphere structure. Springer Verlag, 486–500.

Cull, J.P., O'Reilly, S.Y. & Griffin, W.L., 1991. Xenolith geotherms and crustal models in eastern Australia. Tectonophysics, 192, 359–366.

Ellis, D.J. & Green, D.H., 1979. An experimental study of the effect of Ca upon garnet–clinopyroxene Fe–Mg exchange equilibria. Contributions to Mineralogy and Petrology, 71, 13–22.

Ferguson, J., Ellis, D.J. & England, R.N., 1977. Unique spinel–garnet lherzolite inclusion in kimberlite from Australia. Geology, 5, 278–280.

Finlayson, D.M. (editor), 1990, The Eromanga–Brisbane Geoscience Transect, a guide to basin development across Phanerozoic Australia in Southern Queensland: Bureau of Mineral Resources, Australia, Bulletin 232, 261 pp.

Finlayson, D.M., Prodehl, C, & Collins, C.D.N., 1979. Explosion seismic profiles, and implications for crustal evolution in southeastern Australia: BMR Journal of Australian Geology & Geophysics, 4, 243–252.

Finlayson, D.M., Owen, A., Johnstone, D. & Wake-Dyster, K.D., 1993. Moho and petrologic crust–mantle boundary coincide under southeastern Australia. Geology, 21, 707–710.

Gaul, O.F., 1993. Mantle xenoliths from the Snowy Mountains and Mogo Hill and implications for the thermal profile of the lithosphere. Bsc Hons thesis, Macquarie University, Sydney (unpublished).

Griffin, W.L., 1993. Lithosphere mapping with garnets and spinels. IAVCEI General Assembly, September 1993, Ancient volcanism & modern analogues, Abstract, 42.

Griffin, W.L. & O'Reilly, S.Y., 1986. The lower crust in eastern Australia: xenolith evidence. In: Dawson, J.B., Carswell, D.A., Hall, J. & Wedepohl, K.H. (editors), The nature of the lower continental crust. Geological Society of London, Special Publication 24, 363–374.

Griffin, W.L. & O'Reilly, S.Y., 1987. Is the continental Moho the crust/mantle boundary? Geology, 15, 241–244.

Griffin, W.L. & Ryan, C., 1995. Trace elements in indicator minerals: area selection and target evaluation in diamond exploration. Journal of Geochemical Exploration, 53, 311–357.

Griffin, W.L., Cousens, D.R., Ryan, C.G., Sie, S.H. &

Suter, G.F., 1989. Ni in chrome pyrope garnets: a new thermometer. Contributions to Mineralogy and Petrology, 103, 199–203.

Griffin, W.L., Kaminsky, F.V., Ryan, C.G., O'Reilly, S.Y., Win, T.T. & Ilupin, I.P., 1996b. Thermal state and composition of the lithospheric mantle beneath the Daldyn kimberlite field, Yakutia. Tectonophysics, 262, 19–33.

Griffin, W.L, O'Reilly, S.Y., Ionov, D. & Ryan, C.G., 1996a. Secular evolution of sub-continental mantle. 13th Australian Geological Convention, Canberra, February 1996, Abstracts, 167.

Griffin, W.L., O'Reilly, S.Y., Ryan, C.G., Gaul, O. & Ionov, D., in press. Secular variation in the composition of subcontinental lithospheric mantle. American Geophysical Union, Monograph.

Griffin, W.L., Sutherland, F.L. & Hollis, J.D., 1987. Geothermal profile and crust–mantle transition beneath east-central Queensland: vulcanology, xenolith petrology and seismic data. Journal of Volcanology and Geothermal Research, 31, 177–203.

Griffin, W.L., Wass, S.Y. & Hollis, J.D., 1984. Ultramafic xenoliths from Bullenmerri and Gnotuk maars, Victoria, Australia: petrology of a sub-continental crust–mantle transition. Journal of Petrology, 25, 53–87.

Hamilton, R. & Rock, N.M.S., 1990. Geochemistry, mineralogy and petrology of a new find of ultramafic lamprophyres from Buljah Pool, Nabberu Basin, Yilgarn Craton, Western Australia. Lithos, 24, 275–290.

Ionov, D.A., Aschepkov, I.V., Stosch, H.-G., Witt-Eickschen, G. & Seck, H.A., 1993. Garnet peridotite xenoliths from the Vitim volcanic field, Baikal region: the nature of the garnet–spinel peridotite transition zone in the mantle. Journal of Petrology, 34, 1141–1175.

Irving, A.J., 1974. Geochemical and high pressure studies of garnet pyroxenite and pyroxene granulite xenoliths from the Delegate basaltic pipes, Australia. Journal of Petrology, 15, 1–40.

Jaques, A.L., O'Neill, H.St.C., Smith, C.B., Moon, J. & Chappell, B.W., 1990. Diamondiferous peridotite xenoliths from the Argyle (AK1) lamproite pipe, Western Australia. Contributions to Mineralogy and Petrology, 104, 255–276.

Jones, A.P., Smith, J.V., Dawson, J.B. & Hansen, E.C., 1983. Metamorphism, partial melting, and K metasomatism of garnet–scapolite–kyanite granulite xenoliths from Lashaine, Tanzania. Journal of Geology, 91, 143–165.

Jordan, T.H., 1988. Structure and formation of the continental tectosphere. Journal of Petrology, Special Volume, 1988, 11–38.

Kohler, T.P. & Brey, G.P., 1990. Calcium exchange between olivine and clinopyroxene calibrated as a geothermobarometer for natural peridotites from 2 to 60 kb with applications. Geochimica et Cosmochimica Acta, 54, 2375–2388.

Kopylova, M.G., O'Reilly, S.Y., & Genshaft, Yu. S. 1995. Thermal state of the lithosphere beneath Central Mongolia: evidence from deep-seated xenoliths from the Shavaryn-Saram Volcanic Centre in the Tariat Depression, Hangai, Mongolia. Lithos, 36, 243–256.

Lachenbruch, A.H. & Morgan, P., 1990. Continental extension, magmatism and elevation; formal relations and rules of thumb. Tectonophysics, 174, 39–62.

Mayhew, M. & Johnson, B.D., 1987. An equivalent layer magnetization model for Australia based on MAGSAT data. Earth and Planetary Science Letters, 83, 167–174.

McGuire, A.V., Dyar, M.D. & Ward, K.A., 1989. Neglected Fe^{3+}/Fe^{2+} ratios — a study of Fe^{3+} content of megacrysts from alkali basalts. Geology, 17, 687–690.

McKenzie, D. & Bickle, M.J., 1988. The volume and composition of melt generated by extension of the lithosphere. Journal of Petrology, 29, 625–679.

Muirhead, K.J. & Drummond, B.J., 1990. The base of the lithosphere under Australia. Geological Society of Australia, Special Publication 17, 23–40.

Nickel, K.G., 1986. Phase equilibria in the system SiO_2–MgO–Al_2O_3–CaO–Cr_2O_3 (SMACCR) and their bearing on spinel/garnet lherzolite relationships. Neues Jahrbuch fur Mineralogie Abhandlung, 155, 259–287.

Nickel, K.G. & Green, D.H., 1985. Empirical geothermobarometry for garnet peridotites and implications for the nature of the lithosphere, kimberlites and diamond. Earth and Planetary Science Letters, 73, 158–170.

O'Neill, H.St.C., 1981. The transition between spinel lherzolite and garnet lherzolite, and its use as a geobarometer. Contributions to Mineralogy and Petrology, 77, 185–194.

O'Reilly, S.Y., 1989. Nature of the east-Australian lithosphere. In: Johnson, R.W. (editor), Intraplate volcanism in eastern Australia and New Zealand. Cambridge University Press, 290–297.

O'Reilly, S.Y. & Griffin, W.L., 1985. A xenolith-derived geotherm for southeastern Australia and its geophysical implications. Tectonophysics, 111, 41–63.

O'Reilly, S.Y. & Griffin, W.L., 1990. Geophysical and petrologic properties of the crust/mantle boundary region, eastern Australia: relevance to the Eromanga–Brisbane transect. Bureau of Mineral Resources, Australia, Bulletin 232, 203–212.

O'Reilly, S.Y. & Griffin, W.L., 1994. Moho and petrologic crust–mantle boundary coincide under southeastern Australia. Comment. Geology, 22, 666–667.

O'Reilly, S.Y. & Griffin, W.L., 1996. 4-D lithosphere mapping: methodology and examples. Tectonophysics, 262, 1–18.

O'Reilly, S.Y. & Zhang, M., 1995. Geochemical characteristics of lava-field basalts from eastern Australia and inferred sources: connections with the subcontinental lithospheric mantle. Contributions to Mineralogy and Petrology, 121, 148–170.

O'Reilly, S.Y., Chen, D.G., Griffin, W.L. & Ryan, C.G., 1997. Trace elements compositions of mantle olivines: implications for geothermobarometry. Mineralogical Magazine, 61, 257–269.

O'Reilly, S.Y., Griffin, W.L. & Pearson, N.J., 1989. Xenoliths of crustal origin. In: Johnson, R.W. (editor), Intraplate volcanism in eastern Australia and New Zealand. Cambridge University Press, 275–288.

O'Reilly, S.Y., Griffin, W.L. & Stabel, A., 1988. Evolution of Phanerozoic eastern Australian lithosphere: isotope evidence for magmatic and tectonic underplating. Journal of Petrology, Special Volume, 89–108.

O'Reilly, S.Y., Jackson, I. & Bezant, C., 1990. Seismic and thermal parameters of upper mantle rocks from eastern Australia: implications for seismic modelling. Tectonophysics, 185, 67–82.

Pearson, N.J. & O'Reilly, S.Y., 1991. Thermobarometry and P–T–t paths: the granulite to eclogite transition. Journal of Metamorphic Geology, 9, 349–359.

Pearson, N.J., O'Reilly, S.Y. & Griffin, W.L., 1991a. Heterogeneity in the thermal state of the lower crust beneath eastern Australia. Exploration Geoscience, 22, 295–298.

Pearson, N.J., O'Reilly, S.Y. & Griffin, W.L., 1991b. The granulite to eclogite transition beneath the eastern margin of the Australian craton. European Journal of Mineralogy, 3, 293–322.

Pearson, N.J., O'Reilly, S.Y. & Griffin, W.L., 1995. The crust–mantle boundary beneath cratons and craton mar-

gins: a transect across the southwest margin of the Kaapvaal craton. Lithos, 36, 257–288.
Pollack, H.N. & Chapman, D.S., 1977. On the regional variation of heat flow, geotherms and lithospheric thickness. Tectonophysics, 38, 279–296.
Rudnick, R.L., McDonough, W.F., McCulloch, M.T. & Taylor, S.R., 1986. Lower crustal xenoliths from Queensland, Australia: evidence for deep crustal assimilation and fractional crystallization of continental basalts. Geochimica et Cosmochimica Acta, 50, 1099–1155.
Ryan, C.G., Griffin, W.L. & Pearson, N.J., 1996. Garnet geotherms: derivation of P–T data from heavy-mineral concentrates. Journal of Geophysical Research, 101, 5611–5625.
Sachtleben, T. & Seck, H.A., 1981. Chemical control of Al-solubility in orthopyroxene and its implications for pyroxene geothermometry. Contributions to Mineralogy and Petrology, 78, 157–165.
Sass, J.H. & Lachenbruch, A.H., 1979. Thermal regime of the Australia continental crust. In: McElhinny, M.W. (editor), The Earth, its origin, structure and evolution. Academic Press, London, 301–352.
Shee, S.R., Wyatt, B.A. & Griffin, W.L., 1993. Major and trace element mineral chemistry of peridotite xenoliths from the Wesselton kimberlite, South Africa. IAVCEI 1993 General Assembly, Abstracts, p. 98.
Stacey, F.D., 1992. Physics of the Earth. Brookfield Press, Brisbane.
Striker Resources, 1995. Beta Creek Diamond Project. Quarterly Report to the Australian Stock Exchange, January 1995.
Wasilewski, P. J., 1987. Magnetic properties of mantle xenoliths and the magnetic character of the crust–mantle boundary. In: Nixon, P.N. (editor), Mantle xenoliths. Wiley, 577–588.
Witt-Eickschen, G. & Seck, H.A., 1991. Solubility of Ca and Al in orthopyroxene from spinel peridotite: and improved version of an empirical geothermometer. Contributions to Mineralogy and Petrology, 106, 431–439.
Wood, B.J., 1974. Solubility of alumina in orthopyroxene coexisting with garnet. Contributions to Mineralogy and Petrology, 46, 1–15.
Wyllie. P.J., 1988. The origin of kimberlite. Journal of Geophysical Research, 85, 6902–6910.
Xu, X., O'Reilly, S. Y., Griffin, W. L. & Zhou, X., 1996. A xenolith-derived geotherm and the crust/mantle boundary at Qilin, southeast China. Lithos, 38, 41–62.
Xu, X., O'Reilly, S. Y., Griffin, W. L. & Zhou, X., in press. The nature of the Cenozoic lithosphere beneath Nushan, East Central China. In: Flower, M.F.J., Chung, S.L., Lo, C.H. & Lee, T.Y. (editors), Mantle dynamics and plate interactions in east Asia. AGU Special Geodynamics Series (invited submission).

AGSO Journal of Australian Geology & Geophysics, 17(1), 73–92

Australian Neoproterozoic palaeogeography, tectonics, and supercontinental connections

M.R. Walter[1,2] & J.J. Veevers[1,3]

Increasingly precise stratigraphic resolution by biostratigraphy, isotope stratigraphy, and sequence analysis in the Neoproterozoic allows more convincing palaeogeographic reconstructions than hitherto possible, so that the original connections amongst structural basins can be demonstrated. The Neoproterozoic stratigraphy of Australia can now be analysed in terms of four supersequences, with finer subdivision possible in the Ediacarian or 'Terminal Proterozoic'. The palaeogeography of Australia during eight time intervals within the Neoproterozoic is assessed, with varying degrees of confidence.

Our interpretation follows previous models of the Adelaide Rift Complex as a Neoproterozoic intracratonic rift between the Gawler and Curnamona cratons. The rift is at a high angle to the associated east–west elongated epicratonic sag of the Centralian Superbasin. In the earliest Cambrian the Flinders zone of the Adelaide Rift Complex was transformed to a failed arm or aulacogen by continental breakup along its southern part. Sedimentation ceased before the Late Cambrian–Early Ordovician (500 Ma) Delamerian Orogeny. The Rift Complex underwent two phases of onlap (= extension) accompanied by the deposition of (Sturtian and Marinoan) glacials. A third phase of onlap represented by the Billy Springs Formation occurred during right-lateral shearing (Petermann Ranges orogeny) that caused thrusting and the emergence of the east–west oriented Musgrave Block in the middle of the Superbasin. The Superbasin was finally dismembered by the rise of the southern Arunta Block between the Amadeus and Ngalia structural basins during the mid-Carboniferous Alice Springs Orogeny.

According to the SWEAT hypothesis, Australia was joined in the Neoproterozoic with India, Antarctica, and Laurentia, so that the Tasman Line faced the Canadian-Wyoming cordilleran line. The configuration of the north–south trending Adelaide Rift Complex and the east–west trending Centralian Superbasin was mirrored by the basins in Laurentia to form a T, which split at the end of the Neoproterozoic by growth of a precursor of the Pacific Ocean.

Introduction

Currently a global effort is extending to the latest Proterozoic the insights that detailed stratigraphy has provided in the Phanerozoic, and at the same time focusing on the tectonic, climatic and biological changes that mark the transition from the Proterozoic to the Phanerozoic (Knoll & Walter 1992).

Stratigraphic resolution within the Neoproterozoic is generally poor, though great strides have been taken in the 'Terminal Proterozoic' (Ediacarian). Conventional lithostratigraphic analyses have in the past been supplemented by stromatolite biostratigraphy and to some extent magnetostratigraphy. In recent years new finds of metazoan trace and body fossils have offered promise of stratigraphic utility (Jenkins 1995, Narbonne & Aitken 1995). Sequence stratigraphy is sharpening intrabasinal analysis (Christie-Blick et al. 1995). Where zircon-bearing tuffs have been found, substantial improvements in resolution have resulted from precise dating (e.g. Compston et al. 1987, Grotzinger et al. 1995). Big advances of recent years have resulted from the application of isotope chemostratigraphy (e.g. Knoll et al. 1986, Kaufman & Knoll 1995, Knoll et al. 1995) and acritarch biostratigraphy (e.g. Vidal & Knoll 1983, Zang & Walter 1992). In combination, all these approaches now allow a degree of precision in the analysis of Neoproterozoic history previously beyond our reach.

The Australian continent is a large segment of the Earth's crust and on it Neoproterozoic rocks are particularly abundant, making this a good place to elucidate the history of the eon. There is a long history of such studies, with notable early work including Chewings (1914) and Mawson (1925) on stromatolites, Mawson on tillites (1949) and Sprigg (1947) and Glaessner (1984) on the Ediacara Fauna of early metazoans (amongst many other studies). Our approach has been to focus on isotope chemostratigraphy and acritarch biostratigraphy, within a framework of sequence stratigraphy, lithostratigraphy and tectonic analysis; projects by K. Grey (acritarchs) and C.R. Calver (isotopes and sedimentology) on the Ediacarian interval have recently been completed, and others by S.W. Grant (acritarchs and isotopes on the interglacial succession), K. Cotter (palaeobiology of the early Neoproterozoic), A. Hill (isotope geochemistry of the early Neoproterozoic), and P. Gorjan (sulphur isotope geochemistry) are continuing. Field work in the Savory and Amadeus Basins and the Adelaide Rift Complex and the study of cores from the Georgina, Amadeus and Officer Basins, the Stuart Shelf and the Adelaide Rift Complex, with a compilation of existing data, have lead to the recognition and describing of the Centralian Superbasin (Fig. 1) (Walter & Gorter 1994, Walter et al. 1995). It has proven possible to analyse the Neoproterozoic stratigraphy of the Superbasin in terms of four supersequences (Fig. 2), and to produce isopach maps for these (Walter et al. 1995). The stratigraphic interpretations in those papers form the basis for what follows here, and we have added new information from the Paterson Orogen and from the Kimberley region and other areas of northern Australia. We have not included Tasmania because it was probably isolated from the rest of the continent at that time.

In this paper we combine our work with that of Preiss (1987, 1993), Zang (1995), Christie-Blick et al. (1995) and many others (e.g. in Jenkins et al. 1993) to make a preliminary attempt to portray the palaeogeography of the Australian continent during the Neoproterozoic. We have selected eight time intervals to portray on palaeogeographic maps. Some represent 'moments' in geological history, others are poorly resolved. Specifically, stratigraphic resolution within the early Neoproterozoic Supersequence 1 (Fig. 2) is extremely poor, that in the overlying Supersequence 2 is moderately good, and that in the Ediacarian-aged Supersequences 3 and 4 is, in a Proterozoic context, very good. The stratigraphic framework is taken from Preiss (1987, 1993), Walter & Gorter (1994) and Walter et al. (1994, 1995), and the reader is referred to those publications for detailed discussions. Preiss (1987, 1993) has attempted a much more detailed analysis of the palaeogeography of the Adelaide Rift Complex and we have adopted his interpretations. Some aspects of the hydrocarbon prospectivity that follow from this analysis are discussed in Walter & Gorter (1994) and Bradshaw et al. (1994).

Our palaeogeographic maps follow in Figure 3, after an outline of our correlation scheme and its radiometric calibration. A synthesis of Neoproterozoic palaeogeography and tectonics of Australia-Antarctica and Laurentia is presented by Veevers et al. (1997).

[1] Australian Plate Research Group
[2] Centre for Ecostratigraphy and Palaeobiology, School of Earth Sciences, Macquarie University, North Ryde NSW 2109, Australia
[3] Key Centre for the Geochemical Evolution and Metallogeny of Continents, Macquarie University, North Ryde NSW 2109, Australia

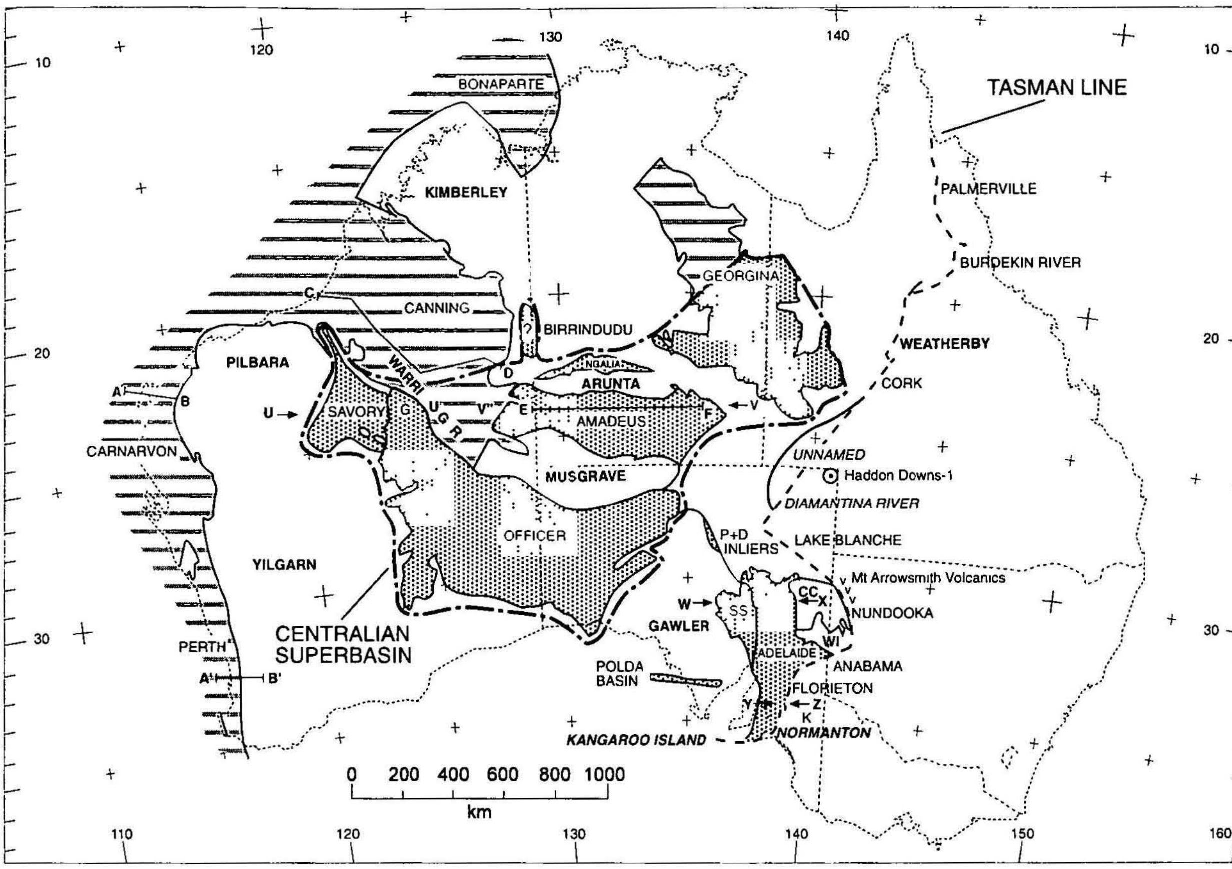


Figure 1. Centralian Superbasin (dot-and-dashed outline) and constituent structural basins (stippled) (amended from Walter et al. 1995) (see below). Stuart Shelf (SS), Adelaide Rift Complex and adjacent structures: CC = Curnamona Craton; WI = Willyama Inliers; K = Kanmantoo Fold Belt (Scheibner 1993). Proterozoic is delimited by the Tasman Line (Murray et al. 1989), with components labelled as faults (e.g. PALMERVILLE), an undifferentiated structure (WEATHERBY), a lineament (*DIAMANTINA RIVER*) and a shear zone (*NORMANTON*). Phanerozoic basins on west shown by ruled pattern. A–B, C–D, E–F locate the 500–45 Ma time-space diagrams (Fig. 4) and cross sections (Fig. 5); and U–V, W–X, Y–Z locate the 900–445 Ma time-space diagrams (Fig. 4) and cross sections (Fig. 5). The Centralian Superbasin is modified from Walter et al. (1995, fig. 1) in these ways: (a) the Savory Basin is expanded to the north and northwest by including the Tarcunyah Group, correlated with Supersequence 1 (Bagas et al. 1996); (b) to the east, the northern limit of the Superbasin passes about 22°S towards the Ngalia Basin, north of the Gibson (G) Sub-basin (with Supersequence 1 seen in diapirs) (Wells 1980) across the Warri Gravity Ridge (GR) into an area of >8 km deep basement beneath the thin Phanerozoic Canning Basin (Walter & Gorter 1994, fig. 2). Basement <2 km deep in this area could be a continuation of the Petermann Ranges uplift of the Musgrave Block; c) the southeastern outline of the Officer Basin is modified from Preiss et al. (1993, fig. 6.1). The Polda Basin and the Peake & Denison (P+D) Inliers are added from Preiss et al. (1993, fig. 6.2).

Correlation and calibration (Figs 2, 4). There is a 200 m.y. gap between latest Mesoproterozoic orogeny and the next recorded events, mafic volcanism (Volcanic Interval I–V/I) of the Amata mafic suite (AMS), Gairdner dyke swarm (GDS), and Willouran volcanic province in the Callanna Group, including the Rook Tuff (Preiss, 1987), all dated ca 800 Ma (Zhao & McCulloch 1993, Zhao et al. 1994). This was at least partly coeval with the deposition of Supersequence 1, including consanguineous lavas in the Bitter Springs Formation and sills in the Jillyili and Coondra Formations. Time-slices A and B are arbitrarily assigned dates younger than these 800 Ma events. Formations within Supersequence 1 are correlated by means of lithostratigraphy and stromatolite and preliminary acritarch biostratigraphy (Walter et al. 1995, Zang 1995; K. Grey, Geological Survey of Western Australia, pers. comm. 1996).

A gap of unknown duration is followed by Supersequence 2, conventionally assigned a date of ca 700 Ma, with Volcanic Interval II (V/II) represented by possible tuff and agglomerate in the Appila Tillite of the Adelaide Rift Complex (Preiss 1987, p. 364) and possibly the Wantapella Volcanics (option '**a**', Preiss 1987, 1993). The evidence for volcanism is weak, as the volcanics in the Appila Tillite could be re-worked from an older unit (Preiss, Geological Survey of South Australia, pers. comm. 1996) and the Wantapella Volcanics are more likely to be younger (see below); we have included them here to draw attention to the possibility of volcanism at this time. We agree with many authors that the Tapley Hill Formation and correlatives constitute a post-glacial stratigraphic marker. In the Adelaide Rift Complex, the Umberatana Group includes the Marinoan glacials, correlated with the Varanger glaciation, which is considered to be 590–610 Ma (Knoll & Walter 1992). Within this interval, the Ediacara fauna is about 560–570 Ma, apparently about the same age (564 ± 40 Ma by the Rb/Sr method, but with many uncertainties) as the Table Hill Volcanics (Compston 1974), and probably the Antrim Plateau Volcanics; with the probably somewhat older Wantapella Volcanics (the more likely option '**b**') this comprises Volcanic Interval III (V/III).

Supersequence 1

We attempt to portray the palaeogeography at two intervals during the deposition of Supersequence 1. Time-slice 1A (Fig. 3) is during the deposition of the basal sand sheet, and time-slice 1B, the overlying succession of carbonate, evaporite and fine-grained siliciclastics. Without an objective time correlation of these rocks and in the absence of sequence

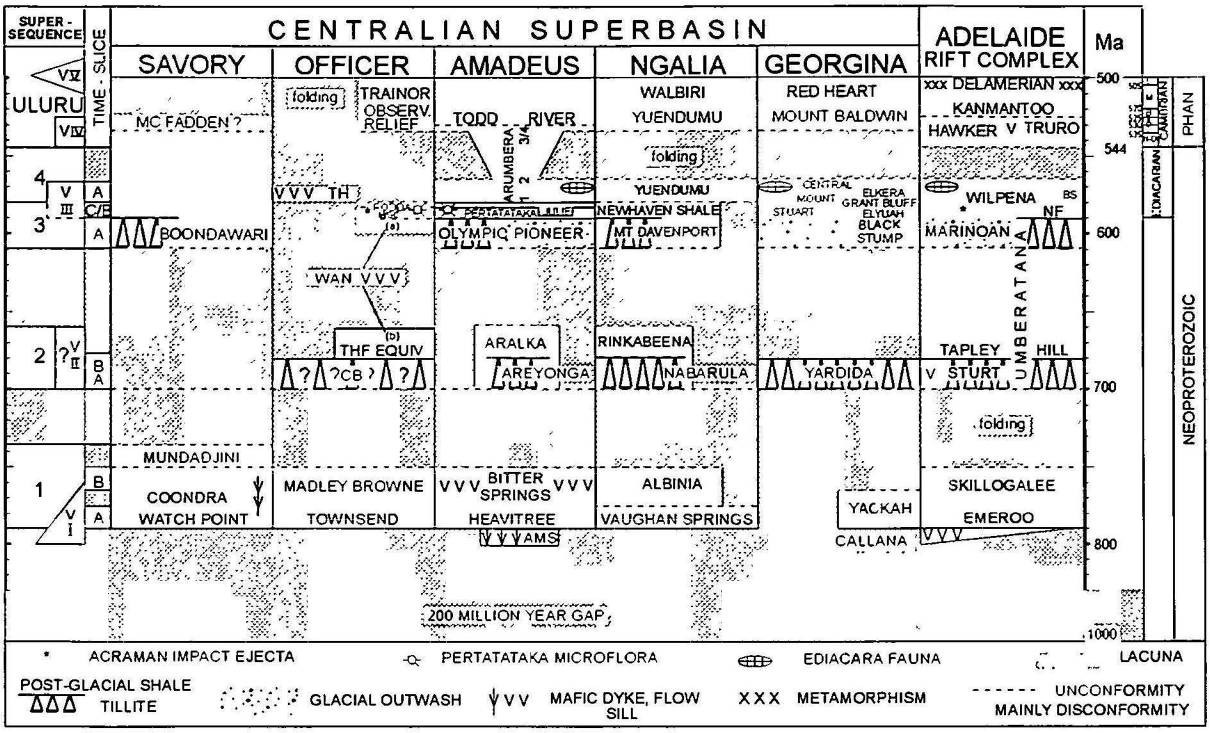


Figure 2. Correlation chart of the Centralian Superbasin and Adelaide Rift Complex, with ages calibrated in Ma. AMS = Amata mafic suite. BS = Billy Springs Formation. CB = Chambers Bluff Tillite. M/GC = Murnaroo Formation/Giles Creek Mudstone. NF = Nuccaleena Formation. TH = Table Hill volcanics. THF equiv. = Tapley Hill Formation equivalent ? V = times of volcanism. WAN = Wantapella Volcanics. After Walter et al. (1995), which should be seen for detailed discussion.

analyses in most structural basins we rely on lithostratigraphy and stromatolite and preliminary acritarch biostratigraphy.

Time-slice 1A

No geological record of the interval after the 1.1 Ga Musgrave Ranges and other orogenies and slightly younger dyke swarms is known until the 800 Ma start of Volcanic Interval I (V/I in Fig. 2).

We have assumed that the basal sand sheet of the Centralian Superbasin is at least roughly the same age everywhere, and that it correlates with the lower part of the Emeroo Sub-group of the Adelaide Rift Complex (time intervals T-1 and T-2 of Preiss 1987). Preiss (1987) interpreted T-2 as a time of relative high sea level, consistent with our assumption of widespread marine sedimentation in the adjacent Centralian Superbasin. No correlatives of the underlying Callanna Group of the Adelaide Rift Complex are known with any confidence from elsewhere in Australia, although Preiss (1987) and Zang (1995) recognise a possible correlative in the eastern Officer Basin. Correlations of the Callanna Group and the overlying Burra Group in the Adelaide Rift Complex with units in the Centralian Superbasin are still uncertain and controversial. An alternative interpretation to that presented here is that the Callanna Group belongs to Supersequence 1 and the Burra Group is younger (Bagas et al. 1996).

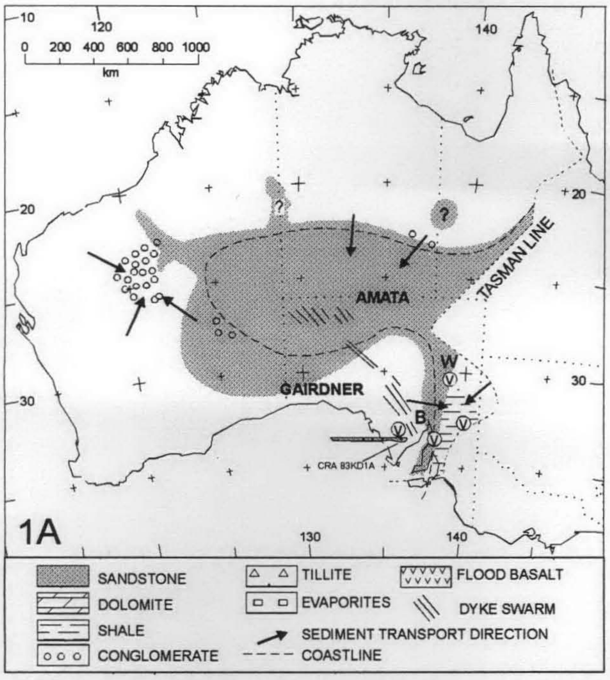
In the southern Georgina Basin, the sand sheet wedges out against several internal basement highs and against granitic basement (Walter 1980). At the locations of such wedge-outs carbonates are interbedded with the sands, and so the sands and carbonates of this supersequence are partly coeval.

As yet there are no detailed sedimentological studies of the sand sheet in the Centralian Superbasin, but most observers consider that it is a mixed fluvial and shallow marine succession. In the Heavitree Quartzite of the Amadeus Basin, palaeocurrent directions and isopachs suggest a provenance from the northeast to north-northeast. This is consistent with the observation that the correlative unit in the Georgina Basin, the basal Yackah beds, thins to the north and wedges out against granitic basement. Fluvial facies predominated, but parts were intertidal to very shallow marine (Clarke 1976). Lindsay (1991) recognised cycles which begin with a deeply channelled erosion surface, on which were deposited conglomerate and sandstone from braided streams, or laminated shale deposited in lakes. The basal beds are overlain by laminated and cross-bedded sandstone of mixed fluvial and dune origin, and pass upwards into well-stratified, cross-bedded, tidal sandstones. The cycle terminates with well-laminated or parallel-bedded, fluvial sandstone, indicating a sheet-flood event. The Vaughan Springs Quartzite of the Ngalia Basin has been interpreted as the deposits of alluvial fans overlain by fluvial, tidal flat and marine sandstone and conglomerate (Clarke 1976). The Townsend Quartzite of the Officer Basin was interpreted by Jackson & van de Graaff (1981) as shallow marine to deltaic.

In the Savory Basin the Jilyili Formation formed a westerly and northwesterly prograding delta, fed by the fluvial system of the Brassey Range Formation, and emptying into the shallow-marine depocentre of the Glass Spring Formation (Williams 1992). The possibly correlative Watch Point and Coondra Formations were sourced from the northwest and southwest, and include beds of conglomerate. The Watch Point Formation is interpreted as a rapidly prograding, coarse-grained delta (Williams 1992). The sandstone of the Googhenama Formation of the Tarcunyah Group occurs on the northern margin of the Savory Basin (Bagas et al. 1996).

The Redcliff Pound Group and possible equivalents, all in the Birrindudu Basin northwest of the Ngalia Basin, consist of poorly outcropping and poorly known siliciclastics and minor carbonates which have been suggested to correlate with Supersequence 1 (Blake et al. 1979, K. Grey in GSWA 1990). This correlation is based only on the common presence of siliciclastics and carbonates in the Redcliff Pound Group and Supersequence 1. The Redcliff Pound Group is overlain by a thin outlier of the Antrim Plateau volcanics.

In the Adelaide Rift Complex 'deposition probably took



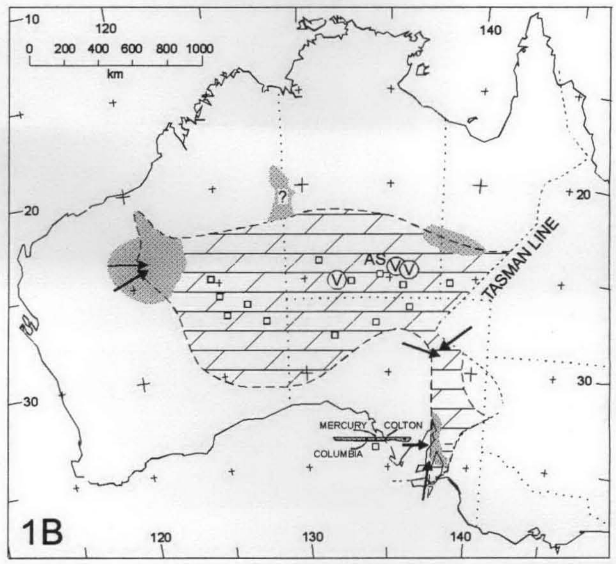


Figure 3. Palaeogeographic maps. Time-slice 1A, lower Supersequence 1. Also shown are the underlying Amata and Gairdner dykes and the underlying Beda (B) and Willouran (W) volcanics of the Callanna Group. In the Polda Trough, sediment of this age is associated with basalt in CRA 83KD1A drill-hole (Flint et al. 1988). Time-slice 1B, upper Supersequence 1. Drill-holes in the Polda Trough penetrate sediment of this age (Flint et al. 1988). AS = Alice Springs. Time-slice 2A, lower Supersequence 2. Time-slice 2B, upper Supersequence 2. Time-slice 3A, lower Supersequence 3. Time-slice 3B, middle Supersequence 3. Time-slice 3C, upper Supersequence 3. Time-slice 4A, lower Supersequence 4. Short arrows indicate presence of Table Hill Volcanics in seismic sections, solid black is outcrop, encircled Vs indicate inferred extent (Cook 1988).

place in proximal alluvial fans ... grading basinward to braided streams and perhaps meandering streams' (Preiss 1987, p. 332). Subsequent marine transgression resulted in the deposition offshore of fine sand, silt and mud. Transgression from the southeast, inferred by Preiss (1987), cannot be observed because of lack of outcrop.

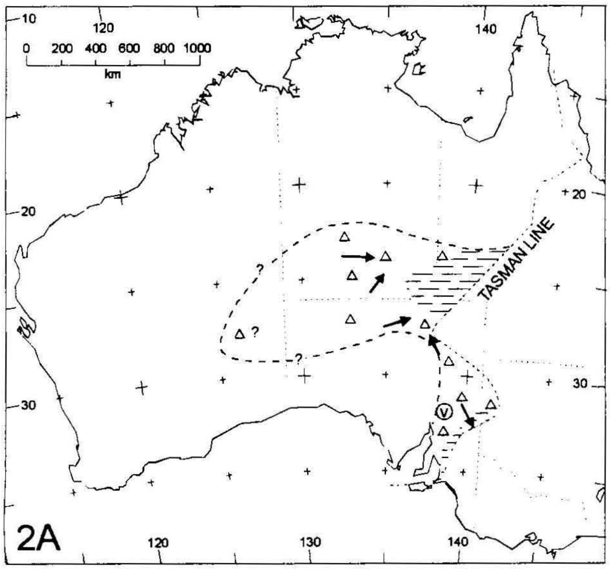
Time-slice 1B

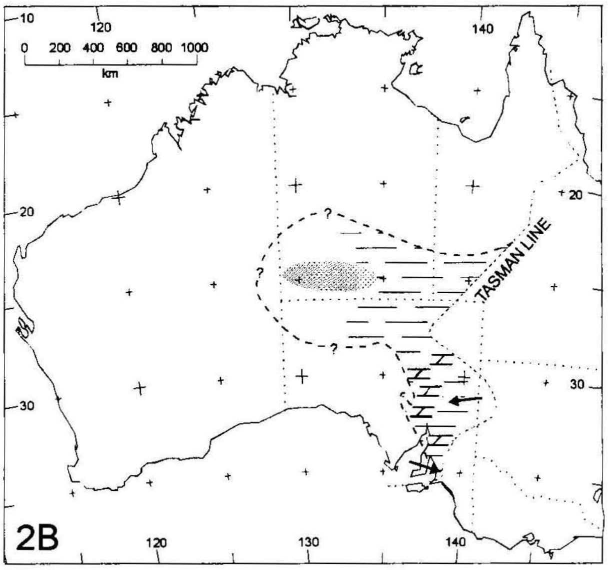
The upper part of Supersequence 1 is characterised by interbedded stromatolitic carbonates and evaporites including halite and anhydrite (Fig. 3, 1B). The lithostratigraphic correlations within the Centralian Superbasin are convincing because of the presence of distinctive lithologies, continuity in outcrop or in seismic records, and the presence of a distinctive assemblage of columnar stromatolites (Walter et al. 1995). Correlation to the Adelaide Rift Complex, based on lithological similarities and the common presence of the stromatolite *Baicalia burra* (Walter et al. 1995), is contentious.

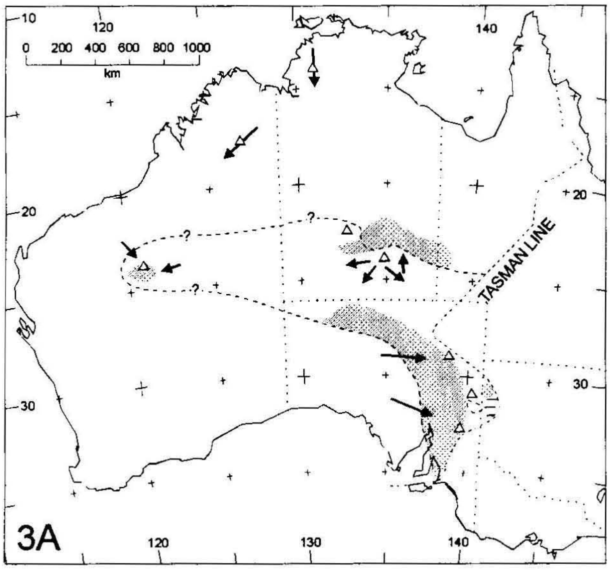
In the Amadeus Basin, the lower (Gillen) member of the Bitter Springs Formation contains thick units of halite and anhydrite as well as dolostone and fine siliciclastics. Stewart (1979) suggested that the evaporites formed in a marine barred basin, but Lindsay (1987a) favoured deposition in two major, poorly circulated anoxic sub-basins of the saline giant type, with carbonates and sulphates developed around basin margins, and halite and possibly potassium salts deposited in the sub-basin centres. A significant period of erosion occurred after deposition of the Gillen Member (Southgate 1989, 1991). The environment of the upper (Loves Creek) member began with deeper water, quiet marine conditions and then shallowed upwards to an emergent lacustrine setting (Southgate 1989, 1991). The central stromatolite-dominated unit represents an initial transgressive stage, with domical stromatolites alternately buried and exposed by the movement of ooid grainstones. Following the transgression, water conditions were deeper and quieter. There was then an abrupt reversal in stromatolite growth patterns and conditions became cyclic and upward shallowing, with evidence of emergence at the top of each cycle. The uppermost part of the Loves Creek Member is interpreted as an environment of shallow metahaline to hypersaline lakes in a flood plain (Southgate 1986, 1991). The stromatolite development indicates gradual shallowing culminating in emergence, indicated by desiccation cracks, indurated surfaces, karst deposits, and halite pseudomorphs and other evidence of brine concentration in ponded areas. Cyanobacterial communities inhabited the shallow metahaline to hypersaline lakes and were preserved by saline groundwaters and silica precipitation. Mafic lavas in the Loves Creek Member (Wells et al. 1967, Shaw & Wells 1983) are consanguineous with the Amata and Gairdner dyke swarms and volcanics in the Callanna Group (Zhao et al. 1994) and mark the end of Volcanic Interval I. A long period of exposure resulted in the weathering and erosion of the top of the carbonate units.

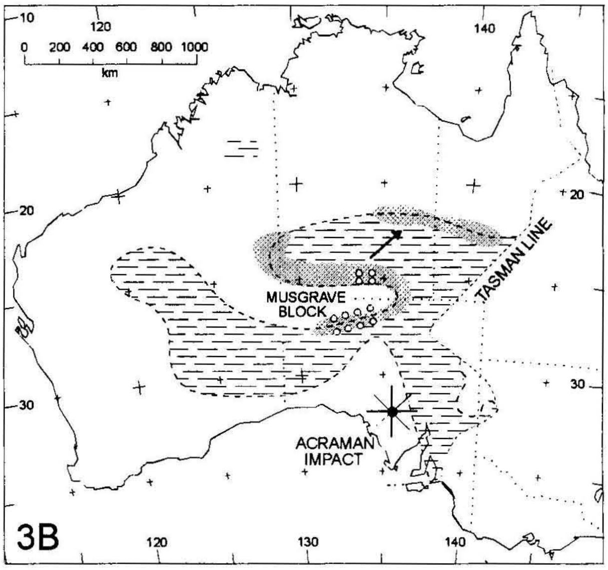
The Albinia Formation of the Ngalia Basin and the Yackah beds of the Georgina Basin are both siltstone and shale interbedded with stromatolitic cherty dolostone. Little is known about either because of very poor outcrop. Correlative units in the Officer Basin are known by many different names, including the Alinya, Browne, Madley, Kanpa, Hussar and Steptoe Formations (Walter et al. 1995, Zang 1995). These comprise stromatolitic dolostone, anhydrite, and halite, with interbedded sandstone, siltstone and shale. Seismic interpretation suggests that at least 4000 m are present in the Yowalga Sub-basin (Townson 1985). There are columnar stromatolites and chert microfossils closely comparable with those in the Bitter Springs Formation; the rock types are also very similar, and a similar set of palaeoenvironments can be envisaged, although there have been no detailed studies.

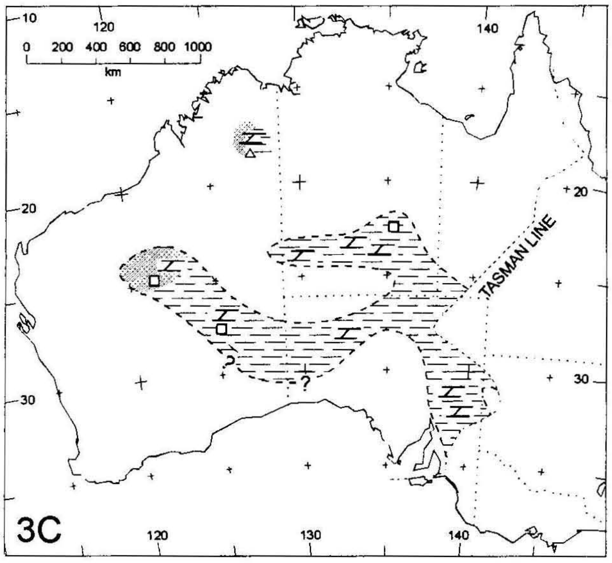
In the southeast Savory Basin the Skates Hills Formation consists of stromatolitic dolostone, sandstone, siltstone, shale, local thick conglomerate, and minor chert (Williams 1992). Evaporitic conditions are indicated by the presence of cauliflower chert and crystal voids after gypsum. Both the rock types and the contained stromatolites are closely comparable with those of the Bitter Springs Formation (Grey 1995). The Mundadjini Formation is considered by Williams (1992) to

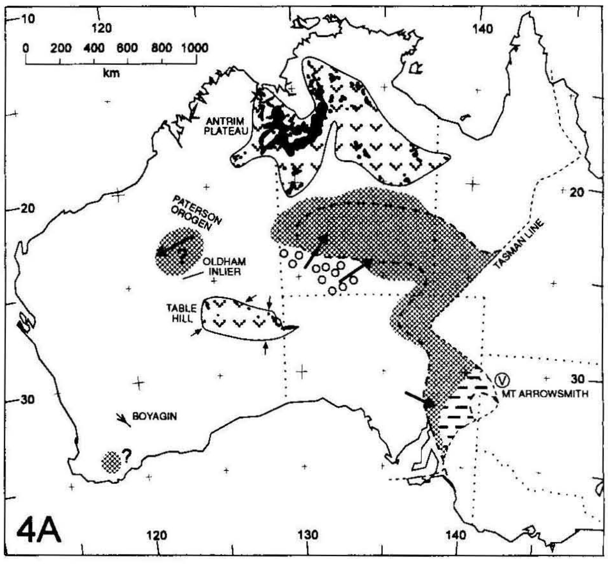

2A
TASMAN LINE
0 200 400 600 800 1000
km


2B
TASMAN LINE
0 200 400 600 800 1000
km


3A
TASMAN LINE
0 200 400 600 800 1000
km


3B
TASMAN LINE
MUSGRAVE BLOCK
ACRAMAN IMPACT
0 200 400 600 800 1000
km


3C
TASMAN LINE
0 200 400 600 800 1000
km


4A
ANTRIM PLATEAU
PATERSON OROGEN
OLDHAM INLIER
TABLE HILL
BOYAGIN
MT ARROWSMITH
TASMAN LINE
0 200 400 600 800 1000
km

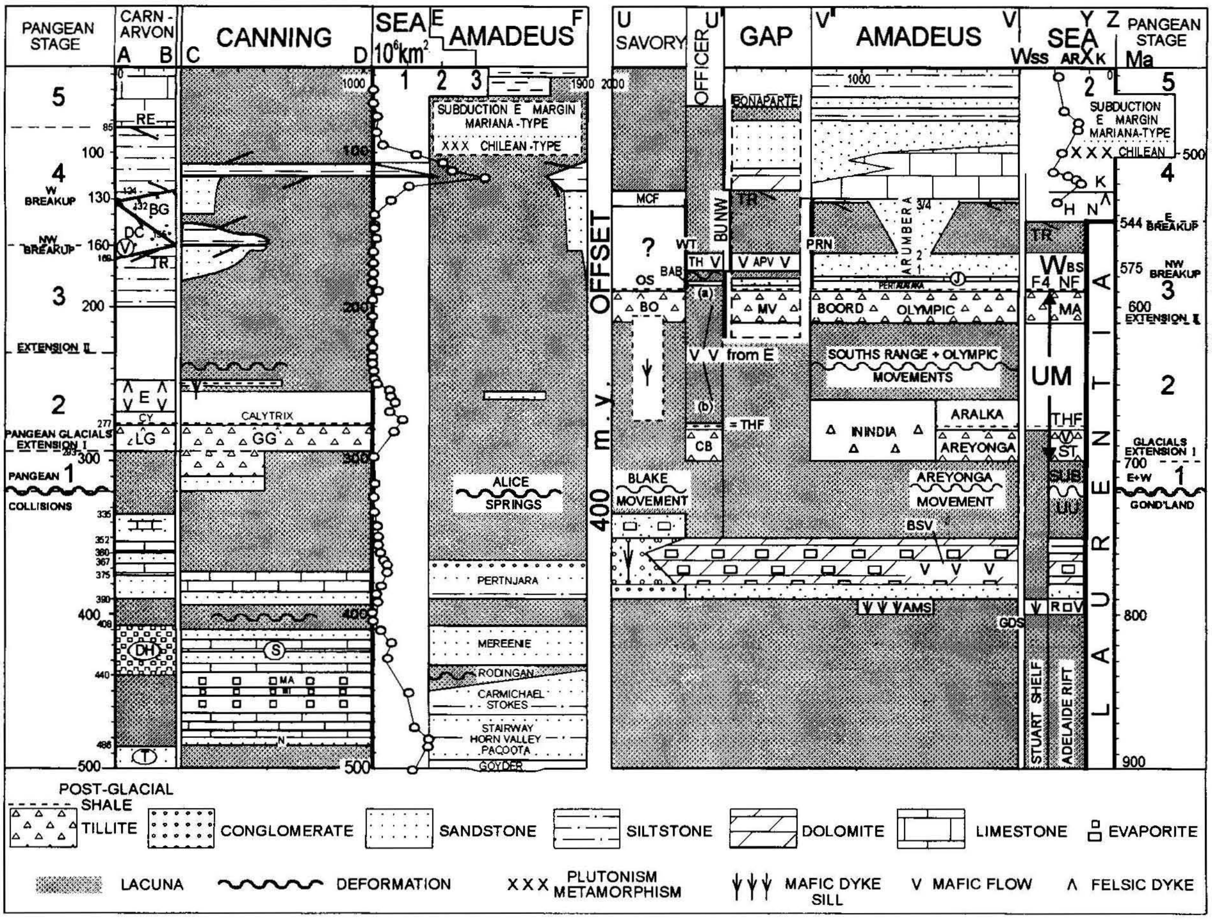


Figure 4. Time-space diagrams from 500 to 45 Ma along the 1900 km long line of A–B, C–D, E–F, and from 900–445 Ma along the 2000 km long line of U–V, W–X, Y–Z, both located in Fig. 1.

500–45 Ma AB–CD–EF (left side of diagram).

Pangean stages from Veevers (1990); BU = breakup. Pangea-forming events at the start of stage 1 (ca 320 Ma) are the collision of Laurussia and Gondwanaland, marked by the Sudetic (330–320 Ma) phase of the Variscan orogeny of Europe-Africa (Ziegler 1990, Veevers et al. 1994), the collision of Siberia with Baltica (Sengor et al. 1993) on one side, and possibly with Greater Australia on the other, as suggested by palaeomagnetism (Klootwijk 1995).

AB: Carnarvon Basin (Barber 1988, Gorter et al. 1994). BG = Barrow Group; CY = post-glacial Carrandibby Formation succeeding the LG = Lyons Group glacial sediment; DC = Dingo Claystone; DH = Dirk Hartog Group; E = Edell volcanics (Veevers & Tewari 1995); RE = regression; T = Tumblagooda Group; TR = transgression. The encircled V at 160 Ma denotes the Wandagee Province (WP) of picritic diatremes dated by U-Pb of zircon as 16010 Ma, coincident with the age of breakup on the northwest (GSWA 1990, pp. 566, 587).

CD: Canning Basin (Kennard et al. 1994, Gorter et al. 1994). Details of post-glacial Calytrix Formation on diamictite of the Grant Group (GG) from Redfern & Millward (1994) and Foster & Waterhouse (1988). The mafic sills at ca 250 Ma are from Veevers & Tewari (1995). MA = Mallowa Salt; MI = Minjoo Salt; N = Nambeet Formation; S = Sahara Formation.

Sea: area of Australian platform flooded by the sea (Veevers 1995).

EF: Amadeus (BMR Palaeogeographic Group 1990, Gorter et al. 1994). The region was inverted at the 320 Ma Alice Springs orogeny, dotted with nonmarine sediment in the Late Permian (255 Ma) and intermittently in the Cenozoic on either side of an Oligocene lacuna (Senior et al. 1995), and encroached by the sea at the peak of the Aptian (115 Ma) transgression (Veevers 1995). Rodingan Movement from Oaks et al. (1991, p. 84). In box near top: Subduction eastern margin, change from Chilean- to Marianas-type subduction in Cenomanian (95–90 Ma)(Veevers 1984, 1991).

900–445 Ma: UV–WX–YZ (right side of diagram).

UV (i) Savory Basin (Walter et al. 1994). MCF = McFadden Formation. BO = Boondawari Formation diamictite and OS = overlying siltstone (Walter et al. 1995, p. 186); mafic sills are 'vesicular and amygdaloidal fine-grained basalt ... near the contact of the Watch Point and Coondra Formations ... emplaced at a shallow depth early in the history of the basin' (Walter et al. 1994, p. 535), possibly at the same time as the eruption of the Bitter Springs volcanics in the Amadeus Basin. 'A preliminary Rb-Sr isochron date of c.640 Ma for a coarse-grained dolerite that intrudes the glacial Boondawari Formation' (Williams 1994, p. 843) plots below the ca 600 Ma presumed age of the glacials but possibly falls within the (unstated) range of error. The Blake Movement took place soon after the deposition of the Coondra Formation, and involved inversion of the basin along its northwest margin in the Blake Fault and Fold Belt (Williams 1994, p. 847).

(ii) Officer Basin (Walter et al. 1994). 'VV from E' = Wantapella Volcanics, projected from the east, overlie the Chambers Bluff Tillite (CB), which is overlain by a thin laminated dolomite and siltstone, possibly equivalent to the Tapley Hill Formation (THF) (Preiss 1987, p. 201–203). Preiss (1993) groups the Wantapella Volcanics with the overlying strata, option (a), but they could be much older, option (b).The Babbagoola Formation (BAB) was deformed by thrusting and stripped before being overlain by the Table Hill Volcanics (TH). On the northern and eastern edge of the Officer Basin, the Musgrave Block was uplifted at 590 Ma during movement along the Woodroffe Thrust (WT) in the south and the Petermann Range Nappe (PRN) in the north, as indicated by a widespread sheet of sand on the south, west, and north (ABC Quartzite, Murnaroo Formation, Cyclops Member, Grant Bluff Formation) to 550–530 Ma, the age of syntectonic muscovite in the Musgrave Block (Maboko et al. 1992).

(iii) Gap in information of 445–750 Ma rocks occupied by column of Bonaparte Basin and East Kimberley projected from the north. Data from Cook (1988), Shergold (1995), BMR Palaeogeographic

be the western lateral equivalent of the Skates Hill Formation, although it is predominantly siliciclastic. Evaporite pseudomorphs are abundant. Its presence on the tectonically active western margin of the superbasin is consistent with the interpretation of a facies change westwards from carbonates to sandstones (Fig. 3, 1B). On the northern margin of the Savory Basin the Waroongunyah and Waltha Woora Formations of the Tarcunyah Group consist of stromatolitic carbonates, red beds and evaporites (Bagas et al. 1996).

The Burra Group is considered to be the correlative in the Adelaide Rift Complex; this correlation is tenuous and is based on lithological similarities and the possible common presence of a distinctive stromatolite (Walter et al. 1995). During times T-5 to T-12 of Preiss (1987) there was widespread carbonate deposition in the shallower areas, giving way to black muds in the southeast. Oolitic, stromatolitic and intraclastic facies suggest deposition in marginal marine environments. Magnesite was deposited in lagoons. Sandy deltas spread from marginal areas so that towards the end of this period siliciclastic deposition was dominant. Late in these times organic-rich muds accumulated in relatively deep offshore environments, below wave-base.

Supersequence 2

Supersequence 2 followed after a substantial lacuna, spanning at least 50 Ma.

Time-slice 2A

We assume that all the glacial sediments that mark the base of this supersequence are approximately coeval. They are known from every basin in the Centralian Superbasin except the Savory Basin, but are patchy in their distribution and frequently thin. The patchy distribution and rapid lateral thickness changes result, at least in part, from syndepositional faulting. In the Kimberley region, the Landrigan Tillite, deposited on a pavement striated from the east, may be of this age (Coats & Preiss 1980), although recent work suggests that it is Marinoan (Plumb 1996, Corkeron et al. 1996). The thickest accumulations are in the Adelaide Rift Complex.

In the Amadeus Basin, the Areyonga Formation is preserved as erosional remnants on the Bitter Springs unconformity (Areyonga Movement, Fig. 4). Small steep-sided valleys are cut into the main erosional surface. The valleys are filled with large, randomly oriented blocks. Elsewhere the formation consists of diamictite, conglomerate and sandstone with rare dropstones and minor shale. Clasts are very varied and have both intrabasinal and extrabasinal sources. The lower, probably marine, part of the succession passes upwards into a fluvial succession. The overlying diamictites are thinner and probably represent subglacial and ice-margin deposits. The final phase of sedimentation consisted of shallow-marine ice-proximal deposits indicated by poorly bedded diamictites and sandstone bodies with abundant soft-sediment deformation (Lindsay 1989).

The Inindia beds of the southern Amadeus Basin are probably lateral equivalents of the Areyonga Formation. They comprise massive sandstone with siltstone interbeds, with a diamictite at several localities (Wells et al. 1966).

Group (1990), and Walter et al. (1994). MV = Moonlight Valley Tillite and overlying 15 m thick dark-red shale with sparse pebbles (Coats & Preiss 1980, p. 185; Preiss 1987, photo on p. 206). APV = Antrim Plateau Volcanics; BU NW indicates breakup of a continental sliver off northwest Australia (Veevers 1988).

(iv) Amadeus. **Phanerozoic from Wells et al. (1970) and Gorter et al. (1994), showing facies change from Cambrian sandstone in the west to carbonate in the east; Early Cambrian Arumbera 3/4 and underlying Neoproterozoic rocks from Walter et al. (1995), including post-glacial shale on tillite at 680 Ma (Aralka/Areyonga [AREY]) and 590 Ma (Pertatataka/Olympic). The Areyonga Movement produced uplift in the region, as indicated by clasts of the Bitter Springs Formation, the Heavitree Quartzite, and older rocks of the Arunta Block in the Areyonga Formation of the Amadeus Basin, and resulted in a radical change in the palaeogeography; the Souths Range Movement (Wells et al. 1970, p. 129) and the Olympic Movement (Oaks et al. 1991, p. 81) also produced uplift. The basin was disrupted in the south by uplift of the Musgrave Block at 590 Ma during movement along the Woodroffe Thrust (WT) and the Petermann Range Nappe (PRN), as described above for the Officer Basin. Dykes of the Amata mafic suite (AMS) ca 800 Ma from Sm-Nd mineral isochron dates of 790 ± 40 Ma and 797 ± 49 Ma (Zhao & McCulloch 1993; Zhao et al. 1994); and mafic lavas in the middle of the Bitter Springs Formation (Bitter Springs Volcanics, BSV) have a comparable isotopic signature (Zhao et al., 1994) but, by superposition, are younger. J = Julie Formation.**

WX: (i) Sea **(top, Phanerozoic, only), area of sea on Australian platform (Veevers 1995), continued down to 534 Ma from Sea column on left.**

(ii) Stuart Shelf (SS) and Adelaide Rift Complex (AR) **at 31°S across the Central Flinders Zone. Information from Preiss (1987, 1990, 1993). Sequences of post-glacial shale on tillite at 680 Ma are Sturtian/Tapley Hill Formation (THF) and at 590 Ma facies 4 = shale and carbonate of the Nuccaleena Formation (F4 NF)/Marinoan glacials (MA). The dykes of the Gairdner Dyke Swarm (GDS) underlie the Tapley Hill Formation (THF), and are dated by Zhao & McCulloch (1993) as ca 800 Ma from Sm-Nd mineral isochron dates of 802 ± 35 Ma and 867 ± 47 Ma. The Rook Tuff (R) in the northern Flinders area contains a lenticular porphyritic dacite, and zircons from the dacite are dated by the SHRIMP U-Pb method as 802 ± 10 Ma. Zircon-bearing rock of this age possibly extended 350 km southwest to the Acraman area of the Gawler Block whence it was ejected during a bolide impact and deposited in the Bunyeroo Formation. A zircon (grain 18) from the ejecta layer has an indistinguishable, possibly original, date of 804 ± 9 Ma by the SHRIMP U-Pb method (Compston et al. 1987, p. 444). The underlying Willouran volcanic province (WVP), including the Wooltana Volcanics, is grouped in the same ca 800 Ma range of ages from the common trace-element composition that suggests it belonged to a plume-related magmatic event (Zhao et al. 1994). Another ca 800 Ma date comes from zircon in a mafic granulite xenolith in a kimberlite pipe at the Calcutteroo locality, about 230 km north of Adelaide, that intrudes the folded Burra and Umberatana Groups. Chen et al. (1994) found by the SHRIMP method a date of ca 780 Ma, which they interpret as registering mantle-lower crustal magmatism shortly before 780 Ma. The regional unconformity beneath the Umberatana Group (UM)(SUB-UU) is 'more intense in the Central Flinders Zone, where several diapirs became active and the Burra Group was stripped from many areas. Angular unconformities ... suggest tilting of fault blocks and perhaps minor compressive deformation. ... Locally the Burra Group was folded relatively tightly adjacent to the Bungarider Fault in the Willouran Ranges' (Preiss 1987, p. 265). BS = Billy Springs Formation; GDS = Gairdner Dyke Swarm; H = Hawker Group; R = Rook Tuff; UM = Umberatana Group; W = Wilpena Group.**

YZ: The Kanmantoo Fold Belt (K) **(Scheibner 1993), as seen in the Kanmantoo Trough about 35°S, comprises an Early Cambrian succession of basal sandstone and carbonate and mixed carbonate/clastics of the Normanville Group (N) and presumably equivalent basalt, porphyritic andesite, and flow-banded trachyte of the Truro Volcanics (encircled V) (Preiss 1987, p. 268), overlain by the Early–Middle Cambrian Kanmantoo Group including turbidite fans that prograded towards the east or southeast (Jenkins 1989). All were intensely deformed and overthrusted, and injected with granite during the Delamerian orogeny about 500 Ma (crosses). 'Subduction eastern margin, change from Chilean- to Marianas-type', latest Cambrian (ca 500 Ma) (Powell 1984). Continent-backed lithosphere, probably Laurentia, was adjacent to Australia along the Tasman Line until breakup about 544 Ma (von der Borch 1980). K = Kanmantoo Group; N = Normanville Group; encircled V = Truro Volcanics.**

Interpreted 700–544 Ma Pangean stages, **modified from Veevers (1990). The deformation at the start of stage 1 (ca 720 Ma) is that reported from East Africa, interpreted as the collision of East and West Gondwanaland (E+W G) (Maboko et al. 1985, Stern 1994).**

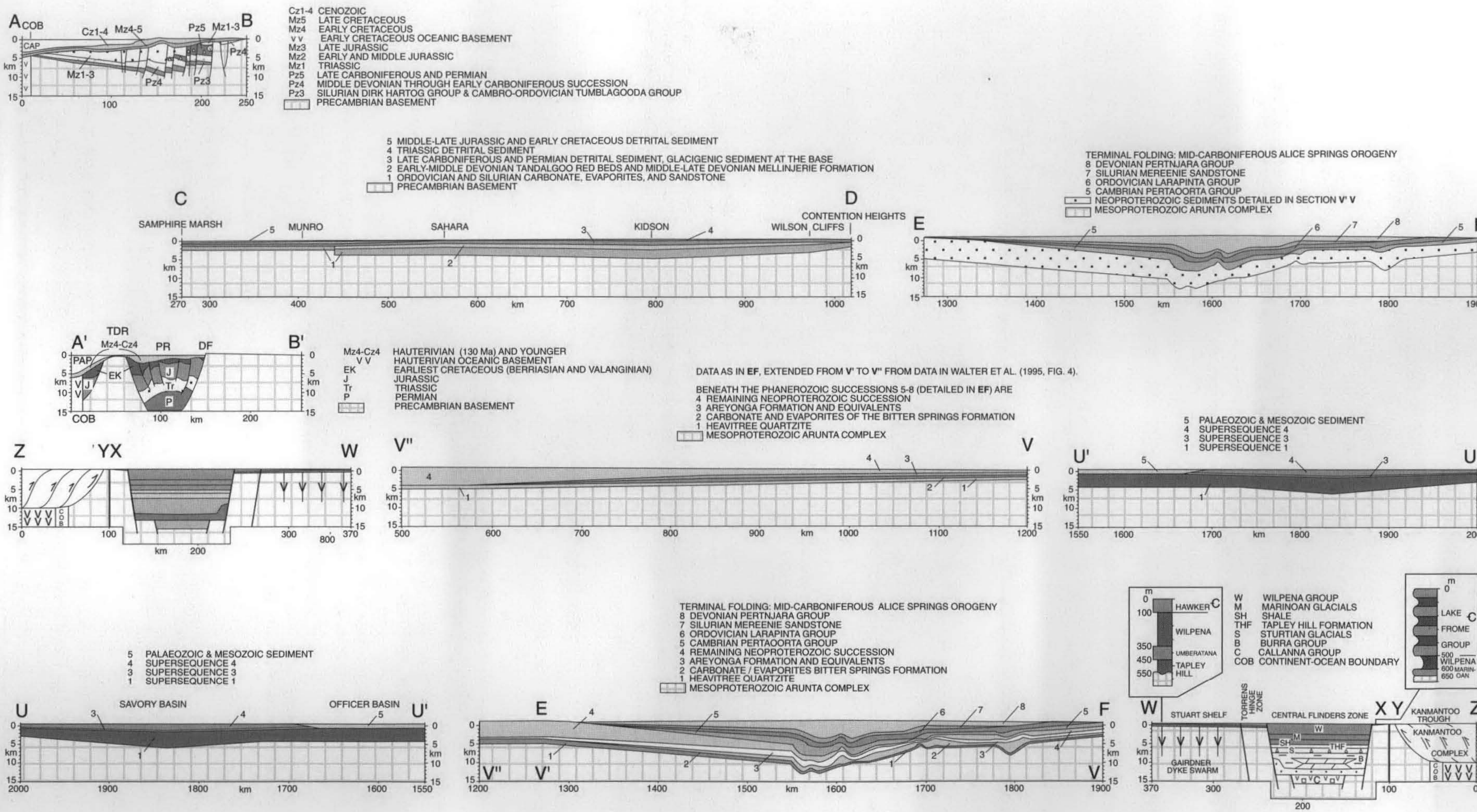


Figure 5. Cross sections, located in Fig. 1, along the 1900 km long line of A–B, C–D, E–F from 500 to 45 Ma, and along the 2000 km long line of U–V,W–X, Y–Z from 900–445 Ma. V:H = 4.

The Naburula Formation of the Ngalia Basin consists of a basal diamictite, shale and 'cap dolomite'. At the type section the diamictite is only about 2 m thick (Wells & Moss 1983). The Yardida Tillite of the Georgina Basin consists of light to dark green-grey diamictite and laminated siltstone with infrequent fine to very coarse-grained sandstone and arkose. The equivalent Mount Cornish Formation consists of blue-green diamictite with clasts, up to 12 m in diameter, of gneiss, pegmatite, granite, dolerite, sandstone, and dolostone, with interbeds of green varve-like siltstone (Walter 1980).

In the eastern Officer Basin, the Chambers Bluff Tillite consists of pebbly, silty diamictite and calcareous diamictite, a middle unit of silty diamictite containing abundant clasts, and an upper thin unit of sandstone and minor limestone. At the base of the succession are graded sandy siltstones that may be glacial varves. Clasts comprise a wide variety of igneous, sedimentary and metamorphic rocks. There is no internal evidence for the age of the Chambers Bluff Tillite, but it is suggested by Preiss (1993) that the grey dolomite and silty shale and the overlying sandstone near the top of the succession may be correlatives of the Sturtian Tapley Hill Formation and Marinoan glacial sandstones of the Adelaide Rift Complex, respectively. So a Sturtian age for the tillite is most likely (see Preiss 1987 and 1993 for discussion). The diamictite is probably of marine-glacial origin (Preiss 1987). The environment may have been initially lacustrine in the Chambers Bluff area, and the succession may have transgressed westward on to basement. Deeper marine, iron-rich, sediments were deposited in the east.

All the glaciogenic sediment described above is assumed here to correlate with the most extensive of the Sturtian glacial units of the Adelaide Rift Complex, at time S-5 of Preiss (1987). Such a correlation is consistent with the correlation of the superjacent 'lower marker cap dolomite', which occurs in all basins except the Kimberley region. In the Adelaide Rift Complex, glaciogenic sediment seems to have been derived from the west (Gawler Craton) and the northeast (Curnamona Cratonic Nucleus and Muloorina Ridge). Highlands in these areas collected ice which fed glaciers in the major valleys, and these coalesced in the lowlands to form a continuous thick ice-sheet over the shelf regions (Preiss 1987). Agglomerate and tuff are reported from the Appila Tillite (Preiss 1987), but these may be reworked from older volcanics (Preiss pers. comm. 1996).

Time-slice 2B

Widespread deposition in a shallow epeiric sea of a thick succession of silt and mud is interpreted as reflecting a major post-glacial eustatic rise in sea level. The succession shallows up and in places there are peritidal carbonates and sands.

The Aralka Formation of the Amadeus Basin consists of evenly bedded green siltstone and shale with subordinate sandstone. Near the base, the siltstone and shale are dark grey and contain dolostone concretions (the 'lower marker cap dolomite'), as elsewhere. In the northeastern part of the basin, the Ringwood Member consists of stromatolitic dolostone, limestone and siltstone, and the Limbla Member consists of siltstone, shale, sandy calcarenite and festoon cross-laminated sandstone. Little work has been carried out on the environmental interpretation of the Aralka Formation. The dominant rock types, laminated siltstone and shale with some ripple marks, were presumably deposited in relatively deep water, but still

Figure 5 (details). TOP ROW

***AB*. Northern Carnarvon Basin, at about 22°S. From west to east, the section crosses the easternmost Cuvier Abyssal Plain (CAP) and the continent-ocean boundary (COB) (from Veevers & Cotterill, 1978), the main part of the Carnarvon Basin to the Darling Fault System, and finally the thin cover over the Precambrian basement (from GSWA 1990, p. 459). The breakup unconformity in the west (W in Fig. 4) is marked by the base of Mz4. Breakup in the northwest (outer columns of Fig. 4) is reflected by the encircled V at 160 Ma (Fig. 4) that denotes the Wandagee Province of picritic diatremes dated by U-Pb of zircon as 160 ± 10 Ma (GSWA 1990, pp. 566, 587). Poorly dated mafic flows along the northwest margin, penetrated in wells at Ashmore Reef, Scott Reef, and near Yampi, are believed to be the same age (Veevers 1984, pp. 187, 188).**

***A'B'*. Section through Perth Rift Complex System at latitude of Perth, from eastern edge of Perth Abyssal Plain (PAP) across the continent-ocean boundary (COB), a presumed outer rift, the Turtle Dove Ridge (TDR), the inner, Perth Rift Complex (PR), the Darling Fault (DF) and the Yilgarn Block. From Veevers & Hansen (1981, fig. 7).**

***CD*. Onshore Canning Basin, from Forman & Wales (1981, plate 3), ages modified from Kennard et al. (1994) and Gorter et al. (1994).**

***EF*. Amadeus Basin, at about 24°S. AA[1] of Lindsay (1987a, fig. 7), subdivided and extended to east and west from information in Wells et al. (1970). Above the Mesoproterozoic Arunta Complex are the <800 Ma (1) Heavitree Quartzite and (2) carbonate and evaporites of the Bitter Springs Formation; Lindsay (1987a) detailed the growth of salt cores beneath the major structures; (3) Areyonga Formation and equivalents; (4) the rest of the Neoproterozoic sediments; (5) the Cambrian Pertaoorta Group; (6) the Ordovician Larapinta Group; (7) the Silurian Mereenie Sandstone; and (8) the Devonian Pertnjara Group, all terminally folded during the mid-Carboniferous Alice Springs orogeny.**

BOTTOM ROW

***UU'*. Neoproterozoic Savory Basin and, additionally, Phanerozoic Officer Basin. Data from Walter et al. (1995, fig. 4).**

***V"V'V (EF)*. Amadeus Basin, at about 24°S. AA[1] of Lindsay (1987, fig. 7), subdivided and extended to east and west from information in Wells et al. (1970). Above the Mesoproterozoic Arunta Complex are the <800 Ma (1) Heavitree Quartzite and (2) carbonate and evaporites of the Bitter Springs Formation; Lindsay (1987) detailed the growth of salt cores beneath the major structures; (3) Areyonga Formation and equivalents; (4) the rest of the Neoproterozoic sediments; (5) the Cambrian Pertaoorta Group; (6) the Ordovician Larapinta Group; (7) the Silurian Mereenie Sandstone; and (8) the Devonian Pertnjara Group, all terminally folded during the mid-Carboniferous Alice Springs orogeny.**

***WX*. Stuart Shelf, Central Flinders Zone, Curnamona Craton at about 32°S. Compiled from crustal cross-section (Preiss 1987, fig. 93, p. 260), with columns of the Stuart Shelf and Curnamona Craton (Preiss 1986). Thicknesses in the section of the Central (and projected North) Flinders Zone (Preiss 1987):**

UNIT	***km***
Wilpena Group	**3**
Marinoan glacials	**1**
shale	**1.5**
Tapley Hill Formation	**1**
Sturtian glacials	**1**
Burra Group	**6**
Callanna Group	**2.7**
Total	**16.2**

***YZ*. Kanmantoo Trough. From Jenkins' (1989) interpretive section of the Mount Lofty Ranges as a set of imbricate thrusts, with our interpretation of the continent-ocean boundary (COB) and Early Cambrian oceanic basement.**

MIDDLE ROW

Sections ZW and U'U are mirror images of WZ and UU' in the bottom row. VV" is derived from V"V by removing Phanerozoic strata 5–8 to simulate the end-Neoproterozoic state, and then orienting it (as also ZW and U'U) so that east is on the left, to afford direct comparison with AB-CD-EF above.

above storm wave-base. The abundant ooids, intraclasts and stromatolites of the Ringwood Member indicate a littoral environment. The top of the Naburula Formation of the Ngalia Basin comprises dark grey to black siltstone like that of the Aralka Formation (Wells & Moss 1983). Locally at the top of the Yardida Tillite of the Georgina Basin there is dark grey, laminated, dolomitic shale with, in its lower half, abundant lenses of dolostone (Walter 1980).

These formations are closely comparable with their presumed correlative in the Adelaide Rift Complex, the Tapley Hill Formation, of time interval S-6, interpreted by Preiss (1987) as marine and deposited mostly below wave base. A substantial amount of the pyrite in the formation is extremely depleted in ^{32}S ($\delta^{34}S$ up to +50, Lambert et al. 1984, Hayes et al. 1992), possibly indicating that access to the open ocean was restricted. The Tapley Hill Formation is the first unit to transgress westward from the Adelaide Rift Complex across the Stuart Shelf (Fig. 5, column above section WX).

Supersequence 3

Sequence analysis, chemostratigraphy and biostratigraphy all provide a basis for precise correlation within this supersequence, and offer the promise of narrow temporal resolution (Christie-Blick et al. 1995, Jenkins 1995, Walter et al. 1995). Because most of the detailed studies of chemostratigraphy and biostratigraphy have not yet been published, we analyse only three broadly defined time-slices.

Time-slice 3A

Glacial units lie at the base of the supersequence, and were deposited in environments of the same kind as Supersequence 2 (Fig. 3, 3A). Again, we have assumed that the glacial sediments are approximately coeval. Diamictites are not known from the Georgina Basin, nor with certainty from the Officer Basin, and elsewhere are patchy in their distribution. Arkose, conglomerate and arkosic sand are more widespread, and are interpreted as glacial outwash deposits.

In the eastern part of the Amadeus Basin, the Olympic Formation contains a distinctive reddish diamictite, considered to be tillitic in part, and containing striated and faceted clasts and dropstones. For the most part, the formation consists of red and green mudstone and siltstone, with intercalated sandstone, conglomerate and dolostone (Field 1991; Preiss et al. 1978, table 1). Rare lonestones occur in siltstone beds above mass-flow conglomerates. Most clasts are carbonate, others are quartz, quartzose sandstone, granite, and gneiss. Field (1991) interpreted the formation as indicative of a periglacial environment rather than a continental ice sheet. He regarded it as partly nonmarine and partly lacustrine or marine, with shoreface and foreshore deposits marginal to an ice-sheet of fringing glaciers, perhaps a trough between two areas of glacial activity.

The Pioneer Sandstone has been traced from the north-central Amadeus Basin east into the Olympic Formation (R.J.F. Jenkins, University of Adelaide, pers. comm.). It consists of cross-bedded feldspathic sandstone (Field 1991), except in the east, where much of the unit is conglomerate (Preiss et al. 1978). The upper part of the Pioneer Sandstone consists of two intertidal deposits: a cross-bedded unit with centimetre-amplitude, bimodal, tabular foresets; and an overlying unit of tidal channel-fill sands (Field 1991). The formation is interpreted as glacial outwash from the tillitic Olympic Formation (Preiss et al. 1978, Shaw & Wells 1983). Field (1991) interpreted the conglomerate as mainly a mass-flow deposit, and the lonestones as dropstones. He recognised non-marine, paralic and shallow-marine to basinal settings. A shoreface environment is inferred for the lower part of the formation, and a fluvial to paralic environment for the upper part.

In the Georgina Basin the Black Stump and Oorabra Arkoses form deposits up to 1000 m thick in half-grabens, and consist of arkose, pebbly arkose, conglomerate, siltstone and shale (Walter 1980) They are proximal deposits in a regime of rapid erosion and mass-transport, interpreted as glacial outwash deposits (Walter 1980).

The Mount Davenport Member, the lower part of the Mount Doreen Formation of the Ngalia Basin, comprises poorly sorted boulder, cobble and pebble conglomerate, diamictite and arkose. Clasts are subrounded, striated, and faceted, up to 4 m in diameter, and include igneous and metamorphic rocks, sandstone, and dolostone (Preiss et al. 1978, Wells & Moss 1983).

The Moonlight Valley Tillite and equivalents of northern Australia (Dow & Gemuts 1969, Edgoose 1986; Fig. 3, 3A) rest on glaciated pavements and are overlain by 'cap dolomites' very similar to those above the Marinoan glacials in the Adelaide Rift Complex. Recent work by Plumb (1996) and Corkeron et al. (1996) supports this correlation and disputes the alternative proposed by Coats & Preiss (1980).

The lower part of the Boondawari Formation of the Savory Basin consists of diamictite, rhythmite, sandstone, pebbly sandstone and conglomerate (Williams 1992, Walter et al. 1994). Clasts are angular, subrounded or wedge-shaped pebbles, cobbles, and boulders. They are polished, striated, and faceted, and include metamorphic, igneous and a wide variety of sedimentary rocks, resembling those of the Nabberu Basin to the southwest. The middle Boondawari Formation consists mainly of sandstone. Williams (1992) interpreted the diamictites as glaciogenic, indicating a glacial shallow-marine environment, distant from the ice-source. Clasts were derived from a wide variety of sources from outside the Savory Basin, particularly from the west (Williams 1992). The presence of glauconite suggests a marine environment. The middle part of the formation has been interpreted as a high energy, sandy shelf environment that was glacially influenced (Williams 1992). No correlative units are recognised with confidence in the Officer Basin, though a thin sandstone in the eastern part of the basin could date from this time interval (Preiss 1993), as could the diamictites of the Lupton and Turkey Hill Formations in the western part of the basin (Jackson & van de Graaff 1981).

The Marinoan glaciation of the Adelaide Rift Complex encompasses time intervals M6–M8 of Preiss (1987). On the Stuart Shelf to the west periglacial conditions produced patterned ground (Williams & Tonkin 1985) and dunefields. In the Adelaide Rift Complex, paralic sandstone of the Elatina Formation is inferred to tongue out laterally to the east and northeast into diamictites of the Mount Curtis and Pepuarta Tillites. On the basis of observed gradients in the concentrations of boulders, Preiss (1987) inferred that there were only local sources of ice rather than an extensive ice sheet. Rhythmite in the Elatina Formation has been interpreted as the distal deposits of glacial lakes (Williams 1983, Preiss 1987). Siltstone in the southeast is considered by Preiss (1987) to mark an opening to a fully marine basin in that direction. Palaeomagnetic studies of the Elatina Formation indicate deposition in near-equatorial latitudes, at 2.7° ± 3.7°N (Embleton & Williams 1986, Schmidt & Williams 1995). Evidence for grounded ice at low altitudes, in the form of glaciated pavements, is known only from the Moonlight Valley Tillite and equivalents of northern Australia (Dow & Gemuts 1969, Coats & Preiss 1980, Edgoose 1986; Figs 3, 3A), currently some 15° of latitude north of the Elatina Formation site.

Time-slice 3B

As for the earlier glaciation, a major eustatic rise in sea level followed deglaciation, resulting in the deposition of a thick succession of silt and shale in a shallow epeiric sea. The Musgrave Block emerged and shed coarse sediment to the north and south (Fig. 3).

The potential for fine subdivision of this time interval is illustrated particularly by the work of Preiss (1987), Jenkins

(1995), and Christie-Blick et al. (1995) in the Adelaide Rift Complex. We have decided not to work at this fine scale until the detailed chemostratigraphic and biostratigraphic studies of C. Calver and K. Grey, which allow comparable fine-scale subdivision of the Ediacarian successions in the Centralian Superbasin, have been published. In time-slice 3B we have generalised the palaeogeography of times equivalent to the lower and middle Wilpena Group of the Adelaide Rift Complex (M10–14 of Preiss 1987). If the implied correlations in Fig. 2 are correct, our map (Fig. 3, time 3B) can be taken to indicate the palaeogeography at time M14 of the Acraman meteorite impact on the Gawler Craton, west of the Adelaide Rift Complex, which spread an ejecta blanket north into the Officer Basin (Rodda beds) and east into the Adelaide Rift Complex (Bunyeroo Formation of the Wilpena Group). Fairly precise correlation to the Amadeus Basin with acritarch biostratigraphy and isotope chemostratigraphy indicates that the Acraman impact occurred soon after the deposition of the sands of the Cyclops Member in the middle of the Pertatataka Formation.

The Pertatataka Formation, including sandstones of the Cyclops and Waldo Pedlar Members, is present throughout the northeastern part of the Amadeus Basin. The formation consists of grey-green and purple-brown, laminated, micaceous siltstone and shale with thin interbeds of sandstone and rare limestone. In places it contains glauconite and clay pellets. Korsch (in Kennard et al. 1986) described one area of outcrop as shale with about 30% distal sandstone turbidite beds. Palaeocurrents are north to northeast directed. Wave-rippled tops suggest deposition within storm wave-base. In seismic sections the formation is distinguished by weak discontinuous parallel reflectors with, in places, large north-prograding clinoforms (e.g. Kennard & Lindsay 1991).

With a diverse assemblage of acritarchs, presumably marine plankton (Zang & Walter 1989, 1992; Grey 1993), the Pertatataka Formation is considered to be predominantly marine, and part of a single major upward-shallowing sequence which includes the carbonate of the overlying Julie Formation. Pertatataka deposition started during a sudden deepening of the basin so that the water depth of the basin was at its maximum. Pelagic muds and turbidites were deposited from turbidity currents travelling northward in an outer submarine fan to a basin plain that rose to a strandline in the northeast (Kennard et al. 1986).

The Winnall beds in the southwest of the Amadeus Basin consist of siltstone, sandstone, pebbly sandstone, dolostone and limestone (Wells et al. 1970). Conglomerate indicates uplift to the south with erosion of the Bitter Springs Formation and metamorphic and igneous rocks. Shallow marine conditions, indicated by glauconite and phosphate, in a subsiding depression then predominated.

The equivalent units in the Georgina Basin are the Elyuah and Grant Bluff Formations, and parts of the Elkera and Central Mount Stuart Formations (Walter 1980). The Elyuah Formation consists mainly of laminated, grey, green or red, fissile shale. The Grant Bluff Formation is mainly undulose laminated to thin-bedded, fine-grained quartz arenite. The lower Elkera Formation consists of interbedded siltstone, sandstone and shale. The sandstone locally contains anhydrite and pseudomorphs of anhydrite nodules and gypsum. The trace fossil *Planolites ballandus* occurs rarely. There are no detailed sedimentological studies, but trace fossils, stromatolites, ripple marks and sulphate evaporites suggest a shallow marine environment. Further west these units grade into the Central Mount Stuart Formation, which has a higher proportion of sandstone.

In the Ngalia Basin the Newhaven Shale Member of the Mount Doreen Formation consists of a uniform succession of red shale (Wells & Moss 1983). In the lower part of the Boondawari Formation of the Savory Basin, diamictite is overlain by a rhythmite (OS in Fig. 2). Above the rhythmite the dominant lithology is coarse to fine-grained sandstone containing scattered pebbles, cobbles and occasional boulders. Planar and trough cross-bedding are common. Polymict conglomerate fills penecontemporaneous scour-channels in sandstone (Williams 1992, Walter et al. 1994).

In the Kimberley region, the glacial units and the succeeding 'cap dolomites' are overlain by a succession of shale, siltstone, greywacke and sandstone (Coats & Preiss 1980 and earlier references therein).

In the eastern Officer Basin an unnamed siltstone is overlain by the Murnaroo Sandstone (M) and by the Rodda beds, which contain ejecta from the Acraman impact near the base (Wallace et al. 1989). The Murnaroo Formation is widespread in the eastern Officer Basin and consists of sandstone, which is fine to coarse-grained and rarely conglomeratic. The Rodda beds consist of grey-green siltstone, which is frequently calcareous and dolomitic. They include grey and minor pink, brown and purple limestone and dolostone beds, feldspathic and calcareous sandstone, and pebble–cobble conglomerate beds. The upper part of the succession consists of grey-green siltstone with thin beds of very fine sandstone. Sukanta et al. (1991) recognised five sequences in the Rodda beds; a canyon-forming event cuts down into the underlying Murnaroo Sandstone. Well-preserved acritarchs in the Rodda beds (Jenkins et al. 1992, K. Grey pers. comm.) are closely comparable with those in the upper Pertatataka Formation of the Amadeus Basin. The Rodda beds were deposited in a deep-water slope-and-basin environment by mass flow, turbidity currents and hemipelagic processes (Brewer et al. 1987).

The Babbagoola Formation of the western Officer Basin consists of shale, siltstone, sandstone, anhydrite, gypsum and conglomerate. In Hussar-1, conglomerate is overlain by interbedded sandstone, claystone and siltstone (Phillips et al. 1985). Jackson & van de Graaff (1981) and Jackson & Muir (1981) divided the succession into three units: a lower unit of fissile grey to green laminated siltstone and claystone about 100 m thick, a middle unit of dolostone or dolomitic sandstone (less than 10 m thick), and an upper unit of reddish-brown poorly sorted sandstone and siltstone. Recent work indicates that some of what has been called Babbagoola Formation is part of Supersequence 1 (K. Grey, Geological Survey of Western Australia, pers. comm. 1996).

In the Adelaide Rift Complex, mud and silt of the Bunyeroo Formation, including the Acraman impact ejecta layer (Compston et al. 1987), were being deposited in what Preiss (1987) interpreted as a moderately deep marine environment. Several diapirs formed small islands.

Time-slice 3C

In each region the post-glacial siliciclastic successions shallow upward to become rich in carbonate, locally with evaporites, ooid grainstones, and stromatolites (Fig. 3). The Julie Formation of the Amadeus Basin is a succession of dolostone, limestone and siltstone with lenses of sandstone. The dolostone is frequently oolitic, and locally contains stromatolites. The Julie Formation is the upper part of a single major upward-shallowing sequence which began with rapid deepening to form the Pertatataka Formation. Lithologies indicate a shallow marine environment with ooid shoals. The upper part of the Boord Formation in the western Amadeus Basin consists of oolitic, stromatolitic calcilutite and calcarenite, which are interbedded with siltstone and shale. No equivalent units are known from the Ngalia Basin.

The Elkera Formation of the Georgina basin consists of interbedded siltstone, stromatolitic dolostone, sandstone, and shale. The sandstone locally contains anhydrite and pseudomorphs of anhydrite nodules and gypsum. Some argillaceous beds contain halite pseudomorphs. Trace fossils, stromatolites, ripple marks, and sulphate evaporites suggest a shallow marine environment. The formation apparently grades west into sandstone of the Central Mount Stuart Formation, interpreted as deltaic (Shaw & Warren 1975, Walter 1980).

Recent reinterpretations of the stratigraphy of the Neoproterozoic succession in the Kimberley region suggest that the glacial diamictites of the Egan Formation represent a third, younger, glaciation not recognised in other basins in Australia (Plumb 1996, Corkeron et al. 1996). On the basis of the occurrence of the stromatolite *Tungussia julia* the Egan Formation has been correlated by these authors with the Julie Formation, and the included diamictites are suggested to result from a local mountain glaciation. Also included in the formation are dolomite, limestone, and fine to coarse-grained siliciclastics. Overlying this is the Yurabi Formation of dolomitic sandstone.

The upper Boondawari Formation of the Savory Basin is an argillaceous-carbonate association of upward-coarsening shale and siltstone to fine and coarse-grained sandstone. Halite casts occur in sandstone. Some sandstone and siltstone horizons have graded bedding, occasional mud-cracks, and load and flute casts. The dolostone locally is oolitic, pisolitic and stromatolitic.

In the Officer Basin, the middle Rodda beds include limestone and dolostone, as well as laminated siltstone and shale, and feldspathic and calcareous sandstone, and are correlated with the Julie Formation. As mentioned above, the Rodda beds were deposited in a deep-water slope and basin environment by mass flow, turbidity currents and hemipelagic processes (Brewer et al. 1987).

In the Adelaide Rift Complex at time M-15 of Preiss (1987), lime mud and fine calcareous sand and silt of the Wonoka Formation were deposited in a deep shelf to slope environment. Divergent carbon-isotopic compositions of kerogen and carbonate imply a restricted basinal environment (C. Calver, Macquarie University, pers. comm. 1995). Canyons up to 1 km deep were eroded into the shelf, implying a water

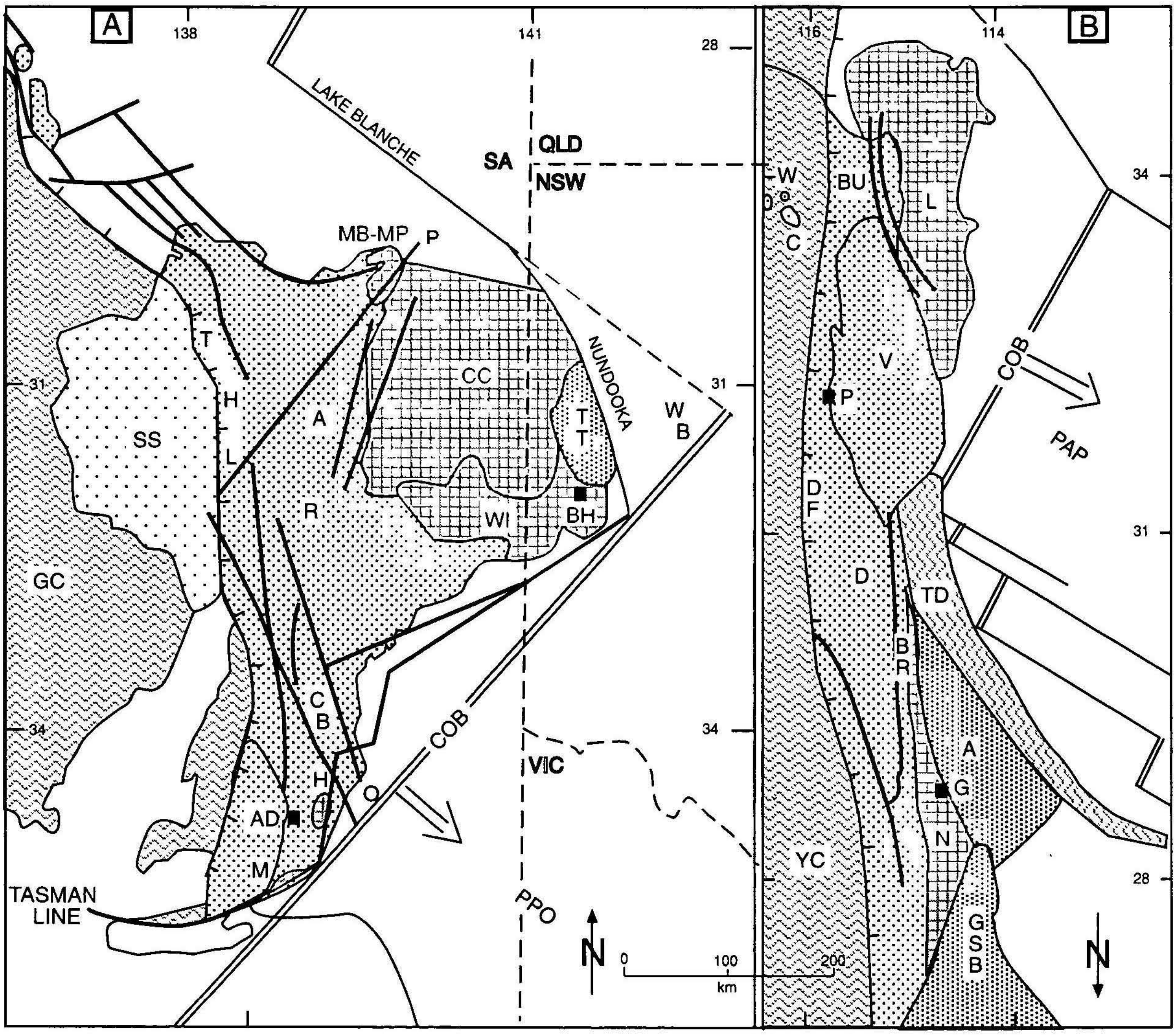


Figure 6. Plan view of Adelaide and Perth Rift Complexes at same scale, oriented such that the passive continental margin is on the right. (Documented below). Adelaide: AD = Adelaide BH = Broken Hill CB = Crystal Brook lineament CC = Curnamona Craton GC = Gawler Craton H = Houghton Inlier O = Oororoo Fault P = Paralana Fault PPO = Paleo-Pacific Ocean SS = Stuart Shelf THL = Torrens Hinge Line TT = Torrowangee Trough WB = Wonaminta Block WI = Willyama Inliers. Perth A = Abrolhos Sub-Basin BR = Beagle Ridge BU = Bunbury Trough C= Collie Basin D = Dandaragan Trough G = Geraldton GSB = Gascoyne Sub-Basin L = Leeuwin Complex N = Northampton Complex P = Perth PAP = Perth Abyssal Plain TD = Turtle Dove Ridge V = Vlaming SubBasin W = Wilga Basin YC = Yilgarn Craton.

A: Adelaide Rift Complex. **From Preiss (1987). Palinspastic view in the Cambrian, shortly after breakup, oriented with north up. Broad arrow indicates presumed direction of seafloor spreading and progradation of the Kanmantoo submarine fans; continent-ocean boundary (COB) (Tasman Line) restored to straight-line segments by eliminating the foldbelt arc.**

B: Perth Rift Complex. **From Harris et al. (1994).**

depth of at least that much; controversy continues as to whether the canyons were cut in a subaerial or a subaqueous environment (Christie-Blick et al. 1995). For the first time, the basin deepened to the north as well as to the south from a shallow zone in the central Flinders Ranges (Preiss 1987).

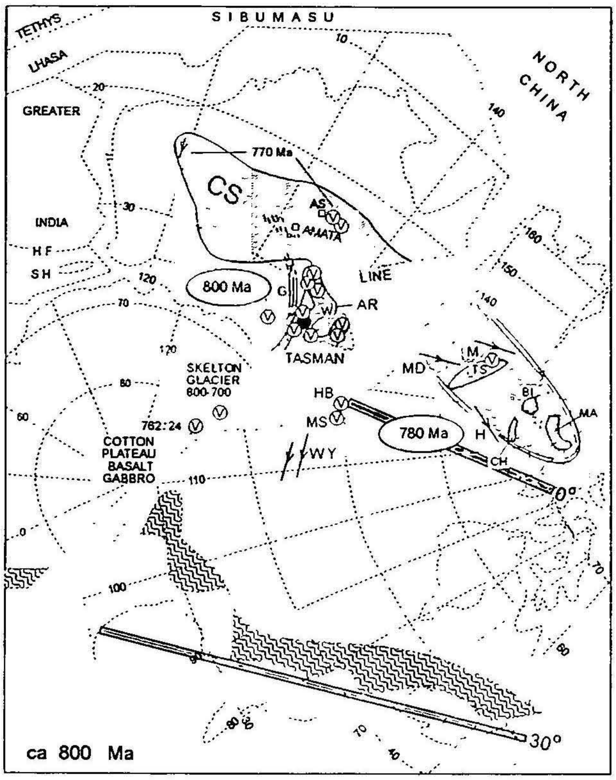


Figure 7. Australia–Antarctica and Laurentia reconstruction at 800 Ma. HF = Himalayan Front, SH = Shillong Hills.

Base as in the cartoons of Moores (1991), Borg & De Paulo (1994) and Park et al. (1995), but from the quantitative reconstruction, on a Mercator projection, by Powell et al. (1993). Greater India, Lhasa, Sibumasu from data in Veevers & Tewari (1995); North China from McKerrow et al. (1992, fig. 1). Dotted (Tasman) line on eastern and northern margins of Australia (present coordinates) marks limit of known Precambrian rocks; the light dotted line on the western side of Laurentia marks the limit of Proterozoic strata in the cordillera (Hoffman 1989).

Laurentia*: average trend of 780 Ma mafic dykes of the Outer Fold Belt of the Mackenzie Mountains (M) and the Wyoming Province (WY), and of the 780 Ma Hottah sheets (H), from Park et al. (1995, fig. 4). Outline (heavy broken line) of Mackenzie Mountains Supergroup in Mackenzie Mountains (M), and equivalents in the Brock Inlier (BI), Coppermine Homocline (CH), and Minto Arch (MA), from Narbonne and Aitken (1995, p. 104). Note mirror-image symmetry of the E–W trend of the Centralian Superbasin along the Amadeus Transverse Zone coming off the re-entrant of the Tasman Line and the SW–NE trend of the basins with the Mackenzie Mountains Supergroup, outlined by the double line, at a right-angle to the subsequent cordillera. The comparable 700 Ma outline (broken line) also contains the SW–NE trend. This is the zero-thickness line of Neoproterozoic and Early Cambrian (Stewart 1972), extended around Victoria Island (Kaufman & Knoll *1995, p. 42).

***Australia*: the ca 800 Ma Amata and Gairdner (G) mafic dyke swarms (Zhao et al. 1994, fig. 1), mafic volcanics, mainly flows, of the Adelaide Rift Complex, including the Wooltana Volcanics (W) (Preiss 1987, fig. W-3, p. 373), and shoreline of the overlying Supersequence 1A and B, which includes the Bitter Springs volcanics, east of Alice Springs (AS). According to Zhao et al. (1994, fig. 5), a postulated plume head (large black dot) could have resulted in domal uplift and subsidence (and presumably minor volcanism) of an area with radius of 4000 km, and Park et al. (1995) postulate that the Laurentian dykes and sheets (and the Amata and Gairdner dykes) may represent subswarms of a giant radiating dyke swarm centred on the Adelaide Rift Complex.**

Supersequence 4

Time-slice 4A

Most of the southern part of the Superbasin, including much of the Officer Basin and the central Amadeus Basin, may have been emergent (Fig. 3, time 4A). In the northern part of the Superbasin, time-slice 4A began with a deepening, during which possibly turbiditic sands were deposited locally in a basin fringed by fluvial and shallow marine deltaic complexes. Coarse sediment was again shed from the Musgrave Block. An extensive flood of basalt covered the western Officer Basin (Table Hill Volcanics) and northern Australia (Antrim Plateau Volcanics), and also the Mount Arrowsmith block (Scheibner 1993). However, the ages of both the Table Hill Volcanics and the Antrim Plateau Volcanics are very poorly known.

In the Amadeus Basin, the Arumbera Sandstone consists of red-brown and white sandstone and minor siltstone, shale, conglomerate and carbonate. The lower Arumbera Sandstone (units 1 and 2) contains *Charniodiscus* and other metazoan body fossils of the latest Proterozoic Ediacara fauna, but no definite trace fossils have been found (Walter et al. 1989). The formation was deposited in a shallow marine and deltaic or coastal plain setting (Lindsay 1987b), with sediment supply from the southwest carried by braided streams. The Amadeus Basin was modified as a result of the Petermann Ranges orogeny to form a shallow, east–west trending basin along the northern margin, separated from a shallower southern shelf area by a broad central ridge. Subsequent sedimentation was controlled by varying rates of subsidence in each tectonic subdivision. Deposition was largely progradational, as major deltaic complexes developed on the southern and southwestern margins of the sub-basins. The Mount Currie Conglomerate in the southern part of the basin is a presumed proximal correlative of the Arumbera Sandstone, though whether of the lower, Neoproterozoic part, or of the upper, Cambrian part, or both, is unknown. The same applies to the Sir Frederick Conglomerate and its correlative the Ellis Sandstone, in the western part of the basin (Wells et al. 1964).

The red-brown and green-grey sandstone and siltstone of the upper part of the Central Mount Stuart Formation of the Georgina Basin include soft-bodied metazoan fossils and trace fossils of the Ediacara fauna (Walter 1980), which indicate a marine environment. To the east, the sandstones of the uppermost Elkera Formation and to the west the predominantly fine- to coarse-grained, red-brown sandstone of the Yuendumu Sandstone of the Ngalia Basin may be this age

The siltstone, sandstone and conglomerate of the upper Louisa Downs Group of the Kimberley region probably correlate with Supersequence 4, but there is no definitive evidence as yet. Preliminary searches for fossils have not been successful.

In northern Australia the Antrim Plateau Volcanics extend over an area of 425 000 km^2, and comprise several hundred metres of tholeiitic flows with interbeds of nonmarine sandstone, chert, and limestone (Bultitude 1976, Cook 1988). They are conventionally regarded as being of Early Cambrian age, but the only direct evidence, stromatolites in interbedded cherts, indicates a latest Neoproterozoic age (Walter 1972, p. 49). The Table Hill Volcanics cover an area of 90 000 km^2 in the Officer Basin southwest of the Musgrave Block (Jackson & van de Graaff 1981, Cook 1988, GSWA 1990 p. 377). They are tenuously dated by Rb/Sr at 563 $\pm$ 40 Ma (recalculated from Compston 1974) or latest Neoproterozoic, probably the same age as the Antrim Plateau Volcanics.

The McFadden Formation of the Savory Basin, also possibly of this age, consists of fine to coarse-grained sandstone, feldspathic sandstone, minor pebble and granule conglomerate, and siltstone. Conglomerate occurs as thin interbeds or penecontemporaneous channel fills. Large cross-bed sets, up to 10 m thick, are typically flaggy and show grading. They may be point-bar deposits formed in large meandering channels in delta build-up areas, delta fronts, and migrating ridges or giant channel ripples (Williams 1992). Current lineation, ripple mark, and dewatering structures are also present. Grain-size coarsens northwards towards the basin margin. Palaeocurrent directions show a strong west-southwesterly directed flow. Sediment is immature and coarse-grained, suggesting that the source was probably local, most probably from the Paterson Orogen, with local input from the Oldham Inlier (Fig. 3, time 4A).

In the Adelaide Rift Complex the base of Supersequence 4 is likely to be at or near sequence boundary 10 of Christie-Blick et al. (1995), in the upper Wonoka Formation. The overlying Bonney Sandstone was deposited 'as a prograding coastal sandflat or tidal delta' and above this the Rawnsley Quartzite as a 'prograding shoreface, barrier and tidal channel facies complex' at times M-16 and M-17 (Preiss 1987). The Ediacara fauna of soft-bodied metazoans is preserved (in the Rawnsley Quartzite) in muddy sands, perhaps the deposits of submarine channels. There appear to be deeper basinal sands to the southeast and northeast, the latter being in part the Billy Springs Formation. We interpret this deepening as a third phase of onlap.

What seems to have been a eustatic fall in sea level occurred near the time of the Proterozoic–Cambrian boundary, exposing the continent (except that part of the Amadeus Basin with continuous deposition of the Arumbera Sandstone) to subaerial conditions. Fluvial and shallow marine deposition resumed in the Early Cambrian (Fig. 4) (Cook 1982, 1988).

Tectonics

The part of the Adelaide Rift Complex north of 32°S that outcrops in the Flinders Ranges is confined between the internal Gawler Craton on the west and the external Curnamona Craton on the east (Fig. 1). Preiss (1990, fig. 3) compared this part of the Adelaide Rift Complex with the failed arm of the Mesozoic–Cenozoic Bass Basin of southeastern Australia, confined by the internal mainland craton and the external Tasmanian craton; the Fleurieu-Nackara (Delamerian) arc on the southeast is matched by the passive margin ('successful arms') of offshore western Victoria and western Tasmania.

Our interpretation follows von der Borch's (1980) model of the Adelaide Rift Complex as a Neoproterozoic (800–544 Ma) intracratonic rift between the Gawler and Curnamona cratons transformed in the earliest Cambrian (544 Ma) to a failed arm (Flinders zone) by continental breakup along its southern part. Analysis of the subsidence history of the Neoproterozoic and Cambrian basins supports this interpretation (Lindsay et al. 1987). Sedimentation ceased before the Late Cambrian–Ordovician (500 Ma) Delamerian orogeny.

Powell et al. (1994, fig. 8) also view the Adelaide Rift Complex as an intracratonic rift basin but for a much shorter time interval, from 800 to 720 Ma, and thereafter as a passive continental margin. The need for an earlier breakup came from Powell et al.'s (1993) postulate from palaeomagnetic data that Laurentia and Australia began to separate after 720 Ma during the rapid growth of the Pacific Ocean such that East Gondwanaland and Laurentia were about 8000 km apart by 580 Ma. Accordingly, Powell et al. (1994) interpreted the onlap of the post-Sturtian glacial Tapley Hill Formation as reflecting breakup at 700 Ma, and the Adelaide Rift Complex as a passive continental margin from about 700 Ma to the 500 Ma Delamerian Orogeny. Continental separation was dated by the rapid divergence of the apparent polar wander paths of Australia and Laurentia from 720 Ma. But another interpretation of the palaeomagnetic evidence suggests the radically different possibility of Laurentia remaining fixed alongside Australia–Antarctica at least to 600 Ma, entailing a 100 m.y. later birth of the Pacific Ocean (Meert & van der Voo 1994, fig. 7). Birth of the Pacific Ocean towards the end of the Neoproterozoic would be consistent with there being no known oceanic facies older than Phanerozoic.

Furthermore, breakup in eastern Australia is indicated by the change in sediment provenance and the southeastward shift in the depocentre represented by the Cambrian Kanmantoo Group (von der Borch 1980, Flottmann et al. 1996, Foden 1996), and the appearance of Early Cambrian (525 Ma) mafic lava and felsic tuff (Shergold 1995) along the newly formed margin and of 530 Ma ophiolite in the palaeo-Pacific ocean floor (Aitchison et al. 1992). Breakup no older than latest Neoproterozoic is consistent with the geochemical evidence outlined above for restricted water circulation in the Adelaide Rift Complex during the deposition of the Tapley Hill Formation (about 700 Ma) and the lower and middle Wonoka Formation (about 570–580 Ma). It is notable that Preiss (1987) indicated a substantial change in palaeogeography in the Adelaide Rift Complex at the time of deposition of the Wonoka Formation. As noted above, the subsidence history of the Neoproterozoic and Cambrian basins also suggests breakup near the base of the Cambrian (Lindsay et al. 1987).

A comparative stratigraphic-tectonic study of Phanerozoic basins also bears on the age of continental breakup. We compare the Neoproterozoic Adelaide Rift Complex and Centralian Superbasin with the Palaeozoic–Mesozoic Perth Rift Complex and Canning and Amadeus Basins. The Canning and Amadeus Basins and much of the Centralian Superbasin occupy the Amadeus Transverse Zone (ATZ). The comparison is made in plan (Fig. 1), in time-space (Fig. 4), in section (Fig. 5), and, in detailed plan (Fig. 6) between the Adelaide Rift Complex and Perth Rift Complex. The ages of these structures are 400 m.y. apart, which is the period of the Pangean cycle (Veevers 1990).

Time-space

Notable comparisons (Fig. 4) are:

(a) supercontinental collisions of Gondwanaland and Laurussia at 320 Ma and of East and West Gondwanaland at 720 Ma;
(b) post-collision lacunas at 320–300 Ma and 720–700 Ma (assumed age);
(c) Extension I generating new basins for glaciogenic sediment at 293 Ma and 700 Ma (assumed age), followed by
(d) Extension II making new (coal) basins at 230 Ma, and new (glaciogenic) basins at 610 Ma (assumed age);
(e) continental breakup of northwest and west Australia at an observed 160 Ma and 130 Ma (west of Perth Rift Complex), and a postulated breakup at 575 Ma of northwest Australia and 544 Ma on the east (east of Adelaide Rift Complex), all accompanied by mafic volcanics. Major marine transgressions follow continental breakup, with a peak in the Aptian (115 Ma) and peaks in the Early Cambrian (520 Ma) and Early Ordovician (480 Ma); likewise, the subduction on the eastern margin changed from Chilean type to Mariana type at 90 Ma and 490 Ma.

Cross-section

In cross section (Fig. 5), the Adelaide Rift Complex mimics the salient features of the Perth Rift Complex (Fig. 5, ZY–XW and A'B') as do the contemporary basins occupying the Amadeus Transverse Zone. The only significant difference is the thick infilling of the Amadeus Basin by the Devonian Pertnjara Group, which arose from compression associated with the Alice Springs Orogeny, itself a distant effect of the Variscan collision of Gondwanaland and Laurussia.

Plan

The Neoproterozoic Adelaide Rift Complex was compared by

Preiss (1987, fig. 3) with aulacogens and geosynclines in Laurussia and with the Mesozoic continental margin and failed arm of the Bass Basin of southeastern Australia; by von der Borch (1980) with the southern margin of Australia; and by Eyles (1993) with the Gulf Coast of the USA. We believe a more cogent comparison is with the Permian and Mesozoic Perth Rift Complex (Fig. 6), also shown in time-space (Fig. 4) and in section (Fig. 5). Essential features, in bands of similar size and shape from craton to ocean, are as follows (see Appendix for explanation of abbreviations):

Adelaide	*Perth*	*Feature*
GC	YC	Archaean craton
SS	C, W	epicratonic sediment
THL	DF	hinge-line
AR	BU, V, D	inner basins of thick epicratonic sediment including glacials
CC, WI	L, BR, N	inner ridge of pre-basinal rocks
TT	A, GSB	outer basins
?WB	TD	outer ridge

Supercontinental connections

According to the SWEAT hypothesis (Moores 1991, Dalziel 1991), Neoproterozoic Australia (with India and Antarctica) was joined to Laurentia so that the Tasman Line faced the Canadian–Wyoming cordillera, and the Grenville Orogen of Laurentia continued into Antarctica.

Events common to Australia and Laurentia include:

a) ***ca 800 Ma*** (Fig. 7). According to Zhao et al. (1994, fig. 5), a postulated plume head (black dot in Fig. 7) could have resulted in domal uplift and subsidence (and presumably minor volcanism) of an area with radius of 4000 km, and Park et al. (1995) postulated that the Laurentian dykes and sheets, together with the Amata and Gairdner dykes, represent subswarms of a giant radiating dyke swarm centred on the Adelaide Rift Complex. The tholeiitic magma intruded pre-Neoproterozoic rocks except the early Neoproterozoic Mackenzie Mountains Supergroup and equivalents, which occupied a SW–NE-trending depocentre. Following the magmatism in Laurentia, sediment continued to accumulate in a narrower SW–NE-trending depression and newly accumulated along the cordilleran area. In Australia, sediment initially accumulated with the igneous rock in the Adelaide Rift Complex and then succeeded it, as sediment in the Centralian Superbasin rested over the Amata dyke swarm and was interlayered with the Bitter Springs volcanics. The configuration of S–N-trending Adelaide Rift Complex and E–W-trending Centralian Superbasin is a mirror image of the basins in Laurentia, such that the combined system forms a T.

b) ***ca 700 Ma.*** The 723 Ma Franklin igneous events in northern Canada and the emplacement of the 725 ± 35 Ma Cooee Dolerite in Tasmania denote extension of the continental crust soon after the ca 720 Ma collision of East and West Gondwanaland by the closing of the Mozambique ocean, possibly reflected in Tasmania by the 730 Ma Penguin Orogeny. Further extension trapped the copious glaciogenic deposits produced during the early Pangean icehouse. In Laurentia, glacial deposits stretched from 30°N to 65°N, and in Australia from the Adelaide Rift Complex to the Centralian Superbasin. This succession of events constitutes the first stage of a Pangean cycle, involving crustal extension driven by the first release of heat from the supercontinental insulator to provide accommodation space for the glaciogenic sediment produced by the icehouse climate (Veevers, 1990). This anticipates by some 400 m.y. the train of events starting with the 320 Ma collision of Gondwanaland and Laurussia, followed by 300 Ma extension and fill of Gondwanan glaciogenic sediment. Powell et al. (1994) and Young (1995) interpreted the extension that produced the rifts with glacial sediments as reflecting supercontinental breakup; instead, we believe it reflects supercontinental crustal extension. Incidentally, Eyles (1993) also interpreted the Sturtian glaciogenic strata as resulting from the break-out of Laurentia from the Neoproterozoic supercontinent, as well as deposits preserved within mobile belts along compressional plate margins.

c) ***ca 600 Ma.*** A second extensional event (cf. 230 Ma Pangean extension by rifting) provided accommodation space for a second set of glaciogenic sediments, the Marinoan Elatina Formation and Pepuarta Tillite in Australia and Ice Brook and Mount Vreeland formations in Laurentia. The Marinoan saw the sea cross the Curnamona Craton, as shown in Fig. 5, column at east end of WX (Preiss 1987, p. 393–399). The corresponding facies in what had become the Mesozoic greenhouse climate was coal measures.

d) ***600–565 Ma.*** Mafic volcanism in Australia, including the voluminous Antrim Plateau (tholeiitic) Volcanics of the north and the Table Hill Volcanics of the centre and dykes in the southwest and Tasmania, accompanied the right-lateral shearing (Petermann Ranges Orogeny) that was transformed to seafloor spreading at the end of the Neoproterozoic. A 580 Ma trachyte in Utah (Christie-Blick & Levy 1989) is the only known possible equivalent in Laurentia.

e) ***ca 544 Ma = Neoproterozoic–Cambrian boundary.*** Breakup of Laurentia from Australia–Antarctica is interpreted in the cordilleran area of Laurentia from the pattern of subsidence (Bond et al. 1984, Kominz 1995) and in the Mackenzie Mountains from a spectacular regional angular unconformity in the latest Neoproterozoic (Narbonne & Aitken 1995). In Mexico, latest Neoproterozoic basalt in an otherwise non-igneous succession (Stewart et al. 1984, McMenamin & McMenamin 1990) may indicate breakup. The evidence of breakup in Australia is given above.

Conclusions

Palaeogeography

After a hiatus of some 200 million years, the Australian Neoproterozoic record started with ca 800 Ma mafic magmatism from a plume-head near Adelaide, which produced mafic volcanics interbedded with carbonates and evaporites in the Callanna Group of the Adelaide Rift Complex and dyke swarms to the northwest in the Gairdner and Amata areas. Correlatives of these rocks are not known elsewhere in Australia. This rift-volcanic event was followed by widespread subsidence and accumulation of Supersequence 1. The first sediments, mixed fluvial and shallow marine, were followed by interbedded stromatolitic carbonates and evaporites, including halite and anhydrite, deposited in peritidal to very shallow marine settings, and locally accompanied by mafic volcanism.

At about 700 Ma, Supersequence 2 started with the Sturtian glaciation, followed by thick widespread post-glacial silt and mud in an epeiric sea that shallowed up to peritidal carbonate and sand. Supersequence 3 started at about 610 Ma with the second (Marinoan) glaciation, with environments as for Supersequence 2. Diamictites were replaced by arkose, conglomerate, and arkosic sand, interpreted as glacial outwash deposits. A major eustatic rise in sea level at about 590 Ma followed deglaciation, with deposition of thick silt and shale in an epeiric sea. The Musgrave Block emerged from the middle of the Centralian Superbasin and shed coarse sediment to the north and south. The supersequence shallows up to peritidal carbonates, including extensive ooid and intraclast shoals, and evaporites in the southern Georgina Basin.

Supersequence 4 (about 580–544 Ma) began with a flooding event, during which turbiditic sands were deposited locally and coarse sediment was again shed from the Musgrave Block, bypassing the emergent southern part of the Superbasin and the central Amadeus Basin to form fluvial and shallow marine deltaic complexes in the north. Extensive tholeiitic lava covered the western Officer Basin and northern Australia (though the ages of these units are poorly known). What seems to have been a eustatic fall in sea level near the Proterozoic–Cambrian boundary exposed most of the Superbasin to subaerial conditions until fluvial and shallow marine deltaic conditions were re-established in the Early Cambrian.

Tectonics

The Adelaide Rift Complex is a Neoproterozoic intracratonic rift between the Gawler and Curnamona cratons, and the Centralian Superbasin an associated epicratonic sag elongated east–west at a high angle to the rift. The combined structures compare closely with the Palaeozoic–Mesozoic Perth Rift Complex and Canning and Amadeus Basins. During the Ediacarian, right-lateral shearing (Petermann Ranges Orogeny) caused thrusting and the emergence of the Musgrave Block, oriented east–west through the middle of the Superbasin.

Continental breakup in the northwest is indicated by the eruption of flood basalt during the Ediacarian, and at the Proterozoic–Cambrian transition in the southeast by a shift in provenance of the sediments, and volcanism on the newly formed margin. At this latter time the Flinders zone of the Adelaide Rift Complex was transformed to a failed arm. Sedimentation ceased before the Late Cambrian–Ordovician (500 Ma) Delamerian Orogeny.

According to the SWEAT hypothesis, Neoproterozoic Australia (with India and Antarctica) was joined to Laurentia, so that the Tasman Line faced the Canadian–Wyoming cordillera. The configuration of the north–south-trending Adelaide Rift Complex and the east–west-trending Centralian Superbasin is mirrored in the basins in Laurentia to form a T, which split at about 544 Ma.

Acknowledgments

This is Publication 55 of the Key Centre for the Geochemical Evolution and Metallogeny of Continents. Our work has been supported by the Australian Research Council and Macquarie University and, at various times, by provision of core material from the Geological Surveys of South Australia, Western Australia, Northern Territory, and the Australian Geological Survey Organisation and later the Bureau of Resource Sciences. The Geological Survey of the Northern Territory and the Geological Survey of Western Australia have helped with field work logistics. Our colleagues Clive Calver and Kath Grey contributed through many discussions; their detailed work will be published elsewhere. Reviews of the manuscript by Marita Bradshaw, Kath Grey, John Lindsay and Wolfgang Preiss are appreciated.

References

Aitchison, J.C., Ireland, T.R., Blake, M.C.Jr & Flood, P.G., 1992. 530 Ma zircon age for ophiolite from the New England Orogen: oldest rocks known from eastern Australia. Geology, 20, 125–128.

Bagas, L., Grey, K. & Williams, I.R., 1996. Reappraisal of the Paterson Orogen and Savory Basin. Geological Survey of Western Australia Annual Review for 1995, 55–63.

Barber, P.M., 1988. The Exmouth Plateau deep water formation: a case history. In: Purcell, P.G. & Purcell, R.R. (editors), The North West Shelf, Australia. Proceedings of the Petroleum Exploration Society of Australia Symposium, Perth, 173–187.

Blake, D.H., Hodgson, I.M. & Muhling, P.C., 1979. Geology of the Granites–Tanami region. Bureau of Mineral Resources, Australia, Bulletin 197.

BMR Palaeogeographic Group, 1990. Australia, evolution of a continent. Bureau of Mineral Resources, Australia, 97 pp.

Bond, G.C., Nickeson, P.A. & Kominz, M.A., 1984. Breakup of a supercontinent between 625 Ma and 555 Ma: new evidence and implications for continental histories. Earth and Planetary Science Letters, 70, 325–345.

Borg, S.G. & De Paulo, D.J., 1994. Laurentia, Australia, and Antarctica as a Late Proterozoic supercontinent: constraints from isotopic mapping. Geology, 22, 307–310.

Bradshaw, M.T., Bradshaw, J., Murray, A.P., Needham, D.J., Spencer, L., Summons, R.E., Wilmot, J. & Winn, S., 1994. Petroleum systems in west Australian basins. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 93–118.

Brasier, M., Cowie, J. & Taylor, M., 1994. Decision on the Precambrian–Cambrian boundary stratotype. Episodes, 17, 3–8.

Brewer, A.M., Dunster, J.N., Gatehouse, C.G., Henry, R.L. & Weste, G., 1987. A revision of the stratigraphy of the eastern Officer Basin. South Australia, Geological Survey, Quarterly Geological Notes, 102.

Bultitude, R.J., 1976. Flood basalts of probable Early Cambrian age in northern Australia. In: Johnson, R.W. (editor), Volcanism in Australia. Elsevier, Amsterdam, 1–20.

Chen, Y.D., O'Reilly, S.Y., Kinny, P.D. & Griffin, W.L., 1994. Dating lower crust and upper mantle events: an ion microprobe study of xenoliths from kimberlitic pipes, South Australia. Lithos, 32, 77–94.

Chewings, C., 1914. Notes on the stratigraphy of Central Australia. Transactions of the Royal Society of South Australia, 38, 41–52.

Christie-Blick, N. & Levy, M. (editors), 1989. Late Proterozoic and Cambrian tectonics, sedimentation, and record of metazoan radiation in the western United States. American Geophysical Union, Washington DC, 28th International Geological Congress, Field Trip Guidebook, T331, 1–113.

Christie-Blick, N., Dyson, I.A. & von der Borch, C.C., 1995. Sequence stratigraphy and the interpretation of Neoproterozoic earth history. Precambrian Research, 73, 3–26.

Clarke, D., 1976. Heavitree Quartzite. In: Wells, A.T. (editor), Geology of the Late Proterozoic–Palaeozoic Amadeus Basin. 25th International Geological Congress, Excursion Guide, 48A, 26–28.

Coats, R.P. & Preiss, W.V., 1980. Stratigraphic and geochronological reinterpretation of Neoproterozoic glaciogenic sequences in the Kimberley region, Western Australia. Precambrian Research, 13, 181–208.

Compston, W., 1974. The Table Hill Volcanics of the Officer Basin: Precambrian or Palaeozoic? Geological Society of Australia, Journal, 21, 403–412.

Compston, W., Williams, I.S., Jenkins, R.J.F., Gostin, V.A. & Haines, P.W., 1987. Zircon age evidence for the Late Precambrian Acraman ejecta blanket. Australian Journal of Earth Sciences, 34. 435–445.

Cook, P.J., 1982. The Cambrian palaeogeography of Australia and opportunities for petroleum exploration. APEA Journal, 22(1), 42–64.

Cook, P.J., 1988. Palaeogeographic atlas of Australia. Volume 1. Cambrian. Bureau of Mineral Resources, Canberra.

Corkeron, M., Grey, K., Li, Z.X. & Powell, C.McA., 1996. Neoproterozoic glacial episodes in the Kimberley Region, northwestern Australia. Geological Society of Australia, Abstracts 41, 13th Australian Geological Convention, Canberra, 97.

Dalziel, I.W.D., 1991. Pacific margins of Laurentia and East Antarctica-Australia as a conjugate rift pair: evidence and implications for an Eocambrian supercontinent. Geology, 19, 598-601.

Dow, D.B. & Gemuts, I., 1969. Geology of the Kimberley Region, West Australia, The East Kimberley. Western Australia Geological Survey, Bulletin 120.

Edgoose, C.J., 1986. First report of Late Proterozoic glacigene sediments and striated pavements, Litchfield Province, Northern Territory, Australia. Abstracts, 12th International Sedimentological Congress, Canberra 1986, 90–91.

Embleton, B.J.J. & Williams, G.E., 1986. Low palaeolatitude of deposition for Late Precambrian periglacial varvites from South Australia: implications for palaeoclimatology. Earth and Planetary Science Letters, 79, 419–430.

Eyles, N., 1993. Earth's glacial record and its tectonic setting. Earth-Science Reviews, 35, 1–248.

Field, B.D., 1991. Paralic and periglacial facies and contemporaneous deformation of the Late Proterozoic Olympic Formation, Pioneer Sandstone and Gaylad Sandstone, Amadeus Basin, central Australia. In: Korsch, R.J. & Kennard, J.M. (editors), Geological and geophysical studies in the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 236, 127–136.

Flint, R.B., Fanning, C.M. & Rankin, L.R., 1988. the Late Proterozoic Kilroo Formation of the Polda Basin. South Australia Geological Survey Quarterly Geological Notes, 106, 16–23.

Flottmann, T., Haines, P., James, P. & Belperio, A.P., 1996. The tectonic setting and internal structure of the Cambrian Kanmantoo Basin, southeast Australia. Geological Society of Australia, Abstracts 41, 13th Australian Geological Convention, Canberra, 144.

Foden, J., 1996. Provenance of Neoproterozoic and early Palaeozoic sediments east Australia: implications from Nd isotope and zircon studies. Geological Society of Australia, Abstracts 41, 13th Australian Geological Convention, Canberra, 146.

Forman, D.J. & Wales, D.W., 1981. Geological evolution of the Canning Basin, Western Australia. Bureau of Mineral Resources, Australia, Bulletin 210, 91 pp., 36 plates.

Foster, C.B. & Waterhouse, J.B., 1988. The *Granulatosporites confluens* Oppel-zone and Early Permian marine faunas from the Grant Formation on the Barbwire Terrace, Canning Basin, Western Australia. Australian Journal of Earth Sciences, 35, 135–157.

Glaessner, M.F., 1984. The dawn of animal life: a biohistorical study. Cambridge University Press, Cambridge, 244 pp.

Gorter, J.D., Nicoll, R.S. & Foster, C.B., 1994. Lower Palaeozoic facies in the Carnarvon Basin, Western Australia: Stratigraphy and hydrocarbon prospectivity. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 373-396.

Grey, K., 1993. Acritarch biostratigraphy of the Ediacarian of the Centralian Superbasin. In: Jenkins, R.J.F., Lindsay, J.F. & Walter, M.R. (editors), Field guide to the Adelaide Geosyncline and Amadeus Basin, Australia. Australian Geological Survey Organisation, Record 1993/35, 133 pp.

Grey, K., 1995. Neoproterozoic stromatolites from the Skates Hill Formation, Savory Basin, Western Australia, and a review of the distribution of *Acaciella australica*. Australian Journal of Earth Sciences, 42, 123–132.

Grotzinger, J.P., Bowring, S.A., Saylor, B.Z. & Kaufman, A.J., 1995. Biostratigraphic and geochronologic constraints on early animal evolution. Science, 270, 598–604.

GSWA, 1990, Geology and mineral resources of Western Australia. Geological Survey Western Australia Memoir 3, 827 pp.

Harris, L.B., Higgins, R.I., Dentith, M.C. & Middleton, M.F., 1994. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 801–809.

Hayes, J.M., Lambert, I.B. & Strauss, H,. 1992. The sulfur-isotopic record. In: Schopf, J.W. & Klein, C. (editors), The Proterozoic biosphere: a multidisciplinary study. Cambridge University Press, 129–132.

Hoffman, P.F., 1989. Precambrian geology and tectonic history of North America. In: Bally, A.W. & Palmer, A.R. (editors), The geology of North America: an overview. Boulder, Colorado, Geological Society of America, The Geology of North America, v. A, 447–512.

Jackson, M.J. & Muir, M.D., 1981. The Babbagoola Beds, Officer Basin, Western Australia: correlations, micropaleontology and implications for petroleum prospectivity. BMR Journal of Australian Geology & Geophysics, 6, 81–93.

Jackson, M.J. & van de Graaff, W.J.E., 1981. Geology of the Officer Basin, Western Australia. Bureau of Mineral Resources, Australia, Bulletin 206, 102 pp.

Jenkins, R.J.F., 1989. The Adelaide Fold Belt: Tectonic reappraisal. In: Jago, J.B. & Moore, P.S. (editors), The evolution of a late Precambrian–early Palaeozoic rift complex: the Adelaide Geosyncline. Geological Society of Australia Special Publication 16, 396–420.

Jenkins, R.J.F., 1995. The problems and potential of using animal fossils and trace fossils in Terminal Proterozoic biostratigraphy. Precambrian Research, 73, 51–69.

Jenkins, R.J.F., Lindsay, J.F. & Walter, M.R., 1993. Field guide to the Adelaide Geosyncline and Amadeus Basin, Australia. Australian Geological Survey Organisation, Record 1993/35, 133 pp.

Jenkins, R.J.F., McKirdy, D.M., Foster, C.B., O'Leary, T. & Pell, S.D., 1992. The record and stratigraphic implications of organic-walled microfossils from the Ediacaran (terminal Proterozoic) of South Australia. Geological Magazine, 129, 401–410.

Kaufman, A.J. & Knoll, A.H., 1995. Neoproterozoic variations in the C-isotopic composition of seawater: stratigraphic and biogeochemical implications. Precambrian Research, 73, 27–49.

Kennard, J.M. & Lindsay, J.F., 1991. Sequence stratigraphy of the latest Proterozoic. Cambrian Pertaoorrta Group, northern Amadeus Basin, central Australia. In: Korsch, R.J. & Kennard, J.M. (editors), Geological and geophysical studies in the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 236.

Kennard, J.M., Jackson, M.J., Romine, K.K., Shaw, R.D. & Southgate, P.N., 1994. Depositional sequences and associated petroleum systems of the Canning Basin, W.A. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 657–676.

Kennard, J.M., Nicoll, R.S. & Owen, M. (editors), 1986. Late Proterozoic and Early Palaeozoic depositional facies of the northern Amadeus Basin, central Australia. 12th International Sedimentological Congress, Field Excursion 25B. Bureau of Mineral Resources, Canberra, 125 pp.

Klootwijk, C., 1995. Palaeomagnetism suggests mid-Carboniferous convergence between Greater Australia and Altaids. AGSO Research Newsletter, 22, 14–17.

Knoll, A.H. & Walter, M.R., 1992. Latest Proterozoic stratigraphy and earth history. Nature, 356, 673–678.

Knoll, A.H., Grotzinger, J.P., Kaufman, A.J. & Kolosov, P., 1995. Integrated approaches to terminal Proterozoic stratigraphy: an example from the Olenek Uplift, northeastern Siberia. Precambrian Research, 73, 251–270.

Knoll, A.H., Hayes, J.M., Kaufman, A.J., Swett, K. & Lambert, I.B., 1986. Secular variation in carbon isotope ratios from Upper Proterozoic successions of Svalbard and East Greenland. Nature, 321, 832–838.

Kominz, M., 1995. Thermally subsiding basins and the insulating effect of sediment with application to the Cambro-Ordovician Great Basin sequence, western USA. Basin Research, 7, 221–233.

Lambert, I.B., Knutson, J., Donnelly, T.H., Etminan, H. & Mason, M.G., 1984. Genesis of copper mineralisation, Myall Creek Prospect, South Australia. Mineralium Deposita, 19, 266–273.

Lindsay, J.F., 1987a. Upper Proterozoic evaporites in the Amadeus basin, central Australia, and their role in basin tectonics. Geological Society of America Bulletin, 99, 852–865.

Lindsay, J.F., 1987b. Sequence stratigraphy and depositional controls in Late Proterozoic–early Cambrian sediments of Amadeus Basin, central Australia. American Association of Petroleum Geologists Bulletin, 71, 1387–1403.

Lindsay, J.F., 1989. Depositional controls on glacial facies associations in a basinal setting, Late Proterozoic, Amadeus Basin, central Australia. Palaeogeography, Palaeoclimatology, Palaeoecology, 73, 205–232.

Lindsay, J.F., 1991. New evidence for ancient metazoan life in the Late Proterozoic Heavitree Quartzite, Amadeus Basin, central Australia. In: Korsch, R.J. & Kennard, J.M. (editors), Geological and geophysical studies in the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 236, 91–95.

Lindsay, J.F., Korsch, R.J. & Wilford, J.R., 1987. Timing the breakup of a Proterozoic supercontinent: evidence from Australian intracratonic basins. Geology, 15, 1061–1064.

Maboko, M.A.H., Boelrijk, N.A.I.M., Priem, H.N.A. & Verdurmen, E.A.T., 1985. Zircon U-Pb and biotite Rb-Sr dating of the Wami River granulites, Eastern Granulites, Tanzania: evidence for approximately 715 Ma old granulite-facies metamorphism and final Pan-African cooling approximately 475 Ma ago. Precambrian Research, 30, 361–378.

Maboko, M.A.H., McDougall, I., Zeitler, P.K. & Williams, I.S., 1992. Geochronological evidence for ~530–550 Ma juxtaposition of two Proterozoic metamorphic terranes in the Musgrave Ranges, central Australia. Australian Journal of Earth Sciences, 39, 457–471.

Mawson, D., 1925. Evidence and indications of algal contributions in the Cambrian and Pre-cambrian limestones of South Australia. Transactions of the Royal Society of South Australia, 49, 186–190.

Mawson, D., 1949. The Elatina glaciation. Transactions of the Royal Society of South Australia, 73, 117–121.

McKerrow, W.S., Scotese, C.R., & Brasier, M.D., 1992. Early Cambrian continental reconstructions. Journal of the Geological Society of London, 149, 599–606.

McMenamin, M.A.S., & McMenamin, D.L.S., 1990. The emergence of animals, the Cambrian breakthrough. New York, Columbia University Press, 217 pp.

Meert, J.G. & van der Voo, R. 1994. The Neoproterozoic (1000–540 Ma) glacial intervals: no more snowball earth? Earth and Planetary Science Letters, 123, 1–13.

Moores, E.M., 1991. Southwest U.S.-East Antarctic (SWEAT) connection: a hypothesis. Geology, 19, 425–428.

Murray, C.G., Schiebner, E. & Walker, R.N., 1989. Regional geological intrepretation of a digital coloured residual Bouguer gravity image of eastern Australia with a wavelength cut-off of 250 km. Australian Journal of Earth Sciences, 36, 423–449.

Narbonne, G. & Aitken, J.D., 1995. Neoproterozoic of the Mackenzie Mountains, northwestern Canada. Precambrian Research, 73, 101–121.

Oaks, R.Q.Jr., Deckelman, J.A., Conrad, K.T., Hamp, L.T., Phillips, J.O. & Stewart, A.J., 1991. Sedimentation and tectonics in the northeastern and central Amadeus Basin, central Australia. In: Korsch, R.J. & Kennard, J.M. (editors), Geological and geophysical studies in the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 236, 73–90.

Park, J.K., Buchan, K.L. & Harlan, S.S., 1995. A proposed giant radiating dyke swarm fragmented by the separation of Laurentia and Australia based on paleomagnetism of ca.780 Ma mafic intrusions in western North America. Earth and Planetary Science Letters, 132, 129–139.

Phillips, B.J., James, A.W. & Philip, G.M., 1985. The geology and hydrocarbon potential of the north-western Officer Basin. APEA Journal, 25, 52–61.

Plumb, K.A., 1996. Revised correlation of Neoproterozoic glacial successions from the Kimberley region, northwestern Australia. Geological Society of Australia, Abstracts 41, 13th Australian Geological Convention, Canberra, 344.

Powell, C.McA., 1984. Uluru regime. In: Veevers, J.J. (editor), Phanerozoic earth history of Australia. Oxford, Clarendon, 290–340.

Powell, C.McA., Li, Z.X., McElhinny, M.W., Meert, J.G. & Park, J.K., 1993. Paleomagnetic constraints on timing of the Neoproterozoic formation of Gondwana. Geology, 21, 889–892.

Powell, C.McA., Preiss, W.V., Gatehouse, C.G., Krapez, B. & Li, Z.X., 1994. South Australian record of a Rodinian epicontinental basin and its mid-Neoproterozoic breakup (~700 Ma) to form the Palaeo-Pacific Ocean. Tectonophysics, 237, 113140.

Preiss, W.V. (compiler), 1986. Adelaide Geosyncline and Stuart Shelf: Precambrian and Palaeozoic geology (with special reference to the Adelaidean). 1:600 000 scale map, second edition. South Australia, Department of Mines and Energy, Adelaide.

Preiss, W.V., 1987. The Adelaide Geosyncline — Neoproterozoic stratigraphy, sedimentation, palaeontology and tectonics. Bulletin of the Geological Survey of South Australia, 53, 438 pp.

Preiss, W.V., 1990. A stratigraphic and tectonic overview of the Adelaide Geosyncline, South Australia. In: Jago, J.B. & Moore, P.S. (editors), The evolution of a late Precambrian–early Palaeozoic rift complex: the Adelaide Geosyncline. Geological Society of Australia, Special Publication 16, 1–33.

Preiss, W.V., 1993. Neoproterozoic. In: Drexel, J F., Preiss, W.V. & Parker, A.J. (editors), The geology of South Australia, V. 1. The Precambrian geological survey of South Australia, Bulletin 54, 171–203.

Preiss, W.V., Walter, M.R., Coates, R.P. & Wells, A.T., 1978. Lithological correlations of Adelaidean glaciogenic rocks in parts of the Amadeus, Ngalia, and Georgina Basins. BMR Journal of Australian Geology

& Geophysics, 3, 45–53.

Redfern, J. & Millward, E., 1994. A review of the sedimentology and stratigraphy of the Permo–Carboniferous Grant Group, Canning Basin, Western Australia. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 753–756.

Scheibner, E., 1993. Structural framework of New South Wales. Geological Survey of New South Wales Quarterly Notes, 93, 1–36.

Schmidt, P.W. & Williams, G.E., 1995. The Neoproterozoic climatic paradox: equatorial palaeolatitude for Marinoan glaciation near sea level in South Australia. Earth and Planetary Science Letters, 134, 107–124.

Sengor, A.M.C., Natalin, B.A. & Burtman, V.S., 1993. Evolution of the Altaid tectonic collage and Palaeozoic crustal growth in Eurasia. Nature, 364, 299–307.

Senior, B.R., Truswell, E.M., Idnurm, M., Shaw, R.D. & Warren, R.G., 1995. Cainozoic sedimentary basins in the eastern Arunta Block, Alice Springs region, central Australia. AGSO Journal of Australian Geology & Geophysics, 15, 421–444.

Shaw, R.D. & Warren, R.G., 1975. Alcoota, N.T., 1:250,000 Geological Series sheet SF/53-10. Explanatory notes. Bureau of Mineral Resources, Australia.

Shaw, R.D. & Wells, A.T., 1983. Alice Springs (Second Edition) Northern Territory, 1:250,000 Geological Series — Explanatory notes. Bureau of Mineral Resources, Australia.

Shergold, J.H., 1995. Timescales. 1. Cambrian. Australian Geological Survey Organisation, Record 1995/30.

Southgate, P.N., 1986. Depositional environment and mechanism of preservation of microfossils, upper Proterozoic Bitter Springs Formation, Australia. Geology, 14, 683–686.

Southgate, P.N., 1989. Relationship between cyclicity and stromatolite form in the Neoproterozoic Bitter Springs Formation, Australia. Sedimentology, 36, 323–339.

Southgate, P.N., 1991. A sedimentological model for the Loves Creek Member of the Bitter Springs Formation, northern Amadeus Basin. In: Korsch, R.J. & Kennard, J.M. (editors), Geological and geophysical studies in the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 236, 113–126.

Sprigg, R.C., 1947. Early Cambrian (?) jellyfishes from the Flinders Ranges, South Australia. Transactions of the Royal Society of South Australia, 71, 212–224.

Stern, R.J., 1994. Arc assembly and continental collision in the Neoproterozoic East African Orogen. Annual Review of Earth and Planetary Sciences, 22, 319–351.

Stewart, A.J., 1979. A barred basin marine evaporite in the Upper Proterozoic of the Amadeus Basin, central Australia. Sedimentology, 26, 33–62.

Stewart, J.H., 1972. Initial deposits in the Cordilleran geosyncline: evidence of a late Precambrian (R m.y.) continental separation. Geological Society of America Bulletin, 83, 1345-1360.

Stewart, J.H., McMenamin, A.S. & Morales-Ramirez, J.M., 1984. Upper Proterozoic and Cambrian rocks in the Caborca region, Sonora, Mexico: physical stratigraphy, biostratigraphy, paleocurrent studies, and regional relations. U.S. Geological Survey Professional Paper 1309, 36 pp.

Sukanta, U., Thomas, B., von der Borch, C. & Gatehouse, C.G., 1991. Sequence stratigraphic studies and canyon formation, South Australia. PESA Journal, 19, 68–73.

Townson, W.G., 1985. The subsurface geology of the western Officer Basin — results of Shell's 1980–1984 petroleum exploration campaign. APEA Journal, 25, 34–51.

Veevers, J. J. (editor), 1984. Phanerozoic Earth history of Australia. Oxford, Clarendon, 418 pp.

Veevers, J.J., 1988. Morphotectonics of Australia's northwestern margin — a review. In: Purcell, P.G. & Purcell, R.R. (editors), The North West Shelf, Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 19–27.

Veevers, J.J., 1990. Tectonic-climatic supercycle in the billion-year plate-tectonic eon: Permian Pangean icehouse alternates with Cretaceous dispersed-continents greenhouse. Sedimentary Geology, 68, 1–16.

Veevers, J.J., 1991. Mid-Cretaceous tectonic climax, Late Cretaceous recovery, and Cenozoic relaxation in the Australian region. Geological Society of Australia, Special Publication 18, 1–14.

Veevers, J.J., 1995. Emergent long-lived Gondwanaland vs submergent short-lived Laurasia: Supercontinental and Pan-African heat imparts long-term buoyancy by mafic underplating. Geology, 2, 1131–1134.

Veevers, J.J. & Cotterill, D., 1978. Western margin of Australia: evolution of a rifted arch system. Geological Society of America Bulletin, 89, 337–355.

Veevers, J.J. & Hansen, L., 1981. Volcanism in the rift-valley system that evolved into the western margin of Australia. Journal of the Geological Society of Australia, 28, 377–384.

Veevers, J.J. & Tewari, R.C., 1995. Permian–Carboniferous and Permian–Triassic magmatism in the rift zone bordering the Tethyan margin of southern Pangea. Geology, 23, 467–470.

Veevers, J.J., Clare, A. & Wopfner, H., 1994. Neocratonic magmatic–sedimentary basins of post-Variscan Europe and post-Kanimblan eastern Australia generated by right-lateral transtension of Permo-Carboniferous Pangea. Basin Research, 6, 141–157.

Veevers, J.J., Walter, M.R. & Scheibner, E, 1997. Neoproterozoic tectonics of Australia-Antarctica and Laurentia and the 560 Ma birth of the Pacific Ocean reflect the 400 m.y. Pangean supercycle. Journal of Geology, 105, 225–242.

Vidal, G. & Knoll, A.H., 1983. Proterozoic plankton. Geological Society of America Memoir, 161, 265–277.

von der Borch, C.C., 1980. Evolution of late Proterozoic to early Palaeozoic Adelaide Foldbelt, Australia: comparisons with post-Permian rifts and passive margins. Tectonophysics, 70, 115–134.

Wallace, M.W., Gostin, V.A. & Keays, R.R., 1989. Discovery of the Acraman impact ejecta blanket in the Officer Basin and its stratigraphic significance. Australian Journal of Earth Sciences, 36, 585–587.

Walter, M.R., 1972. Stromatolites and the biostratigraphy of the Australian Precambrian and Cambrian. Special Papers in Palaeontology, 11, Palaeontological Association London, 190 pp.

Walter, M.R., 1980. Adelaidean and Early Cambrian stratigraphy of the southwestern Georgina Basin: correlation chart and explanatory notes. Bureau of Mineral Resources, Australia, Report 214, BMR Microform MF92.

Walter, M.R. & Gorter, J.D., 1994. The Neoproterozoic Centralian Superbasin in Western Australia: the Savory and Officer Basins. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Proceedings of Petroleum Exploration Society of Australia Symposium, Perth, 851–864.

Walter, M.R., Elphinstone, R. & Heys, G.R., 1989. Proterozoic and Early Cambrian trace fossils from the Amadeus and Georgina Basins, central Australia. Alcheringa, 13, 209–256.

Walter, M.R., Grey, K., Williams, I. & Calver, C.R., 1994. Stratigraphy of the Neoproterozoic to early Palaeozoic Savory Basin, Western Australia, and correlation with the Amadeus and Officer Basins. Australian Journal of Earth Sciences, 41, 533–546.
Walter, M.R., Veevers, J.J., Calver, C.R. & Grey, K., 1995. Neoproterozoic stratigraphy of the Centralian Superbasin, Australia. Precambrian Research, 73, 173–195.
Wells, A.T., 1980. Evaporites in Australia. Bureau of Mineral Resources, Australia, Bulletin 198, 104 pp.
Wells, A.T. & Moss, J.F., 1983. The Ngalia Basin, Northern Territory: stratigraphy and structure. Bureau of Mineral Resources, Australia, Bulletin 212, 88 pp.
Wells, A.T., Forman, D.J. & Ranford, L.C., 1964. Geological reconnaissance of the Rawlinson–MacDonald 1:250 000 sheet areas, Western Australia. Bureau of Mineral Resources, Australia, Report 65, 35 pp.
Wells, A.T., Forman, D.J., Ranford, L.C. & Cook, P.J., 1970. Geology of the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 100, 222 pp.
Wells, A.T., Ranford, L.C., Stewart, A.J., Cook, P.J. & Shaw, R.D., 1967. The geology of the northeastern part of the Amadeus Basin, Northern Territory. Bureau of Mineral Resources, Australia, Bulletin 113, 97 pp.
Wells, A.T., Stewart, A.J. & Skwarko, S.K., 1966. Geology of the south-eastern part of the Amadeus Basin, Northern Territory. Bureau of Mineral Resources, Australia, Report 88, 59 pp.
Williams, G.E., 1983. Precambrian varves and sunspot cycles. In: McCormac, B.M. (editor), Weather and climate responses to solar variations. Colorado Associated University Press, Boulder, 517–533.
Williams, G.E. & Tonkin, D.G., 1985. Periglacial structures and paleoclimatic significance of a late Precambrian block field in the Cattle Grid Copper Mine, Mt. Gunson, South Australia. Australian Journal of Earth Sciences, 32, 287–300.
Williams, I.R., 1992. Geology of the Savory Basin, Western Australia. Western Australia Geological Survey, Bulletin 141, 115 pp.
Williams, I.R., 1994. The Neoproterozoic Savory Basin, Western Australia. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Western Australian branch of the Petroleum Exploration Society of Australia Limited, Perth, 841–850.
Young, G.M., 1995. Are Neoproterozoic glacial deposits preserved on the margins of Laurentia related to the fragmentation of two supercontinents? Geology, 23, 153–156.
Zang, W., 1995. Early Neoproterozoic sequence stratigraphy and acritarch biostratigraphy, eastern Officer Basin, South Australia. Precambrian Research, 74, 119–175.
Zang, W. & Walter, M.R., 1989. Latest Proterozoic plankton from the Amadeus Basin in central Australia. Nature, 337, 642–645.
Zang, W. & Walter, M.R., 1992. Neoproterozoic and Cambrian microfossils and biostratigraphy, Amadeus Basin, central Australia. Association of Australasian Palaeontologists, Memoir 12, 132 pp.
Zhao, J.-X & McCulloch, M.T., 1993. Sm-Nd mineral isochron ages of Late Proterozoic dyke swarms in Australia evidence for two distinctive events of mafic magmatism and crustal extension. Chemical Geology, 109, 341–354.
Zhao, J.-X., McCulloch, M.T. & Korsch, R.J., 1994. Characterisation of a plume-related ~800 Ma magmatic event and its implications for basin formation in central-southern Australia. Earth and Planetary Science Letters, 121, 349–367.
Ziegler, P.A, 1990. Geological atlas of western and central Europe. 2nd edition. Shell Internationale Petroleum Maatschappij B.V., The Hague, 239 pp, 56 encl.

AGSO Journal of Australian Geology & Geophysics, 17(1), 93–99

The mantle dynamical repertoire: plates, plumes, overturns and tectonic evolution

Geoffrey F. Davies[1]

In the present mantle, plates and plumes are the main active components. However, calculations of the thermal evolution of the mantle based on purely thermal convection with plates and plumes yield very smooth changes, which are difficult to reconcile with the apparently episodic accumulation of continental crust and with observed changes in tectonic style through Earth history. The effects of composition on the density of the lithosphere, both at the surface and after subduction, can change the dynamics of plates and mantle convection, resulting in a repertoire of mantle behaviour that may better explain the observed tectonic evolution. Thus major mantle overturns may have occurred during the Archaean and possibly the Proterozoic, with dramatic tectonic and magmatic effects, and plate tectonics may not have worked in its modern form when the mantle was hotter. Plumes are not an alternative to plates: they come from a different thermal boundary layer. Plumes have probably played a significant but usually secondary role throughout Earth history.

Introduction

The Earth's tectonic regime is governed by the way the mantle gets rid of its internal heat. Currently, this is by plate tectonics, which is a form of thermal convection: the plates and their associated 'plate-scale' flow are driven by the negative thermal buoyancy of the plates, and the cycle of plate formation at a spreading centre, cooling, subduction and reheating in the mantle accounts for most of the heat lost from the mantle. Thus the plate-scale flow is the dominant form of mantle convection at present.

The density of the plates, and hence their 'negative buoyancy', is affected by composition as well as by temperature. There are two ways in particular in which this might affect the dynamics of the plates and the mantle. One is through the lower density of the oceanic crust, which makes plates positively buoyant until they are about 15 m.y. old. The other is through effects on phase transformations in the mantle transition zone, which might enhance or inhibit the penetration of mantle convection through the transition zone.

The result of the first effect may be that plate tectonics, in its modern form, has not always been viable. The lithosphere would then have had to behave in a different way, which would result in a different form of tectonics. The second effect may have caused the mantle to become layered, either intermittently or for long periods. The breakdown of such layering, if it then occurred, would probably have had dramatic tectonic and magmatic consequences.

The role of plumes may also have changed, though perhaps not dramatically. However if the mantle was ever layered, then plume heads would have been smaller and plumes probably more numerous.

Plumes cannot be a substitute for plate tectonics, because plumes arise from a lower thermal boundary layer, whereas plates involve the upper thermal boundary layer. A change in the way the upper thermal boundary layer operates does not imply any change in the way plumes operate. If plate tectonics is not available as a way to remove heat from the mantle, then the upper thermal boundary layer will have to find another behaviour that does.

The present mantle

Robust inferences about mantle convection can be made from well-established observations, principally the large-scale topography of the sea floor and heat flow through the sea floor. Here I will simply summarise the main points of the arguments about the present mantle; the details can be found elsewhere (Davies & Richards 1992). The seafloor heat flow is well explained as due to cooling and thickening of the oceanic lithosphere by conduction (Fig. 1). This heat comes from the passive upwelling of hot mantle under spreading centres. The cooled lithosphere is eventually subducted, and ultimately reheated from the internal heat of the mantle, which is renewed by radioactivity. The net result of this process of formation of lithosphere, cooling, subduction and reheating is that heat is removed from the mantle. The ultimate source of this motion is the negative buoyancy of the cooled lithosphere (forces such as 'slab pull' and 'ridge push' exist because the lithosphere is cold and dense). The process is therefore a form of convection; in other words, heat is transported by fluid motions that are driven by buoyancy.

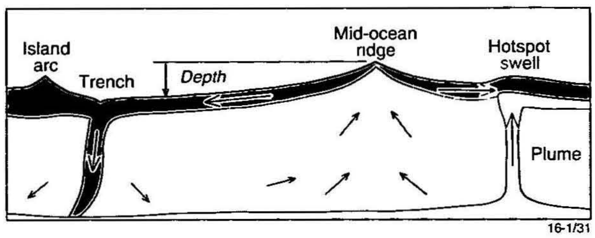


Figure 1. Sketch of the present dynamical system of the mantle, as deduced from topography and heat flow observations (Davies & Richards 1992). The active components are plates and plumes. The negative buoyancy of plates drives a large-scale flow that involves a passive return flow under mid-ocean ridges. Plumes are driven independently by the buoyancy of material from a hot thermal boundary layer at the base of the mantle. Each active component generates characteristic topography. The mid-ocean ridge topography is due to the cooling and thermal contraction of the plates. Hotspot swells are due to the buoyancy of plume material arriving under the lithosphere. The plates and the plate-scale flow are dominant in terms of heat flow, mass flow and topography, and the plumes are secondary, corresponding to about 10% of the Earth's heat budget, which is consistent with them coming from the base of the mantle rather than from the transition zone.

The plates strongly control the structure of mantle convection, since upwelling clearly must occur under spreading centres and subduction zones are the dominant source of active downwelling: the strength of the plates inhibits 'dripping' off their lower surface away from subduction zones. If there were a significant amount of such dripping, it would be evident through widespread topographic and gravity lows, and no such signals are evident.

The depth of the sea floor increases in proportion to the square root of its age (Fig. 1), and this can be explained by thermal contraction due to conductive cooling of the lithosphere. The observed decline of heat flux in inverse proportion to the square root of seafloor age supports this model. A clear implication is that mantle upwelling under normal mid-ocean ridges is passive. If there were active, buoyant upwellings under ridges, the ridge crest would be higher, and the sea

[1] Research School of Earth Sciences, Australian National University, Canberra, ACT 0200, Australia

floor would subside more steeply as it migrated off the buoyant upwelling. (This is observed where there are plumes rising under ridge crests, but elsewhere it is not true.) The implication is that the upwelling under 'normal' ridges is a passive 'return flow' complementing the active downwelling at subduction zones.

Hotspot swells are the only clearly identified manifestation of active, buoyant upwelling (Fig. 1). The size of the swells reflects the amount of buoyant mantle maintaining them. Taking account of the rates of plate motion over plumes, it is possible to calculate the flux of heat transported by plumes, and this turns out to be less than 10% of the Earth's heat budget (Davies 1988, Sleep 1990). This estimate is not dependent on assumptions about the temperature of plumes or the viscosity of the mantle. Also, it is consistent with the amount of heat expected to be conducting from the core into the base of the mantle. This heat would generate the hot thermal boundary layer that gives rise to plumes.

The topography of the sea floor provides constraints also on the possible layering of the mantle. If the mantle were layered at present, then there should be much stronger buoyant upwellings reaching the base of the lithosphere, and these would generate topography rivalling the mid-ocean ridge system in scale and magnitude. This argument proceeds as follows. There is very little radioactivity in the upper mantle (Jochum et al. 1986), so most of the heat emerging at the surface must come from below the upper mantle. Cooling of the upper mantle can account for only a small fraction of the total. Therefore, if there is a barrier to flow in the mantle transition zone, most of the Earth's heat budget must conduct through it. This will generate a hot thermal boundary layer at the base of the upper mantle, and from this will arise hot upwellings. When these reach the base of the lithosphere, they will raise it and so generate topography. There is a proportionality between heat flux, buoyancy flux and topography, so the uplifts would be comparable to the mid-ocean ridges (which are the expression of the same heat *leaving* the upper mantle).

The mantle may be close to the condition where flow through the transition zone is blocked. Seismic tomography seems to show some subducted slabs penetrating the transition zone and others not (Grand 1994, van der Hilst et al. 1991, 1997), although some of the slab contortions can be explained by local trench motions (van der Hilst 1995, van der Hilst & Seno 1993). Numerical models show that old, stiff slabs and large plume heads can probably penetrate a phase transition barrier, but that young, thin slabs and some plume tails may not (Davies 1995a).

The picture of the present mantle that emerges from these arguments is that there is a dominant 'plate-scale' mode of mantle convection, a secondary plume mode driven by a lower thermal boundary layer, and substantial flow between the upper mantle and the lower mantle, though possibly with localised and transient interruptions.

There have been arguments made that geochemical evidence requires the mantle to be layered (Allegre et al. 1987, Jacobsen & Wasserburg 1979, O'Nions 1987). The evidence certainly requires the source of mid-ocean ridge basalts (which must be the shallow mantle) to be different from sources represented by hotspots, which must be deeper. However, there is little else that can be concluded solely on the basis of the geochemistry. It is plausible, though not proven, that the geochemical heterogeneity of the mantle can be explained by slow mantle mixing at depth due to higher viscosity, separation of the heavy oceanic crust component at the base of the mantle, and inhibition of flow through the transition zone (Christensen & Hofmann 1994, Davies 1990, 1995a). Neither the geochemistry (Hofmann 1997) nor the geophysics (above) supports a simplistic picture with a depleted upper mantle and a primitive lower mantle.

Thermal evolution of the mantle: simple smooth case

It is possible to calculate the rate of change of mantle temperature, and thus to calculate the thermal history of the mantle. The results depend on how the mantle is being cooled. It is thus possible to calculate the implications of different assumptions, and to use the record of the Earth's tectonic history to test the models. Conversely, we will see here that the calculations suggest possibilities that have not been apparent until recently, and these may provide alternative ways to interpret the rock record. The objective is to gain the kind of fundamental understanding of the Earth's past tectonic regime(s) that we think we have for its present regime of plate tectonics.

I will not present equations here. These are available in the references. The rate of decrease of mantle temperature depends on the net rate of cooling. This is the cooling due to the action of plate tectonics, as described above, minus the heat input due to radioactivity and due to heat coming from the core. The latter is assumed here to be evident in the heat carried by plumes, which assumes that plumes come from the base of the mantle and carry the heat conducted from the core into the base of the mantle.

The key to these calculations is knowing how much heat is transported by convection (i.e. by plates and plumes). There is a simplified 'boundary layer theory' of convection that relates the rate of heat transport to the temperature difference between the surface (top or bottom) and the interior fluid (Davies 1993, Stacey & Loper 1984). The resulting expression also involves the mantle viscosity (which obviously affects the rate of convection), and this is a strong function of temperature: an increase of 100°C causes a viscosity decrease by about a factor of ten.

An example of the results of this kind of calculation is shown in Figure 2 (after Davies 1993). The top panel shows the temperatures at the top of the mantle (below the lithosphere) and at the top of the core. A schematic temperature profile through the mantle and core is shown in Figure 3. The core temperature is much higher because there is an adiabatic increase of temperature through the mantle of about 1000°C, as well as the difference in temperature between the deep mantle and the top of the core. The calculation in Figure 2 started with the lowest mantle at the same temperature as the core. The mantle cools rapidly, because it is initially hot and has a low viscosity, so convective cooling is fast. After about 0.5 Ga, the rate of mantle cooling slows, because of the effect of radioactive heating.

The lower panel of Figure 2 shows the contributions to the heat budget of the mantle. The heat loss ('plates') drops rapidly at first, like the mantle temperature, until it approaches the rate of radioactive heat generation ('heat gen'). Thereafter it tracks the heat generation. The mantle temperature drops only slowly in this phase (top panel). This is because of the strong temperature-dependence of the viscosity: it takes only a modest change in temperature to accommodate a large change in heat transport.

Initially there is no temperature difference between the deep mantle and the top of the core, by assumption. This means there will be no heat conduction from the core into the mantle, so there will be no hot thermal boundary layer at the bottom of the mantle and no plumes generated. Thus, initially, plumes transport no heat. As the mantle cools, the bottom thermal boundary layer becomes established and plumes can begin. In Figure 2, the lower panel shows that the heat transported by plumes rises rapidly from zero early in the calculation. Thereafter the plume flux is fairly constant. This is because both the mantle and the core are cooling only slowly, so the temperature difference between them does not change much, and the rate of plume generation does not change much.

The results in Figure 2 illustrate important features of the Earth's internal thermal system in a fairly simple case. An important point is that the cooling of the mantle is governed mainly by the decline of radioactive heating. This changes by about a factor of three since 3.5 Ga, but is accommodated by a change in mantle temperature of only about 200°C.

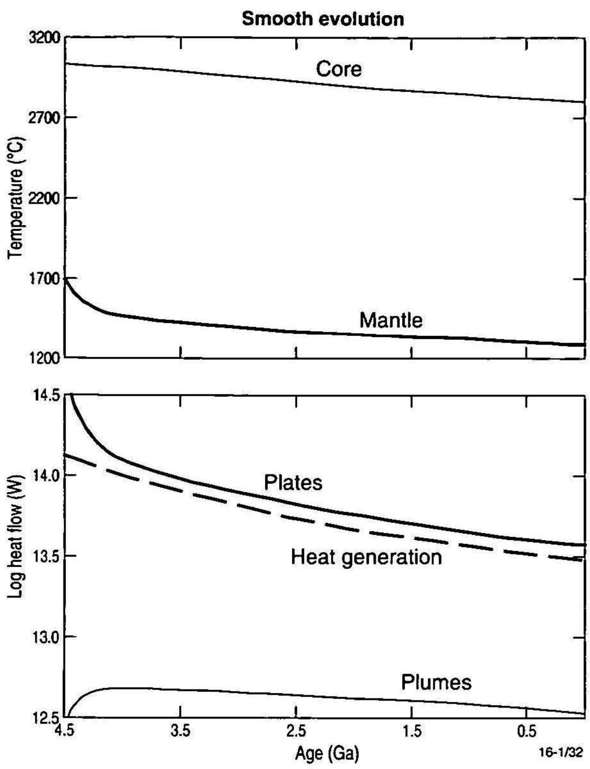


Figure 2. Thermal evolution of the core and mantle, assuming whole mantle convection, plate tectonics, and plumes coming from the core-mantle boundary. Top panel: temperatures at the top of the mantle (below the lithosphere) and at the top of the core. Bottom panel: heat flows in and out of the mantle. The 'plates' curve is the rate of heat loss from the mantle. 'Heat gen' is the rate of radioactive heat generation, and 'plumes' is the rate of heat input into the mantle in the form of plumes from the core-mantle boundary (from Davies 1993).

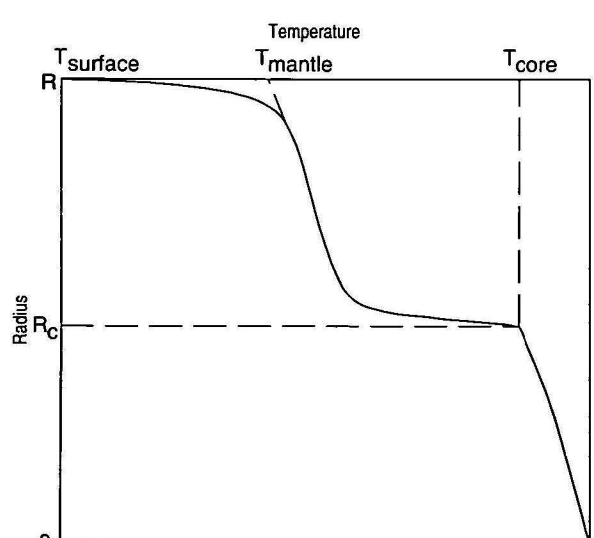


Figure 3. Schematic temperature profile through the mantle and core, illustrating the thermal boundary layers at the top and bottom of the mantle, the adiabatic temperature increase through the mantle, and the temperatures used to characterise the evolution of the mantle and core in the model of Fig. 2.

Mantle phase transformations

The mantle transition zone, between about 400 km and 670 km depth, is inferred to be a region in which upper mantle minerals go through a series of pressure-induced transformations to denser structures. Because the pressure at which a transformation occurs depends on temperature, the depth of the transformation is different in cold subducted lithosphere from the depth in surrounding hot mantle.

The transformation of Mg_2SiO_4 from the spinel structure to $MgSiO_3$ (perovskite structure) plus MgO (sodium chloride structure) occurs at a depth of about 660 km, and experiments indicate that its transformation pressure decreases as the temperature increases (Akaogi & Ito 1993). This is often expressed by saying the transformation has a negative Clapeyron slope, the Clapeyron slope (or more strictly, the Clausius-Clapeyron slope) being the temperature derivative of the transformation pressure.

The effect of this negative slope in subducted lithosphere is illustrated schematically in Figure 4. Within the slab, the lower temperature causes the phase boundary to occur deeper, where the pressure is higher. As a result there is a region (shaded) where the low-pressure phase exists within the slab at the same depth as the high pressure phase exists outside the slab. This region has a lower density, and so it is buoyant. This buoyancy (broad arrow) opposes the descent of the subducted lithosphere. If this buoyancy were sufficiently large, compared with the negative thermal buoyancy of the slab, it might prevent the slab from penetrating the phase transformation zone.

It was first shown by Machatel & Weber (1991) that this could cause intermittent layering in numerical models of a convecting fluid. Whether this might be true in the mantle has remained an open question because of the simplifications in many numerical models. More recent, somewhat more realistic models suggest that it is not so likely in the present mantle (Davies 1995a), which would be consistent with other evidence, summarised above, that the mantle is not layered at present.

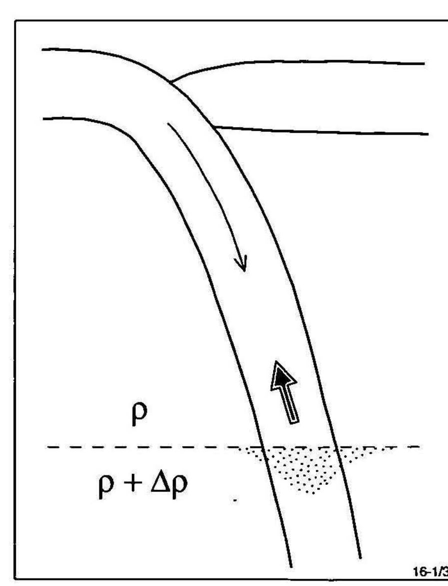


Figure 4. Sketch showing the deflection of a phase transformation boundary to greater depth in the subducting lithosphere because of the lower temperature in the subducting lithosphere and the temperature-dependence of the transformation pressure. In the zone of deflection (shaded) the less-dense low-pressure phase exists, and this region is buoyant as a result (broad arrow).

However, this mechanism would have been more able to induce layering in the past, when the mantle was hotter. Because the plates would have moved faster in that case, they would have been younger and thinner when subducted, and less able to penetrate a phase barrier. If layering was induced in this way, the plate cycle would cool the upper mantle. As the upper mantle cooled, there would come a point at which a plate was thick enough to penetrate the phase barrier, and the layering might then break down.

Numerical models indicate that the layering can also be broken down by another mechanism, which amounts to there being a limit on the temperature difference that can exist between the layers. As the upper mantle cools relative to the lower mantle, thermal boundary layers are formed on either side of the interface between them as heat conducts through the interface. These thermal boundary layers generate convection, and if this becomes sufficiently vigorous it might pull material through the interface. This becomes more likely as the temperature difference between the layers increases, for two reasons: first, the local convection is more vigorous, and second, thermal contraction of the upper layer increases the likelihood that the cooler upper mantle material can break through into the lower layer (and *vice versa* for the lower layer).

The effect of these rather complicated ideas is that layering will break down if one of two conditions is reached: (a) the upper mantle cools to a critical temperature (at which plates begin to break through); (b) the temperature difference between the upper mantle and the lower mantle reaches a maximum value (at which internal breakthrough may occur). The actual value of the critical temperatures is far from certain at present, but they can be estimated roughly, well enough for possible consequences to be explored.

One example of a possible mantle thermal evolution with these effects included is shown in Figure 5 (Davies 1995b). A smooth ('whole mantle') evolution similar to that shown in Figure 2 is included for comparison. The first impression of the effect of the phase barrier must be that it can produce dramatically different behaviour: the temperatures of the upper mantle and lower mantle undergo large swings through the earlier parts of Earth history, ultimately to die out and approach the smooth "whole mantle" curve. Before looking at the details of this model, two main aspects of the character of the behaviour can be noted. First, the model is episodic, more like the apparently episodic history of additions to the continental crust. Second, there are three long phases of behaviour, reminiscent of the three tectonic eras of Earth history.

The model starts with the upper mantle and the lower mantle both at a high temperature. In this condition layering occurs. The upper mantle cools, because of the cycle of plate formation and subduction. The lower mantle heats, because it is heated by radioactivity but not cooled by the plates. As their temperatures diverge there is some heat transfer between them, involving convection driven by the internal thermal boundary layers (see Davies 1995b for details). However it is not sufficient to stop the divergence, and a breakdown of the layering occurs when the temperature difference reaches the critical value. At this point it is assumed, for the purposes of the calculation, that hot material from the lower mantle replaces the cooled material, and the latter is mixed with the remaining lower mantle material. Thus the temperature of the upper mantle is reset to the temperature the lower mantle had reached, while the temperature of the lower mantle is reset to the mean value of the former upper mantle and the remaining lower mantle. The evolution is then continued.

In this model, four such mantle overturns occur, the last at about 2.7 Ga. The following overturn, nearly as large, is part of the second series (discussed below; Fig. 5). After each overturn, the upper mantle cools to a lower temperature. At

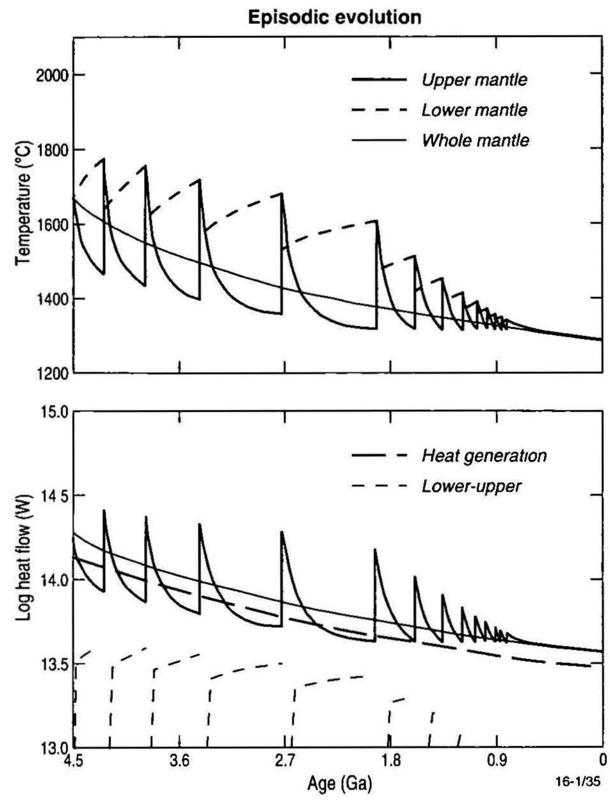


Figure 5. An example of a thermal evolution in which a phase transformation barrier causes transient layering of the mantle, which then convects separately in the upper mantle and the lower mantle. An evolution assuming whole-mantle convection, similar to that in Fig. 2, is included for comparison. When the mantle is layered, the upper mantle and lower mantle temperatures diverge. When the layering breaks down (see text), the cooled upper mantle material is replaced by hot lower mantle material, and the remaining lower mantle material is assumed to mix with the cool material displaced from the upper mantle. Lower panel shows heat flows, including the heat generated by radioactivity and the heat transferred across the interface between the upper mantle and the lower mantle ('lower-upper'). After (Davies 1995b). For this case, the transition zone heat transfer parameter is b_z = 0.6 and plate penetration occurs at an age of 72 m.y. (b_u = 3.25).

about 1.9 Ga in this model, it has cooled to the critical temperature at which plates can start to penetrate the phase barrier, and another overturn is assumed to occur. There follows a second series of overturns triggered by plate penetration. These occur after progressively smaller intervals and with progressively small temperature variations, because the temperature of the lower mantle is being progressively reduced. Eventually, by about 700 Ma, the lower mantle temperature is reduced to the same as the upper mantle, and whole mantle convection is assumed to occur thereafter.

There are other possible courses of this type of thermal evolution, and in fact the parameters of this one, although plausible, have been adjusted to correspond to particular features of the Earth's tectonic history. Some of the other possibilities are illustrated in Figure 6. In Figure 6(a), the ability of subducted lithosphere to penetrate the phase barrier has been increased by assuming it can penetrate by the time it is 25 m.y. old (the 'buckling parameter' b_u = 5.5 (Davies 1995b)), compared with penetration at 72 m.y. (b_u = 3.25) in Fig. 5. In this case the sequence of overturns triggered by plate penetration begins and finishes much earlier, and all overturns are confined to the Archaean.

Figure 6(b) shows the effect of more efficient heat transfer between the upper and lower mantles (parameter b_z = 0.8, compared with b_z = 0.6 in case (a); Davies 1995b). In this

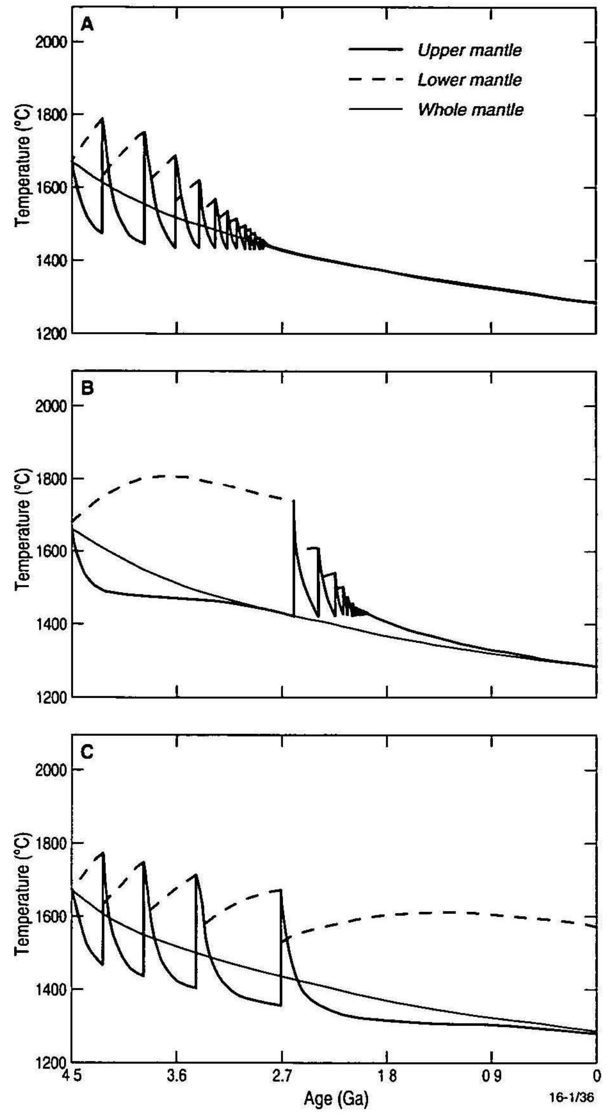


Figure 6. Further examples of thermal evolution models with episodic layering. (a) Plates penetrate at an age of 25 m.y. ($b_u = 5.5$). (b) Heat transfer between layers is more efficient ($b_z = 0.8$; $b_u = 5.5$). (c) Plate penetration is assumed not to occur (with $b_z = 0.6$). This model is less plausible (see text).

case sufficient heat is transferred to keep the temperature difference between the upper and lower mantles from reaching the critical value, and the internally triggered overturns never occur. The mantle is stably layered for a long period through the Archaean. Overturns triggered by plate penetration begin at 2.6 Ga and whole-mantle convection is achieved by 1.9 Ga.

Figure 6(c) is like the case in Figure 5, except that plate penetration is, in effect, turned off. This again shows a long period of stable layering, after the internally triggered overturns have ceased. However this model seems less plausible, since it is necessary to assume that plates as old as 100 m.y. cannot penetrate the phase barrier.

The most important point here is not the details of a particular calculation, but the overall character of this type of thermal history.

Implications of mantle overturns

The task of assessing whether the Earth has undergone an evolution of this type may not be easy. A preliminary discussion has been given elsewhere (Davies 1995b). Here, only some key points will be mentioned. The potential geological implications of the kind of mantle overturn that occurs in the model of Figure 5 are difficult to predict in any detail, but it is easy to show that they could be dramatic.

The arrival in the upper mantle of material 200°C or more hotter than the material it displaces would generate a major magmatic event. Flood basalts are believed to be generated by the arrival of a plume head 100–200°C hotter than ambient upper mantle (Campbell & Griffiths 1990, Richards et al. 1989), but a typical flood basalt province covers less than 1% of the Earth's surface. Thus a mantle overturn event might be like 100 flood basalt events occurring simultaneously. There would probably be substantial tectonic effects too, as mantle flow rates would be much larger during an overturn and the hotter mantle would have lower viscosity: both factors would be likely to cause plates to move much faster It is plausible that an overturn event could generate and aggregate large amounts of mafic crust, and that such crust would be substantially reworked either in the same or subsequent overturn events (Davies 1995b).

Whether such an event might correspond to one of the major crust-forming episodes, such as the 2.7 Ga or 1.9 Ga episodes, will not be easy to determine. The main point here is to demonstrate the possibility, so that it can be considered along with other hypotheses. At present, the main hypotheses for episodic events would seem to be mantle overturns, large plume heads (Campbell & Griffiths 1990, Campbell & Hill 1988), or continental breakup and aggregation (Condie 1995, Hoffman 1988, 1991).

Buoyancy of the oceanic crust

Because the oceanic crust is less dense than the mantle (2.9 g/cm^3 compared with 3.3 g/cm^3), the oceanic lithosphere must cool for about 15 m.y. before its negative thermal buoyancy overcomes the buoyancy of the oceanic crust. The mantle residue from melting at mid-ocean ridges is also less dense than normal mantle (Ringwood & Irifune 1988) (Figure 7a). In combination, these compositional buoyancies mean that oceanic lithosphere has a net positive buoyancy until it is about 20 m.y. old. Since it is the negative buoyancy of the lithosphere that ultimately drives plate tectonics, this means that plates younger than about 20 m.y. could not be driven.

If the mantle was hotter in the past, two things would be different, and each would amplify this effect. First, a hotter

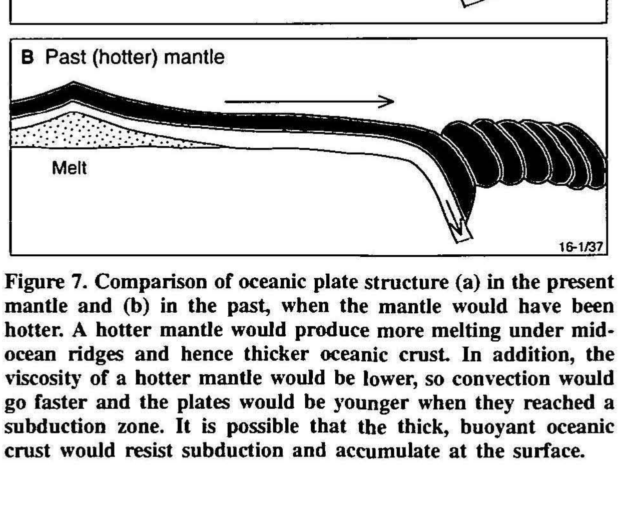


Figure 7. Comparison of oceanic plate structure (a) in the present mantle and (b) in the past, when the mantle would have been hotter. A hotter mantle would produce more melting under mid-ocean ridges and hence thicker oceanic crust. In addition, the viscosity of a hotter mantle would be lower, so convection would go faster and the plates would be younger when they reached a subduction zone. It is possible that the thick, buoyant oceanic crust would resist subduction and accumulate at the surface.

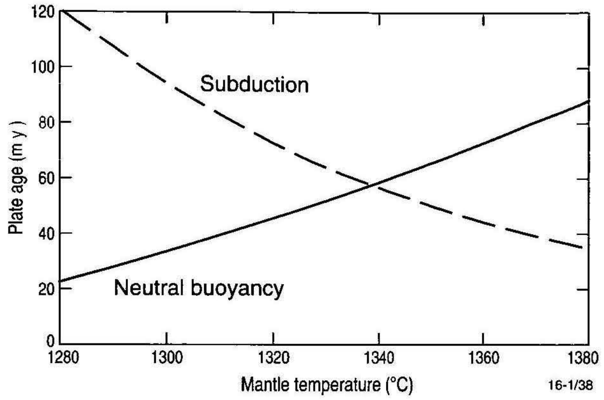


Figure 8. Age at which oceanic lithosphere becomes neutrally buoyant (solid) and mean age of plates upon subduction (dashed) as functions of mantle temperature. If the mantle were hotter, the resulting thicker oceanic crust (Fig. 7) would make the lithosphere positively buoyant until the age given by the solid curve, and the lithosphere would arrive at a subduction zone sooner (Fig. 7), when its age was that given by the dashed curve. At temperatures about 60°C above the present mantle temperature the curves cross, meaning the plates would, on average, be only neutrally buoyant when they arrived at a subduction zone, so there would be no force to pull them down. At higher temperatures, the plates would have to move more slowly in order to be old enough to subduct (following the solid curve), or they might delaminate the crust and mantle parts, the crust accumulating at the surface and the mantle part subducting (Fig. 7). In either case they would remove less heat from the mantle than was necessary to cool it (after Davies 1992).

mantle would generate a thicker oceanic crust, due to more melting at mid-ocean ridges (Fig. 7b). This would increase the compositional buoyancy. Second, the mantle viscosity would be lower, so mantle convection and plates would go faster, and plates would be younger on average when they reached a subduction zone. This would reduce their negative thermal buoyancy. Figure 8 shows how these two effects vary as a function of mantle temperature (Davies 1992). It shows that if the mantle is about 60°C hotter than at present, the oldest plates would only be neutrally buoyant when they reached a trench.

At higher mantle temperatures, plate tectonics could still be driven, but more slowly than is implicit in calculating the 'subduction' curve in Figure 8. The result would be that plate tectonics would no longer remove heat fast enough to cool the mantle. This leads to a paradox: it implies that the mantle would be hot and getting hotter, and could not have cooled to its present temperature.

The paradox can be avoided only by assuming that the plates were not behaving as they do today, or that something other than plate tectonics was operating. Two possibilities can be mentioned here. One is that the transformation of the basaltic oceanic crust to dense eclogite might keep plates moving. This would depend on the transformation proceeding at a sufficient rate, which in turn depends on what controls the transformation. If it is limited by reaction kinetics at low temperature, then it might proceed only slowly, and might not be viable. If it is possible to force the transformation by overdriving the pressure, then it might proceed more rapidly. In any case, there is a threshold pressure, equivalent to roughly 60 km depth, which the crust must reach before it can transform.

A vigorously subducting plate might maintain its subduction, through the deeper part pulling the shallow, buoyant part after it. However, if such subduction were ever interrupted, it might be difficult to restart. This might involve accumulating a large pile of basaltic crust until its root became deep enough and hot enough to transform and sink. The result might be a more episodic form of plate tectonics.

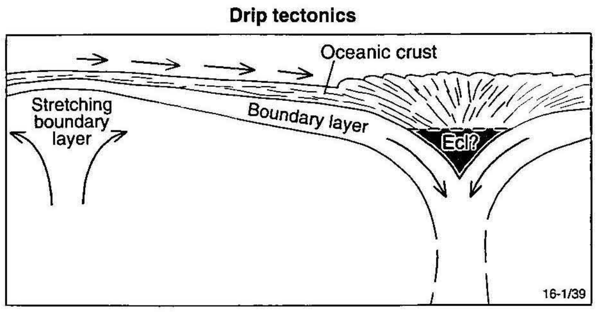


Figure 9. Sketch of possible non-plate behaviour of the mantle's upper thermal boundary layer that might have occurred when the mantle was hotter. The mantle part of the boundary layer is too thin and weak to behave as a rigid plate, so it stretches over a broad zone of upwelling and descends symmetrically under a convergence zone. Thick oceanic crust piles up over the convergence zone. The lower part of this pile might transform to eclogite ('Ecl') and sink episodically (after Davies 1992).

A second possibility, if conventional subduction did not work, might be that the crustal component of the lithosphere separated from the mantle component and accumulated at the surface. A scenario along these lines is illustrated in Figure 9 (Davies 1992). Unfortunately, this mode of tectonics would be less efficient at cooling the mantle than plate tectonics, because the coldest part is left at the surface. It is not clear that this is a viable way of cooling the mantle.

It is not immediately obvious how these possibilities can be quantified in order to estimate the resulting plate velocities and rates of mantle cooling, and this has not yet been done, hence the qualitative and conjectural nature of this discussion. It is clear, however, that there are good reasons of physics to expect that tectonics would have operated differently when the mantle was hotter. What is not clear is whether the tectonics would have been a modified version of plate tectonics or something quite different, perhaps like that sketched in Figure 9.

This means that the thermal evolution calculations presented earlier may not be accurate for those times when the mantle was substantially hotter than at present. What can be said is that the mantle clearly has been able to cool to its present temperature, that the rate of cooling is likely to depend strongly on mantle viscosity and hence on mantle temperature, and that as a result the character of the calculations presented above is likely to be retained as our understanding improves, though details are subject to substantial revision.

Plumes

The existence of mantle plumes can be inferred with little uncertainty from the existence of the volcanic hotspots and their associated hotspot swells (Davies 1988, Davies & Richards 1992). The swells are not due to thickened crust, and the lithosphere is not strong enough to hold up such broad features (about 1000 km across), so the only remaining possibility is that they are held up by buoyant mantle under the lithosphere. As noted above, it can be inferred from the swells that mantle plumes carry no more than about 10% of the Earth's heat budget. The only plausible source of plumes is a hot thermal boundary layer at depth. This is taken here to be at the base of the mantle, for the reasons given earlier.

As shown earlier, if whole mantle convection applied throughout Earth history, then the plume flux would probably not have changed very much (Fig. 2). However, if there were periods in which the mantle was layered, then, by assumption, plumes from the base of the mantle would not have reached the surface. Instead, a thermal boundary layer would form at the base of the *upper* mantle, and smaller, more numerous plumes would probably arise from this. The flux of such upper mantle plumes could be greater than the present plume flux. This can be seen from Figure 5, in which the 'lower-upper' heat flow is the heat passing through the interface between the upper mantle and the lower mantle. It is this heat that would be transported by the upper mantle plumes.

Plumes were not included in the calculations of episodic thermal evolution presented above. This was justified as a first approximation by noting that the plumes transport only a fraction of the Earth's heat budget and thus will not exert the dominant control on mantle temperatures (Davies 1995b). However, it is possible that plumes may be able to play a role in triggering mantle overturns. This possibility has not yet been explored. It is one more reason that the details of Figures 4 and 5 are still conjectural.

It is sometimes suggested that 'plume tectonics' could substitute for plate tectonics, either in the Earth's earlier history or on other planets, particularly Venus. This betrays a misunderstanding of the role of plumes. The existence and vigour of plumes depends on the presence and strength of a hot thermal boundary layer at the base of the convecting layer. It follows that the existence of plate tectonics does not directly affect the role of plumes. Plate tectonics is the current expression of the upper, cool thermal boundary layer of the mantle. If this boundary layer operated differently in the past, then something other than plate tectonics may have prevailed, but plumes would have existed anyway, so long as the mantle was cooler than the core, or the upper mantle was cooler than the lower mantle. Plumes and plates play different roles. Plumes are an expression of heat entering the mantle from below, while plates are an expression of heat leaving the mantle at the top.

References

Akaogi, M. & Ito, E., 1993. Refinement of enthalpy measurement of MgSiO3 perovskite and negative pressure-temperature slopes for perovskite-forming reactions. Geophysical Research Letters, 20, 1839–1842.

Allegre, C.J., Hamelin, B., Provost, A. & Dupre, B., 1987. Topology in isotopic multispace and origin of mantle chemical heterogeneities. Earth and Planetary Science Letters, 81, 319–337.

Campbell, I.H. & Griffiths, R.W., 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth and Planetary Science Letters, 99, 79–83.

Campbell, I.H. & Hill, R.I., 1988. A two-stage model for the formation of the granite-greenstone terrains of the Kalgoorlie-Norseman area, Western Australia. Earth and Planetary Science Letters, 90, 11–25.

Christensen, U.R. & Hofmann, A.W., 1994. Segregation of subducted oceanic crust in the convecting mantle. Journal of Geophysical Research, 99, 19,867–19,884.

Condie, K.C., 1995. Episodic ages of greenstones: a key to mantle dynamics? Geophysical Research Letters, 22, 2215–2218.

Davies, G.F., 1988. Ocean bathymetry and mantle convection, 1. Large-scale flow and hotspots. Journal of Geophysical Research, 93, 10467–10480.

Davies, G.F., 1990. Mantle plumes, mantle stirring and hotspot chemistry. Earth and Planetary Science Letters, 99, 94–109.

Davies, G.F., 1992. On the emergence of plate tectonics. Geology, 20, 963–966.

Davies, G.F., 1993. Cooling the core and mantle by plume and plate flows. Geophysical Journal International, 115, 132–146.

Davies, G.F., 1995a. Penetration of plates and plumes through the mantle transition zone. Earth and Planetary Science Letters, 133, 507–516.

Davies, G.F., 1995b. Punctuated tectonic evolution of the earth. Earth and Planetary Science Letters, 136, 363–379.

Davies, G.F. & Richards, M.A., 1992. Mantle convection. Journal of Geology, 100, 151–206.

Grand, S.P., 1994. Mantle shear structure beneath the Americas and surrounding oceans. Journal of Geophysical Research, 99, 11,591–11,621.

Hoffman, P.F., 1988. United plates of America, the birth of a craton. Annual Review of Earth and Planetary Sciences, 16, 543–603.

Hoffman, P.F., 1991. Did the breakout of Laurentia turn Gondwanaland inside-out? Science, 252, 1409–1412.

Hofmann, A.W., 1997. Mantle chemistry: the message from oceanic volcanism. Nature, 385, 219–229.

Jacobsen, S.B. & Wasserburg, G.J., 1979. The mean age of mantle and crustal reservoirs. Journal of Geophysical Research, 84, 7411–7427.

Jochum, K.P., Hofmann, A.W., Ito, E., Seufert, H.M. & White, W.M., 1986. K, U and Th in mid-ocean ridge basalt glasses and heat production. Nature, 306, 431–436.

Machetel, P. & Weber, P., 1991. Intermittent layered convection in a model mantle with an endothermic phase change at 670 km. Nature, 350, 55–57.

O'Nions, R.K., 1987. Relationship between chemical and convective layering in the earth. Journal of the Geological Society, London, 144, 259–274.

Richards, M.A., Duncan, R.A. & Courtillot, V.E., 1989. Flood basalts and hot-spot tracks: plume heads and tails. Science, 246, 103–107.

Ringwood, A.E. & Irifune, T., 1988. Nature of the 650-km discontinuity: implications for mantle dynamics and differentiation. Nature, 331, 131–136.

Sleep, N.H., 1990. Hotspots and mantle plumes: some phenomenology. Journal of Geophysical Research, 95, 6715–6736.

Stacey, F.W. & Loper, D.E., 1984. Thermal histories of the core and mantle. Physics of the Earth and Planetary Interiors, 36, 99–115.

van der Hilst, R., 1995. Complex morphology of subducted lithosphere in the mantle beneath the Tonga trench. Nature, 374, 154–157.

van der Hilst, R. & Seno, T., 1993. Effects of relative plate motion on the deep structure and penetration depth of slabs below the Izu-Bonin and Mariana island arcs. Earth and Planetary Science Letters, 120, 395–407.

van der Hilst, R., Engdahl, R., Spakman, W. & Nolet, G., 1991. Tomographic imaging of subducted lithosphere below northwest Pacific island arcs. Nature, 353, 37–43.

van der Hilst, R.D., Widiyantoro, S. & Engdahl, E.R., 1997. Evidence for deep mantle circulation from global tomography. Nature, 386, 578–584.1.

AGSO Journal of Australian Geology & Geophysics, 17(1), 101–111

Australia's buoyancy inherited from Gondwanaland[1]

J.J. Veevers[2,3]

In company with the other components and continental successors of the Neoproterozoic to Mesozoic Gondwanaland supercontinent, Phanerozoic Australia has a buoyant cratonic platform characterised by non-marine facies, in contrast to the marine facies of the components of depressed Laurasia. As a supercontinent, Gondwanaland lasted much longer than Laurasia, and was therefore hotter from its more effective insulation of internal heat. Moreover, the Pan-African orogenic cycle, confined to Gondwanaland, augmented the heat supply, and I postulate that Pan-African heat generated a permanently buoyant lower crust by mafic underplating. A crustal layer in the Australian Proterozoic shield with subhorizontal reflectors and velocity (V_p) > 7.5 km s^{-1} is interpreted as the product of mafic underplating beneath latest Neoproterozoic flood basalt. The Pan-African terrane in East Africa also contains evidence of mafic underplating, and most of Gondwanaland (but not Laurasia) was affected by terminal Pan-African (0.5 Ga) uplift and cooling. In equivalent Late Cretaceous and Late Ordovician stages of the 400 m.y. Pangean supercycle, the Australian (and possibly the South American) platform deviated from the global norm by rising faster than eustatic sea level. In Australia, plate boundary events — steeper subduction, mafic underplating of a lower plate along a divergent back-arc boundary — explain uplift in the east, but not that in the west, which relaxed to its natural buoyant state.

Introduction

Fischer (1984) and Veevers (1990, 1994) interpreted the first-order features of the sea-level curve (Vail et al. 1977) as due to the two states of continent and ocean that alternate in a 400 m.y. supercycle:

(1) in the dispersed state, a long mid-ocean ridge displaces water upward to flood the many continents which have low-lying extended margins and low mean elevations; and

(2) in the Pangean state, a short mid-ocean ridge combines with shorter low-lying extended margins to produce an emergent Pangea.

Anderson's (1982) recognition of the Atlantic–African geoid high as a relic of Triassic Pangea pointed to heat generated in Pangea as the driving force of the supercycle. Authors such as Gurnis (1988) elaborated possible deep-seated mechanisms. Here I interpret the difference in buoyancy between Gondwanaland and Laurasia as due to preferential underplating of the lower crust of the long-lived Gondwanaland, and provide cogent information from Australia.

Flooding of the continental platform

From a line of descent through Gondwanaland, Australia and its siblings have inherited a buoyant lithosphere, indicated by the small area of the platform covered by the sea (Fig. 1; Veevers 1995). The Gondwanaland data come from Africa, Arabia, Indostania, South America (Ronov 1994), and Australia (BMR Palaeogeographic Group 1990, modified here).

In Australia, uncertainties about areas covered by the Palaeozoic sea in the present offshore Tasman Fold Belt System and New Guinea were avoided by excluding these areas from consideration. Accordingly, the area of the sea covering the continental platform was measured within the bounds of the present coastline of the mainland, extended to Tasmania from the Permian, and on the eastern margin from the Cainozoic shelf edge and the Cretaceous and older magmatic arc or orogen.

Except for an overlap in the Late Silurian (S2 in Fig. 1) and Early Devonian (D1), the curves of platform areas flooded by shallow seas (Fig. 1C) grossly parallel each other, but the larger area of Gondwanaland minus Antarctica (G in Fig. 1C; 53x10^6 km^2) has only half the area of the sea over the smaller Laurasia (L; 43x10^6 km^2); the percentage area flooded on each supercontinent has the same trend: G ranges from 2 to 24 per cent, and L from 6 to 46 per cent. The first-order curves of flooded areas also grossly parallel the number of continents (Fig. 1D), in sympathy with the curve of global sea level (Figs. 1A, 2).

Departures in Australia (Fig. 1A) from the sea-level curve are Australian peaks in Early/Middle Cambrian (C1/2; 520 Ma) and Early Ordovician (O1; 480 Ma) times versus Late Ordovician (O3; 450 Ma) time, and in Aptian (K1; 116.5 Ma) versus Campanian (K2; 80 Ma) time. An explanation of these departures is attempted below.

The contrast in the size of the flooded areas in Australia (present land area 7.7x10^6 km^2) and South America (17.8x 10^6 km^2) compared with North America (22.0x10^6 km^2) is extreme, in particular during the Gondwanan (Carboniferous through Jurassic) interval of negligible flooding. For interior sites in Australia (dotted circle in Fig. 3A) and in the Amazon Basin of South America, and for mid-continental North America, the total intervals covered by the sea (heavier shading in Figs 1A, B) provide the same contrast: Australia for 70 m.y., or 13 per cent of the 544 m.y. period surveyed; South America for 85 m.y., or 16 per cent, and North America for 275 m.y., or 51 per cent.

This fundamental difference in wetness is exemplified by the distribution of marine carbonate: major in Laurasia, minor in Gondwanaland. Neoproterozoic and Phanerozoic marine platforms in all but high latitudes are characterised, if not dominated, by carbonate sediment, and non-marine platforms by siliciclastic sediment. Abundant marine carbonate in the almost continuous Neoproterozoic and Phanerozoic sections of Laurasia means that studies of stable isotopes in Laurasian rocks have been conducted almost exclusively on carbonate; in Australia, where carbonate is only intermittent, such studies rely on the kerogen preserved in Neoproterozoic (e.g. Calver 1995) and Permian (e.g. Morante 1995) siliciclastic rocks.

Pangean supersequence

Pangean tectonics developed through the following ideas:

- Holmes (1931) proposed monsoon-like convection currents from the juxtaposition of ocean floor and radioactive continental blocks.
- Anderson (1982) interpreted the present positive geoid as a relic of Triassic Pangea.
- Gurnis (1988) made numerical simulations of large-scale mantle convection and the aggregation and dispersal of supercontinents.
- Veevers (1990, 1994) found evidence for a 400 m.y. period of Pangean tectonics from the distinctive Pangean supersequence deposited on the supercontinental platform since 320 Ma (cycle A), earlier from 720 Ma (cycle B), and initially, from 1120 Ma (cycle C).

[1] Publication 54 from the Key Centre for Geochemical Evolution and Metallogeny of the Continents (GEMOC)
[2] Australian Plate Research Group
[3] Key Centre for the Geochemical Evolution and Metallogeny of the Continents, School of Earth Sciences, Macquarie University, N Ryde, NSW 2109, Australia

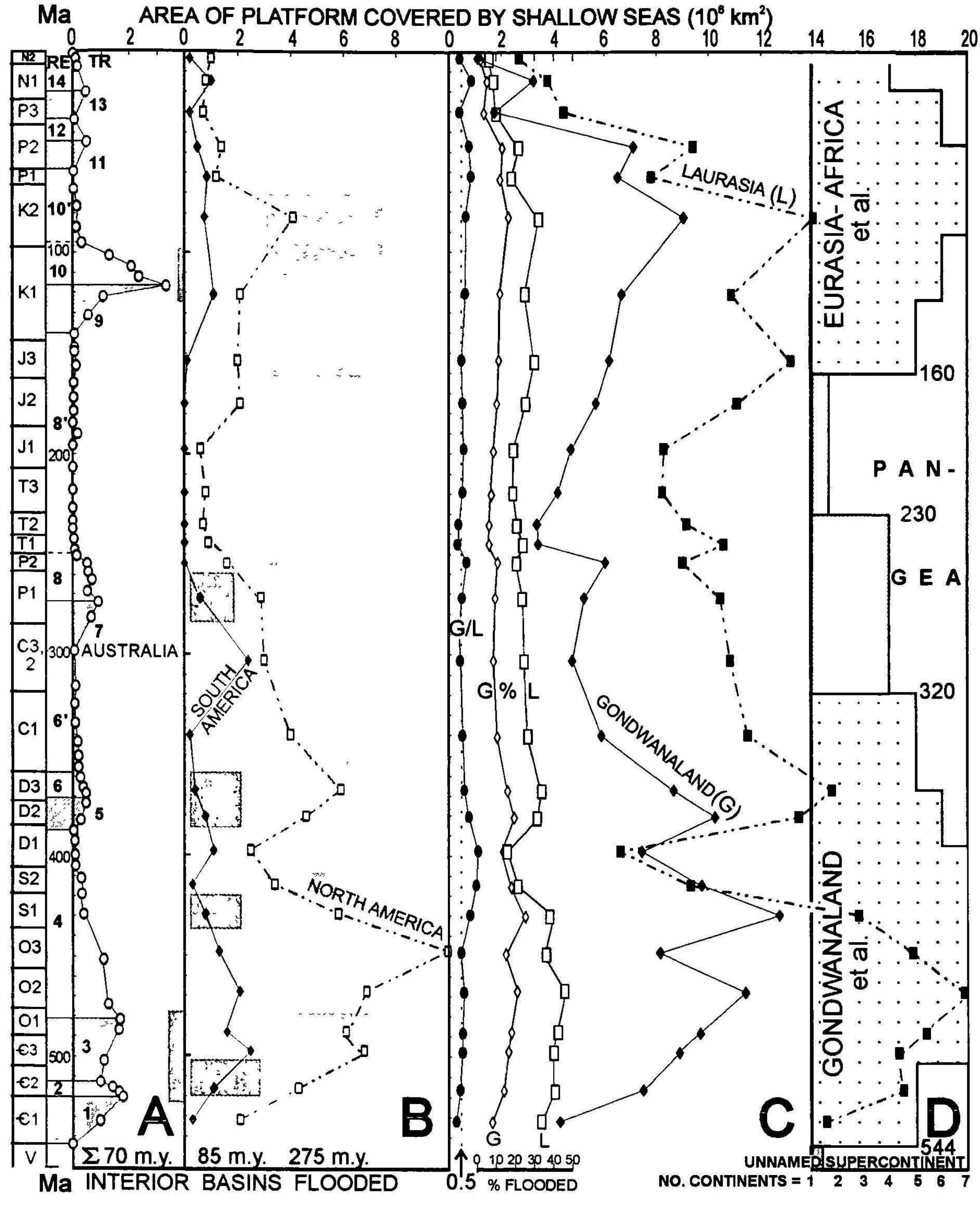


Figure 1. Area of platform covered by shallow seas during the Phanerozoic eon. A: Australia, 70 intervals (from BMR Palaeogeographic Group 1990) grouped in alternately shaded transgressions (TR: odd numbers, 1–13) and unshaded regressions (RE: even numbers, 2–14; including almost total exposure — 6', 8', 10'); dark shading indicates time intervals (total = 70 m.y.) of shallow seas in Georgina–Eromanga basin (circled area in Fig. 3A). B: South America and North America, 28 intervals (from Ronov 1994) representing shallow seas in three Brazilian cratonic basins (Soares et al. 1978) for a total of 85 m.y. and in central mid-continental United States (Bunker et al. 1988) for 275 m.y. C: Laurasia and Gondwanaland (Ronov 1994, modified by Australian data in A) and their ratio (G/L); percentage area flooded (scale below) is shown by open symbols. D: number of continents (scale below), from single Vendian (V) 'unnamed supercontinent' of Dalziel et al. (1994) through Gondwanaland et al. (as in Fig. 2) to initial Pangea at 320 Ma, rifting from 230 Ma, and breakup at 160 Ma to form Eurasia–Africa et al. (Scotese & Denham 1988; Veevers 1990). Diagram from Veevers (1995), but modified to show the Early Silurian (Llandovery) sea covering a minimum area of 0.33 × 10⁶ km² in the Carnarvon Basin (represented by the middle part of the Ajana Formation; Gorter et al. 1994) and in the Canning Basin (represented by the parts of the Sahara Formation preserved after the Prices Creek Movement; Kennard et al. 1994, p. 662; Romine et al. 1994).

Gondwanaland inheritance

Veevers (1995) asserted that, because the supercontinent Gondwanaland lasted much longer than Laurasia, it was therefore hotter from its more effective insulation of internal heat and its own production of heat from radioactivity. Moreover, the Pan-African orogenic cycle, confined to Gondwanaland, augmented the heat supply. Both effects possibly generated a permanently buoyant lower crust by underplating, revealed today by a high-velocity ($V_p > 7.5$ km s^{-1}) lowermost crust.

Evidence of underplating of this age in Australia is found in the area south of the AUS 2 (Tennant Creek–Mount Isa) traverse, under the Georgina Basin. Here the lower crust has a $V_p = 7.53$ km s^{-1} (Rudnick & Fountain 1995), and contains short discontinuous vertical seismic reflection horizons, which Finlayson & Mathur (1984) interpreted as the effects of mafic underplating during one or more phases of extension. Another area of high-velocity ($V_p = 7.61$ km s^{-1}) lowermost crust is the Nullarbor Plain (AUS 3 traverse; Finlayson 1982).

The Georgina Basin contains the late Vendian (ca 570 Ma) 500 m thick tholeiitic Antrim Plateau Volcanics, which cover an area of 400 000 km^2. Together with the coeval tholeiitic Table Hill Volcanics, which cover an area of 50 000 km^2 (Veevers 1984, p. 282), they have a total preserved volume of 0.1×10^6 km^3. Xenoliths are unknown, probably because the geochemistry of the basalt (Bultitude 1976) indicates its unsuitability for sampling deep layers (S.Y. O'Reilly, Macquarie University, personal communication 1995). I postulate that the basalt is the surface expression of a lower crustal mafic underplate represented by the high-velocity zone in the lower crust in the AUS 2 and 3 traverses, and that the mafic underplate provided long-term buoyancy during subsequent Phanerozoic time. Buoyancy would have been effected by the injection of a 5 km thick mafic layer of density 3.0 $t.m^{-3}$ displacing 3.3 $t.m^{-3}$ mantle to induce an isostatic rise of 0.5 km.

Lower layers in South Australia at the eastern end of the AUS 3 (Nullarbor) traverse are dated by the U–Pb method on zircon from xenoliths; the data indicate episodic heating events at 1.6 to 1.5, 0.78, 0.62, and 0.33 Ga (Chen et al. 1994). The Neoproterozoic (0.78 and 0.62 Ga) events are within the 0.95–0.45 Ga Pan-African thermotectonic cycle (Rogers 1993). Evidence of Neoproterozoic mafic underplating is also preserved in the lower crust of Sudan and East Africa (Stern 1994). The many 0.5 Ga dates in the Australian platform belong to the terminal Pan-African event: 'Widespread thermal rejuvenation, shear zone activation and anorogenic magmatism peaked at 0.5 Ga and affected most of West Gondwana and localised regions of Gondwana' (Unrug et al. 1994, p. 440); they were followed by rapid uplift and cooling by denudation. Most of Gondwanaland (but not Laurasia) was affected by this terminal Pan-African (0.5 Ga) uplift and cooling. These and other contrasts with Laurasia are shown in Table 1.

Pangean cycle A tectono-sedimentary stages of subsidence

The Pangean stages (Fig. 2, columns X [encircled numbers 1 to 5] and XI [sequences on the Gondwanaland platform]) are as follows:

- *stage 1 (320–290 Ma),* represented on the platform by a stratigraphic gap or lacuna, reflects the initial accumulation of heat beneath the Pangean insulator and thermal uplift;
- *stage 2 (290–230 Ma),* represented by the epi-Carboniferous to mid-Triassic early Gondwana sequence and equivalents, marks the local thinning of the Pangean crust to form broad basins or sags and, locally (on orogens), bimodal volcanic rifts generated by the initial withdrawal of heat from beneath Pangea;
- *stage 3 (230–160 Ma),* represented by the mid-Triassic (Carnian) to mid-Jurassic rift sequence, reflects a faster withdrawal of heat and concomitant crustal thinning along the incipient rifted margins of Pangea;
- *stage 4 (160–85 Ma),* represented by the mid-Jurassic to mid-Cretaceous drift sequence, reflects the wholesale loss of Pangean heat by fast sea-floor spreading; and
- *stage 5 (85 Ma to present),* represented by the Late Cretaceous and younger drift sequence, reflects a slower loss of heat from the depleted Pangean heat store.

Stages 3 to 5 (boxed numbers) of the previous cycle are sketched in column X of Figure 2.

Comparative Pangean episodes 200–0 Ma and 600–400 Ma

Figure 3 shows in plan the Cretaceous transgressive–regressive episode (events 9, 10, and 10' in Fig. 1A) in Pangean cycle A, and Figure 4 the early Palaeozoic events (1–4 in Fig. 1A) of the preceding 400 m.y. cycle B. These and other events are listed in Table 2 and shown in Figure 5, the time–space diagrams of 200–0 Ma and, offset 400 m.y., 600–400 Ma. The episodes constitute parts of Pangean stages 3 (extension II), all of 4 (fast sea-floor spreading), and parts of 5 (slow sea-floor spreading; Fig. 2, column X) during the transition from Pangea to the dispersed continents.

Gross geography

The Cretaceous episode (Fig. 3) involved an Australian platform wider than the initial state in the Cambrian, owing to the Palaeozoic accretion east of the Tasman Line. The area available for flooding was therefore greater, and the northern and western passive margins became open to the ocean, so that the sea encroached from the north and west as well as

Table 1. Contrasting lithospheric features of Gondwanaland and Laurasia–Eurasia.

Character	*Gondwanaland*	*Laurasia–Eurasia*
Time span	1100–160 Ma	320–160 Ma
Duration as a supercontinent	1000 m.y.	< 200m.y.
Elevation	high	low
Facies	non-marine[a]	marine[b]
Terminal Pan-African event		
Peak K–Ar date	500 Ma	–
Oldest apatite fission-track date	500 Ma	1200 Ma
(A. Gleadow, La Trobe University, personal communication 1995)		(Pan-African event not registered here)
700–500-Ma tectonic events		
Intensity	momentous: Pan-African	slight
Lowermost crustal layer $V_{p(c)}$ ~7.5 km s^{-1} (Veevers 1995, fig. 3)	?common	?rare
Mechanism	mafic underplating	–

[a] Exemplified by rare, intermittently deposited carbonate.
[b] Exemplified by common, almost continuously deposited carbonate.

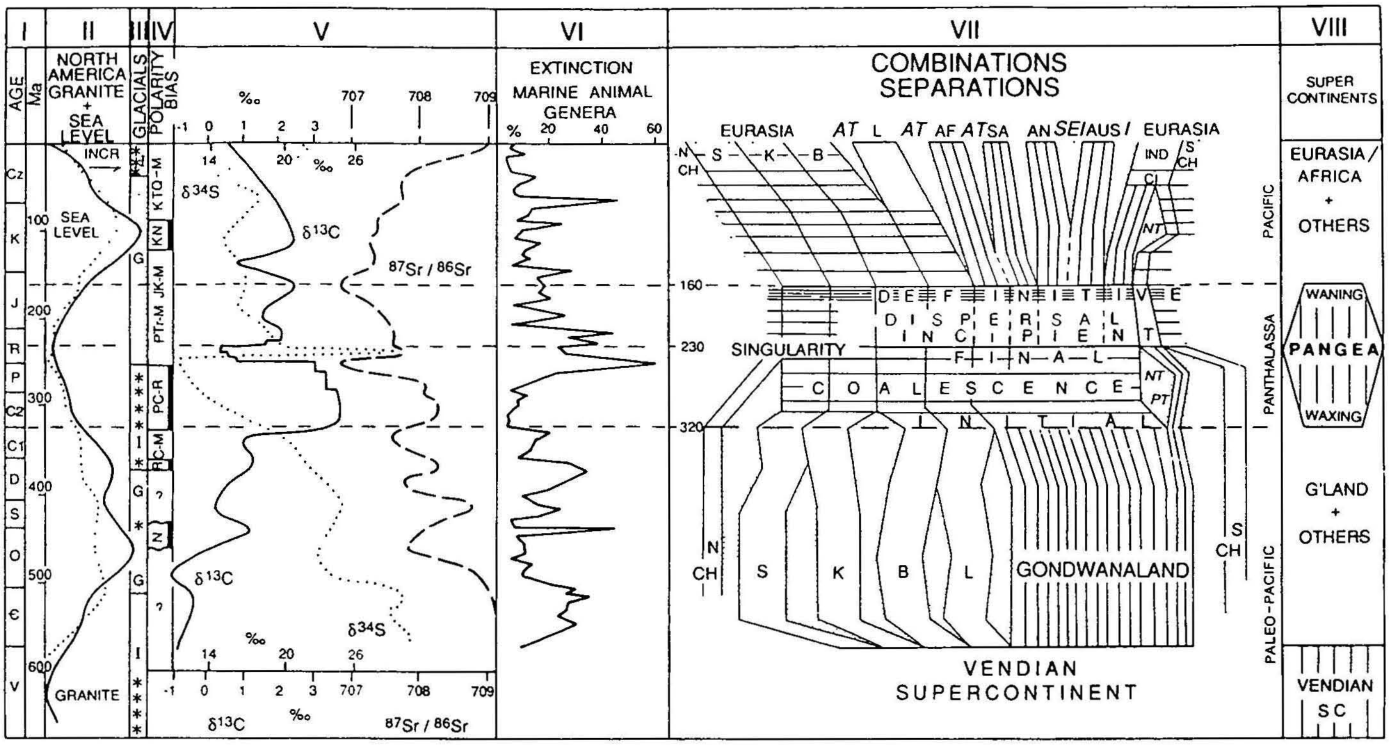


Figure 2. Phanerozoic sedimentary and tectonic indices, combinations and separations of the continents and oceans, supercontinents, flood lava, the Pangean heat anomaly, platform sequences, flooded area, and extension/shortening events, from a modified and augmented

Columns I and XV: **Timescale (Palmer 1983) with the Permian–Triassic boundary changed from 245 to 250 Ma (Veevers et al. 1994a). The Vendian/Cambrian, now regarded as 544 Ma, is referred to its previously accepted age of 570 Ma.**

Column II: **Solid line = ages of North American granites; dotted line = sea level (Fischer 1984).**

Column III: **Glacial occurrences (asterisks) with alternating icehouse (I) and greenhouse (G) states (Fischer 1984); the icehouse/greenhouse boundary is placed at the Permian–Triassic boundary.**

Column IV: **Magnetic polarity bias — N, normal; R, reversed (both with heavy bar); M, mixed (Harland et al. 1990).**

Column V: **Solid line = $\delta^{13}C$ in carbonate, and dotted line = $\delta^{34}S$ (Holser et al. 1988); broken line = $^{87}Sr/^{86}Sr$ (Koepnick et al. 1988).**

Column VI: **Percentage extinction for marine animal genera — < 260 Ma (Raup & Sepkoski 1986), > 260 Ma (Sepkoski 1992).**

Column VII: **Continental and oceanic combinations and separations (after Briden et al. 1974; Ziegler et al. 1979; Bond et al. 1984; and Sengor 1984). Gondwanaland and subsequent fragments marked with vertical and subvertical lines, Pangea and Eurasia with horizontal lines, and rift oceans with stipple; dotted lines alongside Eurasia join on either side. Continents: AF, Africa; AN, Antarctica; AUS, Australia; B, Baltica; CI, Cimmeria; IND, India; K, Kazakhstania; L, Laurentia; N CH, north China; S, Siberia; SA, South America; S CH, south China. Oceans (italicised letters): AT, Atlantic (north, central, south); I, Indian; NT, neo-Tethys; PT, paleo-Tethys; SEI, southeast Indian.**

Column VIII: **Supercontinents (vertical lines), large continents and others.**

Column IX: **Late Palaeozoic and Mesozoic flood lavas: A, European (Wopfner 1984; Ziegler 1988) and eastern Australian (Veevers 1984); B, Siberian traps (Renne & Basu 1991); C, Amazon (Mosmann et al. 1986); D, Karoo (Bristow & Saggerson 1983); E, Transantarctic and Tasmanian (Kyle et al. 1981; Schmidt & McDougall 1977); F, Serra Geral (Schobbenhaus 1984); G, Pacific ridge-crest (Watts et al. 1980); H, Rajmahal (Baksi 1988), J, Pacific mid-plate (Schlanger et al. 1981); K, Deccan (Jaeger et al. 1989).**

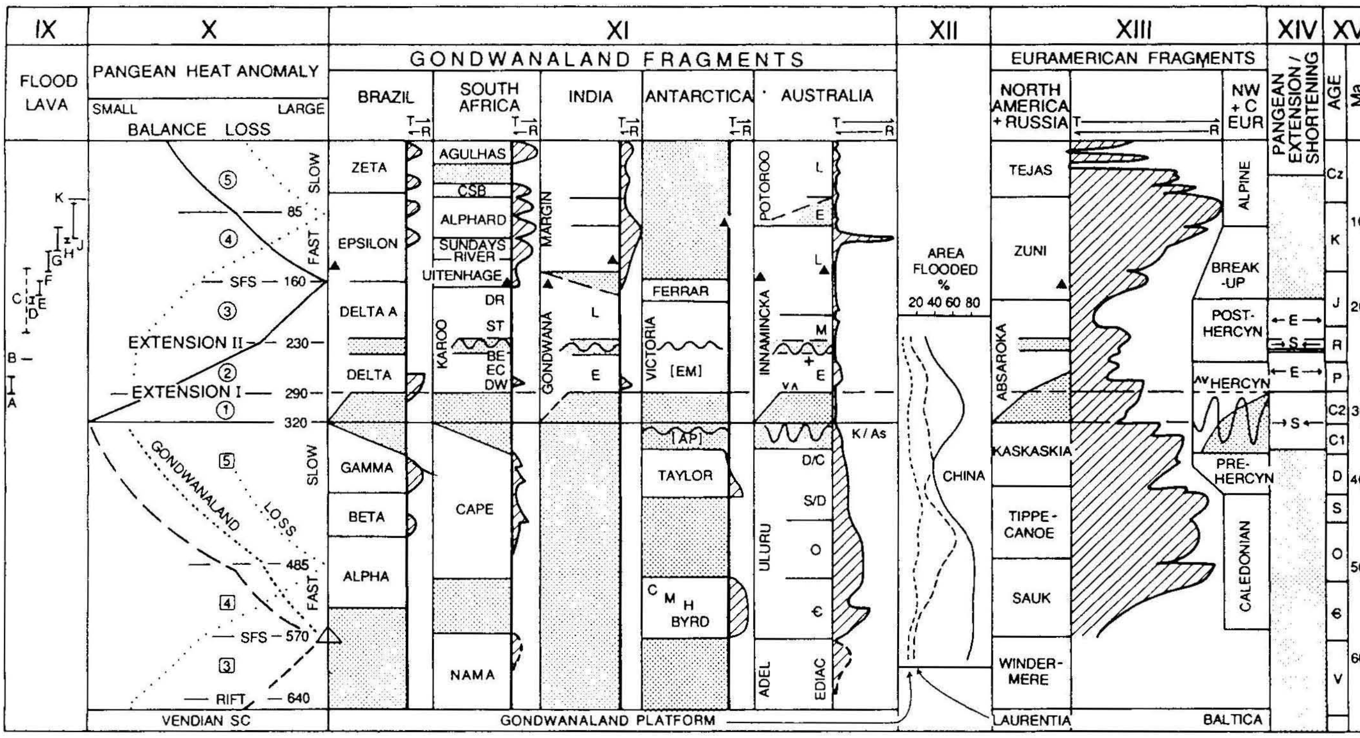
Column X: **320–0 Ma accumulation and dissipation of Pangean heat anomaly in stages 1 through 5 — the dotted line traces the rate of loss of heat that drives plate dynamics, emplaces the volcanics, and initiates basin structure; the solid line represents the residual heat or balance. 650–320 Ma — the broken line traces the accumulation and dissipation of the Vendian supercontinental heat anomaly; the dotted line represents the rate of loss of heat; the double dotted line traces the accumulation of Gondwanaland heat anomaly; the triangle denotes breakup; stages 3 to 5 predicted from Pangean (< 320 Ma) model.**

from the northeastern convergent margin. In the early Palaeozoic (Fig. 4), Australia and New Guinea lay inboard of the continental slivers of Sibumasu and west Burma in the northwest and north China in the north, so that the sea had to cross these margins on its way to the internal Bonaparte, Daly, and Georgina Basins (Fig. 4B), as suggested by the affinity of the faunas in these areas (Metcalfe 1993). A detailed palaeogeography of the regions to the northwest and north, at present contentious, may well show a distribution of land and sea similar to that in the Cretaceous. A remaining difference is the long arm of the Aptian sea in Western Australia (Fig. 3B). This area (western Officer Basin and southwestern Canning Basin), as well as the rest of Western Australia outside the Bonaparte Basin, is represented during the Cambrian by a lacuna (Fig. 5) that almost certainly reflects non-deposition.

The palaeogeography in each cycle (cf. Figs 3 & 4) changed from a marginal transgression behind a newly formed passive margin (A), to a wide transgression of an epeiric sea (B), and finally a regression that exposed almost all the continent (C).

Correlations of Australian stages in Pangean cycles A and B

Correlation of the observed events of cycle A (Jurassic and Cretaceous; Table 2, column 2) with those interpreted for cycle B (Neoproterozoic–Cambrian events; column 3) shows an offset of 400 to 415 m.y. The only exception is the end of the regression (event e), offset an anomalously short 310 m.y. from the Late Cretaceous (90 Ma) to the Early Devonian (400 Ma). Which cycle is anomalous is unknown: it could be



Column XI: **Sequences of the 570–320 Ma Gondwanaland platform and of the < 320 Ma Gondwanaland fragments, and transgressive (T) and regressive (R) curves; lacunas are shown by stipple, the onset of drift by triangles, tectonic shortening by wavy lines, and Early Permian extensional volcanics by 'V's. Brazil (Soares et al. 1978). South Africa (Dingle et al. 1983; Veevers et al. 1994c): BE, Beaufort; CSB, Cape St Blaize; DR, Drakensberg; DW, Dwyka; EC, Ecca; ST, Stormberg. India comprises the Gondwana sequence (Veevers & Tewari 1995), divided by the mid-Triassic lacuna into two parts (E, Early, and L, Late), and the overlying sequence of the eastern margin. Antarctica: H, Heritage; M, Minaret; C, Crashsite (Collinson et al. 1994); the Gondwanide shortening is confined to the Ellsworth Mountains (EM; Grunow et al. 1991) and the mid-Carboniferous shortening to the Antarctic Peninsula (AP; Milne & Millar 1989). Australia (Veevers 1984; Veevers et al. 1994b): the onset of deposition on the Pangean platform is dated at 290 Ma (epi-Carboniferous or Gzelian, east Australian palynological stage 2), and was accompanied by the eruption of thick transtensional volcanics; the cross at the Permian–Triassic boundary signifies the peak eruption of calc-alkaline ignimbrite in eastern Australia; ADEL, Adelaidean; , Cambrian; D/C, Late Devonian/Early Carboniferous; E, Early; EDIAC, Ediacarian; L, Late; M, Middle; O, Ordovician; S/D, Silurian/Middle Devonian; K/As, Kanimblan/Alice Springs deformation.**

Column XII: **Percentage area of continents flooded by the sea (Algeo & Wilkinson 1991); more detailed curves are presented in Figure 1.**

Column XIII: **Sequences of the 570–320 Ma Laurentia and Baltica, of the 320–160 Ma Pangean platform, and of the < 160 Ma Euramerican fragments, and transgressive (T) and regressive (R) curves (Vail et al. 1977); North American sequences are paralleled by those of Russia (Sloss 1976). North America: sequences from Sloss (1988); in the Permian Basin of Texas and New Mexico, the earliest Permian (286 Ma) Wolfcamp Series overlaps the Central Basin Platform (Sloss 1988, p. 265); the base of the Absaroka II subsequence (268 Ma) is marked by greatly accelerated rates of subsidence (Sloss 1988, p. 39). Northwest and central Europe: main phases of plate boundary reorganisation, from Ziegler (1988); in the Permian basins of Europe, subsidence was initiated in the Oslo (volcanic) Graben and renewed in the half-grabens between Norway and Greenland at the same time (290 Ma; Stephanian B or Gzelian) as sagging in the south was accompanied by the eruption of flood lavas and the deposition of the Rotliegende and later successions (Ziegler 1988; Veevers et al. 1994d); HERCYN, Hercynian.**

due to an anomalous second transgressive peak in cycle B at 480 Ma, or to an anomalous Australian continent-wide uplift at 90 Ma cancelling the North American (?eustatic) sea-level peak at 80 Ma (Fig. 1B).

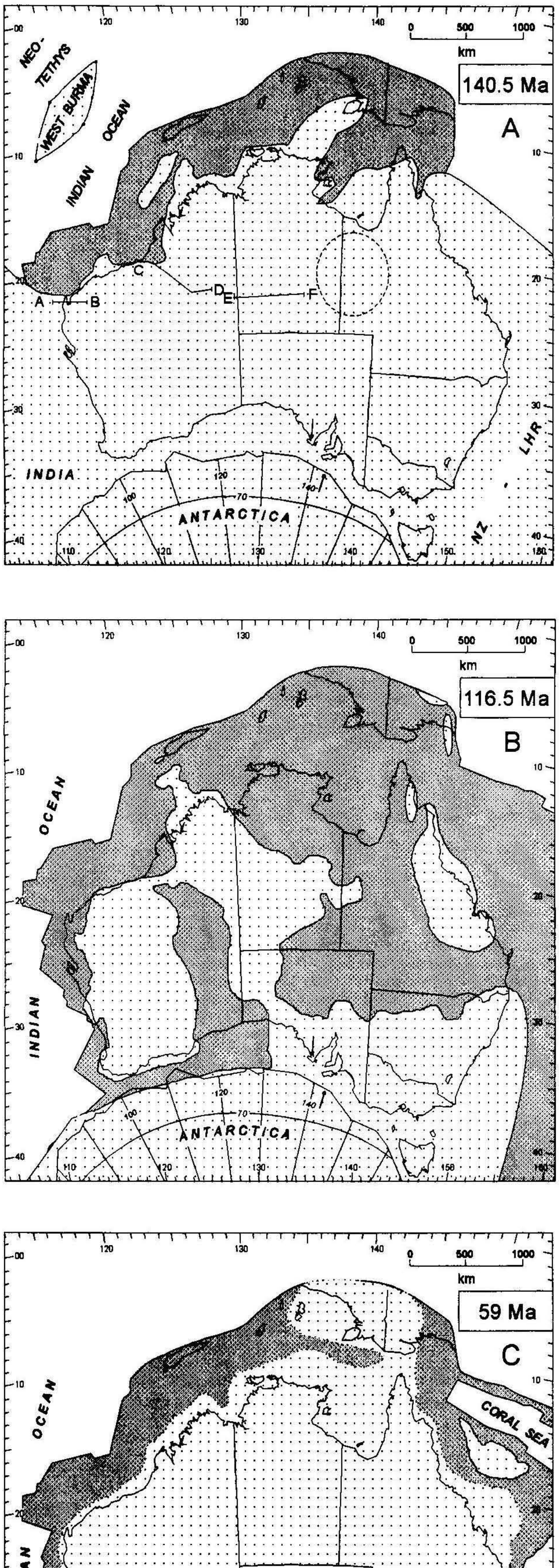
Deviant behaviour of Australia in the Late Cretaceous and Cambrian–Ordovician

Late Cretaceous

The Late Cretaceous major regression in Australia and the minor regression in South America contrast with the peak transgression in North America, seen in plan in Bond (1978, fig. 2), and in the area of platform covered by the sea (Fig. 1 A, B). The sharpest difference is between the Early Cretaceous (Aptian, 116.5 Ma) age of the maximum area of the Australian platform covered by the sea and the Late Cretaceous (Campanian, 80 Ma) age of the widest North American sea, a difference of 36.5 m.y.

Peak plutonic activity in eastern Australia and New Zealand coincided with the Australia-wide Aptian marine transgression (Veevers & Evans 1973, 1975), an example of the Haug effect: 'Times of orogeny are times of transgression of epicontinental seas on the continental interiors'. The flooding of Australia was followed, in the earliest Late Cretaceous or Cenomanian, by an Australia-wide rebound at a time of flooded platforms elsewhere. Russell & Gurnis (1994) explained the Early Cretaceous flooding of eastern Australia by subsidence generated when the dip of the slab decreased. Thereafter, an increase in the dip of the slab, causing uplift, was a consequence of Late Cretaceous back-arc spreading (Mariana-type subduction) replacing Chilean-type subduction off Queensland (Veevers 1991; Fig. 5, 90 Ma). Gallagher et al. (1994) came to a similar conclusion from backstripping and apatite fission-track analysis of eastern Australian basins.

The 3000 km width (between longitudes 152 and 122°E) of the flooded platform and succeeding exposed platform (Fig. 3), however, is not readily attributable to plate boundary effects only. The Early Cretaceous northward tilt of the land surface from the high southern rifted arch reversed to a Late Cretaceous southwesterly tilt from the newly uplifted eastern


140.5 Ma
A
NEO-TETHYS
WEST BURMA
INDIAN OCEAN
INDIA
ANTARCTICA
LHR
NZ
116.5 Ma
B
INDIAN OCEAN
ANTARCTICA
59 Ma
C
INDIAN OCEAN
CORAL SEA
TASMAN SEA
SE INDIAN OCEAN
CD
km

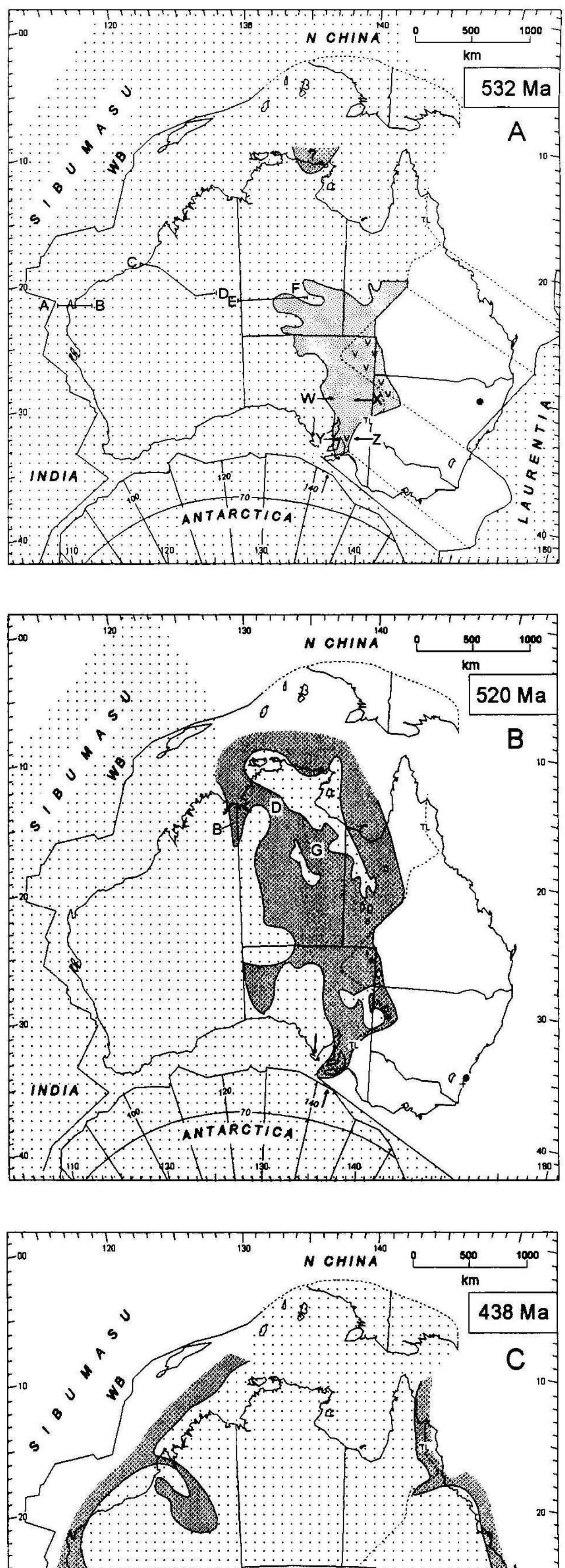

532 Ma
A
N CHINA
SIBUMASU
WB
INDIA
ANTARCTICA
LAURENTIA
520 Ma
B
N CHINA
SIBUMASU
WB
INDIA
ANTARCTICA
438 Ma
C
N CHINA
SIBUMASU
WB
INDIA
ANTARCTICA
km

Table 2. Correlation of events offset 400 Ma.

Event	*Cycle A (320–0 Ma) observed Ma*	*Cycle B (720–320 Ma) interpreted Ma*	*Difference (m.y.)*
a) NW breakup	160	575	415
b) start transgression	140.5	544	403.5
c) W breakup	130		
E breakup		544	414
d) peak transgression	116.5	520	403.5
e) end regression	90	400	310 *
f) change from Chilean to Mariana type subduction	90	490	400

* Offset <<400 m.y.

highlands and southeasterly modestly uplifted (lightly etched) Aptian surface north and west of the head of the Great Australian Bight, so that sediment was funnelled to the Ceduna depocentre (CD in Fig. 3C). Tilting on an east–west axis is not discernible; the dry land of the western half of Australia corresponds to the long-standing exposed shields of the Yilgarn and Pilbara, and the Kimberley and central Northern Territory. The platform therefore subsided and rose, apparently as a unit along a 3000 km east–west axis, across the line of subduction in the east, by a mechanism that extended the Haug effect across the entire platform.

Cambrian–Ordovician

In cycle B, Pangean tectonics possibly led to a similar outcome. Transgressive peaks at 520 and 480 Ma in Australia are offset 70 and 30 m.y. from the 450 Ma peak in North America (Fig. 1A, B). Subduction off southeastern Australia changed from Chilean to Mariana type in the latest Cambrian (490 Ma, f in Fig. 5; Powell 1984, p. 290), some 30 m.y. after the first (520 Ma, **d**, Ordian) peak. This compares with the change in subduction at 90 Ma, some 25 m.y. after the 116.5 Ma (Aptian) peak. A difference (cf. Figs 3 and 4) is that the Cambrian transgression barely extended into Western Australia, and may have been due wholly to the eastern active margin.

Conclusions

The buoyant Australian platform was inherited from Neoproterozoic and early Palaeozoic Gondwanaland, whose buoyancy resulted from mafic underplating of the crust during the complex of Pan-African events between 700 and 500 Ma. Even after mid-Jurassic breakup from the southern Gondwanaland province of Pangea, Australia remained buoyed up (in the Great Western Plateau; Veevers 1984, p. 107) by the underplated crust beneath the western two-thirds of the continent west of the Tasman Line.

Figure 3. (left-hand column) Palaeogeography of the Cretaceous transgression and regression of Australia, modified from BMR Palaeogeographic Group (1990), and extended to New Guinea from information in Audley-Charles (1984), Brown et al. (1980), Dow (1977), Harrison (1969), Pigram & Panggabean (1984), and Pigram & Davies (1987), all located in time in the cycle A column of Figure 5. Land denoted by open stipple, shallow sea by fine screen; ocean floor is clear. Adjacent continents and ocean basins from Veevers et al. (1991). Arrows in coastal Antarctica and South Australia indicate the eastern limit of known Archaean rocks (Oliver et al. 1983). NZ-LHR = New Zealand–Lord Howe Rise. A, start of the Early Cretaceous transgression, Berriasian (140.5 Ma), from BMR Palaeogeographic Group (1990) 'Cretaceous 1': the dotted ellipse about the Northern Territory–Queensland border outlines the area of the Georgina–Eromanga basin covered by the sea for a total of 70 m.y. during the Phanerozoic (column A of Fig. 1); AB–CD–EF denotes the location of Figure 5. B, peak transgression, Aptian (116.5 Ma), from BMR Palaeogeographic Group (1990) 'Cretaceous 4', modified to show the sea entering the Styx Basin on the coast of central Queensland, as indicated by Albian microplankton in the Styx Coal Measures (de Jersey 1960, p. 331–332). C, after the Late Cretaceous regression, Paleocene–early Eocene (59 Ma), from BMR Palaeogeographic Group (1990), 'Cainozoic 1'. CD = Ceduna depocentre, at the focus of centripetal drainage (not shown) in the Late Cretaceous.

Figure 4. (right-hand column) Paleogeography a (400 m.y.) cycle earlier, in the Cambrian–Ordovician–Silurian, modified from BMR Palaeogeographic Group (1990) by showing the continental terranes of southeast Asia (North China, SIBUMASU = SInica, BUrma, MAlaya, SUmatra) in their original position (Metcalfe, 1993, fig. 5), but not showing Tasmania, whose precise position during the early Palaeozoic is obscure. Located in the cycle B column of Figure 5. Land denoted by open stipple, shallow sea by fine screen; ocean floor is clear. Adjacent continents and ocean basins from Veevers et al. (1991). Arrows in coastal Antarctica and South Australia indicate the eastern limit of known Archaean rocks (Oliver et al. 1983). The shoreline transgressed from the east, except in C, where a shallow sea advanced additionally from the west; also shown is the shelf-edge break (eastern edge of shading) on the eastern (oceanward) side, marking the approximate position of the eastward-jumping continent–ocean boundary, initially formed along the Tasman Line (TL) during the earliest Cambrian (544 Ma) breakup of Laurentia from Australia–Antarctica (Bond et al. 1985; von der Borch 1980). TL = Tasman Line. A, start of the Cambrian transgression, Early Cambrian (532 Ma), from BMR Palaeogeographic Group (1990) 'Cambrian 1' or Cook's (1988) 'Cambrian 1a', but modified by regarding (i) the Antrim Plateau Volcanics and equivalents in northern Australia as older than Early Cambrian (= Ediacarian), so that the only rocks north of latitude 20°S which are possibly Early Cambrian are those barren strata beneath the Middle Cambrian (Templetonian) rocks of northern Arnhem Land and Elcho Island (Cook 1988, column 1; Bradshaw et al. 1990, p. 114); and (ii), a modification from Cook (1988), the Ordian stage as Middle Cambrian (cf. Shergold 1995), calibrated as 520 Ma. Early Cambrian formations are given in Walter & Veevers (1996, fig. 2), and include the Chandler Formation in the Amadeus Basin, whose extent is taken from Wells et al. (1970, p. 54); the volcanics (v), from Preiss (1987, fig. 104), were erupted probably in a back-arc basin, on the side of the ocean that opened (at an assumed 7 cm y^{-1}) between Australia–Antarctica and Laurentia; the filled circle denotes an ophiolite with 530 ± 6 Ma zircons from upper Bingara, northeastern New South Wales (Aitchison et al. 1992), brought to this location by subsequent subduction; the conjugate margin of Laurentia is sketched (as dry land) on the other side of the juvenile palaeo-Pacific Ocean. AB–CD–EF, WX, and YZ denote the location of Figure 5. B, peak Cambrian transgression, Ordian (520 Ma), from BMR Palaeogeographic Group (1990) 'Cambrian 2', but modified by eliminating the Babbagoola beds of the western Officer Basin (Cook 1988, column 19), now regarded as Ediacarian (Walter & Veevers 1996, fig. 4). The filled circle denotes Middle Cambrian marine fossils in a basaltic breccia of the Wagonga beds south of Batemans Bay, interpreted as part of a seamount incorporated as an exotic terrane within an accretionary prism during Middle to Late Ordovician subduction (Bischoff & Prendergast 1987); the cross denotes similar but barren rocks at Narooma (Miller & Gray 1995); B, Bonaparte; D, Daly; G, Georgina. C, end of the early Palaeozoic transgression, Early Silurian or Llandovery (438 Ma), modified from BMR Palaeogeographic Group (1990) 'Silurian 1' from information about the intertidal Sahara Formation in the Canning Basin (Romine et al. 1994) and the middle Ajana Formation in the Carnarvon Basin (Gorter et al. 1994).

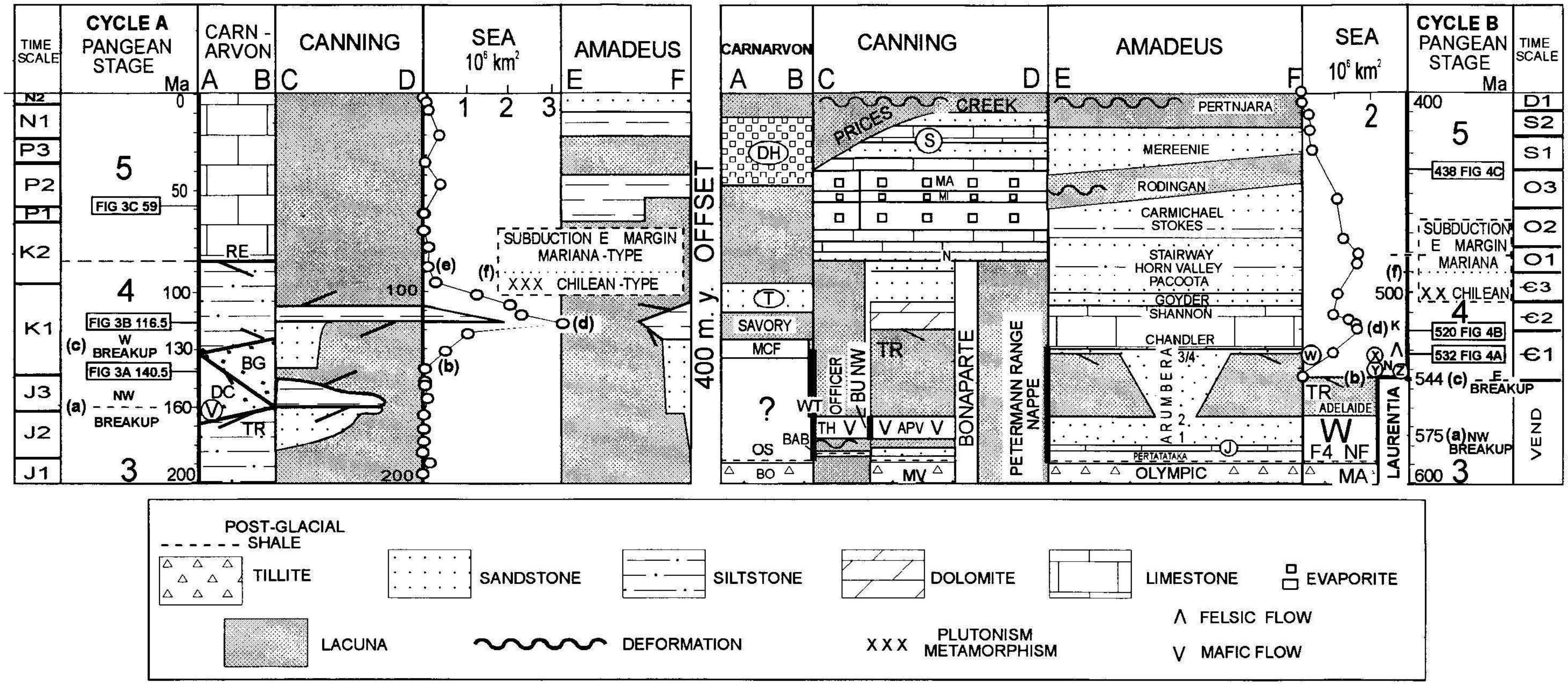


Figure 5. Time–space diagrams from 200 to 0 Ma in Pangean cycle A and, offset 400 m.y., from 600 to 400 Ma in cycle B, along the 1900 km line of AB–CD–EF, and another 100 km along lines WX and YZ (in lower part of 'SEA' column), located in Figures 3A and 4A; modified (and added to) from figure 4 and appendix of Walter & Veevers (1996). Area of platform sea from Figure 1A. In the columns of Pangean cycles and stages (Veevers 1990, 1994), boxed annotations indicate the ages of time-slices represented in Figures 3 and 4, wherein A is at the start and B is at the peak of the transgression and C is at different intervals within the regression. RE = regression; TR = transgression. *Carnarvon Basin (AB), cycle A*: BG = Barrow Group; DC = Dingo Claystone; encircled V at 160 Ma = Wandagee province of picritic diatremes coincident with the age of breakup on the northwest. *Cycle B*: DH = Dirk Hartog Formation; T = Tumblagooda Sandstone; *Savory Basin*: MCF = McFadden Formation; BO = Boondawari Formation diamictite; OS = overlying siltstone. *Canning Basin (CD), cycle A*: shallow transgression at 160 Ma, and peak at 116.5 Ma. *Cycle B*: MA = Mallowa Salt; MI = Minjoo Salt; N = Nambeet Formation; S = Sahara Formation; this sequence succeeds Ordian to earliest Ordovician marine rocks in the *Bonaparte Basin*, which succeed, in turn, the Antrim Plateau Volcanics (APV) and the Moonlight Valley Tillite (MV) and, in the *Officer Basin*, the equivalent Table Hill Volcanics (TH) and Babbagoola beds (BAB); the volcanics are interpreted as indicating breakup to the northwest (BU NW); on the northern and eastern edge of the Officer Basin, the Musgrave Block was uplifted at 590 Ma during movement along the Woodroffe Thrust (WT) in the south and the Petermann Range Nappe, which borders the southwestern part of the Amadeus Basin. *Amadeus Basin (EF), cycle A*: scattered non-marine sediment was deposited intermittently in the Cainozoic on either side of an Oligocene lacuna (Senior et al. 1995); the area was encroached by the sea at the peak of the Aptian (116.5 Ma) transgression (Veevers 1995); in box (bounded by broken line) — change from Chilean to Mariana type subduction in Cenomanian (90 Ma; f). *Cycle B*: J = Julie Formation. *Eastern Australian margin, cycle B*: in box bounded by broken line — change from Chilean to Mariana type subduction at boundary (490 Ma) between Cambrian and Ordovician. *Adelaide, cycle B*: F4 NF at 590 Ma = facies 4 (shale) of the Nuccaleena Formation; K = Kanmantoo Group; MA = Marinoan glacials; N = Normanville Group; inverted 'V' = Truro Volcanics; W = Wilpena Group. *Laurentia* shown juxtaposed against eastern Australia until breakup about 544 Ma.

In equivalent Late Cretaceous and Late Ordovician stages of the 400 m.y. Pangean supercycle, the Australian platform deviated from the global norm by rising faster than eustatic sea level. Plate boundary events — steeper subduction, underplating of a lower plate along a divergent back-arc boundary — explain uplift in the east, which was apparent from the mid-Cretaceous in the Eastern Highlands. The Eastern Highlands were separated by the Central-Eastern Lowlands from the Great Western Plateau (Veevers, 1984, p. 107), which was buoyed up by 700 to 500 Ma underplating.

Acknowledgements

I thank Erwin Scheibner for information on eastern Australia, Geoff Davies and Russell Korsch for their comments on the manuscript, the Australian Research Council for continuing support over the past 25 years, and the Geological Society of America for allowing authors the right to reproduce extracts, including figures (specifically Figs 1 & 2), of their work published by GSA without further permission.

References

Aitchison, J.C., Ireland, T.R., Blake, M.C. Jr & Flood, P.G., 1992. 530 Ma zircon age for ophiolite from the New England Orogen: Oldest rocks known from eastern Australia. Geology, 20, 125–128.

Algeo, T.J. & Wilkinson, B.H., 1991. Modern and ancient continental hypsometries. Journal of the Geological Society of London, 148, 643–653.

Anderson, D.L., 1982. Hotspots, polar wander, Mesozoic convection and the geoid. Nature, 297, 391–393.

Audley-Charles, M.G., 1984. Evolution of the southern margin of Tethys (North Australian region) from Early Permian to Late Cretaceous. In: Audley-Charles, M.G. & Hallam. A. (editors), Gondwana and Tethys. Geological Society of London, Special Publication 37, 79–100.

Baksi, A.K., 1988. Reply to comment on 'Critical evaluation of the age of the Deccan Traps, India: implications for flood-basalt volcanism and faunal extinctions'. Geology, 16, 758–759.

Bischoff, G.C.O. & Prendergast, E.I., 1987. Newly discovered Middle and Late Cambrian fossils from the Wagonga beds of New South Wales, Australia. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen 175, 39–64.

BMR Palaeogeographic Group, 1990. Evolution of a continent. Bureau of Mineral Resources, Australia.

Bond, G., 1978. Speculations on real sea-level changes and vertical motions of continents at selected times in the Cretaceous and Tertiary Periods. Geology, 6, 247–250.

Bond, G.C., Nickerson, P.A. & Kominz, M.A., 1984. Breakup of a supercontinent between 625 Ma and 555 Ma: new evidence and implications for continental histories. Earth and Planetary Science Letters, 70, 325–345.

Bond, G.C., Christie-Blick, N., Kominz, M.A. & Devlin, W.J., 1985. An Early Cambrian rift to post-rift transition in the Cordillera of western North America. Nature, 315, 742–746.

Briden, J.C., Drewry, G.E. & Smith, A.G., 1974. Phanerozoic equal-area world maps. Journal of Geology, 82, 555–574.

Bristow, W. & Saggerson, E.P., 1983. A general account of Karoo vulcanicity in southern Africa. Geologische Rundschau, 72, 1015–1059.

Bradshaw, J., Nicoll, R.S. & Bradshaw, M., 1990. The Cambrian to Permo-Triassic Arafura Basin, northern Australia. APEA Journal, 30, 107–127.

Brown, C.M., Pigram, C.J. & Skwarko, S.K., 1980. Mesozoic stratigraphy and geological history of Papua New Guinea. Palaeogeography, Palaeoclimatology, Palaeoecology, 29, 301–322.

Bultitude, R. J., 1976. Flood basalts of probable Early Cambrian age in northern Australia. In: Johnson, R.W. (editor), Volcanism in Australasia. Elsevier, Amsterdam, 1–20.

Bunker, B.J., Witzke, B.J., Watney, W.L. & Ludvigson, G.A., 1988. Phanerozoic history of the central midcontinent, United States. In: Sloss, L.L. (editor), Sedimentary cover — North American Craton, United States. Geological Society of America, Boulder, Colorado, The Geology of North America, D-2, 243–260.

Calver, C.R., 1995. Ediacarian isotope stratigraphy of Australia. PhD thesis, Macquarie University (unpublished).

Chen, Y.D., O'Reilly, S.Y., Kinny, P.D., & Griffin, W.L., 1994. Dating lower crust and upper mantle events: an ion probe study of xenoliths from kimberlite pipes, South Australia. Lithos, 32, 77–94.

Collinson, J.W., Isbell, J.L., Elliot, D.H., Miller, M.F. & Miller, J.M.G., 1994. Permian–Triassic Transantarctic basin. In: Veevers, J.J. & Powell, C.McA. (editors), Permian–Triassic basins and foldbelts along the Panthalassan margin of Gondwanaland. Geological Society of America, Memoir 184, 173–222.

Cook, P.J., 1988. Palaeogeographic Atlas of Australia, Vol. 1 — Cambrian. Bureau of Mineral Resources, Australia.

Dalziel, I.W.D., Dalla Salda, L.H. & Gahagan, L.M., 1994. Paleozoic Laurentia–Gondwana interaction and the origin of the Appalachian–Andean mountain system. Geological Society of America, Bulletin 106, 243–252.

de Jersey, N.J., 1960. The Styx Coal Measures. Journal of the Geological Society of Australia, 7, 330–333.

Dingle, R.V., Siesser, W.G. & Newton, A.R., 1983. Mesozoic and Tertiary geology of Southern Africa. Balkema, Rotterdam.

Dow, D.B., 1977. A geological synthesis of Papua New Guinea. Bureau of Mineral Resources, Australia, Bulletin 201.

Finlayson, D.M., 1982. Geophysical differences in the lithosphere between Phanerozoic and Precambrian Australia. Tectonophysics, 84, 287–312.

Finlayson, D.M. & Mathur, S.P., 1984. Seismic refraction and reflection features of the lithosphere in northern and eastern Australia, and continental growth. Annales Geophysicae, 2, 711–722.

Fischer, A.G., 1984. The two Phanerozoic supercycles. In: Berggren, W.A. & van Couvering, J.A. (editors), Catastrophes and Earth history. Princeton University Press, New Jersey, 129–150.

Gallagher, K., Dumitru, T.A. & Gleadow, J.W., 1994. Constraints on the vertical motion of eastern Australia during the Mesozoic. Basin Research, 6, 77–94.

Gorter, J.D., Nicoll, R.S. & Foster, C.B., 1994. Lower Palaeozoic facies in the Carnarvon Basin, Western Australia: stratigraphy and hydrocarbon prospectivity. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Petroleum Exploration Society of Australia Ltd, Western Australian Branch, Perth, 373–396.

Grunow, A.M., Kent, D.V. & Dalziel, I.W.D., 1991. New paleomagnetic data from Thurston Island: implications for the tectonics of West Antarctica and Weddell Sea opening. Journal of Geophysical Research, 96, 17935–17954.

Gurnis, M., 1988. Large-scale mantle convection and the aggregation and dispersal of supercontinents. Nature,

332, 695–699.
Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G. & Smith, D.G., 1990. A geological time scale 1989. Cambridge University Press.
Harrison, J., 1969. A review of the sedimentary history of the island of New Guinea. Australian Petroleum Exploration Association, Journal 9, 41–48.
Holmes, A., 1931. Radioactivity and earth movements. Transactions of the Geological Society of Glasgow, 18, 559–606.
Holser, W.T., Schidlowski, M., Mackenzie, F.T. & Maynard, J.B., 1988. Biogeochemical cycles of carbon and sulfur. In: Gregor, C.B., Garrels, R.M., Mackenzie, F.T. & Maynard, J.B. (editors), Chemical cycles in the evolution of the Earth. John Wiley & Sons, New York, 105–173.
Jaeger, J.-J., Courtillot, V. & Tapponnier, P., 1989. Paleontological view of the ages of the Deccan Traps, the Cretaceous/Tertiary boundary, and the India–Asia collision. Geology, 17, 316–319.
Kennard, J.M., Jackson, M.J., Romine, K.K., Shaw, R.D. & Southgate, P.N., 1994. Depositional sequences and associated petroleum systems of the Canning Basin, W.A. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Petroleum Exploration Society of Australia Ltd, Western Australian Branch, Perth, 657–676.
Koepnick, R.B., Denison, R.E. & Dahl, D.A., 1988. The Cenozoic seawater $^{87}Sr/^{86}Sr$ curve: data review and implications for correlation of marine strata. Paleoceanography, 3, 743–756.
Kyle, P.R., Elliot, D.H. & Sutter, J.F., 1981. Jurassic Ferrar Supergroup tholeiites from the Transantarctic Mountains, Antarctica, and their relationship to the initial fragmentation of Gondwana. In: Cresswell, M.M. & Vella, P. (editors), Gondwana five. Balkema, Rotterdam, 283–287.
Metcalfe, I. 1993. Southeast Asian terranes: Gondwanaland origins and evolution. In: Findlay, R.H. et al. (editors), Gondwana eight. Balkema, Rotterdam, 181–200.
Miller, J.M. & Gray, D.R., 1995. Subduction and sediment accretion by underplating in the eastern Lachlan Fold Belt. Geological Society of Australia, Abstracts 40, 109–110.
Milne, A.J. & Millar, I.L., 1989. The significance of mid-Palaeozoic basement in Graham Land, Antarctic Peninsula. Journal of the Geological Society of London, 146, 207–210.
Morante, R., 1995. Permian–Triassic stable isotope stratigraphy of Australia. PhD thesis, Macquarie University (unpublished).
Mosmann, R., Falkenhein, F.U.H., Goncalves, A. & Nepomuceno, F., 1986. Oil and gas potential of the Amazon Paleozoic basins. In: Halbouty, M.T. (editor), Future petroleum provinces of the world. American Association of Petroleum Geologists, Memoir 40, 207–241.
Oliver, R.L., Cooper, J.A. & Truelove, A.J., 1983. Petrology and zircon geochronology of Herring Island and Commonwealth Bay and evidence for Gondwana reconstruction. In: Oliver, R.L., James, P.R. & Jago, J.B. (editors), Antarctic Earth science. Australian Academy of Science, Canberra, 64–68.
Palmer, A.R., 1983. The Decade of North American Geology 1983 geologic time scale. Geology, 11, 503–504.
Pigram, C J & Davies, H.L., 1987. Terranes and the accretion history of the New Guinea orogen. BMR Journal of Australian Geology & Geophysics, 10, 193–211.
Pigram, C.J. & Panggabean, H. 1984. Rifting of the northern margin of the Australian continent and the origin of some microcontinents in eastern Indonesia. Tectonophysics, 107, 331–353.
Powell, C.McA., 1984. Uluru regime. In: Veevers, J.J. (editor), Phanerozoic Earth history of Australia. Clarendon, Oxford, 290–340.
Preiss, W.V., 1987. The Adelaide Geosyncline — Late Proterozoic stratigraphy, sedimentation, palaeontology, and tectonics. Geological Survey of South Australia, Bulletin 53.
Raup, D.M. & Sepkoski, J.J., 1986. Periodic extinction of families and genera. Science, 231, 833–836.
Renne, P.R. & Basu, A.R., 1991. Rapid eruption of the Siberian traps flood basalts at the Permo-Triassic boundary. Science, 253, 176–179.
Rogers, J.J.W., 1993. A history of the Earth. Cambridge University Press.
Romine, K.K., Southgate, P.N., Kennard, J.M. & Jackson, M.J. 1994. The Ordovician to Silurian phase of the Canning Basin WA: structure and sequence evolution. In: Purcell, P.G. & Purcell, R.R. (editors), The sedimentary basins of Western Australia. Petroleum Exploration Society of Australia Ltd, Western Australian Branch, Perth, 373–396.
Ronov, A.B., 1994. Phanerozoic transgressions and regressions on the continents: a quantitative approach based on areas flooded by the sea and areas of marine and continental deposition. American Journal of Science, 294, 777–801.
Rudnick, R.L. & Fountain, D.M., 1995. Nature and composition of the continental crust: a lower crustal perspective. Reviews of Geophysics, 33, 267–310.
Russell, M. & Gurnis, M., 1994. The planform of epeirogeny: vertical motions of Australia during the Cretaceous. Basin Research, 6, 63–76.
Schlanger, S.O., Jenkyns, H.C. & Premoli-Silva, I., 1981. Volcanism and vertical tectonics in the Pacific basin related to global Cretaceous transgressions. Earth and Planetary Science Letters, 52, 435–449.
Schmidt, P.W. & McDougall, I., 1977. Palaeomagnetic and potassium–argon studies of the Tasmanian dolerites. Geological Society of Australia, Journal 24, 321–328.
Schobbenhaus, C. (editor), 1984. Geologia do Brasil. Departamento Nacional da Productaçao Mineral, Brasilia.
Scotese, C.R. & Denham, C.R., 1988. Terra Mobilis: plate tectonics for the Macintosh (computer disk). Earth in Motion Technologies, Austin, Texas.
Sengor, A.M.C., 1984. The Cimmeride orogenic system and the tectonics of Eurasia. Geological Society of America, Special Paper 195.
Senior, B.R., Truswell, E.M., Idnurm, M., Shaw, R.D. & Warren, R.G. 1995. Cainozoic sedimentary basins in the eastern Arunta Block, Alice Springs region, central Australia. AGSO Journal of Australian Geology & Geophysics, 15, 421–444.
Sepkoski, J.J., 1992. Major bioevents during the Paleozoic Era: a view from global taxonomic data bases. International Geological Correlation Program, Project 303, Late Precambrian and Cambrian event stratigraphy, Newsletter 5, 25–26.
Shergold, J.H., 1995. Australian Phanerozoic timescales — biostratigraphic charts and explanatory notes, second series: 1. Cambrian. Australian Geological Survey Organisation, Record 1995/30.
Sloss, L.L., 1976. Areas and volumes of cratonic sediments, western North America and eastern Europe. Geology, 4, 272–276.
Sloss, L.L. (editor), 1988. Sedimentary cover — North American Craton, United States. Geological Society of America, Boulder, Colorado, The Geology of North America, D-2.

Soares, C.S., Landim, P.M.B. & Fulfaro, V.J., 1978. Tectonic cycles and sedimentary sequences in the Brazilian intracratonic basins. Geological Society of America, Bulletin 89, 181–191.

Stern, R.J., 1994. Arc assembly and continental collision in the Neoproterozoic East African Orogen. Annual Review of Earth and Planetary Sciences, 22, 319–351.

Unrug, R., Gresse, P. & Wolmarans, L., 1994. Geodynamic map of Gondwana supercontinent assembly. Geological Society of Australia, Abstracts 37, 440.

Vail, P.R., Mitchum, R.M. & Thompson, S., 1977. Seismic stratigraphy and global changes of sea level; part 4: global cycles of relative changes of sea level. American Association of Petroleum Geologists, Memoir 26, 83–97.

Veevers, J.J. (editor), 1984. Phanerozoic Earth history of Australia. Clarendon, Oxford.

Veevers, J.J., 1990. Tectonic–climatic supercycle in the billion-year plate-tectonic eon: Permian Pangean icehouse alternates with Cretaceous dispersed-continents greenhouse. Sedimentary Geology, 68, 1–16.

Veevers, J.J., 1991. Mid-Cretaceous tectonic climax, Late Cretaceous recovery, and Cenozoic relaxation in the Australian region. Geological Society of Australia, Special Publication 18, 1–14.

Veevers, J.J. 1994. Pangea evolution of a supercontinent and its consequences for Earth's paleoclimate and sedimentary environments. In: Klein, G.D. (editor), Pangea: paleoclimate, tectonics, and sedimentation during accretion, zenith, and breakup of a supercontinent. Geological Society of America, Special Paper 288, 13–23.

Veevers, J.J., 1995. Emergent long-lived Gondwanaland, submergent short-lived Laurasia: supercontinental and Pan-African heat imparts long-term buoyancy by mafic underplating. Geology, 23, 1131–1134.

Veevers, J.J. & Evans, P.R., 1973. Sedimentary and magmatic events in Australia and the mechanism of world-wide Cretaceous transgressions. Nature Physical Science, 245, 33–36.

Veevers, J.J. & Evans, P.R., 1975. Late Palaeozoic and Mesozoic history of Australia. In: Campbell, K.S.W. (editor), Gondwana geology. Australian National University Press, Canberra, 579–607.

Veevers, J.J. & Tewari, R.C., 1995. Gondwana master basin of peninsular India between Tethys and the interior of the Gondwanaland province of Pangea. Geological Society of America, Memoir 187.

Veevers, J.J., Powell, C. McA. & Roots, S.R., 1991. Review of seafloor spreading around Australia: I. Synthesis of the patterns of spreading. Australian Journal of Earth Sciences, 38, 373–389.

Veevers, J.J., Conaghan, P.J. & Shaw, S.E., 1994a. Turning point in Pangean environmental history at the Permian/Triassic (P/Tr) boundary. In: Klein, G.D. (Editor), Pangea: paleoclimate, tectonics, and sedimentation during accretion, zenith, and breakup of a supercontinent. Geological Society of America, Special Paper 288, 187–196.

Veevers, J.J., Conaghan, P.J. & Powell, C.McA., 1994b. Eastern Australia. In: Veevers, J.J. & Powell, C.McA. (editors), Permian–Triassic basins and foldbelts along the Panthalassan margin of Gondwanaland. Geological Society of America, Memoir 184, 11–171.

Veevers, J.J., Cole, D.I., & Cowan, E.J., 1994c. Southern Africa: Karoo Basin and Cape Fold Belt. In: Veevers, J.J. & Powell, C.McA. (editors), Permian–Triassic basins and foldbelts along the Panthalassan margin of Gondwanaland. Geological Society of America, Memoir 184, 223–279.

Veevers, J.J., Clare, A. & Wopfner, H., 1994d. Neocratonic magmatic–sedimentary basins of post-Variscan Europe and post-Kanimblan eastern Australia generated by right-lateral transtension of Permo-Carboniferous Pangea. Basin Research, 6, 141–157.

von der Borch, C.C., 1980. Evolution of Late Proterozoic to early Palaeozoic Adelaide Foldbelt, Australia: comparisons with post-Permian rifts and passive margins. Tectonophysics, 70, 115–134.

Walter, M.R. & Veevers, J.J., 1997. Australian Neoproterozoic palaeogeography, tectonics, and supercontinental connections. AGSO Journal of Australian Geology & Geophysics (this issue).

Watts, A.B., Bodine, J.H. & Ribe, N.M., 1980. Observations of flexure and the geological evolution of the Pacific Ocean basin. Nature, 283, 532–537.

Wells, A.T, Forman, D.J., Ranford, L.C. & Cook, P.J., 1970. Geology of the Amadeus Basin, central Australia. Bureau of Mineral Resources, Australia, Bulletin 100.

Wopfner, H., 1984. Permian deposits of the southern Alps as product of initial alpidic taphrogenesis. Geologische Rundschau, 73, 259–277.

Ziegler, A.M., Scotese, C.R., McKerrow, W.S., Johnson, M.E. & Bambach, R.K., 1979. Paleozoic paleogeography. Annual Review of Earth and Planetary Science, 7, 473–502.

Ziegler, P.A., 1988. Evolution of the Arctic–north Atlantic and the western Tethys. American Association of Petroleum Geologists, Memoir 43.

AGSO Journal of Australian Geology & Geophysics, 17(1), 113–125

Australia and the Melanesian arcs: a review of tectonic settings

Patrick J. Coleman[1]

Until about 1950, the southwest Pacific region was thought by many to be the foundered eastern half of Greater Australia, 'Tasmantis'. The Outer Melanesian arcs defined the northern part of the Tasmantis boundary to the Pacific Ocean basin. Others considered the Inner Melanesian arcs to be the expanded rim of the Australian craton, the outwardly propagated expression of the Tasman Geosyncline. With the surge of exploration after World War II, new data showed that the outer islands could not be matched with island-arc models such as the Sunda Arc, and that New Caledonia, of the inner arcs, seemed increasingly to have Tasman Geosyncline affinities. By 1968, the land and sea data had produced a cluttered and confused picture, ripe for the application of new global tectonic models. After 1968, the Tasman Sea was shown to be a product of Late Cretaceous seafloor spreading and New Caledonia was accepted as part of the expanded rim of eastern Gondwana. The sea floor between the inner and outer chains was relatively young and the product of complex seafloor spreading. Over recent years, the southwest Pacific has in many respects been a showcase of plate tectonic theory. The region is intensely mobile. The obliquely convergent northern Indo-Australian/Pacific plate boundary carries great crustal blocks as allochthons to join others in eastern Indonesia. These crustal blocks also undergo vertical and rotational movement. The outer island chains do not fit the conventional models of an island arc — they are hybrid entities made up of the byproducts of subduction and of exotic terranes; as arcs, they continually change composition, form and configuration, even as they are being built. These outer chains are components of what is here called an 'accommodation boundary'.

Introduction

The Inner and Outer Melanesian arcs were terms proposed first by Carey (1938), and then by Glaessner (1950), for the sub-parallel arc-type strings of islands which border Australia in the southwest Pacific. The Outer Melanesian arcs included northern coastal New Guinea, Bismarck Archipelago, Solomon Islands, New Hebrides and Fiji (Fig. 1). The Inner Melanesian arcs included parts of eastern New Guinea, Rennell–Bellona, and New Caledonia–Norfolk Island. These arc terms have endured and so this paper reviews the broader aspects of geology, geophysics and tectonics of the Melanesian arcs from the days before plate tectonics theory, essentially the 1950s and 1960s, to the present.

Umbgrove wrote his classic *Pulse of the Earth* in 1947 and the Sunda Arc became an island arc model, supplemented by the Japanese Archipelago. By an extension of thought, the outer island chains in the southwest Pacific were also supposed to be island arcs, but with peculiar features, such as the difficulty of putting them into a geosynclinal context. The inner islands, closer to Australia, did not fit any arc model, but it was recognised that these islands lay on one or another of several submarine ridges that seemed stepped-out from the east coast of Australia and connected in some way to the continent. Rifting of bits of continent to give slabs such as New Caledonia was hypothesised. All this was embodied in the concept of a foundered Tasmantis (e.g. Bryan 1944; Fairbridge 1953 — with historical reviews), with a northern and eastern boundary roughly corresponding to the 'Andesite Line'. For those few who espoused continental drift, Tasmantis was eastern Gondwana. The more adventurous geological synthesisers, in the immediate post-World War II period, considered that southern Papua, eastern New Guinea and New Caledonia were the most recent expressions of the continuing, eastern development of the Tasman Geosyncline. Geosynclines were thought to migrate outwards from the core-shield somewhat as a propagating wave-form. All parties would have found much to agree with in the reconstructions shown in Figures 2A (pre-plate tectonics) and 2B (post-plate tectonics), especially Figure 2A. The Tasman Sea was an enigma. A connection with New Zealand was conjectured.

The outer chain of islands — Solomons to Fiji — contains part of the Indo–Australian/Pacific plate boundary, a part here termed the Melanesian Boundary; these islands also make up Fairbridge's Melanesian Border Plateau or Borderlands (Fairbridge 1961). Carey, the major spokesman for the drift or mobilist camp, included them in the Tethyan Shear System (Carey 1958, 1963).

But generally speaking, the post-World War II period up to 1968 was a time of 'fixist' dominance; 'drifters' or 'mobilists' were regarded at best as amiable eccentrics, albeit talented ones, for whom the laws of physics had only passing significance. All this, despite the evidence of differential

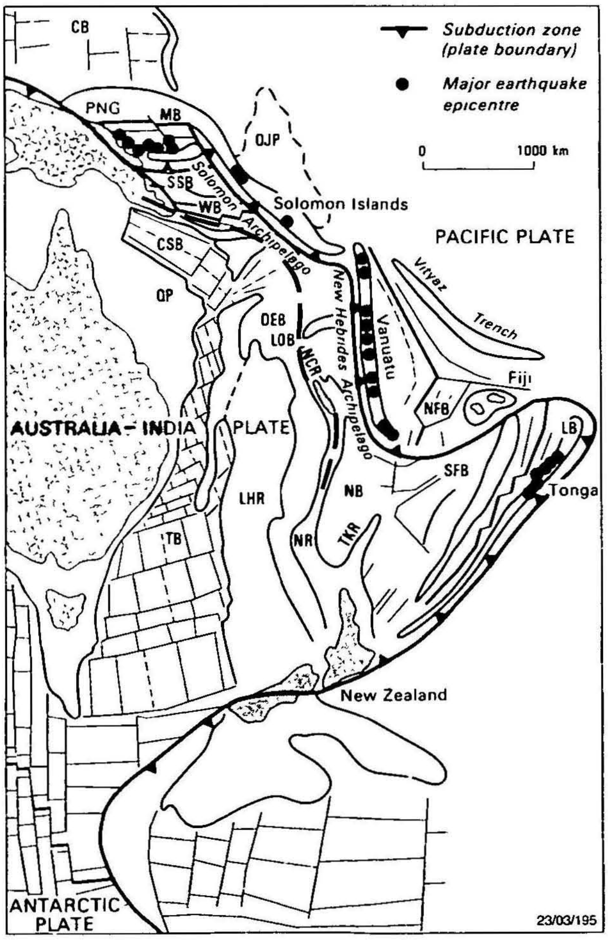


Figure 1. Main physiographic features, locations, and regional tectonic features of the SW Pacific. The Outer Melanesian arcs are defined by the northern line of epicentres New Guinea to Fiji; the Inner Melanesian arcs by the heavy dashed line (after Falvey et al. 1991).

[1] Department of Geology & Geophysics, University of Western Australia, Nedlands, WA 6009, Australia. Present address PO Box 96, Moffat Beach, Qld 4551.

strike-slip movement of several hundred kilometres along the Alpine Fault (there was also the matter of the Great Glen fault on the other side of the world). Horizontal movement of great crustal blocks was perhaps acceptable if it had taken place on land. In this regard, Umbgrove's *Pulse of the Earth* was an electrifying stimulus. Although not espousing continental drift, Umbgrove argued convincingly that the Earth was an extremely vigorous entity and that this vigour is shown in its grand architecture. The southwest Pacific offers prime support to his argument.

This is a review, essentially of tectonic setting and history, and necessarily of limited length. I have not been able to give even a summary of the geology of each and all of the many geographic entities involved. As well, literature on the southwest Pacific is voluminous (for an extensive coverage until 1983 see Jouannic & Thompson 1983). I have had to be selective in citing references. A general review of the geology, structure and tectonics of the southwest Pacific and its component territories is given by Kroenke (1984).

Following the intensive land and marine surveys of the 1950s and 1960s, the cascade of new information has highlighted particular aspects of plate tectonics theory. I have concentrated on these aspects — the notions of subduction flip, horizontal movement (translation) of great blocks, and migration of whole island chains.

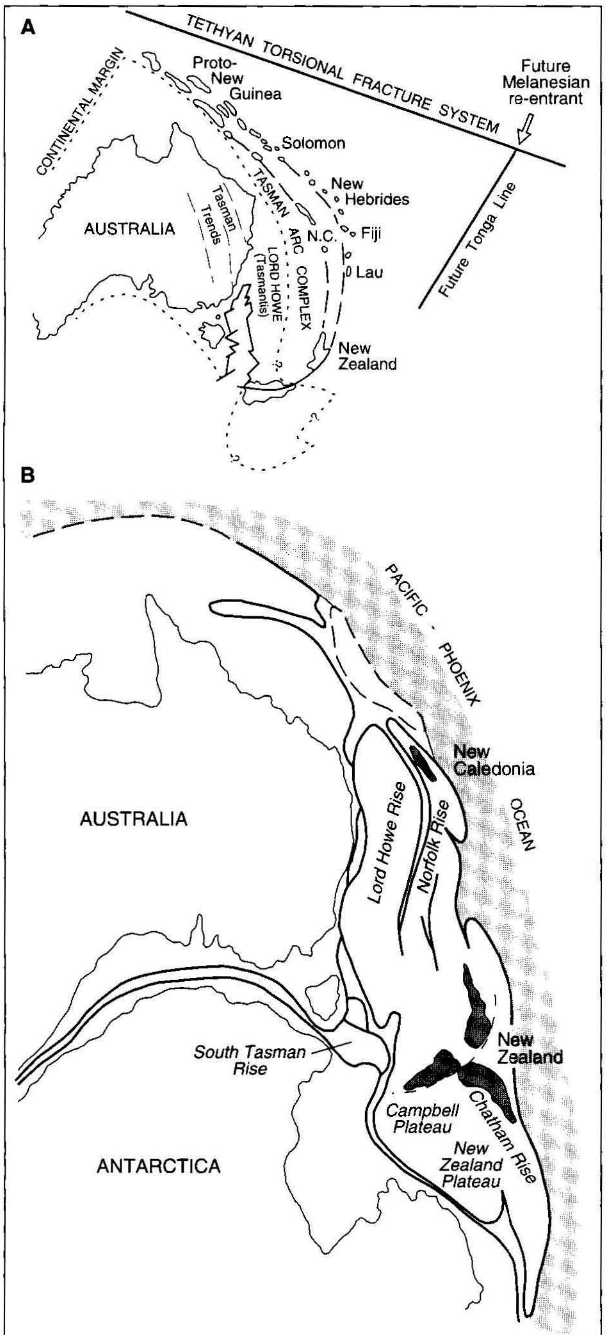


Figure 2A. A pre-plate tectonics attempt (Coleman 1967) to reconcile the land and other elements of the southwest Pacific as they were thought to have been in the earliest Tertiary. The diagram is highly speculative.

Figure 2B. A reconstruction for the Early Cretaceous, mildly speculative, and reflecting the plate tectonics view (after Griffiths 1975).

Before plate tectonics

By the end of the 1950s, land areas of both Inner and Outer Melanesian arcs had been mapped at least to reconnaissance level. Most of this new data was obtained by land surveys of the boot and hammer variety, assisted by wartime aerial photographs. In most areas, there were no maps for these early surveys, not even rough topographic ones. The geologist had to produce his own maps and include in them not only geology but all manner of useful topographic and demographic information.

Land geology

Outer Melanesia. By about 1960, the broader aspects of Solomon Islands geology (Fig. 3) did not match up closely to the Sunda arc. There was no clear distinction between 'inner' and 'outer' arcs (as there was in the East Indies: Umbgrove 1947). The profile was unlike that of the Sunda Arc — the volcanic front was too close to the trench (New Georgia Group) although some volcanoes were too far to the rear (on Choiseul). Forearc rocks were present on Choiseul, but not on Guadalcanal, an island which gave way almost precipitously to the trench. Although seismically active, the spread of hypocentres beneath the chain was diffuse and did not define a zone. The symmetry presented by the Sunda Arc was altogether lacking.

Similarly, in the New Hebrides Arc (a large part of it occupied by Vanuatu), although there was a neatly-defined Wadati-Benioff zone (with easterly dip opposed to that of the Tonga Arc), a well-placed volcanic front or axis, and the arc profile had symmetry, the symmetry was reversed. Later, this reversal was to be explained, of course, by applying the notion of polarity reversal or subduction 'flip'.

Fiji had the longest history of geological exploration, but its geology as known in the mid-1960s was confusing and, again, did not agree with the Sunda model. In particular, there were anomalous contrasts in the sequences of adjacent areas on Viti Levu, which could not be explained in terms of facies or structural controls. The pattern of lineaments was highly evocative of rotation — indeed, the Fiji platform, including the Lau Group to the southeast, was reminiscent of a spiral nebula. Sunda Arc terminology could not be applied to the Fiji platform.

At the other end of the arc system, the Bismarck Archipelago, there were also misfits. New Ireland was clearly an extension of the Solomon arc system. New Britain did not match up closely to the Solomons or to adjacent New Guinea (and still does not). There was no explanation for its volcanic front, the Quaternary Bismarck volcanic arc, which extends to the northwest as far as the Schouten Islands offshore of north coastal New Guinea. The Torricelli–Bewani block, essentially a Neogene arc sequence on northwest coastal Papua New Guinea, was thought even then to be an arc in collision with New Guinea proper, but no mechanism was proposed.

Inner Melanesia. The Louisiade Archipelago (off the tail of New Guinea), had been the target for gold prospectors for

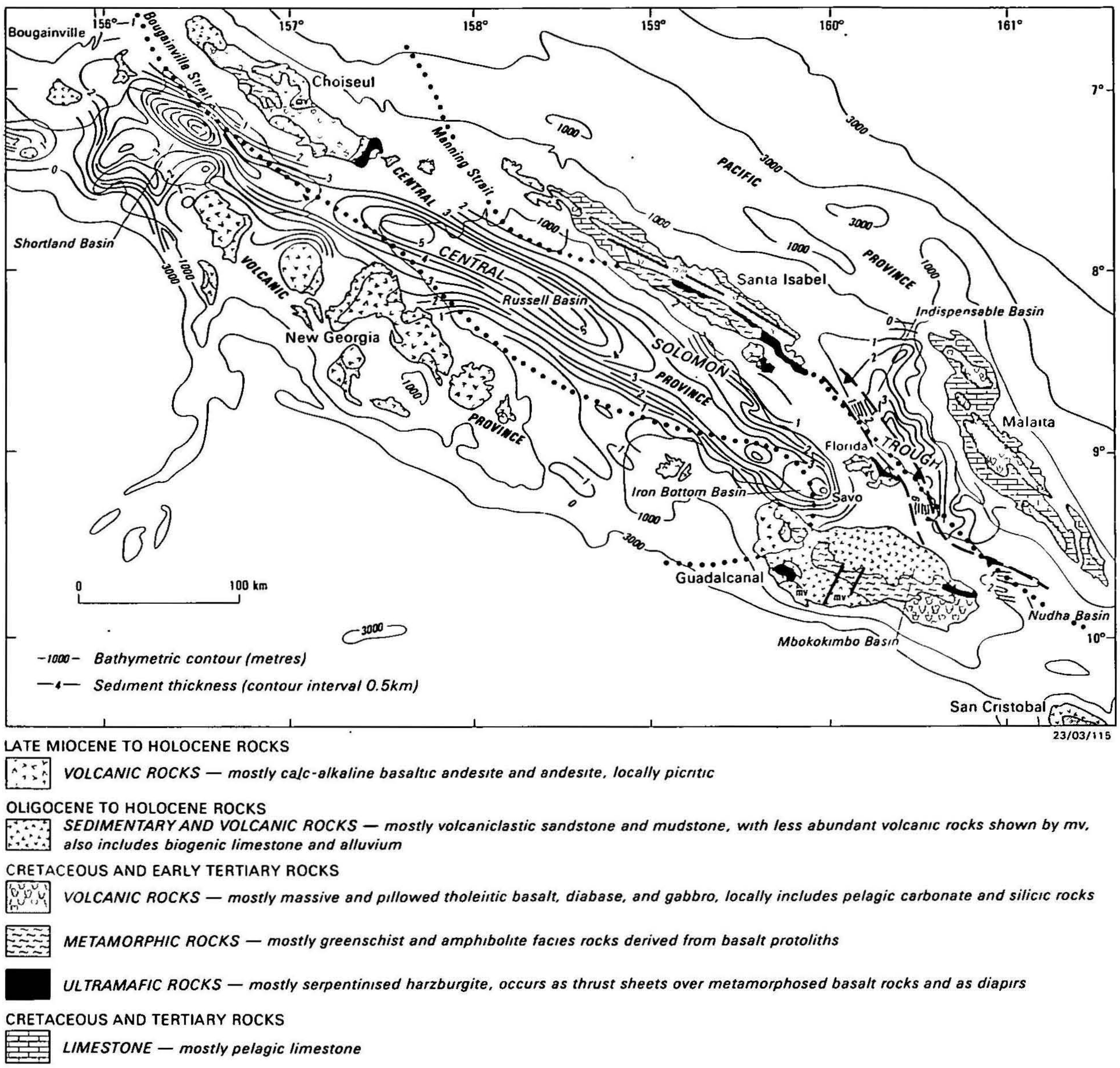


Figure 3. Simplified geology of the Solomon Islands and Solomon Trough sediment thickness (from Falvey et al. 1991).

many years and before 1900 gold was mined at Woodlark, Misima and Sudest, but knowledge of the geology was sketchy (until World War II most of the gold taken in New Guinea was placer). The main conclusion was that the eastern archipelago tied in with the geology of southern Papua New Guinea, but had volcanics and intrusions not easily explained. To the east the large, recently-raised and well-preserved atolls, Rennell and Bellona, crown much larger edifices, part of a large submarine plateau immediately east of the still larger Louisiade Plateau. The nature of these plateaus was unknown.

To the southeast, separated by a complex break, the d'Entrecasteaux–New Caledonia platform carries New Caledonia. For New Caledonia, some few geological reports go back to the last century, but over the inner arcs as a whole, not much solid geological groundwork had been done before 1950. The four prime elements in the geology of New Caledonia had been mapped by the mid-1960s (Fig. 4). These are: over the southeastern half of the island, the great thrust sheets of ultramafics responsible for the presence of mineable nickel; the extensive outpourings of basalt along, and inland from, the northwestern coast; the widespread and thick succession of late Cretaceous and Palaeogene marginal/basinal sediments (metamorphosed in the northeast); and lastly, over the central part, and often shown through windows in the younger sediments, there are older paralic sediments, especially Permian, Triassic and Jurassic, large areas of which had been regionally metamorphosed (Brothers & Lillie 1988, Paris 1981).

Structurally, faulting is intense and many faults show wrench features. The dominant fault systems tended to be parallel to the island axis. The older sediments are orogenic, typically wackes, the younger sediments are typical of marginal paralic basins; the Late Cretaceous in particular was deposited over widespread coastal marginal marine areas and includes extensive minor coal deposits. The Tertiary is a mixed succession and in places includes juxtaposed bathyal sediments, shallow-water platform sediments and occasional reefal deposits. The Miocene carbonates are shallow water reefal in origin.

The absence in the stratigraphic column of Early Cretaceous and Oligocene components, and the uplifting of the vast ultramafic sheets, supported the idea of profound vertical movements not only of individual faults blocks but also of the New Caledonian platform itself. In the late 1950s and in the 1960s, Bureau de Recherches Geologiques et Minieres, Noumea, and ORSTOM pioneered neotectonic and geomorphological studies, results of which anticipated the requirements

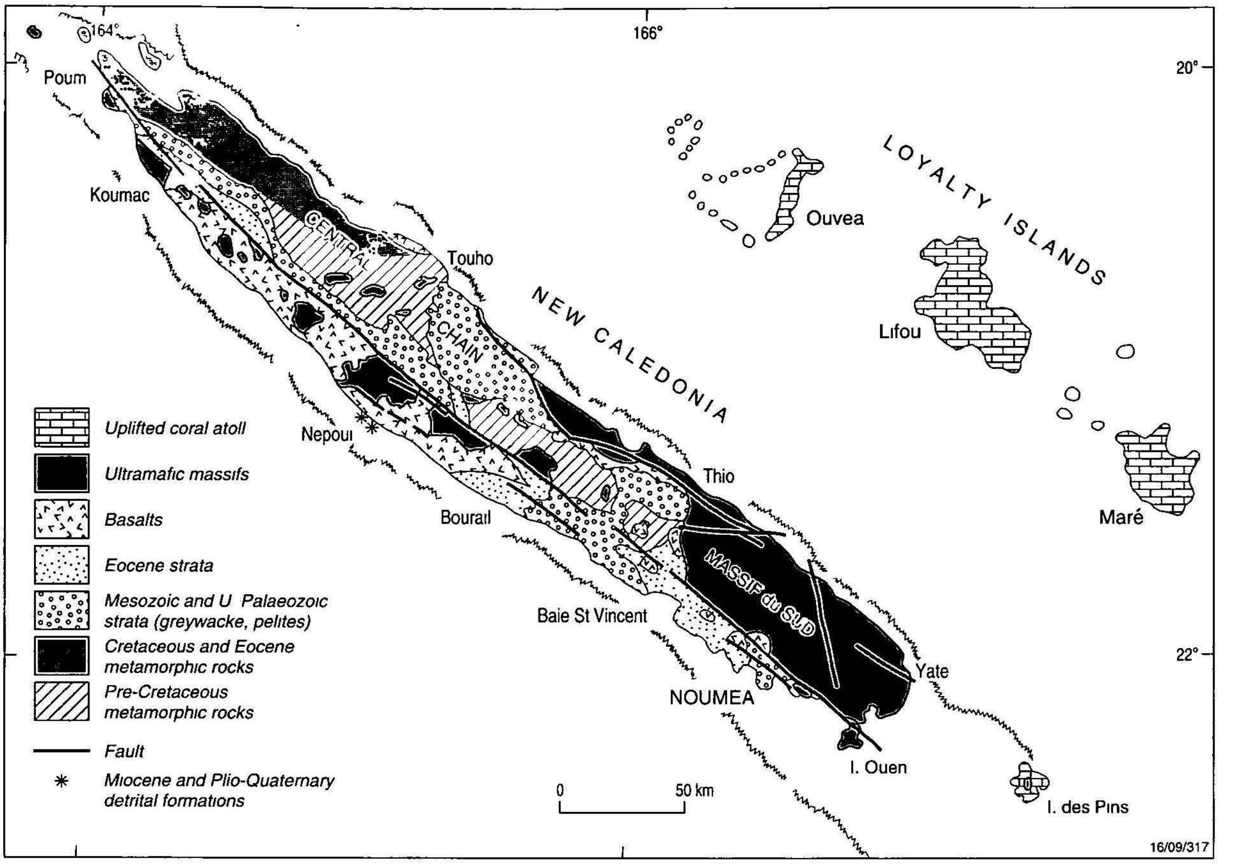


Figure 4. Basic geology of New Caledonia (from Brothers & Lillie 1988).

of plate tectonic theory. For example, the new data showed strong tilting of the New Caledonia platform, and its elevation, poised on what we now recognise as the lithospheric bulge just outboard of the South New Hebrides Trench.

In summary, the geology of New Caledonia was sufficiently well known by the mid-1960s to justify the belief that New Caledonia was the eastern limit to Australia (and, of course, new and old knowledge of the flora and fauna, supported this same message). But the 'why' and the 'how' of the New Caledonia platform remained unknown.

Marine geology

The 1950s and 1960s saw an enormous increase in the knowledge of oceans generally, including the southwest Pacific area. The stimulus arose from the need of military forces with missile-launching submarines to have detailed information about the ocean basins. This knowledge was to include not only detailed bathymetry but also the physics of the ocean floors and crust, especially gravity and magnetics. Much of this knowledge was dramatically new; some of it has only recently been released (e.g. the truly detailed bathymetry of the ocean basins); some of it is still classified. Civilian instrumentalities provided much of this information: they included the Hawaii Institute of Geophysics, Lamont Geological Observatory, Scripps Institution of Oceanography and Russian Academy of Science (with RV *Vityaz*).

In the southwest Pacific region, a major contribution was the revelation of the presence, size and structure of large submarine plateaus and ridges, e.g. the Ontong Java Plateau (Kroenke 1972) and Lord Howe and Norfolk/New Caledonia Ridges (the last two had been roughly mapped previously). The sedimentary caps of these were dredged and cored. Seismic refraction work suggested thick crust of suspected continental origin.

Outer Melanesia. Within the island chains, the presence of basins was securely established, e.g. the Russell Basin in the Solomons (Fig. 3) and the Aoba and Ambrym intra-arc basins in the New Hebrides. However, several of these basins appeared to have three or four kilometres of sediment in them (a minimum figure as it turned out). The presence of these basins gives a severely broken profile. In the Solomon Islands (Fig. 5), vertical relief can be as much as six km over a distance of some 20 km — and more to the structural floors of the basins. The gravity gradients are enormous, as had been suggested earlier by a land gravity survey on northern Guadalcanal (Coleman & Day 1965). Similarly, the Bligh Basin offshore of Viti Levu, Fiji, was shown to be a complex containing several kilometres of sediment (details in Rodda 1994, Johnson 1994).

The severely broken profile of Figure 5 characterises all of the Outer Melanesian arcs. The blocky nature of the crust was revealed by refraction seismic studies (e.g.Wiebenga 1973). Extensive heatflow measurements showed that dilational areas of sea floor, such as the North Fiji Basin, had consistently high flow values. These areas we know today as intra-arc and backarc basins and marginal seas. Older seas had relatively lower values. Mapping of anomalous magnetic lineation patterns was still exploratory.

Inner Melanesia. In the 1950s and 1960s, Inner Melanesian submarine plateaus such as Louisiade and Rennell were discovered and the main ridges delineated. Both New Caledonia and Lord Howe Ridges are broken transversely by defined lines, presumably fracture zones, with offsets suggesting relative horizontal movement. These ridges carry roughly the same amount of sediment, estimated at about 600 m of pelagic oozes. Limited refraction and gravity studies strongly suggested that they were submarine continental crust (Officer 1955). The mystery was, and still is, what kept them down, and how they had achieved isostatic balance so rapidly (even today, a common notion is that high basification of the lower crust has occurred, a kind of 'Beloussovian' oceanisation).

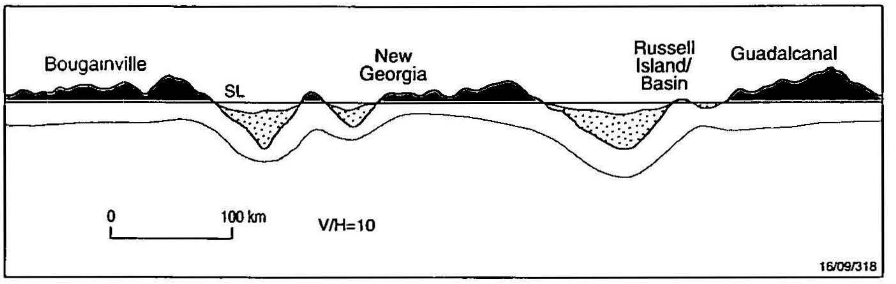


Figure 5. Solomon Islands, longitudinal profile, northwest–southeast. Overall relief is as much as 9 km over 70 km. The deepest basin has 4.5 km of sediment below a water column of more than 2 km. B — Bougainville, SL — sea level, NG — New Georgia, R — Russell Is/Basin, G — Guadalcanal.

New Caledonia was now mapped to at least reconnaissance level, as part of the further delineation of nickel deposits. A synopsis of the geology is contained in Brothers & Lillie (1988), and in the expansive synthesis of Paris (1981). Petroleum prospectivity is summarised by Vially & Mascle (1994). The place of the Louisiade Archipelago as part of the Papuan Orogen (approximates to New Guinea Mobile Belt, e.g. figure 3.3 of Kroenke 1984), marginal to the Australian craton, was confirmed; the large volcanic content arose because of contributions from the Woodlark spreading ridge.

The Tasman Sea is an entity to itself, a classic Careyan sphenochasm. In brief, new bathymetric data showed that the northeastern boundary was not the Lord Howe Ridge, but an ancillary ridge, the Dampier Ridge. The thickness of the pelagic carbonates suggested a relatively young age, guessed to be about Cretaceous. A north–south-trending twin row of seamounts and guyots (Lord Howe is one) was located. The steepness of the continental shelf (sometimes as steep as 20 degrees), the narrowness of the shelf, and the general lack of sediments along the southeastern Australian margin all suggested faulted margins. A pattern of magnetic anomaly lineations was delineated (and later interpreted by Ringis 1972 and Shaw 1978).

Post-plate tectonics

With the advent and widespread acceptance of plate tectonic theory in 1968, the southwest Pacific took on a flawed but recognisable unity (e.g. Packham 1973). The distinction between Inner and Outer Melanesian islands will no longer be stressed in this account; they will be treated as parts of a recognisable whole.

Land geology

In the 1970s and after, new data from land surveys in both Outer and Inner Melanesia, refined existing knowledge rather than adding greatly to it. Gold and copper exploration dominated land activities, and was inspired by the lucky discovery of several epithermal and porphyry copper deposits, e.g. Panguna on Bougainville (Panguna had in fact been recognised as a prospect in the early 50s, but the recognition of its vast promise had to wait — production began in 1972; Lum et al. 1991). By the late 1970s, most islands had been covered by broad-scale geochemical surveys — not very productively as it turned out.

Additional and more detailed land surveys, using satellite-derived imagery, emphasised the prevalence of faults. Some very large faults showed up in satellite imagery, but the ground evidence was scanty or failed to show the true dimensions of these faults. In contrast, as mentioned later, a better appreciation of the size of faults and their intensity was obtained from marine seismic reflection profiles (next section).

Major advances, however, were made in the 1970s in the knowledge of the refined geochemistry of the igneous rocks. Geochemical zonation in lava types shown in the Sunda (and other) volcanic arcs was not demonstrated to the same degree in the outer islands, a partial explanation for this being polarity reversal and the sometimes indiscriminate methods of sampling. The new work emphasised the complex geochemistry of rocks found in the outer islands, e.g. New Georgia onshore (Ramsay et al. 1984) and offshore (Johnson et al. 1987) in the Solomons, and New Hebrides volcanic belt (Crawford et al. 1988). The Solomon Islands, Vanuatu and Fiji furnished material for monographs such as Stanton's *Ore elements in arc lavas* (1994).

Marine work

The dominant activity by far was an increased tempo in marine surveys. Some of this was done by petroleum companies in, for example, the Solomons, Vanuatu and Fiji, but most continued to be done by government and academic institutions. Under the auspices of the Circum-Pacific Council (CPCMER) and the Australia-New Zealand-United States Tripartite Agreement, several major agencies carried out surveys in the Southwest Pacific. The agencies included the Bureau of Mineral Resources, Canberra (now AGSO), New Zealand Department of Scientific and Industrial Research, Hawaii Institute of Geophysics (University of Hawaii) and Scripps Institution of Oceanography (University of California), South Pacific Applied Geoscience Commission (SOPAC) and US Geological Survey. ORSTOM was very active. Additional work was done by German, Japanese and Russian agencies.

Apart from the data of obvious use in the search for petroleum, reflection profiles defined the boundaries of the intra-arc basins and confirmed the presence of enormous thicknesses of volcaniclastics in these basins (thicknesses suggested previously by refraction surveys). Sediment thicknesses in the Russell Basin, central Solomon Islands, had been thought to be overestimated, but were now shown to be considerably underestimated. The source areas for the tremendous volumes of this sediment — epiclastic, not all subaerial — is still a puzzle. Surprisingly, the close spacing of reflection seismic profiles confirmed the fault-ridden nature of the arc terrains, especially in the orogenically active outer islands. The elaborate 'flower structures' and riffle-pattern of large strike-slip faults are readily seen in seismic profiles all along the island chains (Figures 6 & 7). These are of great significance in determining the tectonics of the chain as part of the Indo-Australian/Pacific plates boundary (Ryan & Coleman 1992). Seafloor mapping and seismological analysis showed that convergence was oblique and sinistral (Abers & McCaffrey 1988; Auzende et al. 1994; Cooper & Taylor 1984).

Although most surveys had the assessment of petroleum potential as their primary aim, many surveys had other targets as well. Major efforts were made to map and evaluate manganese nodules (e.g. Skornyakova 1979). Lately, frequent targets have been seafloor hydrothermal systems and associated polymetallic sulphides at spreading centres (as in the North Fiji Basin — Ishibashi et al. 1994; Auzende et al. 1995; Stachelberg & Rad 1990). The Japanese South Pacific seafloor atlas (South Pacific Seafloor Atlas 1995) is a major contribution and covers all sea-bed metals. Methods to obviate the effects of alteration on marine samples of igneous rocks, and the increasing use of refractory components for analysis, constituted a considerable achievement (see Mahoney 1987). Until this time, distribution diagrams based on analyses of marine igneous rocks, using susceptible elements (e.g. FMA diagrams), were of dubious

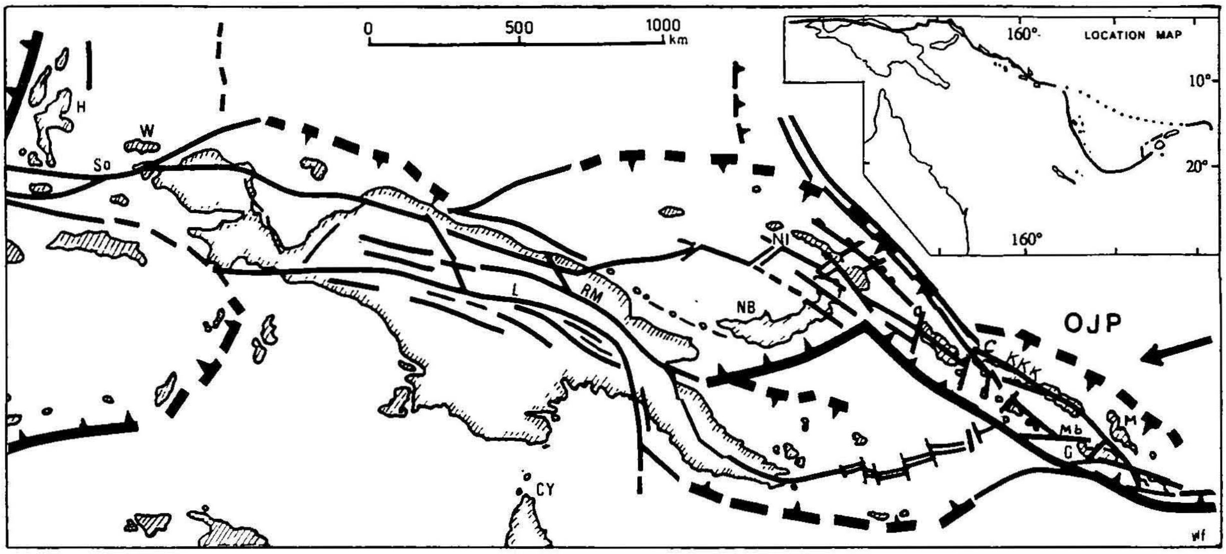


Figure 6. Melanesia and the Melanesian Boundary, an oblique convergence. The three largest fault systems are the Sorong-Ramu (So — RM), Sula — Lagaip (L) and Kia-Kaipito-Korigole (KKK). Both old and new subduction zones tend to transform into faults. OJP is the Ontong Java Plateau. H — Halmaheras; So — Sorong fault; W — Wageo Island; T — Timor; L — Lagaip fault system; RM — Ramu-Markham fault system; CY — Cape York; NB — New Britain; NI — New Ireland; G — Guadalcanal; C — Choiseul; Mb — Mborokua lineament; M — Malaita. Heavy arrow is the convergence vector, Indo-Australian/Pacific plates.

validity. Other work, especially that inspired by SOPAC, looked at the marine-derived geological and pollutant hazards most likely to affect island nations.

As a major side benefit of this varied work, the new data and the wide spectrum of scientists involved led to several programs of deep sea drilling, including the DSDP/ODP Leg 21, sites 203–210 — Tasman-Coral Seas; Leg 30, sites 285–289 — Ontong Java Plateau; Leg 90, sites 587–594 — Tasman Sea; ODP Leg 130, 803–807 — Ontong Java Plateau; Leg 134, 827–833 — Vanuatu; Leg 135, and 834–841 — Lau Basin/Tonga Ridge. The results of these programs included the proving of relative youth for the basins, with basal sediment no older than mid-Cretaceous. The large basins such as the Tasman Sea and New Caledonia Basin had typical open ocean sequences of pelagic oozes, with low terrigenous content and chert layers, no older than late Cretaceous. The ridges such as Lord Howe Ridge had deep-sea sediments no older than latest Cretaceous with thicknesses of 500–700 m. The drill holes within the chains, e.g. the Aoba Basin in Vanuatu (New Hebrides arc), or on the flanks, revealed typically volcaniclastic sequences, including distal turbidites, interspersed with carbonate oozes. These sequences are records of the high tectonism in these chains throughout the Neogene Tertiary. The younger sequences showed very high rates of sedimentation. As with the Solomon Islands basins, the provenance of these sediments is a problem. Oceanic basalts were drilled on the Ontong Java Plateau, and additional refraction results confirmed the great crustal thickness. At present, the plateau is thought to be the result of outpourings of lava from a very slow spreading ridge (Hussong et al. 1979). The overlying pelagic oozes were no older than mid-Cretaceous.

The recording of seafloor marine magnetic anomaly patterns led to striking results. The recognition and interpretation of the magnetic anomaly pattern over the Tasman Sea led almost at once to its recognition as an asymmetric seafloor spreading creation, predominantly of Late Cretaceous age (Ringis 1972); the revised version of Shaw (1978) is still accepted today. Within the Tasman Sea, the Dampier Ridge was a demonstration of spreading ridge jump (Shaw 1978; and see McDougall et al. 1994). The plate motions that led to Tasman seafloor spreading also caused the rifting that led to the separation of Lord Howe and Norfolk ridges (Symonds & Colwell 1992) and the creation of the New Caledonia Basin (Uruski & Wood 1991). Unlike the Tasman Sea, however, the New Caledonia Basin floor shows no formal pattern of magnetic anomalies and so a normal pattern of seafloor spreading seems unlikely. Rifting may have been followed by stretching of continental crust, its thinning and extension (Etheridge et al. 1989), but the mechanism for this is not proven. The twin row of seamounts was due to the passage of the plate over a hot-spot (McDougall & Duncan 1988), seamount ages being from 24 to 6.4 Ma with progressive younging southward of about 7 cm a year.

The North Fiji Basin is complex and the seafloor spreading pattern not easily identified (Hamburger & Isacks 1988, von Stackelberg & von Rad 1990, Auzende et al. 1995). A diffuse pattern of multiple spreading centres was proposed by Hamburger & Isacks (1988); Tanahashi et al. (1991) followed a more conventional model. The North and South Norfolk Basins formed in the early Oligocene.

The Coral Sea was shown to be the result of complex rifting of the continental crust of northeastern Australia (Mutter & Karner 1978). It may have been an offshoot of spreading in the Tasman Sea, but spreading began along a roughly east–west median ridge, the accommodating boundary to the west being a transform along the edge of the Australian craton (Falvey & Taylor 1974, Taylor & Falvey 1977). The southern margin lay along the submarine Queensland Plateau. The northern one lay along the southern edge of the 'tail' of New Guinea (marked by a volcanic sequence).

All of this work had many side benefits. A cruise aimed at mapping of nodules would be accompanied by the running of reflection seismic profiles, and very often, magnetic anomaly profiles. Coring would produce sub-seafloor sediment samples. Swath mapping, to obtain side-scan imagery, is of special use in checking hydrothermal systems and has added enormous detail to bathymetry. Techniques continued to improve — petroleum companies had greatly refined methods of multichannel seismic reflection surveying and the processing of data. The many-faceted nature of this marine work in the southwest Pacific is detailed in Crook et al. (1991).

The data from this surge of varied activity appeared in scientific journals, proceedings of SOPAC- and ORSTOM-inspired workshops (e.g. Anonymous 1977), and also in a set of monographs published by the Circum-Pacific Council for Energy & Mineral Resources. This latter series covered the

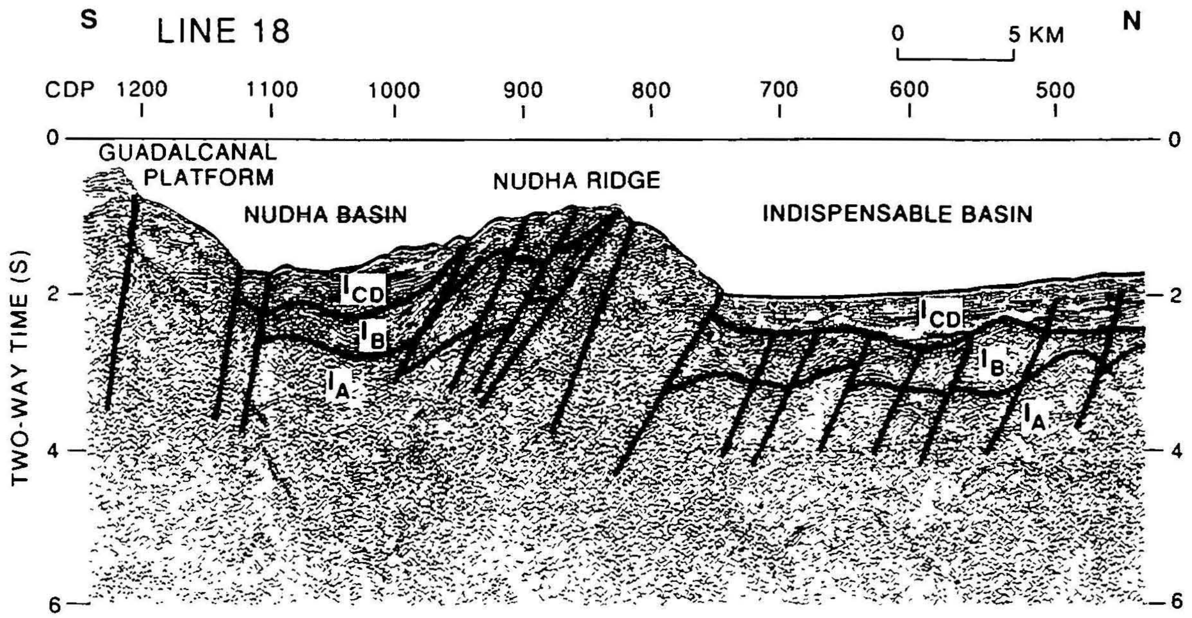


Figure 7. Seismic reflection profile of the easterly extension of the KKK fault complex, off the northeastern coast of Guadalcanal, Solomon Islands (from Vedder & Bruns 1989). The succession is made up mainly of volcaniclastic sediments — Oligo-Miocene (1_A), Late Miocene and Early Pliocene (1_B) and Pliocene/Quaternary (1_{CD}).

Outer Melanesian arcs, and although aimed primarily at petroleum search, contained generally thorough and extensive information (e.g. Taylor & Exon 1987 — Woodlark Basin, Papua New Guinea; Greene et al. 1988 — Vanuatu; Vedder & Bruns 1989 — Solomon Is). The deep sea drilling results were published as part of the DSDP/ODP program.

By about 1980, it was generally accepted that the Inner Melanesian arc system, and hence the old eastern Gondwana rim, had expanded to roughly its present position by the Early Tertiary (see next paragraph). The expansion can easily be followed over the southern part, where the Tasman Sea and the New Caledonia Basin are its measure. The northern part was not so clear, and still is not — the Solomon Sea and d'Entrecasteaux twin ridge complex are two puzzling features.

In 1978, the Outer Melanesian arc system was shown to be a linear set of islands well outside the expanded Inner Melanesian arc. Progressive reconstructions are in Cullen (1970), Packham (1973), Crook & Belbin (1978), Weissel et al. (1977), Falvey (1978), Kroenke (1984 Ch 8), and Walley (1992). The Outer arcs system is shown as building over a southwesterly-dipping subduction zone. The effect of polarity reversal, subduction 'flipping', was demonstrated for both the New Hebrides Arc and the Solomon Islands (Karig & Mammeryx 1972, Cooper & Taylor 1985). For the New Hebrides, subduction reversal resulted in the clockwise rotation about the Santa Cruz platform of the major part of the 'old', pre-7 Ma arc (a few elements were left near the old, Vityaz, trench), accompanied by falling away of slab lithosphere (Moberly 1972) and the seafloor spreading creation of the North Fiji Basin. The prime cause for this activity was the impinging of the great Ontong Java Plateau along the Outer Melanesian arcs while it moved west as part of the Pacific plate (the relative direction of motion of this plate had changed at about 41 Ma — reflected in the Emperor Seamounts/Hawaii 'Bend'). The new subduction direction explains the reverse symmetry of the New Hebrides Arc and the presence of a cluster of very deep earthquakes in the northeast — a deep, remnant expression of the old, abandoned, southwest-dipping slab. The linear Outer chain was now broken and the large sea (?marginal sea) between Inner and Outer chains partly replaced by the North Fiji Basin (see Figure 8 for a model of tectonic evolution).

Just this last paragraph shows the power of plate tectonics theory to coordinate, to bring together and explain, sets of apparently unrelated and contrary 'facts'. In this context, compare again Figure 2A with Figure 2B.

But the Solomons–New Hebrides–Fiji chain still has its peculiar features. It does not conform to arc models such as those of Karig (1974) and Coleman (1978). An alternative suggestion was that this chain had arisen as a linear structure (along a leaky transform?) comparable to that of the Line Islands. The chain was then incorporated as part of the Melanesian Boundary and took on the function and attributes of a young arc in the late Miocene (Coleman 1976). This notion, although venturesome, is not so far removed from the intra-oceanic creation of the same chain in the Yan & Kroenke reconstructions (1993) discussed below.

The Melanesian Outer arcs had been thought to be the expression of a shear zone for many years (Tethyan Shear System of Carey 1958, 1963). With plate tectonic theory, this shearing could be explained in terms of the differing directions of convergence of the Indo-Australian and Pacific plates, that is, the convergence was an oblique one. The Melanesian Boundary, the length of Indo-Australian/Pacific plate boundary along which the islands lie, is a linear mosaic of fault blocks (Fig. 6). The oblique convergence between Pacific and Indo-Australian plates results in intense strike-slip activity between blocks of the mosaic. This activity takes two primary forms (Fig. 9).

In the first, individual blocks can be rotated about vertical and horizontal axes and they can be shifted vertically by large amounts at rapid rates; this last phenomenon is primarily the result of 'ramping' by one block over another (Coleman 1991, Ryan & Coleman 1992). Many examples of anomalous juxtapositions of differing geology can be thus explained, e.g. the presence of bathyal carbonate oozes against shallow-water high energy wackes (Coleman et al. 1988).

In the second form, overall activity is focused along a 'master' fault system, so that the blocks on the outer side of the master fault tend to be carried with the outer plate (the 'translation' of blocks). With continuing oblique convergence the boundary takes on the function of a 'conveyor belt' so that terranes are carried along the boundary as allochthonous 'exotics'. These oblique composite convergences have been termed 'composite transform convergences' (Ryan & Coleman 1992), for they are composite and motion between the opposing plates is accommodated by subduction and by transform faults; but a better and simpler name would be 'accommodation boundaries', for that is what they are (Coleman 1995). Accommodation boundanes are common around the Pacific plate. For example, Alaska is well known as a composite of exotic terranes which were carried north as part of a great accommodation boundary which includes the San Andreas fault complex as a master.

An indirect consequence of the oblique convergence of the Indo-Australian and Pacific plates is shown in the development of the Bismarck Sea and the Woodlark Basin. In the Bismarck Sea, a broken median ridge proved to be a static spreading ridge. In the Woodlark Basin, spreading began about mid-Pliocene and the spreading system has itself been subducting beneath the central Solomons (a contributory cause of the anomalous geochemistry of the New Georgia volcanoes). Both these features can be regarded as ancillary structures accompanying the westward drive of the Pacific plate past the north-moving Indo-Australian plate. The study of the Woodlark system showed that the westward movement of the Solomon block (relative to Indo-Australian plate) has been at a fast rate; at about 3.5 Ma, New Ireland was positioned several hundred kilometres east of its present position (Weissel et al. 1982). Seismological evidence from New Guinea demonstrated strike-slip movement and thrusting (Abers & McCaffrey 1988). Rotation of individual blocks was suspected and then confirmed by palaeomagnetic studies of target blocks (see below).

Mobility

The southwest Pacific is an essay in the motion of great crustal blocks. Mobility is expressed in rotation of blocks about vertical and horizontal axes and by their translational or horizontal movement.

Major rotations

The southwest Pacific shows rotations of large blocks about both vertical and horizontal axes, similar to those from other areas, such as the Aleutians and western North America (Geist et al. 1988, Scholl et al. 1988).

The study of palaeomagnetic declination of ancient samples has a long history (see Green & Pitt 1957). On the basis of such study, the rotation of terranes was put forward long before the acceptance of plate tectonic theory. No mechanism was provided. In the southwest Pacific, evidence of block rotations has supported the conclusions based on other studies.

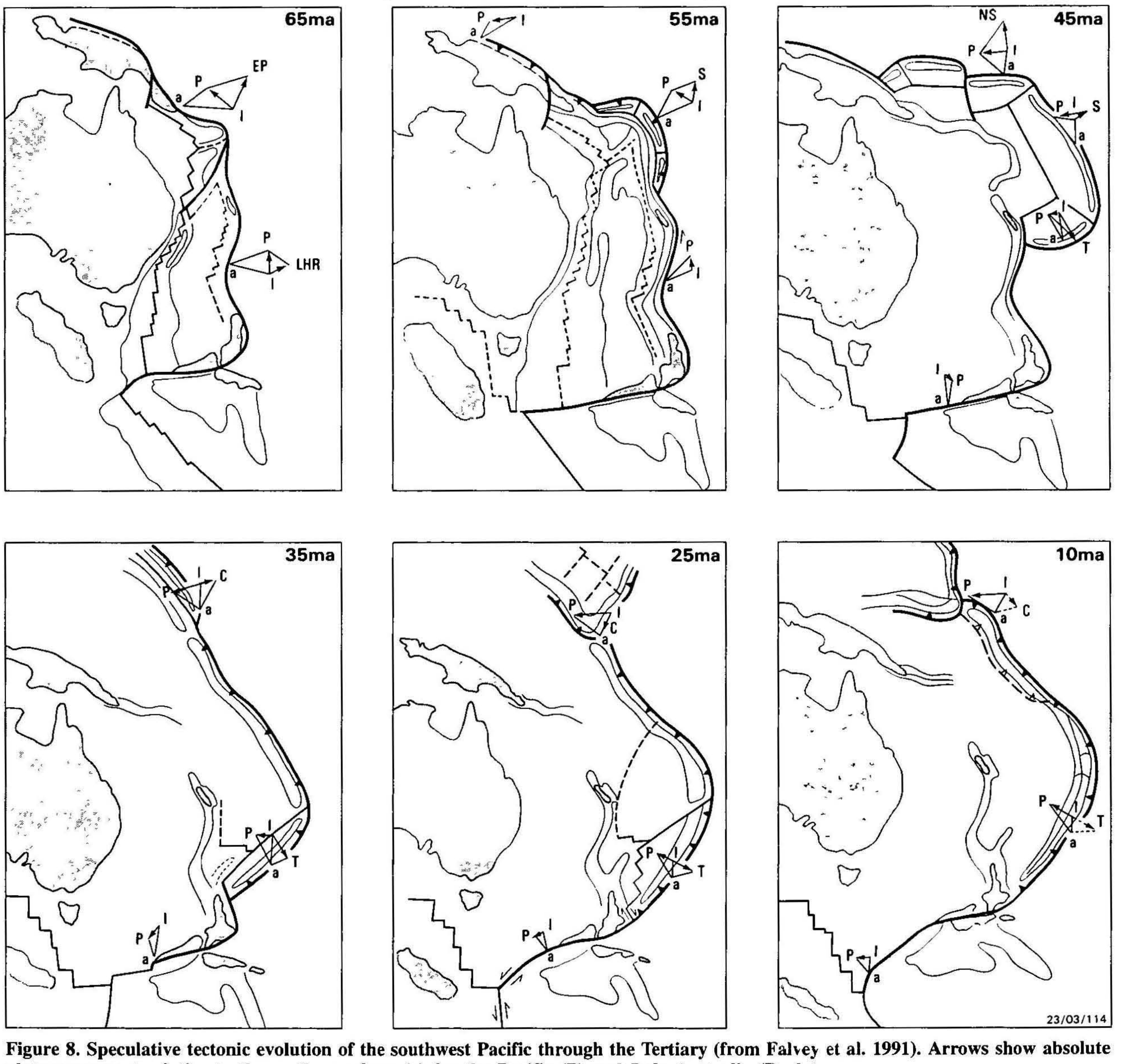


Figure 8. Speculative tectonic evolution of the southwest Pacific through the Tertiary (from Falvey et al. 1991). Arrows show absolute plate movement relative to the aesthenosphere (a) for the Pacific (P) and Indo-Australia (I) plates.

The Fiji results (James & Falvey 1978) suggested anticlockwise rotation of Viti Levu of some 21 degrees since the early Pliocene. Malahoff et al. (1982) supported rotation of Viti Levu: earlier suspicions were thus confirmed. Results from northern Papua New Guinea (Falvey & Pritchard 1984) suggested the westward passage of New Ireland (as part of the Solomon Islands) past New Britain in the late Tertiary (see also Weissel et al. 1982). As well, basins in the Solomon Islands and Vanuatu are of the summit and pull-apart type that have resulted, at least in part, from block rotation, as with those in the Aleutians (Geist et al. 1988).

Movements horizontal and vertical

By the late 1980s, these diverse and various studies in the southwest Pacific had provided, in its essentials, the picture we have today (Figs 1 and 8; and see Kroenke 1984 and Smith 1990). More recent studies have emphasised the speed

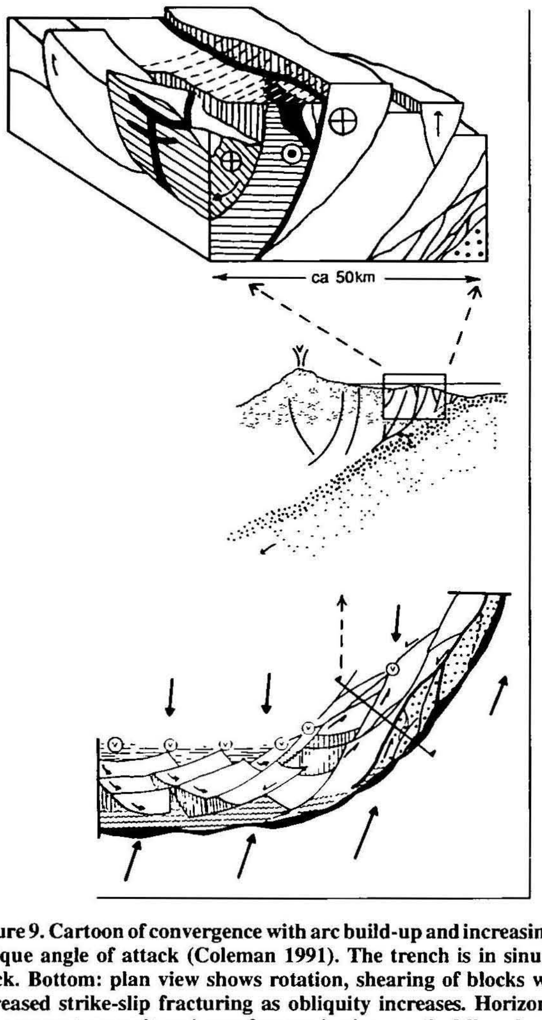


Figure 9. Cartoon of convergence with arc build-up and increasingly oblique angle of attack (Coleman 1991). The trench is in sinuous black. Bottom: plan view shows rotation, shearing of blocks with increased strike-slip fracturing as obliquity increases. Horizontal lines suggest summit or inner forearc basins, vertical lines basins of the pull-apart type, wavy lines near the trench are elements of the accretionary prism. The heavy line suggests the current master fault system. Individual blocks may be carried along the transform complex outboard of the master fault e.g. dotted blocks. Heavy bar is the approximate line of the section above: this is a profile of the arc with deep strike-slip fracturing — a part is enlarged as the block diagram. These profiles stress the undulations in the fracture surfaces both in plan and in profile. Relative motion on these surfaces produces vertical and horizontal movement of blocks and also rotation and tilting. Transtensional and transpressional situations arise along fractures and provide 'plumbing' for diapiric convective hydrothermal systems with caps enriched in metals. These caps can be concealed by debris from a ramped block (centre), tilted to give a misleading outcrop pattern (left) or even removed from the root stock.

of horizontal and vertical movements of crustal slabs within the southwest Pacific.

Rates of vertical movement can be very high, amounting to a centimetre or more per year. Examples include the uplifted terraces of the Huon Peninsula, New Guinea (Chappell 1974a, b); and uplift of bathyal oozes in graben, north central Guadalcanal (Coleman et al. 1988).

On a large scale, the long-standing intensity of horizontal or translational movements in the southwest Pacific was illustrated by Pigram & Davies (1987). In this pioneer paper, they showed that a large part of northern New Guinea consisted of 'exotic' terranes transferred from southern and eastern locations. These results were amplified and supplemented by Struckmeyer et al. (1993). Coleman (1991) and Ryan & Coleman (1992) proposed a mechanism for shifts of this kind (outlined above) and suggested that eastern Indonesia is the future site of a 'New Alaska', a conglomeration of exotic terranes.

Mobility on an even larger scale is demonstrated by Yan & Kroenke (1993). They attempted a palaeogeography of the southwest Pacific going back 100 m.y., using intervals of 0.5 m.y., set in frames defined by today's latitudes and longitudes. This set of reconstructions is the most sophisticated available.

Among major block movements, Yan & Kroenke stipulate a growth-position of the Solomon–Fiji block far to the northeast of Australia in the mid-Eocene, and then show its movement south and west to its present position (Fig. 10). This is in contrast to the older view that the Solomon-Fiji block was part of an arc developed outboard of the Inner Melanesian arc and separated from the latter by a large marginal sea/backarc basin. These differing viewpoints have yet to be reconciled.

On a smaller scale, Yan & Kroenke show rotation eastward of the 'Eua Ridge' from a position south of New Caledonia to its present position as part of the Tonga platform, over the period 40–32 Ma (Fig. 10). The northern part of the Eua Ridge was then 'collected' by the southeastern end of the Solomon–Fiji as it moved westwards, at about 6 Ma (L. Kroenke, University of Hawaii, pers. comm. 1995). This sort of jump of a block across a boundary transform system can be readily explained (Fig. 9). The 'exotic' part is now the Yavuna Group, an anomaly in the general geology of Viti Levu. The Yavuna Group is an allochthon, a Paleogene exotic, with an odyssey that ranged eastwards from near New Caledonia to Tonga and then westward as part of the Fiji platform.

Movements such as these call for an elaborate system of transforms (there is ample evidence for the existence of such a system), together with accommodation boundaries, such as the Melanesian Boundary, that will shift large crustal blocks. Indeed, as discussed earlier, the Melanesian Boundary is studded with assorted crustal blocks. Rotated, raised or lowered, their dominant motion is lateral and to the west.

Such movement also implies not only that arcs may be composite but also that the form and configuration of arcs built along obliquely convergent boundaries is transitory and may change even as the arc is being built. In this sense, to speak of an 'early Miocene' Solomons arc is something of a misnomer.

Summary and conclusions

In the 1950s, the inner arc areas were considered to have been emplaced at the outer edge of mid-Mesozoic 'Australia' (the eastern rim of Gondwana — Fig. 2A). The outer arc areas developed later as bordering oceanic arcs. A genetic connection was implied between these so-called arcs and the Australian continent (Fig. 2A). But the data from the land areas and from the intensive marine studies that began in the 1950s and in the early 1960s did not altogether agree with contemporary models of island arcs.

The new marine data supported the global tectonic theories

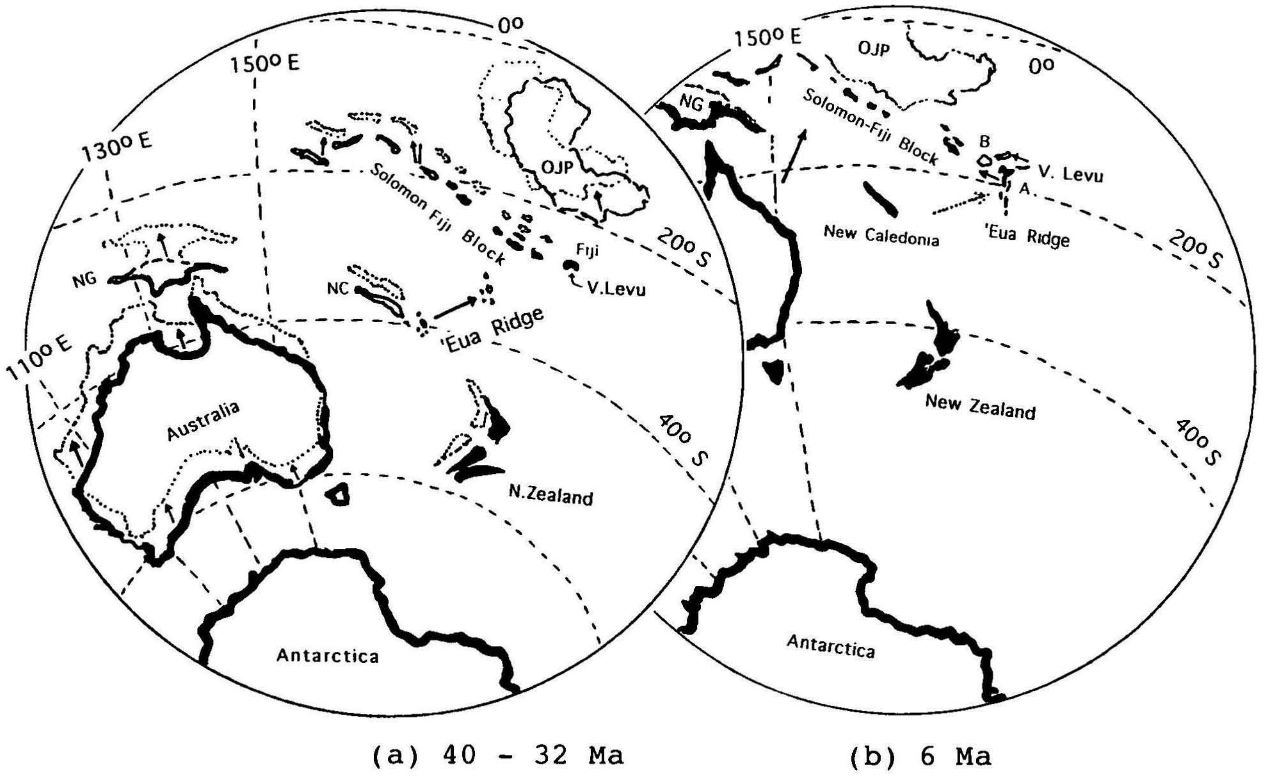


Figure 10. Reconstructions adapted from Yan & Kroenke 1993) — (a) and (b). Figure (a) 40–32 Ma: Now part of Tonga arc massif, the 'Eua Ridge lay south of New Caledonia on the eastern border of Gondwana, and moved eastwards (heavy arrow). Dotted outlines show the 32 Ma positions. In (b), at about 6 Ma, the northern portion of 'Eua Ridge moves westwards as the 'Yavuna Group', part of Viti Levu. 'A' is Viti Levu at 6 Ma, 'B' with open outline is its present position.

of the late 1960s, and in the early 1970s gave rise to new notions of arc behaviour, in particular the idea of arc piles building on the rim of the superposed plate. The geological and geophysical data stressed the mobility of the outer island chains and the pronounced vertical movement of individual crustal blocks. The difficulty of reconciling the early growth stages of the outer chains to conventional arc models was not resolved. Over the same period, new data from the inner arc areas confirmed that they were built up as Andean-type arcs along continental edges.

In the 1980s, the translational movement of discrete crustal blocks was recognised. Pioneer workers within AGSO recognised that much of northern New Guinea was composite, just as much of Alaska had been seen to be composite and made up of 'exotic' terranes. Further work has amplified these results, so we can predict that, in the geological future, eastern Indonesia will become a 'New Alaska' — barring a drastic change in plate geometry.

Most workers would now say that the Outer islands have stepped out and been separated from the inner arcs by dilational seas. But some evidence suggests that these outer islands may have developed independently and been brought to their present connection with old Australian forelands by interplate action. The Outer Melanesian islands mark a special kind of convergent plate boundary — an accommodation boundary. Along this type of convergence, marked by a zone of strike-slip faults and subduction zones, terranes are translated by activity along large-scale strike-slip faults to be mixed with newly formed crust, the by-products of subduction. In brief, the arc being built along such a boundary consists of subduction products plus introduced allochthons. Such hybrid entities include large parts of the Solomon–New Hebrides–Fiji chain. While building, these arc-hybrids continually change their composition, form and configuration.

The southwest Pacific has supplied major contributions in data to support such aspects of plate tectonics as polarity reversal, collision tectonics, the movement of arcs, the splitting of arcs, the broken nature of arc profiles, the importance of strike-slip components along oblique convergences, the transport of allochthonous terranes, the effects of the style of subduction and the digestion of spreading ridges on the genesis of arc rocks, crustal attenuation, the creation of submarine plateaus and the 'oceanisation' of cratonic slices. Researchers have found the southwest Pacific to be an appropriate testing ground for plate tectonic theory. The new data have refined plate tectonic principles; they have added, and will add, to its concepts.

Acknowledgments

I thank Charles Barton for suggesting this contribution and reviewers Ray Shaw and Neville Exon for providing models of the reviewer's skill; I am grateful for their many constructive suggestions and searching commentary.

References

Abers, G. & McCaffrey, R., 1988. Active deformation in the New Guinea fold-and-thrust belt: seismological evidence for strike-slip faulting and basement-involved thrusting. Journal of Geophysical Research, 93(B11), 13332–13354.

Anonymous, 1977. Geodynamics of the Southwest Pacific, International Symposium, Noumea, New Caledonia, August-September 1976. Editions Technic, Paris, 413 pp.

Auzende, J-M., Collot, J-Y., Lafoy, Y., Gracia, E., Gelo, L., Ondreas, H., Eissen, J-P., Larue, M.B., Olisukulu, C., Tolia, D. & Biliki, N., 1994. Evidence for sinistral

strike-slip deformation in the Solomon Island arc. Geo-Marine Letters, 14, 232–237.
Auzende, J-M., Pelletier, B. & Eissen, J-P., 1995. The North Fiji basin — geology, structure and geodynamic evolution. In: Taylor, B. (editor), Backarc basins: tectonics and magmatism. Plenum Press, NY, 139–175.
Brothers, R.N. & Lillie, A.R., 1988. Regional geology of New Caledonia. In: Naim, A.E.M. (editor), The ocean basins and margins, volume 7B — Pacific Ocean. Plenum Press, NY, 325–374.
Bryan, W.H., 1944. The relationship of the Australian continent to the Pacific ocean — now and in the past. Journal & Proceedings of the Royal Society of New South Wales, 78, 42–62.
Carey, S.W., 1938. The tectonic evolution of New Guinea and Melanesia. Unpublished thesis, University of Sydney.
Carey, S.W., 1958. The tectonic approach to continental drift. In: Continental Drift — a symposium. Geology Department, University of Tasmania, Hobart, Special Publication, 177–355.
Carey, S.W., 1963. The asymmetry of the Earth. Australian Journal of Science, 25, 369–383 and 479–488.
Chappell, J., 1974a. Geology of coral terraces, Huon Peninsula, New Guinea: a study of Quaternary tectonic movements and sea-level changes. Bulletin of the Geological Society of America, 85, 553–570.
Chappell, J., 1974b. Upper mantle rheology in a tectonic region: evidence from New Guinea. Journal of Geophysical Research, 79, 390–398.
Coleman, P.J., 1976. A re-evaluation of the Solomon Islands as an arc system. In: Glasby, G.P. & Katz, H.R. (editors), Marine geological investigation in the Southwest Pacific and adjacent areas. Proceedings, IDOE Workshop, Suva Fiji, September 1975. UN ESCAP, CCOP/SOPAC Technical Bulletin 2, 134–140.
Coleman, P.J., 1978. A model of arc development. Bulletin of the Australian Society of Exploration Geophysicists, 9, 159–161.
Coleman, P.J., 1991. Dynamic strike-slip fault systems with respect to the Solomon Islands and their effect on mineral potential. Marine Geology, 98, 167–176.
Coleman, P.J., 1995. Movements along the Melanesian Boundary, southwestern Pacific ocean: Further thoughts on the composite transform convergence (and some economic implications). In: Branagan, D.F. & Facer, R.A. (editors), Mineral search in the south-west Pacific region — a memorial to Ken Glasson. Proceedings of the Eighth Edgeworth David Symposium, September 1995, Earth Resources Foundation, University of Sydney, 25–33.
Coleman, P.J. & Day, A.A., 1965. Petroleum possibilities and marked gravity anomalies in north-central Guadalcanal. British Solomon Islands Geological Record (1959–62), 112–119.
Coleman, P.J., Haig, D.W. & Wilkins, N.A., 1988. Puzzling occurrence of a deepsea ooze in a modern island arc terrain in the Solomon Islands. Search, 19, 90–91.
Cooper, P.A. & Taylor, B., 1984. The spatial distribution of earthquakes, focal mechanisms and subducted lithosphere in the Solomon islands. In: Exon, N.F. & Taylor, B. (compilers), Seafloor spreading, ridge subduction, volcanism and sedimentation in the offshore Woodlark-Solomons region, and Tripartite cruise report for Kana Keoki cruise 82-03-16 Leg 4. UN ESCAP CCOP–SOPAC Technical Report, 34, 43–76. SOPAC Secretariat, Suva, Fiji.
Cooper, P.A. & Taylor, B., 1985. Polarity reversal in the Solomon Islands arc. Nature, 314, 428–430.
Crawford, A.J., Greene, H.G. & Exon, N.E., 1988. Geology, petrology and geochemistry of submarine volcanoes around Epi Island, New Hebrides Island arc. In: Greene, H.G. & Wong, F. (editors), Geology and offshore resources of Pacific Island Arcs — Vanuatu region. Circum-Pacific Council for Energy and Mineral Resources, Earth Science Series, vol. 8, Houston, Texas, 301–327.
Crook, K.A.W. & Belbin, L., 1978. The Southwest Pacific area during the last 90 million years. Journal of the Geological Society of Australia, 25(1), 23–40.
Crook, K.A.W., Exon, N.F. & Coleman, P.J., 1991. A decade of growth in the South Pacific marine geoscience database. Marine Geology, 98, 155–165.
Cullen, D.J., 1970. A tectonic analysis of the Southwest Pacific. NZ Journal of Geology & Geophysics, 13(1), 7–20.
Etheridge, M.A., Symonds, P.A. & Lister, G.S., 1989. Application of the detachment model to reconstruction of conjugate passive margins. Memoir of American Association of Petroleum Geologists, 46, 23–40.
Fairbridge, R.W., 1953. Australian stratigraphy, 2nd edition. University of Western Australia Text Books Board, Nedlands, Western Australia.
Fairbridge, R.W., 1961. The Melanesian Border Plateau, a zone of crustal shearing in the southwest Pacific. Publications du Bureau Central Sismologique International, 1960A, no 22, 137–149.
Falvey, D.A., 1978. Analysis of palaeomagnetic data from the New Hebrides. Bulletin of Australian Society of Exploration Geophysicists, 9(3), 117–123.
Falvey, D.A., Colwell, J.B., Coleman, P.J., Greene, H.G., Vedder, J.G. & Bruns, T.R., 1991. Petroleum prospectivity of Pacific Island arcs: Solomon Islands and Vanuatu. Australasian Petroleum Exploration Association (APEA) Journal, 31(1), 191–212.
Falvey, D.A. & Pritchard, T., 1984. Preliminary paleomagnetic results from northern Papua New Guinea: evidence for large microplate rotations. In: Watson, S.T (editor), Transactions of the Third Circum-Pacific Energy and Mineral Resources Conference, 22–28 August 1982, Honolulu, Hawaii, 623–630.
Falvey, D.A. & Taylor, L.W.H., 1974. Queensland Plateau and Coral Sea Basin: structural and true stratigraphic pattern. Bulletin of the Australian Society of Exploration Geophysics, 5(4), 123–126.
Geist, E.L., Childs, J.R. & Scholl, D.W., 1988. The origin of summit basins of the Aleutian Ridge: implication for block rotation of an arc massif. Tectonics, 7, 327–341.
Glaessner, M.F., 1950. Geotectonic position of New Guinea. Bulletins of the American Association of Petroleum Geology, 34(5), 856–881.
Green, R. & Pitt, R.P.B., 1957. Suggested rotation of New Guinea. Journal of Geomagnetism and Geoelectricity, 19(4), 317–321.
Greene, H.G. & Wong, F. (editors), 1988. Geology and offshore resources of Pacific Island Arcs — Vanuatu region. Circum-Pacific Council for Energy and Mineral Resources, Earth Science Series, Vol. 8, Houston, Texas, 442 pp.
Griffiths, J.R., 1975. New Zealand and the southwest Pacific margin of Gondwanaland. In: Campbell, K.S.W. (editor), Gondwana geology. Australian National University Press, Canberra, 619–637.
Hamburger, M.W. & Isacks, B.L., 1988. Diffuse back-arc deformation in the southwestern Pacific. Nature, 332, 599–603.
Hussong, D.M., Wipperman, L.W. & Kroenke, L.W., 1979. The crustal structure of the Ontong Java and Manihiki plateaus. Journal of Geophysical Research, 84(B11), 6003–6010.

Ishibashi, J-I., Grimaud, D., Nojiri, Y., Auzende, J-M. & Urabe, T., 1994. Fluctuation of chemical compositions of the phase-separated hydrothermal fluid from the North Fiji Basin Ridge. Marine Geology, 116, 215–226.

James, A. & Falvey, D.A., 1978. Analysis of palaeomagnetic data from Viti Levu, Fiji. Bulletin of Australian Society of Exploration Geophysicists, 9(3), 115–117.

Johnson, H., 1994. Structure and petroleum geology of the Bligh Water and Bau Waters Basin. In: Stevenson, A.J., Herzer, R.H. & Ballance, P.F. (editors), Geology and submarine resource of the Tonga–Lau–Fiji region. SOPAC Secretariat, Suva, Fiji, 171–184.

Johnson, R.W., Jaques, A.L., Langmuir, C.H., Perfit, M.R., Staudigel, H., Dunkley, P.N., Chappell, B.W., Taylor, S.R. & Baekisapa, M., 1987. Ridge subduction and forearc volcanism: petrology and geochemistry of rocks dredged from the western Solomon arc and Woodlark Basin. In: Taylor, B. & Exon, N.F. (editors), Marine geology and geophysics of the Woodlark Basin — Solomon Islands region. Circum-Pacific Council for Energy & Mineral Resources, Houston, Texas, Earth Science Series, 7, 155–226.

Jouannic, C. & Thompson, R-M., 1983. Bibliography of geology and geophysics of the southwestern Pacific. CCOP-SOPAC, Technical Bulletin 5, Suva, Fiji, 258 pp.

Karig, D.E., 1974. Evolution of arc systems in the western Pacific. Annals and Reviews of Earth and Planetary Sciences, 2, 51–75.

Karig, D.E. & Mammeryx, J., 1972. Tectonic framework of the New Hebrides island arc. Marine Geology, 12(3), 188–205.

Kroenke, L.W., 1972. Geology of the Ontong Java Plateau. Reports of the Hawaii Institute of Geophysics, HIG 72–5.

Kroenke, L.W., 1984. Cenozoic tectonic development of the southwest Pacific. UN ESCAP, CCOP/SOPAC Technical Bulletin 6, Suva, Fiji, 122 pp.

Lum, J.A., Clark, A.L. & Coleman, P.J., 1991. Gold potential of the Southwest Pacific — Papua New Guinea, Solomon Islands, Vanuatu, Fiji. East West Center (Pacific Islands Development Program), Honolulu.

Mahoney, J.J., 1987. An isotopic survey of Pacific oceanic plateaus: implications for their nature and origin. American Geophysical Union, Monograph 43, 207–220.

Malahoff, A., Hammond, S.R., Naughton, J.J., Keeling, D.L. & Richmond, R.N., 1982. Geophysical evidence for post-Miocene rotation of the island of Viti Levu, Fiji, and its relationship to the tectonic development of the North Fiji Basin. Earth & Planetary Science Letters, 57, 398–414.

McDougall, I. & Duncan, R.A., 1988. Age progressive volcanism in the Tasmantid Seamounts. Earth & Planetary Science Letters, 89, 207–220.

McDougall, I., Maboko, M.A.H., Symonds, P.A., McCullock, M.T., Williams, I.S. & Kudrass, H.R., 1994. Dampier Ridge, Tasman Sea, as a stranded continental fragment. Australian Journal of Earth Sciences, 41(5), 395–406.

Moberly, R.M., 1972. Origin of lithosphere behind island arcs, with reference to the Western Pacific. Bulletin of the Geological Society of North America, 132, 35–55.

Mutter, J.C. & Karner, G., 1978. Cretaceous taphrogeny in the Coral Sea. Bulletin of Australian Society of Exploration Geophysicists, 9(3): 82–87.

Officer, C.B., 1955. Southwest Pacific crustal structure. Transactions of the American Geophysical Union, 36, 449–459.

Packham, G.H., 1973. A speculative Phanerozoic history of the Southwest Pacific. In: Coleman, P.J. (editor), The Western Pacific: island arcs, marginal seas, geochemistry. University of Western Australia Press, Nedlands, Western Australia, 369–388.

Paris, J.P., 1981. Geologie de la Nouvelle Caledonie. Memoire du Bureau Recherche Geologique et Minieres, 113, Orleans, 274 pp, map.

Pigram, C.J. & Davies, H.L., 1987. Terranes and accretionary history of the New Guinea orogen. BMR Journal of Australian Geology & Geophysics, 10(3), 193–211.

Ramsay, W.R.H., Crawford, A.J. & Foden, J.D., 1984. Field setting, mineralogy, chemistry and genesis of arc picrites, New Georgia, Solomon Islands. Contributions to Mineralogy and Petrology, 88, 386–402.

Ringis, J., 1972. The structure and history of the Tasman Sea and the southeast Australian margin. PhD thesis, University of NSW, 338 pp.

Rodda, P., 1994. Geology of Fiji. In: Stevenson, A.J., Herzer, R.H. & Ballance, P.F.(editors), Geology and submarine resource of the Tonga–Lau–Fiji region. SOPAC Secretariat, Suva, Fiji, 131–152.

Ryan, H.F. & Coleman, P.J., 1992. Composite transform-convergent plate boundaries: description and discussion. Marine and Petroleum Geology, 9, 89–96.

Scholl, D.W., Stevenson, A.J., Vallier, T.L., Ryan, H.W. & Geist, E.L., 1988. Arc-massif shearing, block rotation, and transportation — an Aleutian perspective of production and lateral dispersal of arc-tectonostratigraphic terranes. Geological Society of America 1988 Centennial Celebration, Denver, Colorado, Abstract 25225.

Shaw, R.D., 1978. Seafloor spreading in the Tasman Sea: a Lord Howe Rise — eastern Australian reconstruction. Bulletin of Australian Society of Exploration Geophysicists, 9(3): 75–81.

Skornyakova, N.S., 1979. Zonal regularities in occurrence, morphology and chemistry of manganese nodules in the Pacific Ocean. In: Bischoff, J.L. & Piper, D.Z. (editors), Marine geology and oceanography of the Pacific manganese nodule province. Plenum Press, NY, 699–727.

Smith, R.I., 1990. Tertiary plate tectonic setting and evolution of Papua New Guinea. In: Carman, G.J. & Z. (editors), Petroleum exploration in Papua New Guinea. Proceedings of the First PNG Petroleum Convention, Port Moresby, 1990, 229–244.

South Pacific Seafloor Atlas, 1995. Japan–SOPAC Cooperative Study on deepsea mineral resources in the South Pacific, 1985–1994. Japanese International Cooperative Agency/ Metals Minerals Agency Japan/SOPAC (JICA/MMAJ/SOPAC) publication, 24 sheets.

Stanton, R.L., 1994. Ore elements in arc lavas. Oxford University Press, 400 pp.

Struckmeyer, H.I.M., Yeung, M. & Pigram, C.J., 1993. Mesozoic to Cainozoic plate tectonic and palaeogeographic evolution of the New Guinea Region. In: Carman, G.J. & Z. (editors), Petroleum exploration and development in Papua New Guinea. Proceedings of the 2nd PNG Petroleum Convention, Port Moresby, 1993, 261–290.

Symonds, P.A. & Colwell, J.B., 1992. Geological framework of the southern Lord Howe Rise/West Norfolk Ridge region. Australian Geological Survey Organisation, Record 1992/97, 44 pp.

Tanahashi, M., Kisimoto, K., Joshima, M., Lafoy, Y., Honza, E. & Auzende, J.M., 1991. Geological structure of the central spreading system, North Fiji Basin. Marine Geology, 98, 187–200.

Taylor, B. & Exon, N.F. (editors), 1987. Marine geology and geophysics of the Woodlark Basin — Solomon Islands region. Circum-Pacific Council for Energy & Mineral Resources, Houston, Texas, Earth Science Series, 7.

Taylor, L.W.H. & Falvey, D.A., 1977. Queensland Plateau and Coral Sea Basin: stratigraphy, structure and tectonics. Australian Petroleum Exploration Association (APEA) Journal, 17, 13–29.

Umbgrove, J.H.F., 1947. The pulse of the Earth (2nd edition). Nijhof, The Hague.

Uruski, C. & Wood, R., 1991. A new look at the New Caledonia Basin, an extension of the Taranaki Basin, offshore North Island, New Zealand. Marine and Petroleum Geology, 8, 379–391.

Vedder, J.G. & Bruns, T.R. (editors), 1989. Geology & offshore resources of Pacific Island arcs — Solomon Islands and Bougainville, Papua New Guinea regions. Circum-Pacific Council for Energy and Mineral Resources, Earth Science Series, vol. 12, Houston, Texas.

Vially, R. & Mascle, A., 1994. Nouvelle Caledonie: rapport regional d'evaluation petroliere. Regional Reports, Institut Francais du Petrole, Paris.

von Stackelberg, U. & von Rad, U., 1990. Geological evolution and hydrothermal activity in the Lau and North Fiji basins (Sonne Cruise SO-35) — a synthesis. Geologische Jahrbuch, 92, 629–660

Walley, A.M., 1992. Cretaceous–Cainozoic palaeogeography of the New Zealand-New Caledonia region. BMR Record 1992/11 — Palaeogeography 42, Australian Geological Survey Organisation, Canberra.

Weissel, J.K., Hayes, D.E. & Herron, E.M., 1977. Plate tectonic synthesis: the displacements between Australia, Antarctica and New Zealand and Antarctica since the Late Cretaceous. Marine Geology, 25, 231–277.

Weissel, J.K., Taylor, B. & Karner G.D., 1982. The opening of the Woodlark Basin, subduction of the Woodlark spreading system, and the evolution of northern Melanesia since mid-Pliocene time. Tectonophysics, 87, 253–277.

Wiebenga, W.A., 1973. Crustal structure of the New Britain-New Ireland region. In: Coleman, P.J. (editor), The western Pacific — island arcs, marginal seas, geochemistry. University of Western Australia Press, Perth.

Yan, Chun Yeung & Kroenke, L.W., 1993. A plate tectonic reconstruction of the southwest Pacific, 0–100 Ma. In: Berger, W.H., Kroenke, L.W., Mayer, L.A. et al., Proceedings of the Ocean Drilling Program, Scientific Results, Volume 130. US Government Printing Office, Washington DC, 697–709.

Geophysical mapping using the national airborne and gravity datasets: an example focusing on Broken Hill

P.J. Gunn[1], P. Milligan[1], T. Mackey[1], S. Liu[1], A. Murray[1], D. Maidment[1] & R.Haren[1]

The Australian Geological Survey Organisation acquires and maintains extensive databases of airborne magnetic and radiometric surveys, digital elevation models and gravity measurements. These data can be accessed and processed to present information ranging from continental scale to detail at 1:25 000 scale. When data relating to the Broken Hill area of New South Wales are imaged it is possible to infer the distribution of similar rocks elsewhere in Australia and possible tectonic origins for the rocks hosting the Broken Hill Ag-Pb-Zn deposit. Interpretations using 1:100 000 and 1:25 000 scale data facilitate the production of detailed solid geology maps.

Introduction

The Australian Geological Survey Organisation (AGSO) has been acquiring and compiling aeromagnetic, radiometric, gravity and elevation data for over 40 years (Hone et al. 1997). The object of these activities is to assist geological mapping and synthesis, and thereby provide a basis for assessments of mineral and petroleum prospectivity, land use, geohazard identification, and groundwater studies. The national geophysical data sets maintained and distributed by AGSO provide a framework for specific studies by various government entities and private organisations such as exploration companies

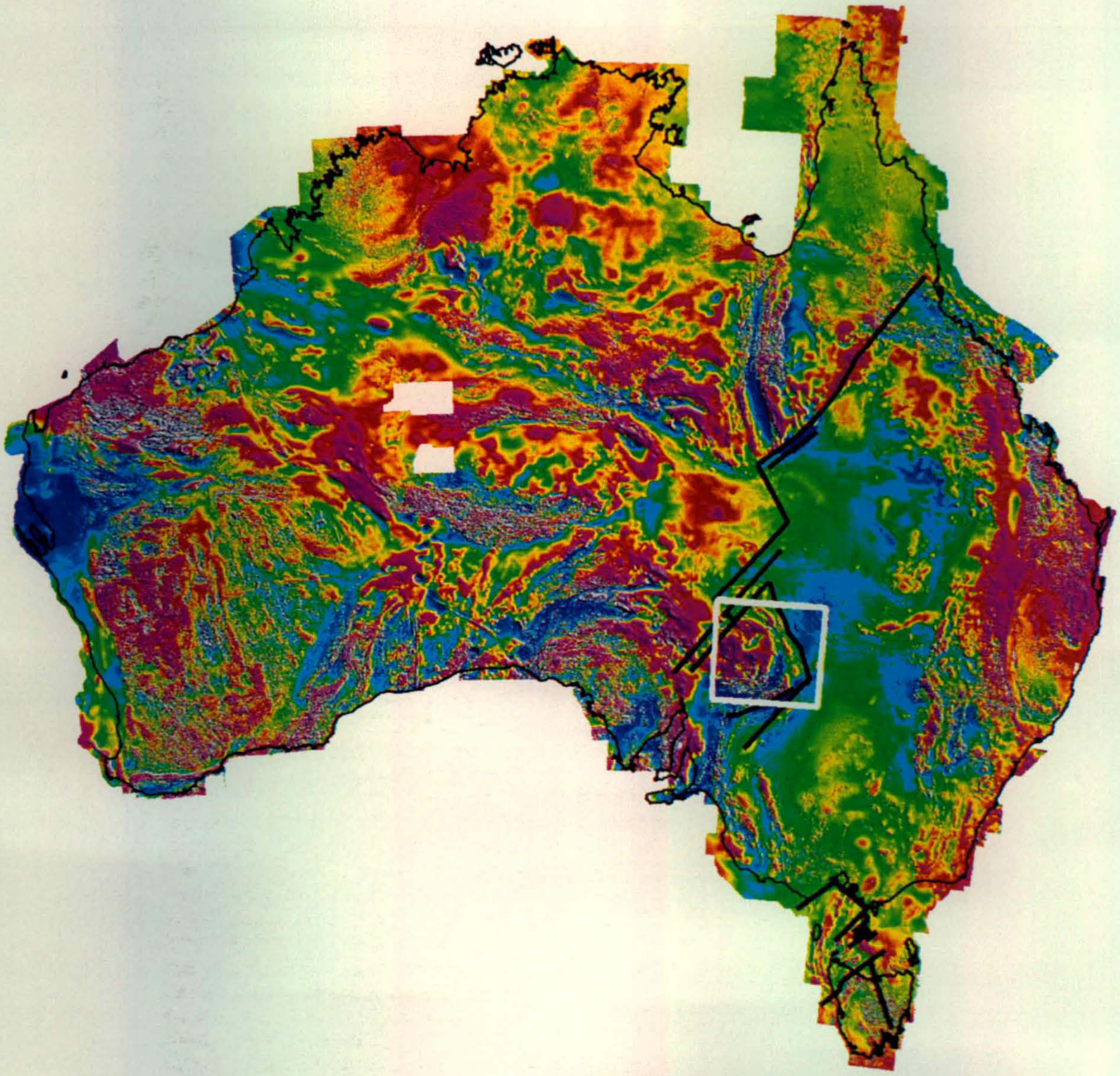


Figure 1. Magnetic anomaly map of Australia, originally displayed at a scale of 1:5 000 000. Red colours indicate higher values of the field. Major interpreted faults and lithological boundaries are indicated. The area outlined in white is the area illustrated in detail in Figures 5, 6 and 7.

[1] Australian Geological Survey Organisation, GPO Box 378, Canberra, ACT 2601, Australia

(Denham 1997).

The purpose of this paper is to outline the national airborne geophysical and gravity databases and, by considering several examples that focus on topics related to the Broken Hill area, to illustrate how such datasets can be successively studied at progressively more detailed scales, which can all assist with the resolution of specific geological problems. The interpretations presented below are based on preliminary assessments of the data and are included primarily to illustrate the types of approaches that can be used to assess such results. As a result, we do not attempt to discuss in detail the concepts presented, which, in any case, are only formative ideas developed during the initial stages of an ongoing program of assessing the regional and detailed geology of the Broken Hill area.

1:5 000 000 magnetic anomaly map of Australia

Figure 1 shows the second edition of the magnetic anomaly map of Australia (Tarlowski et al. 1996b), originally published at a scale of 1:5 000 000. The image is derived from approximately 6 million line km of total magnetic intensity data held in the national airborne geophysical database by AGSO. Most surveys (4.5 million km) were flown by AGSO (formerly the Bureau of Mineral Resources, Geology and Geophysics — BMR), using its own aircraft, as part of the airborne geophysical reconnaissance of Australia, which commenced in 1951. Some surveys were carried out by geophysical contractors operating under contract to AGSO and the State and Northern Territory geological surveys, either separately or in joint projects. A number of other surveys, carried out for the private sector, have been acquired by AGSO. In total, the map comprises the equivalent of 528 1:250 000 map sheet areas.

AGSO surveys before 1990 were flown mainly at an altitude of 150 m above terrain, with flight lines oriented either north–south or east–west and spaced at 1500–3200 m apart, although many of the earliest surveys were acquired with line spacings of more than 3200 m. Data from surveys with line spacings of more than 1600 m are substandard by current standards for line spacing and location accuracy. Since 1990, AGSO surveys have mainly been acquired with line spacings of 400 m or less at an altitude of 100 m or less.

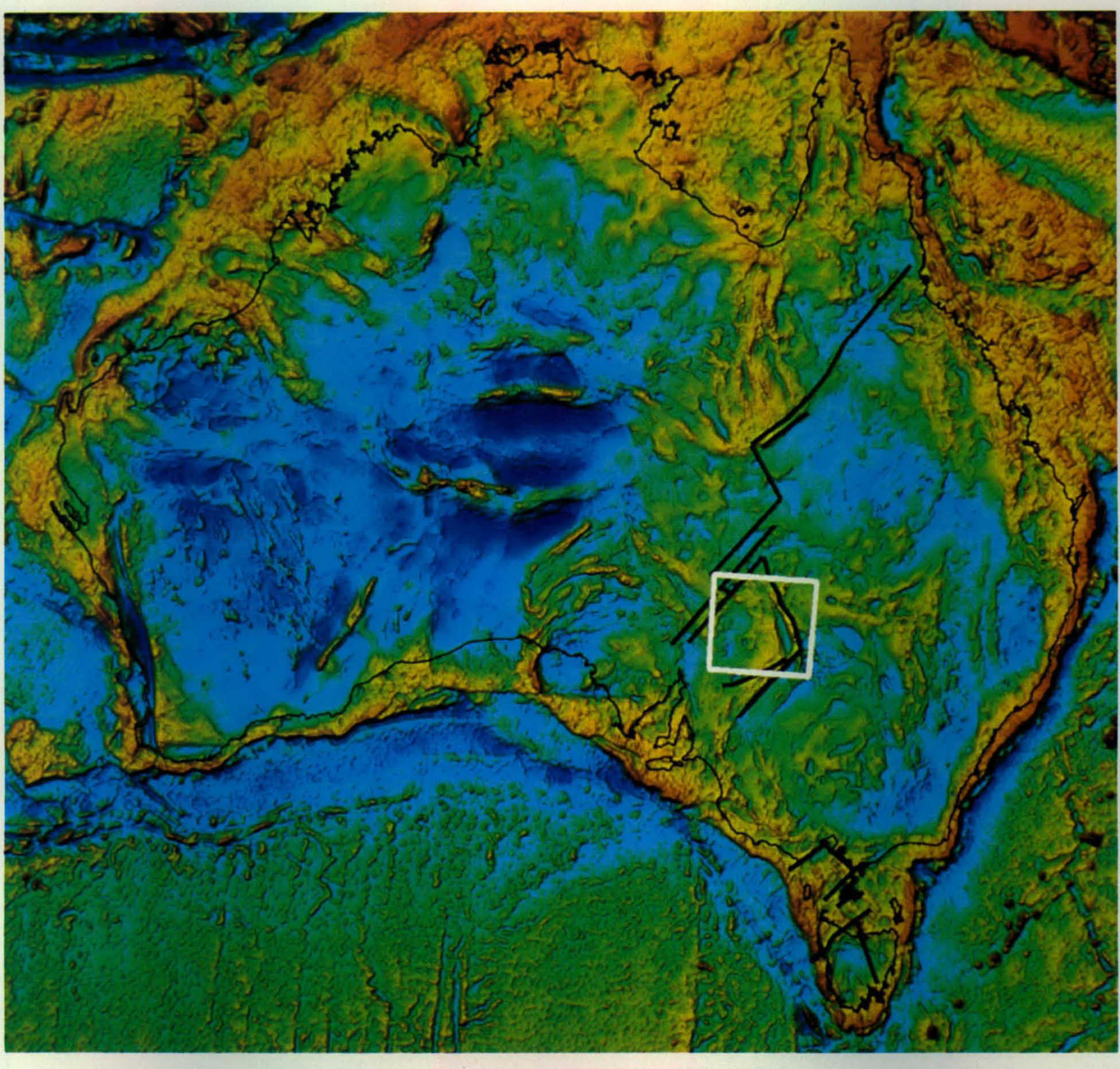


Figure 2. Gravity anomaly map of Australia, originally displayed at a scale of 1:5 000 000. Red colours indicate higher values of the field. Major interpreted faults and lithological boundaries are indicated. The area outlined in white is the area illustrated in detail in Figures 5, 6 and 7.

Data supplied by the Northern Territory Department of Mines and Energy, Mines and Energy South Australia, Tasmania Development and Resources, New South Wales Department of Mineral Resources, and Energy and Minerals Victoria were acquired at an altitude of 100 m or less above terrain along lines spaced at 500 m or less.

Hone et al. (1997) provide detailed index maps of the aeromagnetic database held by AGSO. Profile data for about one-third of the surveys, acquired in the earlier years of the program, were recorded in analogue form and have since been digitised. Other substandard analogue data cover 14 per cent of the land area (mostly the internal sedimentary basins). Coverage for these areas was obtained from the grid of the magnetic map of Australia (BMR 1976), digitised from contour maps. The original cell size of 0.02 degrees has been interpolated to 15 seconds of arc.

The International Geomagnetic Reference Field (IGRF) for an appropriate epoch has been removed from each survey, and a grid created with cell size of 15 seconds of arc (about 400 m), using a minimum curvature method (Briggs 1974). In some cases, the grids were subjected to 'micro-levelling' (Minty 1991).

Most of the individual surveys are approximately delineated by boundaries coinciding with 1:250 000 map sheet areas with a margin of overlap. By gridding according to geographical coordinates, surveys of adjacent sheets have coincident locations of marginal grid rows and columns in the overlap zones. The gridded data for pairs of 1:250 000 map sheets were registered and levelled by minimising (in a least squares sense) the discrepancies between grid values along their common rows or columns. To do this, low-order polynomials were fitted to the differences between values along grid rows (columns), and one grid was then adjusted to match the other grid by using the polynomial. Remaining high-frequency errors were smoothed out (Teskey et al. 1982). Where survey boundaries have complex and irregular shapes, Intrepid software, supplied by Desmond FitzGerald and Associates Pty Ltd, has been used for merging separate areas.

The composite grid across the Australian mainland was corrected for spurious warps introduced during compilation by using two special control lines, flown continuously around Australia early in 1990 (Tarlowski et al. 1996a). This project involved the accurate subtraction of diurnal variations by means of base-station recordings on a continent-wide magnetometer array. A third controlled loop, flown in late 1994 across Bass Strait, tied Tasmanian data to the mainland data. The geodetically based grid has been reprojected to a cell size of 800 m, using a simple conic projection with standard parallels of 18° and 36°S and a central meridian of 132°E.

The image (Fig. 1) was generated from the natural colour palette (magenta high, blue low) using histogram equalisation to maximise the use of the available colour range. Over 80 per cent of grid point values fall within 500 nT of the average value. To emphasise the expression of anomalies attributable to near-surface geology, a Sobel filter (Gonzalez & Woods 1992) was used to produce an artificial sun-angle 'illumination'. The output of this filter was used to modulate the colour intensity and saturation of the initial colour image after transformation to the hue, saturation, value (HSV) colour space (Milligan et al. 1992).

The original grid data with a cell size of 15 seconds of arc (about 400 m) used for the preparation of the image are available from AGSO in digital form for the whole continent, or as individual 1:1 000 000 sheets. The data can also be displayed at larger scales, revealing significantly more detail than the 1:5 000 000 image. AGSO can supply copies of the point located data for most of the surveys used in the compilation.

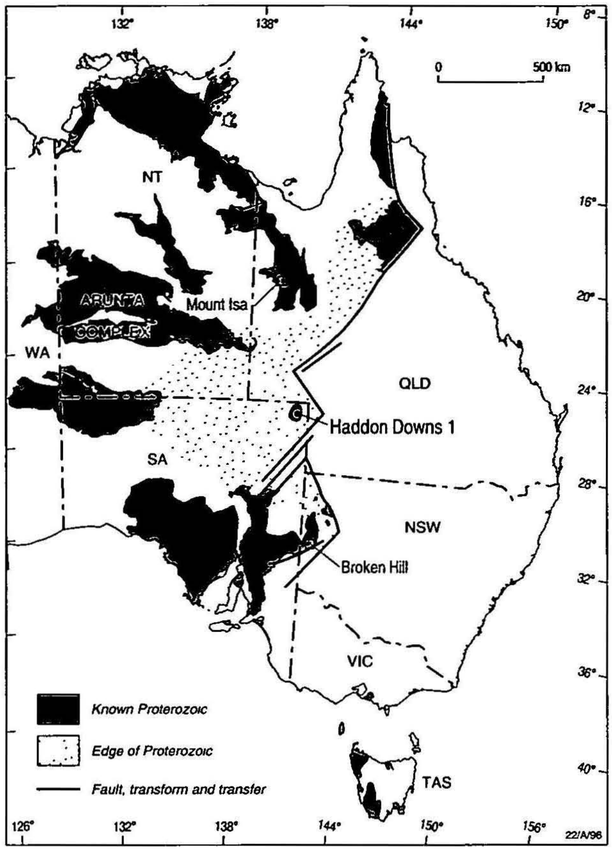


Figure 3. Australian geological setting of the Broken Hill deposit. The faults illustrated have been interpreted from the magnetic and gravity images of Figures 1 and 2. The margin of the Proterozoic, as defined by these faults, is known as the 'Tasman Line'.

1:5 000 000 gravity anomaly map of Australia

The gravity anomaly map of Australia (Milligan et al. 1992), at 1:5 000 000 scale, illustrates a gridded version of more than 800 000 point gravity observations held in the AGSO-maintained national gravity database. The grid mesh interval is 5 km. Bouguer anomalies have been used for onshore areas and free-air gravity for offshore. The map was produced with a 'diffuse' imaging process that combines sun angles in equal proportions from north, northeast, east, and southeast. These directions were chosen to maintain the illusion of 3-dimensional relief, while preventing the suppression of linear features that parallel the direction of a single sun angle. Another benefit of this method is that black regions, which are common with single sun angles, are rare on diffuse sun-angle images because areas that would be in shadow from a single sun angle are illuminated from other directions. The HSV colour space model has been used for the amplitude representation.

All Australia has been covered with a maximum station spacing of 11 km as a result of systematic reconnaissance surveys conducted by AGSO during the 1970s. However, many areas have a denser coverage as a result of more detailed surveys by AGSO, State organisations, and exploration companies. New gravity stations are added to the database as they become available. All data in the national gravity database can be obtained from AGSO. Figure 2 is an image of the most recent update of the database, supplemented from offshore gravity values derived from satellite altimetry (Milligan et al. 1997).

Australian context of the Broken Hill deposit

The Broken Hill Ag-Pb-Zn orebody (Laing et al. 1978, Haydon & McConachy 1987, Stevens et al. 1988, Stevens 1995) in western New South Wales is one of the world's great orebodies; it originally contained over 300 Mt of ore and has yielded more than $70 billion worth of metal since its discovery in 1883. The deposit is hosted by highly metamorphosed Proterozoic rocks of the Willyama Supergroup near the southeastern margin of a basement complex comprising Willyama Supergroup and other Proterozoic assemblages (Fig. 3). Outcrop of this basement complex — the Curnamona Province — is restricted to the southern part of the complex (Fig. 4).

Despite intensive exploration in the region for more than 100 years, Broken Hill remains the sole significant orebody known in the region. One of the great questions of Australian economic geology is whether other deposits of the type and size occur in the region. As rocks of the Willyama Supergroup are variably covered by late Proterozoic, Palaeozoic, Mesozoic, and Cainozoic sediments, the use of regional geophysical data is fundamental to understanding the occurrence of favourable host rocks for such deposits and regional and local-scale structures that may control their emplacement. The 1:5000 000 magnetic and gravity anomaly maps allow a continental-scale evaluation of the setting of the Curnamona Province, which hosts the Broken Hill deposit. A schematic interpretation, which places the Curnamona Province as an isolated block on the eastern margin of the Precambrian assemblages that cover most of central and western Australia, is superimposed on these images (Figs 1, 2).

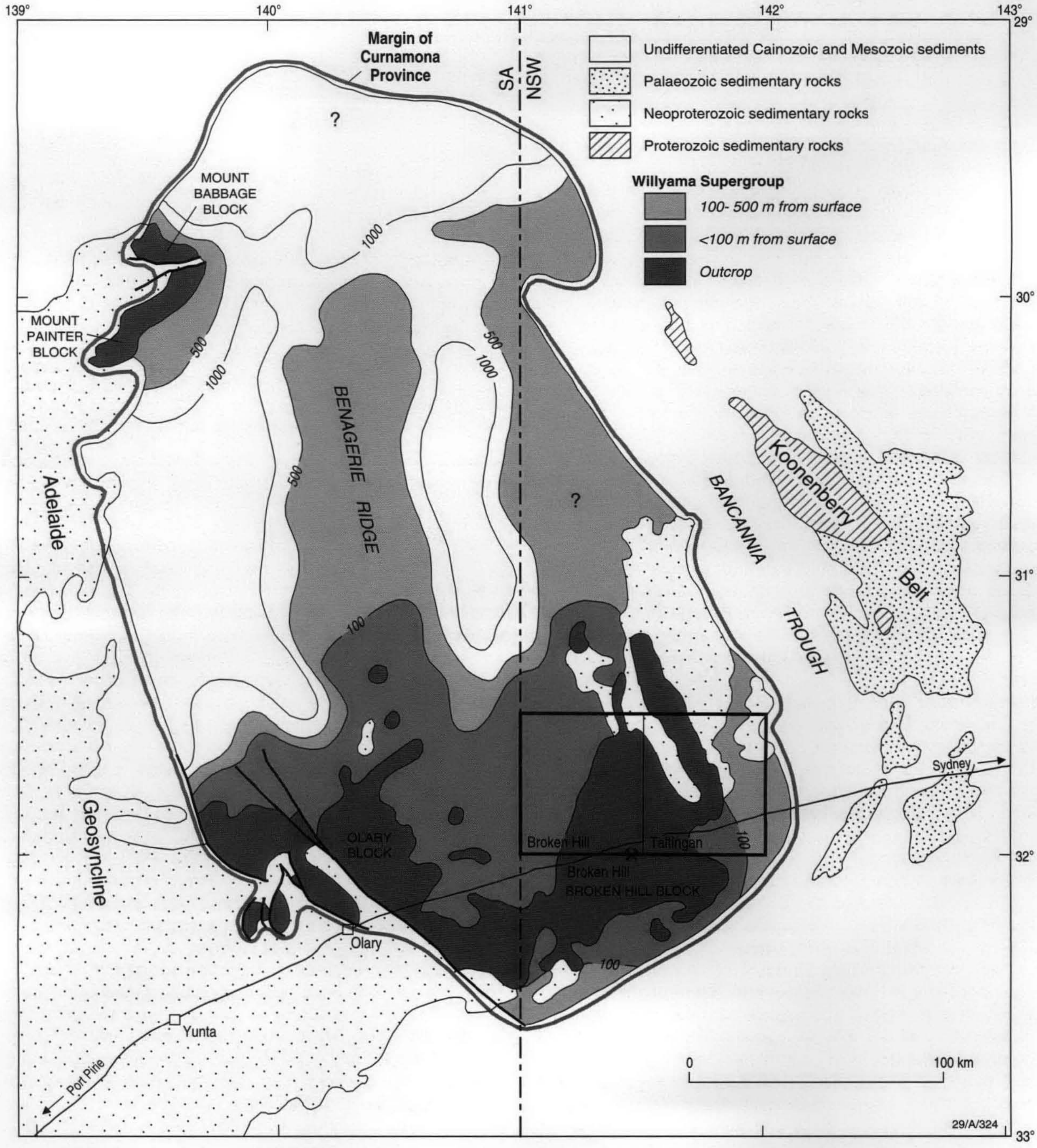


Figure 4. Regional geology of the Curnamona Province. The area outlined covers the area illustrated in Figures 8–11.

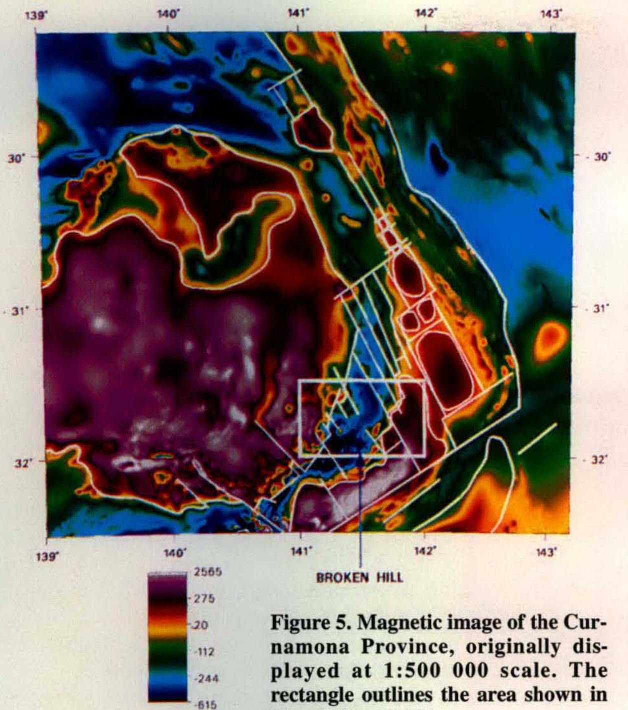


Figure 5. Magnetic image of the Curnamona Province, originally displayed at 1:500 000 scale. The rectangle outlines the area shown in Figs 8, 9 and 10.

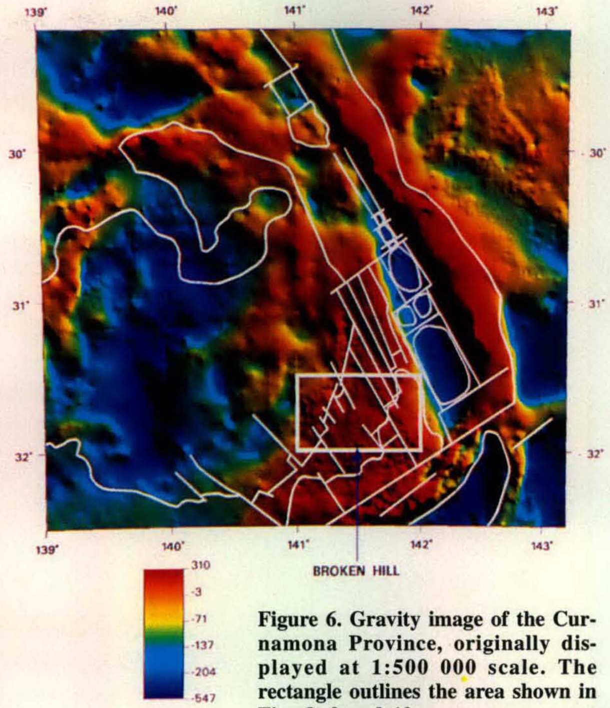


Figure 6. Gravity image of the Curnamona Province, originally displayed at 1:500 000 scale. The rectangle outlines the area shown in Figs 8, 9 and 10.

The abrupt change in magnetic and gravity character east of the Curnamona Province is interpreted as indicating the junction between the Proterozoic terranes of central Australia and the Paleozoic terranes of eastern Australia. The abrupt linear termination of the higher magnetic responses, which characterise much of the Curnamona Province south of the Broken Hill Province, suggests a major fault. This feature corresponds to a major mapped discontinuity called the Redan Fault. The gravity data do not reflect the position of this lineament as clearly, but show a major discontinuity slightly further south with a slightly different trend. The gravity discontinuity, which is also interpreted as indicating a major fault system, is paralleled on the northwestern edge of the Curnamona Province by gravity lineaments, which could also indicate major fault zones. These are not clearly defined in the total magnetic image (Fig. 1), but are more evident in derivative enhancements of the magnetic map of Australia (Tarlowski et al. 1997).

The major gravity discontinuity south of Broken Hill may mark the boundary between the Proterozoic rocks of the Broken Hill area and the Palaeozoic rocks known to the south. This discontinuity is sometimes referred to as the Darling Lineament, as it aligns with the trace of the Darling River, which extends in a northeasterly direction for more than 1000 km across New South Wales. The parallel fault systems along the northern limits of the Curnamona Province may mark a similar junction between the Proterozoic and the Palaeozoic. The western margin of the Curnamona Province is terminated by the Proterozoic rift of the Adelaide Geosyncline.

The junction between the extensive Proterozoic rocks in central and western Australia and the Palaeozoic rocks of eastern Australia, known as the Tasman Line, has been reviewed by Veevers (1984). It is commonly identified as a former divergent continental margin from which fragments of a once more extensive Proterozoic landmass broke away and migrated to an unknown position in the east. Figure 3 shows an interpretation of the geometry of the Tasman Line based on the gravity and magnetic data used to compile Figures 1 and 2. It indicates an irregular continental margin with a semi-rectangular outline formed by a combination of major parallel ruptures, which trend northwesterly, and major transform fault dislocations, which trend northeasterly. The exact trace of the non-transform portion of the boundary is difficult to identify in the area immediately northeast of the Curnamona craton because of the poor quality of the aeromagnetic data in this area and the obscuring gravity effects of various Palaeozoic sedimentary basins. Many workers place the boundary much further west than is shown in Figure 3. The interpretation presented has been influenced by the Haddon Downs 1 well, which has been interpreted as bottoming in Proterozoic rocks similar to those cropping out in the Arunta Complex of the southern Northern Territory (Rankin & Gatehouse 1990), and the magnetic image, which strongly suggests that basement rocks in the Arunta Complex and the Haddon Downs 1 area are similar.

If the above interpretations are correct, the following major tectonic correlations may be proposed regarding the localities of Willyama Supergroup equivalents.

1. As the opening Neoproterozoic Adelaide rift (Drexel 1987) postdates the deposition of the Willyama Supergroup, equivalents of the Willyama Supergroup may exist west of the Adelaide Geosyncline.
2. Eastward continuations of Willyama Supergroup will, if they occur, be terminated at the junction of the Palaeozoic and the Proterozoic.
3. The Mount Isa Inlier (Blake et al. 1990), which contains a series of Proterozoic rifts and which is terminated on its southeastern side by abrupt magnetic and gravity discontinuities (which are interpreted as indicating a major transform fault) may have temporal and genetic affinities with the Broken Hill area. This idea has been proposed by many workers (e.g. McConachie et al. 1993)). However, there is no evidence in the regional images for a connection beneath the broad Great Australian Basin separating the two areas. If Proterozoic units were originally continuous between the two areas, the continuation would now appear to be offset far to the east of the Mount Isa area. The magnetic and gravity data clearly indicate that the rocks of the Mount Isa area continue under cover for several hundred kilometres until they are terminated by the interpreted transfer fault zone.
4. Continuations of Proterozoic lithology south of the Curnamona Craton have drifted to unknown locations to the northeast.

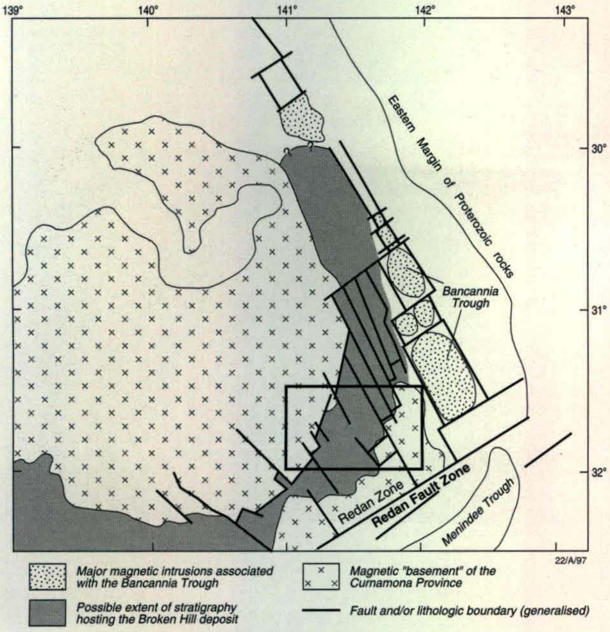


Figure 7. Interpretation of the main lithological and structural elements evident in the 1:500 000 scale magnetic and gravity images shown in Figures 5 and 6.

Magnetic and gravity fields of the Broken Hill area

Figures 5 and 6 are images of magnetic and gravity data over the Curnamona Province (originally produced at a scale of 1:500 000), and allow an appreciation of the distribution of the internal units of the province. A regional interpretation of these data is shown in Figure 7.

The data in Figure 5 have been continued upwards to an elevation of 1000 m above the level of the data of Figure 1, i.e. they show the magnetic field that would have been observed at 1000 m above the level of the original survey flight elevation. The 'upward continuation' process is a filtering operation that suppresses the effects of the shorter wavelength anomalies caused by near-surface magnetic features and, thereby, simplifies the recognition and interpretation of the major magnetic units.

Figure 5 shows that the Broken Hill orebody is situated within an area of low magnetic intensity that appears, from the magnetic data, to extend in an elongated arc northeast and southwest of the deposit. The region of low magnetic intensity is flanked on its northern and southern margins by intensely magnetic units, which appear typical of the core of the Curnamona Province. Northeast of Broken Hill, the basement magnetic character has a lower amplitude and is smoother than over the central portion of the Curnamona Province. This wedge-shaped area, which encompasses several ovoid magnetic highs, exactly overlies the Palaeozoic Bancannia Trough (Evans 1977, Department of Mineral Resources 1993), which has been identified as a Devonian rift. The magnetic high corresponding to the Bancannia Trough appears to be associated with the evolution of the trough, even to the extent that its geometry reflects greater extension in the southeastern part of the trough, and it could possibly be due to a thick sequence of basic volcanics on the floor of the trough. An equally likely explanation is that the anomaly reflects magma generation and intrusion caused by mantle decompression associated with crustal thinning and extension, which could be expected to be associated with the formation of the Bancannia Trough (cf. processes described by Gunn 1997). The line of magnetic anomalies axially distributed along the crest of the Bancannia

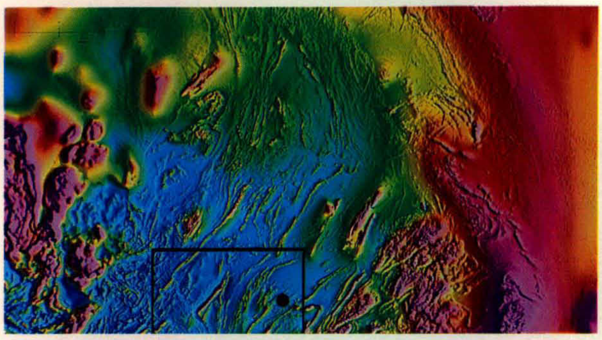


Figure 8. Pixel image of the total magnetic intensity of the combined Broken Hill and Taltingan 1:100 000 scale map sheet areas, originally displayed at 1:100 000 scale. Red colours indicate highest intensity. The area outlined by the rectangle is the area displayed in detail in Figures 12 and 13.

Trough is typical of the progressive development of intrusions associated with crustal thinning. The recognition of these anomalies as being associated with Devonian processes, rather than Proterozoic processes associated with the Curnamona Province, is important, as it confines the distribution of stratigraphic units equivalent to those at Broken Hill.

The gravity image of the Curnamona Province (Fig. 6) shows a regional high in the southeastern part of the province. To fully understand this gravity field, it is again necessary to appreciate the effect of the Bancannia Trough. The pronounced low over the Bancannia Trough can be explained by the Devonian sediments in the rift, as known from outcrop geology, drilling, and seismic reflection studies. The origin of the linear highs that flank this gravity low is intrinsically less obvious, but can be explained in terms of the normal gravity expression of rifts overlying thinned crust. As explained by Gunn (1984, 1997), crustal thinning causes broad gravity highs, owing to the dense mantle material at shallow depths beneath rifts. Such highs normally extend for considerable distances adjacent to the actual rift zone because of the depth of the source. If the rift contains much thickness of low-density sediments, a gravity low will be produced, which will, in general terms, closely overlie its relatively shallow low-density source. The combined effect of these two anomalous responses will be a gravity low overlying the actual rift flanked by gravity highs adjacent to the rift. This is exactly the gravity anomaly pattern observed for the Bancannia Trough. It thus appears necessary to regard the northwestly trending gravity high 60 km northeast of Broken Hill as being unrelated to the gravity response of Willyama Supergroup equivalents.

After isolating the gravity and magnetic responses of the Proterozoic units in the Broken Hill region, it is possible to make specific interpretations of the distribution and origin of the rocks hosting the Broken Hill deposit. As noted above, the rocks hosting the deposit appear to occur in a distinctive zone of low magnetic intensity, i.e. the magnetic low appears to map the distribution of the prospective section which contains the Broken Hill deposit. Conversely, the magnetic data imply that the bulk of the Curnamona Province in the area north and south of Broken Hill is composed of rocks that are magnetically similar, and different from the sequence hosting the Broken Hill deposit.

The magnetic pattern can be interpreted as indicating that the host rocks of the Broken Hill deposit formed in a rift caused by extension splitting rocks originally forming the core of the Curnamona Province. According to this idea, the magnetic area southeast of Broken Hill known as the Redan Zone would have been originally adjacent to the magnetic area northwest of Broken Hill. The magnetic data suggest a series of north-northwest-trending dislocations that may have

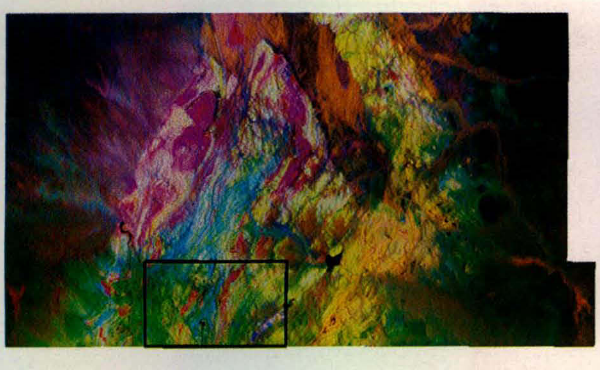


Figure 9. Composite radiometric pixel image of the combined Broken Hill and Taltingan map sheet areas, originally displayed at 1:100 000 scale. Red colours indicate K, blue Th, and green U. The area outlined by the rectangle is the area displayed in detail in Figures 12 and 13.

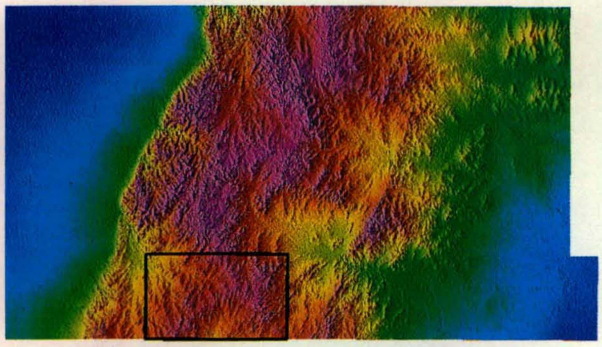


Figure 10. Digital elevation pixel image of the combined Broken Hill and Taltingan map sheet areas, originally displayed at 1:100 000 scale. Red colours indicate highest intensity. The area outlined by the rectangle is the area displayed in detail in Figures 12 and 13.

originated as zones that could have accommodated differential extension during a rifting process. Such faults are clearly defined, albeit with more complex traces than indicated in Figures 5 and 6, by more detailed magnetic images and geological mapping (see below). A broad gravity high over the zone of low magnetic intensity could be evidence of crustal thinning associated with the rifting process. The absence of any major deep magnetic bodies obviously associated with the proposed rift zone does not negate the rift idea, as it has been shown that slow extension during rifting allows mantle cooling without depressurisation and magma generation (White 1992). Not all rifts are associated with significant magma generation.

The above suggestion that the Broken Hill deposit occurs in what was originally a Proterozoic rift accords with the ideas of many contemporary workers (reviewed by Stevens 1995); however, even if the rift interpretation is not correct, the regional analysis of the gravity and magnetic data has indicated the likely distribution of rock units is similar to those hosting the Broken Hill orebody (Fig. 5).

1:100 000 magnetic, radiometric and digital elevation maps of the Broken Hill area

The acquisition and interpretation of detailed airborne geophysical data in the Broken Hill area are components of the ongoing Broken Hill Exploration Initiative, a collaborative National Geoscience Mapping Accord (NGMA) project involving AGSO, New South Wales Department of Mineral Resources, and Mines and Energy South Australia. The aim of the project is to provide a new generation of geoscientific information for the Broken Hill region as a basis for more effective and efficient exploration by the mineral industry.

The total magnetic intensity pixel image for the combined Broken Hill and Taltingan 1:100 000 sheets is reproduced in Figure 8. The image has been produced from data acquired at line spacings of 100–400 m. The data have been 'reduced to the pole', i.e. processed in such a way that asymmetry caused by the inclination of the Earth's magnetic field has been removed and anomalies appear directly above their magnetic sources, except in relatively special situations where rock units have significant permanent magnetisation in a direction other than that of the Earth's field. The image has also been illuminated with a 'sun angle' to enhance northerly trending features. The central zone, which provides maximum resolution, was flown at 100 m line spacing and 60 m flight height.

The magnetic image at this scale reveals several lithostratigraphic units, where anomalies have similar and distinct magnetic texture and appearance. Some of them can be correlated with geologically mapped units. In areas of no outcrop, the lithology causing an anomaly is inferred on the basis of similarity with units for which control is available. Where this is not possible, magnetic units are mapped on the basis of geometry and amplitude without assigning them a specific lithology. Structure is clearly indicated by the traces of the various long, narrow anomalies, and numerous faults are evident as discontinuities and offsets in the magnetic units. Magnetite has been localised along some fault planes and faults and, in such cases, causes linear magnetic highs.

Radiometric responses arise from concentrations of radioactive elements derived from the decay of isotopes of potassium (K), thorium (Th) and uranium (U). Airborne radiometric surveys detect such elements when they occur within 30 cm of the ground surface. The 1:100 000 composite gamma-ray spectrometric image (Fig. 9; K red, Th green, U blue) for the combined Broken Hill and Taltingan Sheet areas provides information on bedrock, structure, weathering, and the character and mobility of the regolith in the region. Individual images, not illustrated here, were produced for the K, Th, and U count rates. The radiometric data enable the identification of different intrusive bodies, reflecting their variable and often distinct chemistry. Bedding in metamorphic units that were originally shales is clearer in the radiometric data than the magnetic data because shale contains high K concentrations that undergo minimal change during metamorphism. When bedrock is covered by overburden, radiometric data reflect contributions of weathered bedrock and transported material. The image clearly indicates the movement of regolith material along drainage channels and fans. Radiometric responses can be used to identify and map regolith material.

Figure 10 shows a digital elevation model (DEM) computed at 1:100 000 scale for the combined Broken Hill and Taltingan map sheets by subtracting radar altimeter values from the true altitude of the survey plane as determined by GPS. DEMs are used to assist mapping of bedrock and regolith. Combinations of gamma-ray radiometric data and DEMs are helpful in defining regolith mobility and calculating the energy associated with regolith distribution.

The 1:100 000 magnetic, radiometric, and DEM maps and images, together with digital vectors of the known outcrop geology and various specialised enhancements of the data set (such as the derivative maps discussed below), have been combined as overlays in a geographic information system (GIS), which has been used as a basis for determining broad scale structural and lithological relationships (Fig. 11). The interpretation has been confined to the area of the Broken Hill and Taltingan 1:100 000 map sheets, between a major fault bounding the prospective stratigraphy in the west (the Mundi Mundi Fault) and the western edge of the Devonian sediments in the east. These major geological boundaries are clearly evident in the images of Figures 8, 9 and 10. More

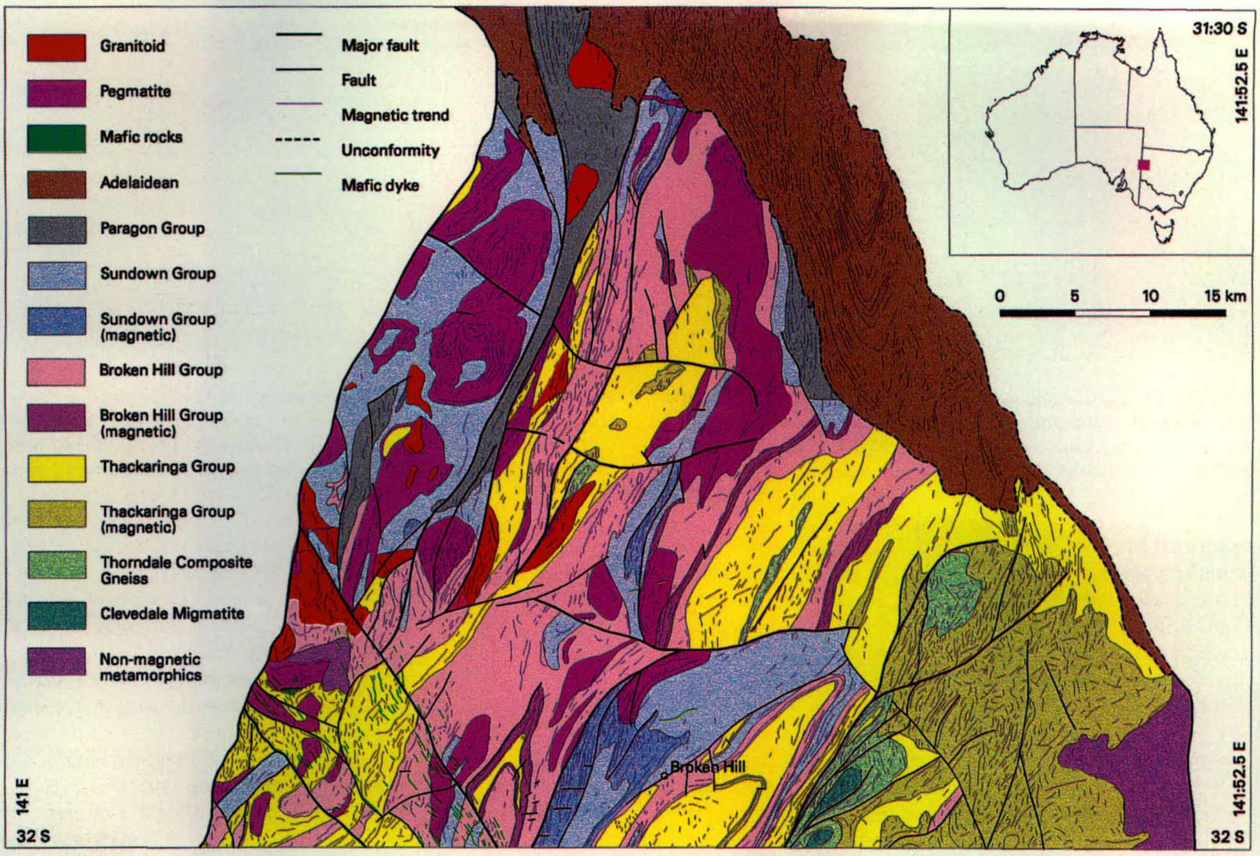


Figure 11. A solid geology interpretation using 1:100 000 scale data covering the central portions of the Broken Hill and Taltingan 1:100 000 map sheets. The object of this map is to provide an overview of the main lithostratigraphic units and structure and thereby provide a basis for more detailed mapping and geophysical interpretation at 1:25 000 scale.

detailed analysis at finer scales is required for a complete assessment of the area. The interpretation illustrated in Figure 11 gives an overview that can be used as a basis for such work.

Detailed interpretations using 1:25 000 maps and images

New South Wales Department of Mineral Resources has carried out geological mapping at 1:25 000 scale in the Broken Hill region for many years. The objective in acquiring new high-resolution aeromagnetic data is to allow refinement of this mapping, particularly in poorly exposed areas. Airborne surveys that provide magnetic data at line spacing of 100 m or less are suitable for analysis at 1:25 000 scale. Magnetic data are measured at 7 m intervals along each profile and with 100 m line spacing it is possible to generate realistic pixel images using cell sizes of 20 m. Gamma-ray radiometric data are integrated over 70 m along profile lines — thus one data point is generated for each 70 m of line, ten times less dense than the magnetic data. Radiometric data collected along 100 m spaced flight lines can be processed with cell sizes of 30 m. This grid size produces blocky images at 1:25 000 scale, but they are still useable for interpretation.

Figure 12 shows the total magnetic intensity image for the 1:25 000 Broken Hill sheet. The city of Broken Hill is located in the region of erratic anomalies just above the southeastern corner of the image. The anomalies contain effects caused by the aircraft flying at higher levels over the city and responses from cultural artefacts in the city and associated mines. No obvious anomaly is associated with the orebody and, if one was originally present, it is likely that the extensive mining activity over the last century would have eliminated or significantly reduced its effects.

Figure 13 shows a greyscale image of the 1:25 000 first vertical derivative of the total magnetic intensity greyscale image for the Broken Hill map sheet area. Derivative images have the advantage of resolving anomalies whose effects coalesce in total magnetic intensity images. In the image of Figure 13, the strong magnetic units are shown as very white zones. Colour versions of this image, not illustrated here, also improved the detail of the total magnetic intensity image. A colour derivative image has the advantage that it clearly shows the relative amplitudes of features.

Fold structures and associated parallel antiforms are evidenced by strong magnetic units, such as banded iron formations, amphibolite, and magnetite-bearing pelitic and psammitic metasediments. The images of Figures 12 and 13 show evidence of a broad-scale dyke suite trending north-northwest on its western side. The broad, high magnetic anomaly cutting across the dyke suite on the upper left-hand side of the image is associated with amphibolites and rock units with abundant magnetite. The long, arcuate magnetic high that starts in the bottom centre of the image and traces a line bending to the northeast is associated with magnetite-rich pelites and psammites. Some retrograde metasediments have very high magnetic susceptibility and, in some cases, the magnetic anomalies are associated with magnetite which has invaded a number of different rock types adjacent to the most magnetic unit. The complex pattern east of the centre of the western margin of the image appears to be associated with a dilation or a tightly folded system. The rocks that produce the high magnetic anomalies are quartz magnetites and magnetite-rich metasediments. Broad north–south-trending retrograde schists with very low magnetic susceptibility have been mapped in this area.

Solid geology interpretations at 1:25 000 scale, based on a combined analysis of the known outcrop geology and the geophysical data set described above, are currently in progress.

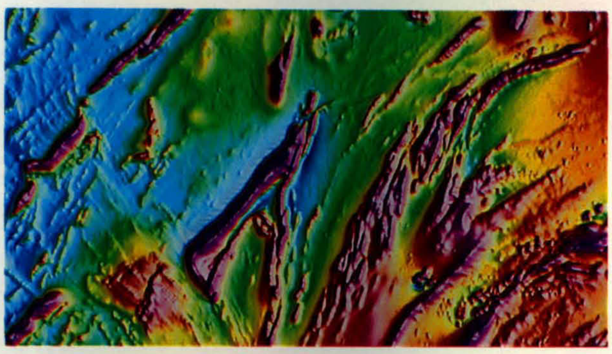


Figure 12. Gradient-enhanced colour pixel image of the total magnetic intensity for the Broken Hill 1:25 000 scale map sheet area, originally displayed at a scale of 1:25 000. Red colours indicate highest magnetic intensity.

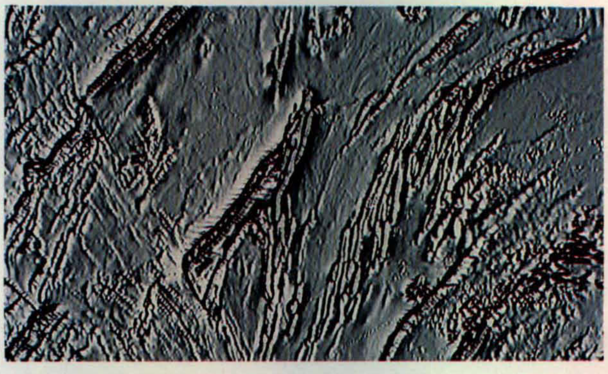


Figure 13. Gradient-enhanced greyscale pixel image of the first vertical derivative of the total magnetic intensity for the Broken Hill 1:25 000 scale map sheet area, originally displayed at a scale of 1:25 000.

The object of this ongoing work is to refine the lithological and structural mapping of the area with the ultimate goal of producing a three-dimensional understanding of the geology of this area. A key part of this project is the integration of detailed rock-property measurements. These are assisting lithological identifications using the geophysical data, and will provide quantitative bases for computer modelling of anomalies to resolve the geometry of their sources. The modelling, incorporating drill data where possible, defines the depth, strike, width, and dip of the magnetic sources.

Conclusion

We have shown how different geophysical data sets at various scales enable geoscientists to provide interpretations that are relevant to broad scale as well as detailed geological studies and exploration strategies. The examples presented range from continental-scale domains and structural zones (1:5 000 000), through a regional assessment of a particular province (1:500 000), to 1:100 000 and ultimately 1:25 000 scale studies of specific units and structures.

The Broken Hill analysis is benefiting from the availability of 100 m line spacing in the airborne data, which allows analysis at 1:25 000 scale. Standard National Geoscience Mapping Accord projects normally involve the acquisition of airborne data with 400 m line spacing. Such line spacing produces data that can only realistically be used at 1:100 000 scale. The Broken Hill Exploration Initiative Project has clearly illustrated the benefits of closer line spacings.

References

Blake, D.H., Etheridge, M.A., Page, R.W., Stewart, A.J., Williams, P. & Wyborn, L.A.I., 1990. Mount Isa Inlier — regional geology and mineralisation. In: Hughes, F.E. (editor), Geology and mineral deposits of Australia and Papua New Guinea. Australasian Institute of Mining and Metallurgy, Monograph 14, 915–925.

BMR, 1976. Magnetic map of Australia, residuals of total intensity, scale 1:2 500 000. Bureau of Mineral Resources, Australia.

Briggs, I.C., 1974. Machine contouring using minimum curvature. Geophysics, 39, 39–48.

Denham, D., 1997. Applications and strategic importance of aeromagnetic and gravity data. AGSO Journal of Australian Geology & Geophysics, 17(2), 3–9.

Department of Mineral Resources, 1993. Petroleum Data Package — Darling Basin. New South Wales Department of Mineral Resources (unpublished).

Drexel, J.F. (editor), 1987. The Adelaide Geosyncline: late Proterozoic stratigraphy, sedimentation, palaeontology and tectonics. Government Printer South Australia, Adelaide, 438 pp.

Evans, P.R., 1977. Petroleum geology of Western New South Wales. APEA Journal, 17, 42–49.

Gonzalez, R.C. & Woods, R.E., 1992. Digital image processing. Addison-Wesley Publishing Company, New York, 716 pp.

Gunn, P.J., 1984. Recognition of ancient rift systems: examples from the Proterozoic of South Australia. Exploration Geophysics, 15, 85–97.

Gunn, P.J., 1997. Regional magnetic and gravity responses of extensional sedimentary basins. AGSO Journal of Australian Geology & Geophysics, 17(2), 115–131.

Haydon, R.C. & McConachy, G.W., 1987. The stratigraphic setting of the Pb-Zn-Ag mineralisation at Broken Hill. Economic Geology, 82, 826–856.

Hone, I.G., Milligan, P.R., Mitchell, J.N. & Horsfall, K.R., 1997. Regional airborne magnetic and gamma-ray databases in Australia. AGSO Journal of Australian Geology & Geophysics, 17(2), 11–21.

Laing, W.P., Marjoribanks, R.W. & Rutland, R.W., 1978. Structure of the Broken Hill mine area and its significance for the genesis of the orebodies. Economic Geology, 73, 1112–1136.

McConachie, B.M., Barlow, M.G., Dunster, J.N., Meaney, R.A. & Schaap, A.D., 1993. The Mount Isa Basin — definition, structure and petroleum geology. APEA Journal, 33, 237–257.

Milligan, P.R., Morse, M.P. & Rajagopalan, S., 1992. Pixel map preparation using the HSV colour model. Exploration Geophysics, 23, 219–224.

Minty, B.R.S., 1991. Simple micro-levelling for aeromagnetic data. Exploration Geophysics, 22, 591–592.

Rankin, L.R. & Gatehouse, C.G., 1990. The northern margin of the Warburton Basin. Quarterly Geological Notes, Geological Survey of South Australia, 113, 14–16.

Stevens, B.J.P., 1995. A review of Broken Hill regional geology. In: Facer, R.A. & Branagan, D.F. (editors), 8th Edgeworth David Symposium: Mineral search in the south-west Pacific region — ma memorial to Ken Glasson. University of Sydney, 37–50.

Stevens, B.J.P., Barnes, R.G., Brown, R.E., Stroud, W.J. & Willis, I.L., 1988. The Willyama supergroup in the Broken Hill and Euriowie Blocks, New South Wales. Precambrian Research, 40/41, 297–327.

Tarlowski, C., Gunn, P.J. & Mackey, T., 1997. Enhancements of the magnetic map of Australia. AGSO Journal of Australian Geology & Geophysics, 17(2), 77–82.

Tarlowski, C., McEwin, A.J., Reeves, C.V. & Barton, C.E., 1996a. De-warping the composite aeromagnetic anomaly map of Australia using control traverses and base stations. Geophysic, 61(3), 696–705.

Tarlowski, C., Milligan, P.R. & Mackey, T., 1996b. Magnetic anomaly map of Australia (2nd edition), scale 1:5 000 000. Australian Geological Survey Organisation, Canberra.

Tarlowski, C., Simonis, F., & Milligan, P., 1993. Magnetic anomaly map of Australia, scale 1:5 000 000. Australian Geological Survey Organisation, Canberra.

Teskey, D.J., Dodds, S.D. & Hood, P.J., 1982. Compilation techniques for 1:1 000 000 magnetic anomaly maps series. Current Research, Part A, Geological Survey of Canada, Paper 82-1A, 351–358.

Veevers, J.J., 1984. Phanerozoic earth history of Australia. Oxford Science Publications, Oxford, 418 pp.

White, R.S., 1992. Magmatism during and after continental break-up. In: Storey, B.C., Alabaster, T. & Pankhurst, R.J. (editors), Magmatism and causes of continental breakup. Geological Society Special Publication 68, 1–16.

AGSO Journal of Australian Geology & Geophysics, 17(1), 137–143

The sesquicentennial of palaeomagnetism and rock magnetism in Australia

P.W. Schmidt[1]

Beginning with some extraordinary observations made by Strzelecki in 1845, a brief history of the development of palaeomagnetism and rock magnetism in Australia is presented. Through a number of factors, ranging from prescient leadership to Australia's special geography and unique geological past, Australia's contribution to these fields surpasses what might otherwise be expected. Although this was particularly so in the early days, it continues to be so today. Australian palaeomagnetic and rock magnetic research currently leads the world in both applied areas, such as mineral exploration, and in fundamental areas, such as geodynamo mechanisms. Australia's high profile in IAGA Palaeomagnetism and Rock Magnetism database development can be traced back to the natural advantage bestowed by the dedication of Irving, and later McElhinny, at ANU in their frequent publication of the Geophysical Journal of the Royal Astronomical Society pole lists. The major Australian laboratories and the contributions that they have made are outlined.

Introduction

> Its colour in the recent fracture is blackish green; on the surface, yellowish brown. The lustre of the paste waxy; that of the hornblende which it contains vitreous; it does not adhere to the tongue, and exhales an argillaceous odour; its streak is dissimilar and dull; its colour a brownish grey; when struck with a hammer, it gives a metallic sound: it is compact, hard, its fracture is somewhat concoidal. The structure is prismatic, the prisms having three, four, five, six, or seven sides. Their diameter varies from three to eight feet; the length of two or three columns, which are still entire, exceeds 100 feet. The clustered columns are sometimes very closely united; sometimes they are only in close contact, and are separated by the fall of the masses. Some of the columns have but a slight influence upon the magnetic needle; and in these the axes range east and west. The columns lying parallel with the meridian, or nearly so, disclose a strong polarity; a phenomenon worth noting, as the property seems to be more dependent on the bearing of the axes of these columns than on their constituents. The discovery of this polarity was consequent upon the anomalous results which the observations of the magnetic intensity furnished me by the prismatic greenstone on Ben Lomond (*Strzelecki 1845 p. 104*).

The above account given by Paul Strzelecki in 1845 is probably the earliest reference to rock magnetism in Australian literature. The 'greenstone' on Ben Lomond he refers to is clearly the Tasmanian Dolerite. Strzelecki recognised and described the dependence of magnetic effects on the orientation of fallen columns. These observations were not simply mundane compass deflections. It would appear that Strzelecki measured the total field anomalies of a number of columns while taking routine magnetic intensity measurements. Indeed, had Strzelecki been distracted further by his curious observations, and had he the fortune of understanding contemporary developments in the research of electromagnetic phenomena, he would have concluded that the Tasmanian Dolerites were magnetised in a steep-upward direction. Of course he would then have deduced that Tasmania was near the magnetic south pole when these rocks formed and wondered if the pole came to Australia or Australia to the pole. By themselves, his observations would not have been enough to differentiate between the two possibilities now known as polar wander and continental drift. Clarification of this momentous question required another century or so and observations from rocks from other continents.

Quite apart from the above, the Tasmanian Dolerite has played a central role in the development of palaeomagnetism, not just in Australia but globally. Another Australian rock formation that has become enshrined in palaeomagnetic history is the Gerringong Volcanics, on the Illawarra coast, NSW, previously known as the Upper Marine Latites. An oriented sample from the Blowhole Flow at Kiama had been collected in the 1920s by Dr W.R. Browne, University of Sydney, and sent to P.L. Mercanton in France. Mercanton's interest was magnetic polarity, and it transpired that these Kiaman rocks were reversely polarised, the first Palaeozoic rocks to be found with this polarity (Mercanton 1926). This is discussed further below.

This paper is a personal perspective on Australian palaeomagnetic and rock magnetic studies. A brief history is given, with an emphasis on the important role these studies have played in the early days. It is difficult today to appreciate the courage that early palaeomagnetic studies took, with almost no-one believing in their interpretation and implications. I have concentrated on contributions of the main laboratories, although it is recognised that several universities have acquired basic measurement equipment, and some of these, it is hoped, may develop significant research programs in the future.

Early days

One of the most exciting developments in geophysics must surely have been the emergence of polar wandering from the mire of early palaeomagnetic observations. Irving's (1956a) confirmation of an ancient axial dipole, that is, a palaeomagnetic dipole field co-axial with palaeolatitudinal zonings gave rise to the axial geocentric dipole (AGD) hypothesis that underpins the whole palaeomagnetic edifice. For reasons unknown, this has since become known as the GAD hypothesis.

The debate that followed on polar wander versus continental drift divided the Earth sciences research community, with some of the more open minded authorities changing their views rapidly as new evidence became available. In comparing results from Europe and the USA, Runcorn (1956a) concluded that 'the agreement is sufficiently close to give strong support to the hypothesis of polar wandering', later converting to 'these seem to require about 20° of displacement of America west from Europe in post Triassic times' (Runcorn 1956b). Revisitation of these judgements conveys the sense of uncertainty that must, initially at least, accompany great discoveries.

Ted Irving arrived in Australia in 1954 from Cambridge to take up a research position with Professor John Jaeger at the forerunner of the Research School of Earth Sciences at the Australian National University (ANU) in Canberra. Runcorn probably learnt of Irving's (1956b) work before publication, so it is not surprising that he quickly changed his mind. The difference between the polar wander paths of Europe and North America that Runcorn had been reporting is one of longitude, because these landmasses had apparently moved latitudinally in the same sense, and at about the same rate for the past few hundred million years. On the other hand, Australia has moved in the opposite sense latitudinally compared with

[1] CSIRO Division of Exploration and Mining, PO Box 136, North Ryde NSW 2113, Australia

Europe and North America, and Irving's pole path for Australia shows this dramatically (Fig. 1). The supremacy of continental drift over pure polar wander necessitated the introduction of the new term 'apparent polar wander path' to de-emphasise the earlier interpretation of the mobility being purely that of the magnetic pole.

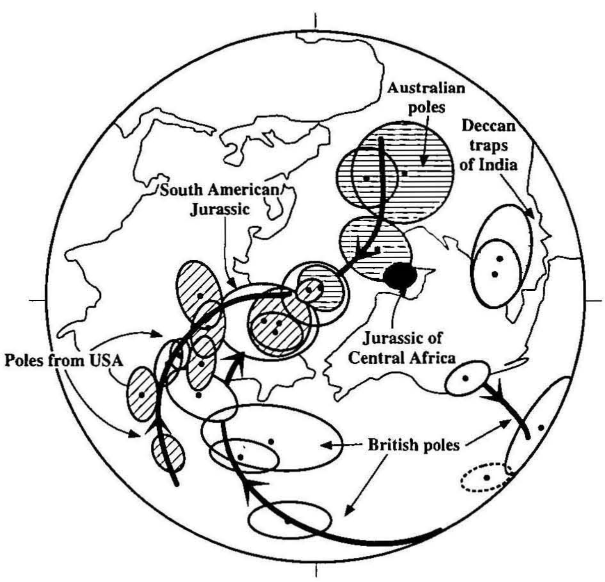


Figure 1. Pole positions relative to the present distribution of continents determined from observations from six continents: Europe, North America, Africa, Australia, Asia and South America (after Jaeger & Irving 1957).

It is ironic that Irving (pers. comm. 1992) originally had not wanted to study the Tasmanian Dolerite because its geological age is only loosely constrained stratigraphically, as indicated in the title of his 1956b paper. Jaeger, however, needed a vehicle in Tasmania for his heat-flow studies, and in any case was interested in the cooling history of the dolerite sills and the secular variation of the geomagnetic field recorded therein. By the early 1960s K-Ar dating fixed the age of the dolerite to a few million years (McDougall 1961). In Irving's own words, 'The old man was right!'. Irving discovered what Strzelecki might have over a century before, that the remanent magnetisation of the Tasmanian Dolerite is directed vertically upward. As an aside to this, Irving tells a marvellous story of his first meeting with Professor Sam Carey, who had agreed to introduce Irving to the dolerites, involving a rope-ladder and a helicopter descending out of the mist that often shrouds the Western Tiers of Tasmania. Irving had become unsettled when Professor Carey's arrival was overdue, initially thinking he has misread his instructions, and was understandably surprised by Carey's mode of transport. Those who knew that Carey was a paratrooper in World War II, and at the time was still an active parachutist, may not have been surprised. Carey was an early and enthusiastic advocate of continental drift and clearly understood the potential that palaeomagnetism held for testing the hypothesis.

One of the first symposia on continental drift was convened by Carey in Hobart in 1956 (Carey 1958), attracting many international notables, some for and some against drift One of these, Chester Longwell, USA, who wrote the epilogue to the proceedings and initially belonged to the latter group, appears to have been swayed somewhat by the arguments put in favour of drift at the symposium. In his symposium contribution Longwell argued that to explain apparent polar wander 'the most plausible suggestion is that ... the body of the Earth has slowly turned' (Longwell, in Carey 1958, p. 6), this being consistent with true polar wander. However, after assessing the persuasive evidence presented in favour of continental drift by Irving at the symposium (Irving, in Carey 1958, pp. 24–61), Longwell notes in the epilogue that 'readers will see much merit in Irving's analysis' (Longwell, in Carey 1958, p. 357). Again there is evidence of receptive individuals in the scientific community changing their opinions rapidly as the evidence mounted.

Irving and his first student, Ron Green, built Australia's first palaeomagnetic laboratory in the middle of the ANU campus, away from other buildings and magnetic interference, and in a surprisingly short time had an outline of the Australian apparent polar wander path (Irving & Green 1958). While some pole positions have since been shown to be younger than first thought, being results of overprinting, this was a remarkable feat considering cleaning techniques had not been developed. Irving also began regular publications of pole lists (in the Geophysical Journal of the Royal Astronomical Society — GJRAS), precursors to today's Global Palaeomagnetic Database. This database is mentioned further below. Because of demands on space, the laboratory was later moved to its present site on Black Mountain in Canberra, where it is now operated by the Australian Geological Survey Organisation (AGSO). Although involved in the design of the Black Mountain laboratory, Irving departed for Canada before it was completed.

Other students of Irving's in the 1960s were Don Tarling, Bill Robertson and Jim Briden. Tarling and Robertson are mentioned further below, in the context of the early polarity time-scale and the CSIRO laboratory respectively. Briden's emphasis was on pre-Carboniferous (Palaeozoic/Late Precambrian) rocks, mostly from the Lachlan and Adelaide Fold Belts. Briden (1965) was the first to appreciate the complexity of magnetic remanence and the prevalence of magnetic overprinting in multiply deformed rocks. Even today, there are only a handful of reliable results that have been gleaned from this fold belt.

In 1956 Frank Stacey joined the ANU group to study rock magnetism and to put palaeomagnetism on a reliable theoretical basis. A fundamental question to be answered at that time was whether the direction of thermoremanent magnetisation acquired by rocks, which may have been under considerable stress, is deflected significantly when the stress is released (either through denudation or drilling). Systematic magnetostriction experiments carried out by Stott & Stacey (1959) discovered one of those quirks of nature; for intrinsically isotropic rocks the deflection accompanying stress release exactly annuls that caused by the stress in the first place, thus justifying the general assumption made in palaeomagnetism that the measured remanence is parallel to the palaeo-field.

Stacey (1960) was the first to apply high-field torque measurements to magnetostriction and magnetic anisotropy studies of weakly magnetic materials such as rocks. Perhaps of greater currency, Stacey (1961) introduced the term 'pseudo-single domain' for that fraction of magnetic remanence carriers that possess single-domain-like behaviour, but are far too large to be truly single domain. While the real nature of pseudo-single domain particles is still a vexed question, empirically they explain why so many rock types yield reliable palaeomagnetic data.

These studies placed palaeomagnetism on a sound physical basis, adding weight to tectonic interpretations and the validity of continental drift. Stacey returned to the UK (Cambridge) in 1961. There he produced a seminal work on his investigations in Australia and summarised those of other pioneers, such as Louis Néel, laying the foundations of theoretical rock magnetism (Stacey 1963). In 1964 Stacey returned to Australia after being appointed Reader in Physics at the University of Queensland. His interests in rock magnetism had not waned and later he and Subir Banerjee (University of Minnesota) produced the first book devoted to theoretical rock magnetism (Stacey & Banerjee 1974). This contribution, describing the principles of rock magnetism, made the subject more accessible and helped palaeomagnetists in developing techniques to improve the quality and reliability of their work. Rock magnetism is now a mandatory adjunct to palaeomagnetic studies.

One of Irving's last studies in Australia before leaving in 1964 was in collaboration with Lin Parry, a physicist from the University of New South Wales. Parry was the first to carry out a systematic study of the size-dependent magnetic properties of annealed and dispersed magnetite particles, such as occur in rocks (Parry 1965). His observations were central to the theory developed by Stacey. Irving and Parry (1963) were the first to recognise that the geomagnetic field maintained a constant reverse polarity throughout the Late Palaeozoic and suggested the term Kiaman Interval after Mercanton's (1926) study of the sample from Kiama provided by Browne (mentioned above). Although this period is now formally known as the Late Palaeozoic Reverse Superchron, the more romantic term 'Kiaman Interval' prevails. Numerous studies on other continents have since shown their hypothesis to be essentially correct, although pin-pointing the boundaries has proved difficult. With Neil Opdyke (University of Florida) and John Roberts (University of New South Wales), Irving is currently refining the lower boundary of this important time interval by revisiting the sequences in New England (NSW), which are now much better mapped and dated. It was a lucky chance that Browne collected from the Blowhole Flow because it is now known that most seaboard rocks have been heavily overprinted during Cretaceous events before Tasman rifting. The rock magnetism work reported by Irving and Parry (1963) shows that the Blowhole Flow possesses ideal palaeomagnetic recorders that have resisted resetting and possess negligible normal polarity magnetisation from the Cretaceous event. It is somewhat incongruous that the term Kiaman remagnetisation has been popularised, referring to remagnetisation during the reverse Permian period, while remagnetisation at Kiama was actually Cretaceous.

Other important early work carried out at ANU related to palaeomagnetism was that of Ian McDougall, who had established a very productive K–Ar laboratory at ANU capable of measuring ages of very young igneous rocks. McDougall and palaeomagnetist Don Tarling, and later Francois Chamalaun, were among the first to erect a polarity time scale for the past few million years (McDougall & Tarling 1964, McDougall & Chamalaun 1966). From a palaeomagnetic viewpoint, apart from contributing to an early magnetostratigraphic framework, this work was instrumental in dismissing the self-reversal explanation for reverse polarity rocks. Chamalaun also showed that polar wander could be used to date rock-types and mineralising events that were difficult to date any other way (Chamalaun & Porath 1968).

Irving spent 10 years in Australia, putting it on the map palaeomagnetically. His research culminated in the first English text book on palaeomagnetism (Irving 1964), an outstanding contribution that remains eminently readable today. Jaeger had succeeded in establishing ANU as a palaeomagnetic/rock magnetic mecca which was to remain influential for decades.

The Black Mountain laboratory

The cruciform shape of the Black Mountain laboratory, and its precise alignment with the magnetic meridian, would, no doubt, have fascinated Eric Von Daniken. However, it is simply a very practical compromise between maximising the number of Helmholtz coil sets and the distances between them, and minimising the size of the laboratory (Fig. 2). Helmholtz coils are required to control the magnetic field, usually to annul it for astatic magnetometers and alternating field or thermal demagnetisation. The sensitivity of the astatic magnetometer could be made very high by reducing the magnetic gradient, but only at the expense of measurement time. In practice, workers concentrated on the more strongly magnetised rocks (igneous and red-beds), which produced reasonable gradients.

Mike McElhinny arrived in 1967 from Salisbury, Rhodesia (now Harare, Zimbabwe) to take over Irving's role. By then, continental drift was becoming more accepted, largely because the reality of seafloor spreading now seemed unavoidable after publication of the compelling interpretation of oceanic magnetic anomalies (Vine & Matthews 1963). Having undertaken his PhD in ionospheric physics before taking up palaeomagnetism in 1959, McElhinny's interests were less geological than were Irving's, and although much geologically and tectonically related palaeomagnetic work was to continue, this era saw a subtle change in research emphasis. New instruments and computers were becoming available, and while these did not increase sensitivity greatly, they certainly reduced the time and effort required. No longer were readings written down and punched onto cards for a centralised computer to process. Results were immediate, greatly expediting the treatment of samples. McElhinny continued the ANU tradition of producing regular GJRAS pole lists, which became inextricably linked to computer systems as time went by.

Natural diversification led to several new avenues opening up in the 1970s such as archaeomagnetism and investigations of the behaviour of the geomagnetic field, palaeointensity, magnetic transitions, excursions and secular variation studies.

Figure 2. Black Mountain laboratory, Canberra. Its original cruciform shape has been altered by the addition of another wing in 1985.

The 1970s saw palaeomagnetic studies blossom at ANU, and at any one time there were usually 4 or 5 PhD students investigating an array of problems, such as archaeo-intensities as recorded by aboriginal fire-hearths, palaeolatitudes of Precambrian glaciations, palaeosecular variation as recorded in lake sediments, oceanic island chains and hot-spot evolution. In 1973 McElhinny published another text book on palaeomagnetism with an emphasis on its role in the development of the new plate tectonic theory, incorporating both continental drift and seafloor spreading. This book (McElhinny 1973) summarised the state-of-the-art, and was received much more readily than was Irving's book.

The development of the SQUID sensor and its adaptation to rock magnetometers in the late 1970s was a dramatic development, opening up the possibility of studying all lithologies The speed with which samples could be processed increased so much that a PhD study, which in the 1960s involved 250–300 samples, in the 1980s usually involved thousands of samples. The ability to measure all lithologies coupled with the ability to process thousands of samples allowed the realisation of magnetostratigraphy as a practical method. This technique has been applied to petroleum and gas exploration with great success.

A number of research fellows were attracted to ANU during McElhinny's time: Brian Embleton in 1971, Chris Klootwijk in 1975 and Phil McFadden in 1980. Embleton carried out studies on Palaeozoic rocks in Australia, particularly central Australia, comparing results from southeast Australia with those from 'cratonic' Australia. On the basis of these studies, McElhinny & Embleton (1974) were the first to hint that the Lachlan Fold Belt was composed of allochthonous terranes. McElhinny & Embleton (1976) also settled an old argument about the position of Madagascar in Gondwanaland Their solution has since been confirmed by the precise mapping of marine magnetic anomalies. Klootwijk, who had studied Indian rocks for his PhD in Utrecht, returned to the Indian subcontinent to undertake a major palaeomagnetic study of the tectonic development of the Himalayan Fold Belt (Klootwijk 1979, 1984). This work was crucial to reconstructing the India-Asia collision and restoring the north-eastern margin of the Indian subcontinent to its former extent before India's collision.

Eminent sabbatical workers were also attracted to ANU during McElhinny's time such as Ron Merrill (University of Washington), David Stone (University of Alaska), Neil Opdyke (University of Florida), Roy Thompson (University of Edinburgh) and Ken Hoffman (California Polytechnic State University). Collaboration between McElhinny and Merrill led to the first attempt to relate palaeomagnetic constraints on the ancient geomagnetic field to dynamo theories (McElhinny & Merrill 1975). Their joint work culminated in the writing of their book (Merrill & McElhinny 1983), the preface of which sets out their aims in relating palaeomagnetism to dynamo theory. Merrill's first visit was in 1974 and he has returned regularly to Australia since then to continue collaboration, initially with McElhinny, but over the last decade, more with McFadden. A series of influential papers have come from the 3Ms, McFadden, Merrill and McElhinny, in various combinations, on the behaviour of the internal field as constrained by palaeomagnetic observations (e.g. McFadden et al. 1988 1991, McFadden & Merrill 1995). A new book by Merrill, McElhinny and McFadden summarising the past decade's collaboration is imminent. In addition to work on the dynamo, McFadden and others (e.g. McFadden 1990; McFadden & Schmidt 1986; McFadden & McElhinny 1990, 1995) have produced a variety of useful statistical tests and methods for palaeomagnetic data.

With McElhinny's departure to the Bureau of Mineral Resources (BMR, now AGSO) in 1982, ANU reassessed its position and after 30 years of world leading research in palaeomagnetism decided that this field was no longer central to its goals. It struck a deal with the BMR to take over the Black Mountain laboratory. This was opportune, since a research component had recently been added to BMR's more traditional role of surveying and database repository. The transfer of palaeomagnetic research from ANU was finally completed when Phil McFadden left to join the BMR in 1983. Although McElhinny resigned in 1988, after a brief few years he found himself transforming the pole lists into a global palaeomagnetic database. Implementing the database was actually much more than simply importing the pole-lists into a suitable software environment, since many more data fields were added. Although instigated by Phil McFadden, development of the database was sponsored by many agencies under the auspices of the International Association of Geomagnetism and Aeronomy, especially the US National Science Foundation, having received strong endorsement from Rob Van der Voo (University of Michigan). With the parallel development of computer hardware and software this database has facilitated palaeomagnetic research beyond expectations and is a superb example of strategic research. Other databases McElhinny has since developed, or is developing (in collaboration with other experts), include SECVR (Secular variation — mainly from lake sediments), PSVRL (Palaeosecular Variation from Lavas), TRANS (Polarity transitions) and MAGST (Magnetostratigraphy). These databases do, or will, provide instant access to a huge amount of data for anyone with modest computer literacy.

Chris Klootwijk, who went to the University of Paris after his time at ANU, and Charles Barton, joined the BMR at this time to stimulate their new research drive. Barton, who was the first to construct a Holocene secular variation chart for the Southern Hemisphere while studying at ANU (Barton & McElhinny 1981), returned to Australia from the University of Rhode Island.

Other important contributions from the AGSO laboratory include Idnurm's (1985) Late Mesozoic to Cenozoic apparent polar wander path and Idnurm & Giddings (1988) Precambrian apparent polar wander path. Their study of the Macarthur Basin sedimentary sequence was a monumental undertaking, foreshadowing the scale of a palaeomagnetic study necessary to disentangle multicomponent magnetisations and extract a complete tectonic/drift history for a major sedimentary basin.

CSIRO North Ryde laboratory

In 1972 the CSIRO established a new Division, based at North Ryde, to assist minerals exploration in Australia. Under the direction of the Chief, Ken McCracken, Bill Robertson set up a palaeomagnetism and rock magnetism laboratory at the North Ryde site in two old farmhouses. As mentioned above, Robertson had studied under Irving at ANU in the 1960s, but had gone to Canada before coming back to Australia to take up this post. Robertson started a number of projects having commercial applications, including studies of iron ore and banded iron formations in Western Australia and intrusions in the Sydney Basin, the latter being sources of industrial aggregate. However, commercial sponsorship of research did not begin in earnest until Embleton took over the laboratory after Robertson resigned in 1977. Embleton, who had worked with McElhinny at ANU from 1971, took up a position with Robertson in 1975. In 1980 a new special purpose laboratory was completed at North Ryde (Fig. 3). Besides offices, this laboratory was designed to accommodate three Helmholtz coil sets, although only two were required since one of the old farmhouses of the original laboratory, which housed the AF demagnetiser, was retained. The other two coil sets provided field control for a cryogenic magnetometer and a computer controlled carousel furnace that automatically heats and cools samples without operator attendance. This furnace has the same sample throughput as the cryogenic magnetometer,

Figure 3. CSIRO North Ryde (Sydney) laboratory, showing Helmholtz coil sets for the cryogenic magnetometer (background) and the furnace (foreground).

optimising efficiency and enabling up to 200 samples to be processed each week.

Extension of the work begun by Robertson in the Sydney Basin led to an important paper showing that significant stratigraphy, probably including Jurassic strata, has been stripped from the Sydney Basin since the Cretaceous (Schmidt & Embleton 1981). Over the years the laboratory has produced many key-poles for the Australian apparent polar wander path. Fundamental contributions from this laboratory include those of Clark & Schmidt (1981), on the thermomagnetic properties of titanomagnetites, Schmidt (1982, 1985) on palaeomagnetic data analysis, and Embleton & Williams (1986) and Schmidt et al. (1991) on the palaeomagnetic evidence for low-latitude glaciations in the Precambrian. Many other contributions are of a more applied nature (Clark 1983, 1984; Schmidt & Clark 1995). Clark (1984) reported a comprehensive rock magnetic study of pyrrhotite, which was sorely needed for applications of rock magnetism to exploration. Many of the results of this laboratory are in CSIRO reports to commercial sponsors. The current focus of the laboratory is magnetic petrology, instrumentation such as differential vector magnetometry and, recently, expansion into applications to coal, petroleum and gas exploration. In 1987 Embleton became Chief of the Division, and the laboratory is currently run by Schmidt. The CSIRO laboratory has been, and is, intimately involved in training students from universities in Sydney, especially Macquarie University. In recent years, most Australian PhDs in palaeomagnetism and rock magnetism have graduated from Macquarie University, under co-supervision by CSIRO staff.

University of Western Australia laboratory

While at Macquarie University, Chris Powell became increasingly interested in palaeomagnetism throughout the 1980s and when he left Macquarie University to take up the chair of geology at the University of Western Australia in 1990, a new palaeomagnetic laboratory was spawned there, using renovated equipment from CSIRO. Zheng-Xiang Li, who gained his PhD from Macquarie University using CSIRO facilities, soon followed to take over the general running of the new laboratory. Since then the laboratory has acquired the latest equipment, including a top-of-the-line cryogenic magnetometer.

Most projects that the University of Western Australia laboratory is working on have to do with Precambrian problems, although the tectonics of South East Asia is also central to their interests. This laboratory has contributed a number of papers, including studies of the formation and breakup of Gondwanaland and Rodinia (Li & Powell 1993, Powell et al. 1993, Li et al. 1995), the supercontinent that begat Pangaea, Rodinia being Russian for 'to beget'. In addition, the laboratory has established itself in applied palaeomagnetism and has attracted industry funding to study banded iron formations and iron ore. This laboratory is the only active university-based laboratory in Australia today.

Concluding remarks

It is clear that Australian palaeomagnetic and rock magnetic research have played an important, and often leading, role in the early growth of these fields. The results from the first ANU laboratory, including those from the Tasmanian Dolerite, gave overwhelming evidence in favour of continental drift. Apart from the fact that about half the books on palaeomagnetism and rock magnetism have emanated from Australian laboratories, Australia's legacy to palaeomagnetic studies is ensured by the various databases.

Throughout the 1970s and early 1980s the momentum in palaeomagnetic research was maintained with AGSO taking over from ANU, and CSIRO taking the initiative of building a special laboratory. Over the past decade, however, these studies have waned somewhat. This is in stark contrast to the huge growth of palaeomagnetic laboratories elsewhere, particularly in the USA. The reasons for this are manifold.

After ANU's withdrawal, there were no universities with mainstream interests in either palaeomagnetism or rock magnetism. Initial productive collaboration between CSIRO and Macquarie University was slowly choked by CSIRO's increasing requirement to tackle applied problems. In fact, the two main laboratories remaining after ANU's exit were both government laboratories (CSIRO and AGSO) that do not have complete freedom to undertake curiosity driven research. Many studies are confidential to sponsors and their publication embargoed.

The new University of Western Australia laboratory should compensate somewhat for the decline in government pure-research, but universities are now actively competing with government laboratories for industry funding. Both government and industry should be alarmed at this situation. In the long term, the maintenance of a reasonable level of curiosity driven research is indispensable for the sustenance of applied research. This problem is not limited to rock magnetic and palaeomagnetic laboratories and is one of the great challenges facing science in Australia today.

Acknowledgments

It is with deep sorrow that as I finish this brief history of Australian palaeomagnetism/rock magnetism I have just heard of Keith Runcorn's tragic death. As David Stone said in a

message circulated to AGU members, 'I am sure that we will all miss him greatly.'

I am indebted to Ted Irving for mentioning the Strzelecki account to me whilst I was a postdoctoral fellow at the Earth Physics Branch, Ottawa in 1978. The liberal interpretations of Strzelecki's observations, however, are purely the result of my artistic licence. Many colleagues suggested pertinent improvements and corrections to an earlier version of the paper, but undoubtedly omissions (hopefully not errors) and personal biases remain, for which I apologise.

References

Barton, C.E. & McElhinny, M.W., 1981. A 10 000 yr geomagnetic secular variation record from three Australian maars. Geophysical Journal of the Royal Astronomical Society, 67, 465–485.

Briden, J.C., 1965. Ancient secondary magnetizations in rocks. Journal of Geophysical Research, 70, 5205–5221.

Carey, W.S. (editor), 1958. Continental drift: a symposium. Geology Department, University of Tasmania, 375 pp.

Chamalaun, F.H. & Porath, H., 1968. Palaeomagnetism of Australian hematite ore bodies. I. The Middleback Ranges of South Australia. Geophysical Journal of the Royal Astronomical Society, 14, 451–462.

Clark, D.A., 1983. Comments on magnetic petrophysics. Bulletin of the Australian Society of Exploration Geophysicists, 14, 49–62.

Clark, D.A., 1984. Hysteresis properties of sized dispersed monoclinic pyrrhotite grains. Geophysical Research Letters, 11, 173–176.

Clark, D.A. & Schmidt, P.W., 1981. Theoretical analysis of thermomagnetic properties, low-temperature hysteresis and domain-structure of titanomagnetites. Physics of the Earth and Planetary Interiors, 30, 300–316.

Embleton, B.J.J. & Williams, G.E., 1986. Low palaeolatitude of deposition for late Precambrian periglacial varvites in South Australia: implications for palaeoclimatology. Earth and Planetary Science Letters, 79, 419–430.

Idnurm, M., 1985. Late Mesozoic and Cenozoic palaeomagnetism of Australia. I. A redetermined apparent polar wander path. Geophysical Journal of the Royal Astronomical Society, 83, 399–418.

Idnurm, M. & Giddings, J.W., 1988. Australian Precambrian polar wander: a review. Precambrian Research, 40/41, 61–88.

Irving, E., 1956a. Palaeomagnetic and palaeoclimatic aspects of polar wandering. Geofisica Pura e Applicata, 33, 23–41.

Irving, E., 1956b. The magnetisation of the Mesozoic dolerites of Tasmania. Papers and Proceedings of the Royal Society of Tasmania, 90, 157–168.

Irving, E., 1964. Palaeomagnetism. John Wiley and Sons, NY, 399 pp.

Irving, E. & Green, R., 1958. Polar wandering relative to Australia. Geophysical Journal of the Royal Astronomical Society, 1, 64–71.

Irving, E. & Parry, L.G., 1963. The magnetism of some Permian rocks from New South Wales. Geophysical Journal of the Royal Astronomical Society, 7, 395–411.

Jaeger, J.C. & Irving, E., 1957. Palaeomagnetism and the reconstructions of Gondwanaland. C.R. du 3e Congrès de la P.I.O.S.A., Tananarive, Section C, 233–242.

Klootwijk, C.T., 1979. A review of palaeomagnetic data from the Indian fragment of Gondwanaland, In: Farah, A. & De Jong, K.A. (editors), Geodynamics of Pakistan. Geological Survey of Pakistan, Quetta, 5–24.

Klootwijk, C.T., 1984. A review of Indian Phanerozoic palaeomagnetism: implications for the India–Asia collision. Tectonophysics, 105, 331–353.

Li, Z.X. & Powell, C.McA., 1993. Late Proterozoic to Early Palaeozoic palaeomagnetism and the formation of Gondwanaland. In: Findlay, R.H., Unrug, R. & Veevers, J.J. (editors), Gondwana 8: assembly, evolution and dispersal. A.A. Balkema, Rotterdam, 9–21.

Li, Z.X. & Powell, C.McA., 1995. South China in Rodinia: part of the missing link between Australia–East Antarctica and Laurentia? Geology, 23, 407–410.

McDougall, I., 1961. Determination of the age of a basic igneous intrusion by the potassium-argon method. Nature 190, 1184–1186.

McDougall, I. & Tarling, D.H., 1964. Dating geomagnetic polarity zones. Nature, 202, 171–172.

McDougall, I. & Chamalaun, F.H., 1966. Geomagnetic polarity scale of time. Nature, 212, 1415–1418.

McElhinny, M.W., 1973. Palaeomagnetism and plate tectonics. Cambridge University Press, Cambridge, 358 pp.

McElhinny, M.W. & Embleton, B.J.J., 1974. Australian palaeomagnetism and the Phanerozoic plate tectonics of Eastern Gondwanaland. Tectonophysics, 22, 1–29.

McElhinny, M.W. & Embleton, B.J.J., 1976. The palaeoposition of Madagascar: remanence and magnetic properties of Late Palaeozoic sediments. Earth and Planetary Science Letters, 31, 101–112.

McElhinny, M.W. & Merrill, R.T., 1975. Geomagnetic secular variation over the past 5 my. Reviews of Geophysics and Space Physics, 13, 687–708.

McFadden, P.L., 1990. A new fold test for palaeomagnetic studies. Geophysical Journal International, 103, 163–169.

McFadden, P.L. & McElhinny, M.W., 1990. Classification of the reversal test in palaeomagnetism. Geophysical Journal International, 103, 725–729.

McFadden, P.L. & McElhinny, M.W., 1995. Combining groups of paleomagnetic directions or poles. Geophysical Research Letters, 22, 2191–2194.

McFadden, P.L. & Merrill, R.T., 1995. Fundamental transitions in the geodynamo as suggested by paleomagnetic data. Physics of the Earth and Planetary Interiors, 91, 253–260.

McFadden, P.L., Merrill, R.T. & McElhinny, M.W., 1988. Dipole/quadrupole family modeling of paleosecular variation. Journal of Geophysical Research, 93, 11583–11588.

McFadden, P.L., Merrill, R.T., McElhinny, M.W. & Sunhee, Lee, 1991. Reversals of the Earth's magnetic field and temporal variations of the dynamo families. Journal of Geophysical Research, 96, 3923–3933.

McFadden, P.L. & Schmidt, P.W., 1986. The accumulation of palaeomagnetic results from multicomponent analyses. Geophysical Journal of the Royal Astronomical Society, 86, 965–979.

Mercanton, P.L., 1926. Inversion de l'inclinaison magnétique terrestre aux âges geologiques. Terrestrial Magnetism and Atmospheric Electricity, 31, 187–190.

Merrill, R.T. & McElhinny, M.W., 1983. The Earth's magnetic field: its history, origin and planetary perspective. Academic Press, London, 401 pp.

Parry, L.G., 1965. Magnetic properties of dispersed magnetite powders. Philosophical Magazine, 11, 303–312.

Powell, C.McA., Li, Z.X., McElhinny, M.W., Meert, J.G. & Park, J.K., 1993. Paleomagnetic constraints on the timing of the Neoproterozoic breakup of Rodinia and the Cambrian formation of Gondwana. Geology, 21, 889–892.

Runcorn, S.K., 1956a. Paleomagnetic survey in Arizona and Utah: preliminary results. Bulletin of the Geological Society of America. 67, 301–316.

Runcorn, S.K., 1956b. Palaeomagnetic comparisons between Europe and North America. Proceedings of the Geological Association of Canada, 8, 77–85.

Schmidt, P.W., 1982. Linearity spectrum analysis of multicomponent magnetization and its application to some igneous rocks from south-eastern Australia. Geophysical Journal of the Royal Astronomical Society, 70, 647–665.

Schmidt, P.W., 1985. Bias in converging great circle methods. Earth and Planetary Science Letters, 72, 427–432.

Schmidt, P.W. & Clark, D.A., 1994. Rock magnetic properties and palaeomagnetism of Hamersley BIFs and iron ores, North Western Australia. Precambrian Research, 69, 133–155.

Schmidt, P.W. & Embleton, B.J.J., 1981. Magnetic overprinting in southeastern Australia and the thermal history of its rifted margin, Journal of Geophysical Research, 86, 3998–4008.

Schmidt, P.W., Williams, G.E. & Embleton, B.J.J., 1991. Early timing of remanence in haematite of the Neoproterozoic Elatina Formation, South Australia. Earth and Planetary Science Letters, 105, 355–367.

Stacey, F.D., 1960. Stress-induced magnetic anisotropy of rocks. Nature, 188, 134–135.

Stacey, F.D., 1961. Theory of the magnetic properties of igneous rocks in alternating fields. Philosophical Magazine, 6, 1241–1260.

Stacey, F.D., 1963. The physical theory of rock magnetism. Advances in Physics, 12, 45–133.

Stacey, F.D. & Banerjee, S.K., 1974. The physical principles of rock magnetism. Elsevier, Amsterdam, 195 pp.

Stott, P.M. & Stacey, F.D., 1959. Magnetostriction and palaeomagnetism of igneous rocks. Nature, 183, 384–385.

Strzelecki, P.E. De, 1845. Physical description of New South Wales and Van Diemen's Land. Longman, Brown, Green and Longmans, London.

Vine, F.J. & Matthews, D.H., 1963. Magnetic anomalies over oceanic ridges. Nature 199, 947–949.

AGSO Journal of Australian Geology & Geophysics, 17(1), 145–155

The Australian National Gravity Database

A.S. Murray[1]

AGSO's Australian National Gravity Database contains data from more than 900 000 point gravity observations on the Australian mainland, over the continental margins, on the Australian Antarctic Territory, and other external territories of Australia. These data have been collected from nearly 1000 gravity surveys dating back to 1937. This repository of gravity information is a valuable national asset with importance to the mineral and petroleum exploration industries, geodesy and the international scientific community.

The Australian Fundamental Gravity Network provides a consistent and accurate set of gravity control points throughout Australia and its territories. These gravity control points, which are fully documented, are spaced at intervals of roughly 150 km over the continent. Ideally, all gravity surveys conducted in Australia should be tied into this network.

Introduction

The Australian Geological Survey Organisation (AGSO) maintains a database containing information on gravity surveys conducted in Australia and over its continental margins. Gravity survey reports, data and maps come from within AGSO, State Geological Surveys, mineral and petroleum exploration companies, universities and overseas organisations.

This paper outlines the history of gravity surveying in Australia, the sources of data, the development of the Australian National Gravity Database, the Australian Fundamental Gravity Network, data standards and future directions.

History

Gravity measurements were first made in Australia in 1819, but the systematic gravity surveying of prospective areas did not commence until the 1930s (Dooley & Barlow 1976). The first reference to gravity work in the National Gravity Database relates to a 1937 survey, while the earliest survey data included in the digital data files were collected in 1938. Interestingly, these surveys were actually carried out on the Fly River delta in Papua. Detailed gravity surveys were conducted over coalfields and oil-shale prospects in every mainland State in the 1940s for exploration and resource assessment to define Australia's indigenous energy supplies (see, for example, Chamberlain 1948).

Regional gravity traversing of the continent began in 1949 with measurements being made every few kilometres along main highways. This could only be a slow and patchy way of collecting data for the whole continent, but it was the foundation stone of the National Gravity Database.

The first systematic observations aimed at creating a gravity control network were made in 1950, using pendulum apparatus borrowed from Cambridge University (Dooley et al. 1961). Measurements were made at 59 points on mainland Australia (i.e. excluding Tasmania). The estimated accuracy of this network was 6.0 $\mu m.s^{-2}$ internally and 20 $\mu m.s^{-2}$ to external international values based on the 1930 Potsdam Reference System.

During the 1950s, large scale detailed and semi-detailed gravity surveys were conducted by mineral and petroleum exploration companies using quartz mechanism gravity meters. These Worden and Sharpe meters were light and easily portable, which made them very suitable for work in remote locations and over rough terrain. The Bureau of Mineral Resources, Geology and Geophysics (BMR, now AGSO) trialled the use of helicopters as a means of access to observation points in unpopulated areas with few roads. Using this method, a regularly spaced grid of stations could be established to achieve regional coverage of the continent (Hastie & Walker 1962). Regional road surveys were continuing during the decade and these would provide tie points to control the helicopter surveys.

The 1960s saw the introduction of LaCoste and Romberg temperature-controlled steel movement gravity meters, which had the great advantage of a worldwide dial range, thus avoiding the need for the imprecise resetting of the range required by quartz meters. These LaCoste meters made the creation of a national gravity base network feasible. The first practical Fundamental Gravity Network known as the Isogal network was established in the years 1964 to 1967; a more detailed description of this is given in the section *Fundamental Gravity Network*.

The period from the late 1950s to the early 1970s was a time of concentrated gravity surveying by private exploration companies encouraged by the Commonwealth Government's subsidy scheme. The *1959 Petroleum Search Subsidy Act* provided for reimbursement of expenses incurred in conducting geophysical surveys, with the aim of promoting discoveries of oil or gas resources in prospective but as yet unexploited areas. One important requirement for being granted a subsidy was the production of detailed reports on the survey method and results. From these reports a wealth of basic gravity survey data was available for addition to the National Gravity Database. Simultaneously with this upsurge in activity, BMR was conducting a systematic regional reconnaissance gravity survey of Australia with stations spaced at approximately 11 km in a regular grid pattern. This surveying, using helicopters for transport, was done in yearly blocks until 1962 by BMR staff and from 1963 onwards by geophysical contractors (Darby & Vale 1969). Reconnaissance coverage of the continent was completed in 1975 (Fraser et al. 1976).

Gravity surveying in marine areas began in 1957, using a North American underwater gravimeter. Surveys were conducted in Port Phillip Bay in 1957 and around the northern Australian coastline in 1958. Systematic surveying of Australia's continental shelf began in 1965, using shipborne gravimeters. These surveys were carried out for BMR by private contractors. In 1984 the RV *Rig Seismic* was chartered by BMR to carry out systematic mapping of Australia's continental shelf, slope and margins, and this program continues to the present (1997). A number of marine gravity traverses have been made by various international bodies, such as the US Navy, Lamont-Doherty Geological Observatory, Woods Hole Oceanographic Institution, Hawaii Institute of Geophysics, Geophysical Service International and the Compagnie Generale de Geophysique.

Computer processing of gravity data began in the early 1960s, and by 1967 had become a regular if occasionally error-prone method of reducing field gravity and barometer observations to meaningful accelerations and heights. The basic data produced were stored on magnetic tapes, which became the precursor to the digital National Gravity Database. Each survey block on each magnetic tape was headed by a line of text more or less descriptive of the data contained therein. These headers listed on printouts became the first index to the database. By 1970 it was clear that a properly

[1] Australian Geological Survey Organisation, GPO Box 378, Canberra, ACT 2601, Australia

planned and structured database was required to cope with the burgeoning quantity of digital data. The development of this new database is described below (see *Database structure*).

Data sources

The BMR was founded in 1946. Gravity surveys were carried out as a major part of the geophysical program from its inception. The first gravity surveys were intended to delineate coal seams and possible oil and gas resources. Systematic regional surveying along highways began in 1949 and continued as a major source of continental coverage until 1961, when reconnaissance helicopter surveying took over this role. Large scale detailed and semi-detailed surveys were not carried out by BMR between 1950 and 1986, as these were considered to be the responsibility of State geological surveys and private companies. Detailed surveys, however, were conducted along seismic lines and in small areas as an adjunct to crustal and metalliferous investigations.

Systematic upgrading of the regional coverage in selected map sheet areas to a 4 km spaced station grid began in 1987. This has proceeded at a rate of one or two 1:250 000 sheet areas per year, the map sheets selected being part of the National Geoscience Mapping Accord (NGMA). Priority is given to map sheets in areas of economic potential.

The reports on subsidised surveys, which first appeared in 1962, contained a vast amount of gravity information in the form of tabulated lists of station values, annotated gravity anomaly maps and copies of the field notebooks. The subsidy scheme continued to provide a flow of gravity information until 1974.

Owing to the large quantity of gravity information generated by the subsidy scheme and other surveys, contracts were awarded between 1974 and 1977 to process or reprocess data from these gravity surveys. Data were put into digital form by data entry from tabulations or by digitising maps, and then adjusted to the currently accepted height, gravity and position reference datum. The total number of stations processed into the digital database between 1974 and 1981 was 240 000 at a total contract expenditure of $360 000 (more than $1m in current terms).

State Government Departments of Mines or Geological Surveys have collected gravity survey data by conducting their own surveys and by requiring exploration companies to submit reports on their survey work in exploration lease or permit areas. South Australia and Tasmania required 7 km regional coverage rather than the less detailed 11 km grid over the rest of the continent, and have processed much of the company data collected in their States. New South Wales and Victoria have systematically improved coverage in their States in the last 15 years, and Queensland and Western Australia have cooperated with AGSO in establishing a 4 km grid in NGMA areas and have conducted surveys in areas of commercial interest. Much gravity information useful for the National Gravity Database still lies in company reports in the various State repositories or is still held by the companies; these data should be assessed for inclusion.

Various universities throughout Australia have, from time to time, conducted gravity surveys as part of departmental research projects or individual theses. Usually these survey results are made available to AGSO in exchange for the regional and existing gravity data in the area of study. The University of Tasmania was particularly active in the 1960s in detailed surveying of settled areas of the State.

The greater part of gravity surveying in Australia, by number of stations, has been carried out by or for private companies engaged in oil or mineral exploration. For the most part these surveys have been quite detailed and over limited areas, usually an exploration lease, but several extensive surveys were carried out in the 1950s and 1960s. The West Australian Petroleum Company (WAPET) surveyed the coastal plains of Western Australia, covering large parts of the Perth, Carnarvon and Canning Basins in 1956 and 1963, and the Hunt Oil Company surveyed a large part of the Officer Basin in 1963. Several other surveys have covered an area greater than a 1:250 000 map sheet. Most private company surveys in the last 20 years have been tied into the Fundamental Gravity Network by a marked base network station or by a tie to a marked benchmark. Direct ties to the Fundamental Gravity Network are preferable, as these stations usually have a more accurate gravity value than a benchmark, which may have been measured with only one reading. Many private company survey results have not been placed in the public domain, particularly those done around active production areas. AGSO will always welcome data from private company gravity surveys which are no longer commercially sensitive, to add to the database for the benefit of Australia; usually a data exchange agreement can be negotiated.

Surveys carried out by overseas organisations in the Australian region have fallen into the following categories:

- marine surveys, usually as part of worldwide or oceanic cruises, forming an irregular network of lines of observations;
- onshore surveys, carried out for mineral or petroleum exploration;
- single point or a small number of point observations, usually in major cities, as part of international ties;
- absolute gravity measurements, carried out in widely spaced locations on the continent.

The statistical breakdown of gravity data sources by year is shown in Table 1. A map showing the current distribution of gravity stations is shown in Figure 1.

The Fundamental Gravity Network

Fundamental gravity networks are essential for the effective use of the gravity method. They provide a uniform reference standard for gravity measurements. The current world gravity reference system consists of a network of absolute gravity values adopted by the International Union of Geodesy and Geophysics, which replaced the 1930 Potsdam Gravity System. This system is known as the International Gravity Standardization Net 1971 and it consists of 1854 stations around the world (Morelli et al. 1973). The datum for this system is provided by absolute measurements, the scale is controlled by both absolute and pendulum measurements, and the internal structure is provided by approximately 24 000 gravimeter observations.

In Australia nearly all gravity measurements made by AGSO, State geological surveys, private industry, and universities are tied to the Australian Fundamental Gravity Network. This network links AGSO's National Gravity Database and the local surveys conducted by State geological surveys and the exploration industry with global networks. The Geodetic Reference System 1967 has been adopted in Australia as the reference surface for the computation of theoretical gravity. The increasing dominance of satellite geo-positioning systems (GPS) in world mapping has caused Australia and other countries to move towards a geocentric datum for mapping. This will lead to the adoption of the World Geodetic System 1984 (WGS84) or its successor as the reference surface for the computation of theoretical gravity. Australia is moving to the Geodetic Datum of Australia which is the WGS adapted to match heights with the Australian Height Datum. The shift to the Geodetic Datum of Australia, will occur in stages from 1995 to 1998.

The worldwide reference systems permit the standardisation of gravity measurements on land and sea. This provides obvious benefits to national and international studies that require gravity data over the Earth's surface, such as in geophysics, geodesy

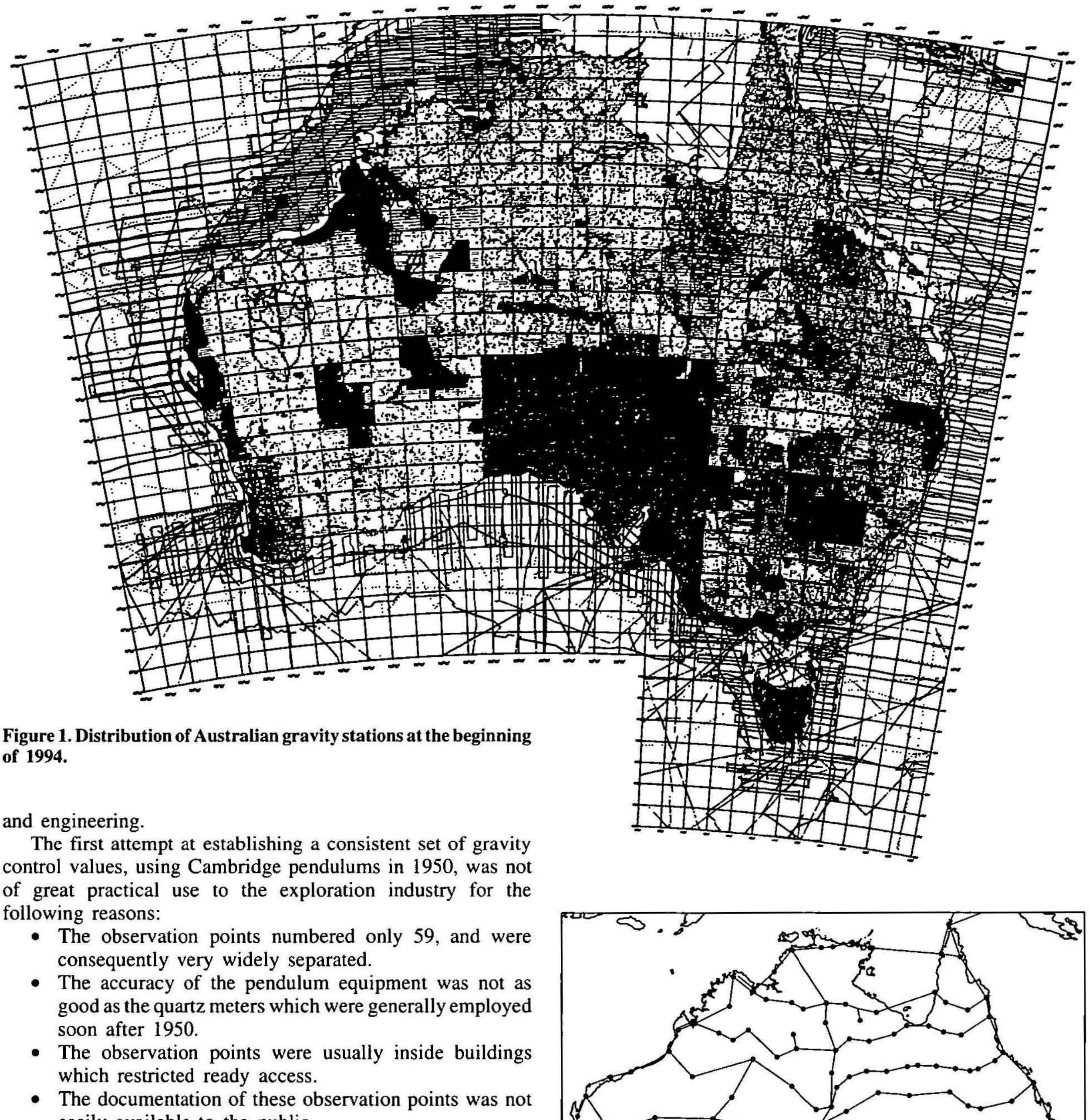


Figure 1. Distribution of Australian gravity stations at the beginning of 1994.

and engineering.

The first attempt at establishing a consistent set of gravity control values, using Cambridge pendulums in 1950, was not of great practical use to the exploration industry for the following reasons:

- The observation points numbered only 59, and were consequently very widely separated.
- The accuracy of the pendulum equipment was not as good as the quartz meters which were generally employed soon after 1950.
- The observation points were usually inside buildings which restricted ready access.
- The documentation of these observation points was not easily available to the public.

In 1960 a series of calibration ranges was established to provide accurate gravity intervals which could be used to check or determine the scale factor of gravity meters (Barlow 1965, 1967). The scale factor defines the amount of change in gravitational acceleration equating to a movement of one scale division on the meter's dial. Eight calibration ranges were established, one at each State capital, and at Alice Springs and Townsville. These intervals were measured using five Worden gravimeters, read at alternate ends of the calibration line a number of times. Quartz type gravimeters, such as the Worden, are designed to have a constant scale factor over the range of their dial, and calibration on an accurate range before and after a survey is recommended. Steel spring meters (LaCoste & Romberg), conversely, have a variable scale factor which depends on the dial position in a slowly varying parabolic relationship, so calibration of these meters on a fixed range is less definitive.

The 1964 to 1967 'Isogal' surveys provided a useful, documented network (Figure 2) of about 600 gravity control stations at 180 localities (Barlow 1970). In 1964 BMR possessed only one LaCoste and Romberg gravity meter (G20), so the bulk of the work was done using Worden and Sharpe quartz meters. These quartz meters have a limited dial range of about 1600 $\mu m.s^{-2}$, which restricts the observer to a relatively narrow band of gravity values if resetting of the meter scale is to be avoided. It is not possible to reset meters and obtain very accurate gravity measurements, as the reset cannot be defined exactly in scale divisions and the meter drift is perturbed as

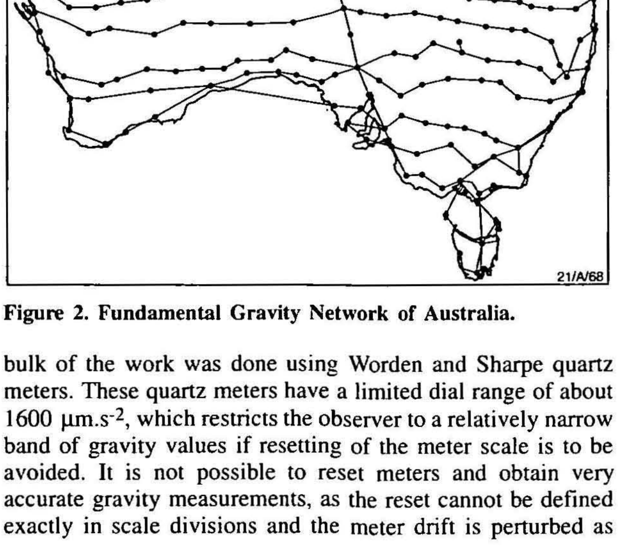


Figure 2. Fundamental Gravity Network of Australia.

the mechanism undergoes mechanical relaxation after the reset. Clearly if quartz meters were to be used to establish a continental network, measurements had to be made along lines of roughly equal gravity, hence the name 'Isogal'. These lines were aligned approximately east–west because the largest variation in gravity occurs with latitude. Some zigzagging of the lines was required to compensate for large local gravity anomalies and elevation effects. Three north–south tie lines were added in 1967. A base network was established in Papua New Guinea in 1967, and was tied to the Australian network. The gravity values calculated from all these surveys were labelled Isogal 65. The accuracy of the Isogal 65 network has been shown by subsequent measurements to be about 2.0 $\mu m.s^{-2}$ with one or two stations in error by 4.0 $\mu m.s^{-2}$.

The easternmost tie line of the Isogal network, from Laiagam in Papua New Guinea down the east coast of Australia to Hobart, Tasmania, was established as the Australian Calibration Line in 1970 (Shirley 1966, Wellman et al. 1974b).

Table 1. Yearly totals of gravity surveys (stations) by source.

Year	*AGSO (BMR)*	*States*	*Private company*	*Other*	*Total*
1939			1 (1143)		1 (1143)
1947	1 (888)		1 (1672)		2 (2560)
1948	1 (57)	1 (171)		1 (13)	3 (241)
1949	2 (246)	1 (327)			3 (573)
1950	1 (241)	3 (322)	2 (816)		6 (1379)
1951	6 (1295)	1 (49)	1 (1122)		8 (2466)
1952	7 (1524)	7 (778)			14 (2302)
1953	6 (576)	3 (516)		1 (145)	10 (1237)
1954	2 (163)	2 (333)			4(496)
1955	3 (771)	10 (828)	2 (1732)		15 (3331)
1956	7 (1683)	3(252)	6 (30 463)		16 (32 398)
1957	4 (1190)	3 (1628)	2 (290)	1 (157)	10 (3265)
1958	6 (1611)	1 (157)	3 (1258)		10 (3026)
1959	12 (2456)	1 (55)	4 (5024)		17 (7535)
1960	7 (3774)	1 (67)	4 (2601)	1 (17)	13 (6459)
1961	6 (4487)	4 (790)	7 (4911)		17 (10 188)
1962	7 (5528)	1 (166)	10 (12 764)	3 (881)	21 (19 339)
1963	9 (8603)	9 (9575)	21 (58 711)	5 (10 680)	44 (87 569)
1964	4 (7565)	11 (2011)	18 (15 788)	5 (1064)	38 (26 428)
1965	2 (7204)	9 (752)	12 (18 644)	3 (517)	26 (27 117)
1966	7 (8173)	8 (2238)	15 (24 857)	2 (616)	32 (35 884)
1967	11 (21 695)	14 (4522)	18 (13 657)	9 (3651)	52 (43 525)
1968	8 (19 140)	10 (6580)	12 (13 711)	7 (10 140)	37 (49 571)
1969	8 (7781)	17 (6778)	16 (15 783)	8 (9336)	49 (39 678)
1970	7 (21 361)	7 (1693)	16 (18 477)	10 (14 929)	40 (56 460)
1971	9 (34 443)	8 (1619)	4 (8966)	14 (28 649)	35 (73 677)
1972	10 (31 610)	9 (2525)	2 (282)	12 (23 408)	33 (57 825)
1973	9 (7825)	1 (1489)	2 (3317)	3 (604)	19 (13 235)
1974	1 (1562)	6 (3680)	1 (741)	1 (112)	9 (6095)
1975	8 (3811)	7 (995)		4 (870)	19 (5676)
1976	3 (556)	4 (901)		6 (4327)	13 (5784)
1977	5 (2958)	3 (2022)	8 (3082)	5 (1626)	21 (9688)
1978	5 (2241)	4 (65 345*)	5 (3593)	1 (2061)	15 (73 240*)
1979	2 (1308)	9 (2359)	4 (2016)	1 (11)	16 (5694)
1980	9 (2840)	15 (4768)	6 (4882)	1 (46)	31 (12 536)
1981	7 (621)	19 (9417)	11 (19 121)		37 (29 159)
1982	9 (5302)	14 (4987)	4 (2876)	2 (751)	29 (13 916)
1983	1 (403)	6 (1824)	2 (741)	2 (4027)	11 (6995)
1984	6 (7188)	7 (1861)	5 (2231)		18 (11 280)
1985	3 (2266)	6 (3456)	6 (19 008)	1 (1296)	16 (26 026)
1986	4 (6416)	6 (1827)	2 (981)	2 (1418)	14 (10 642)
1987	6 (2591)	3 (1678)	1 (2756)		10 (7025)
1988	2 (2778)	3 (841)			5 (3599)
1989	3 (2153)	3 (1255)			6 (3408)
1990	2 (1465)	1 (119)			3 (1584)
1991	5 (3546)	3 (1858)	7 (16 001)	2 (615)	17 (22 020)
1992	3 (3040)	2 (1114)	1 (54)		6 (4108)
1993	10 (4491)	2 (797)			12 (5288)
1994	9 (3086)	7 (3613)	8 (27 775*)		24 (34 474*)

* Indicates surveys from other years are included in the total.

The localities at which measurements were made to establish the Australian Calibration Line are Laiagam, Mount Hagen, Menyamya, Lae and Port Moresby in Papua New Guinea, and Thursday Island, Iron Range, Cooktown, Cairns, Townsville, Mackay, Rockhampton, Maryborough, Brisbane, Grafton, Kempsey, Williamtown (Newcastle), Sydney, Canberra, Albury, Melbourne, Flinders Island, Launceston and Hobart in Australia. The Australian Calibration Line was established in 1970 using 3 LaCoste, 4 Worden and 2 Sharpe gravimeters, followed in 1971 by measurements using 4 LaCoste meters. Numerous other measurements of gravity intervals have been made up and down this line at various times, with all the AGSO LaCoste meters and a number of meters from other countries. These measurements provide a comparison between the scale factors of the Australian meters and those of the foreign meters.

The following international ties define the datum and scale of the Australian network:

- Measurements were made with Japanese GSI pendulums in 1964 at Melbourne, Sydney, Brisbane, Mackay, Townsville and Cairns, and in 1970, tying between Tokyo and Brisbane (Langron 1966, Shirley 1966).
- In 1966, measurements were made along the Western Pacific Calibration Line, which originates in Canada, travels through Japan, Hong Kong, the Philippines and Singapore before looping through eastern Australia and New Zealand.
- Also in 1966 the United States Airforce made measurements at Canberra, Melbourne, Sydney and Perth (Whalen 1966, Shirley 1966).
- In 1969, when two more LaCoste meters had been purchased, measurements were made along the Western Pacific Calibration Line at Point Barrow, Fairbanks and Anchorage in Alaska, at 8 locations in Japan, Okinawa, Taiwan, Hong Kong, Manila, Singapore, Darwin and Sydney. The Sydney station connected the Western Pacific Calibration Line to the new Australian Calibration Line described above (Wellman et al. 1974a).
- In 1972, scientists from the Soviet Union measured the gravity interval between Moscow and Sydney, using OVM pendulums, and in 1973 another team measured along the Australian Calibration Line with eight GAG-2 gravimeters, accompanied by BMR staff with four LaCoste meters (Wellman et al. 1974b).
- In 1973, a tie was made between Sydney, Christchurch in New Zealand and McMurdo base in Antarctica.
- In 1974, pendulum measurements were made by Soviet scientists at Moscow, Port Moresby and Hobart (Gusev 1975, Wellman 1975).
- In 1979, Japanese scientists measured along the Australian Calibration Line with two La Coste meters as part of a project to define an international scale for gravity (Nakagawa & Nakai 1980, Nakagawa et al. 1983).
- Also in 1979, Soviet scientists measured absolute gravity values at Sydney, Hobart, Perth, Alice Springs, Darwin and Port Moresby (Arnautov et al. 1979).

Most offshore Australian territories have been tied to the Australian network (Wellman 1976). Christmas Island was tied to Perth in 1973 and again with Cocos Island to Perth in 1976. Lord Howe and Norfolk Islands were tied to Sydney in 1982 (Williams & Murray 1985).

A major strengthening of the gravity base network was carried out in 1980 using 7 LaCoste and Romberg gravimeters (Wellman et al. 1985a). This survey provided the strong north–south links needed to convert the Isogal network into a true Fundamental Gravity Network in which reliable loop closures could be calculated. The results of this survey and reworking of the 1964–67 measurements (all calculated to IGSN71 datum and international scale) were tabulated in Wellman et al. (1985b). This new set of gravity control values was labelled Isogal 84 with accelerations expressed as micrometres per second squared ($\mu m.s^{-2}$), to distinguish them from the Isogal 65 values (based on the 1930 Potsdam datum), which were expressed as milligals. Wellman et al. (1985b) have a more detailed account of the two reference systems and an approximate formula to convert from one to the other. The estimated accuracy of the Isogal 84 values is 0.2 $\mu m.s^{-2}$ for stations measured in 1980 and 1.0 $\mu m.s^{-2}$ for stations reworked from the Isogal 65 values.

During the 1980s and early 1990s, minor refurbishment of the Fundamental Gravity Network was undertaken in response to requests for values at new localities and information about stations which had been destroyed. New stations were established at Timber Creek, Kununurra, Broome, Cloncurry, Coober Pedy, Mount Hope and Portland. All stations in Tasmania were remeasured in 1985.

Beginning in 1994, a thorough remeasurement and augmentation of the Fundamental Gravity Network was initiated with a precise gravity survey of the Eyre Peninsula and central north of South Australia, using the six AGSO LaCoste meters (Murray & Wynne 1996). In 1995, new stations at Broken Hill, Olary and Mannahill were tied to Mildura (Murray et al. 1995). The network in Victoria was upgraded in 1995 (Murray & Wynne 1995). The accuracy expected from these surveys is a standard deviation of 0.1 $\mu m.s^{-2}$ with a maximum error of 0.25 $\mu m.s^{-2}$ for each segment.

Since the early 1970s, descriptions of all gravity base stations in the Fundamental Gravity Network have been available as hand-drafted line diagrams illustrating the base stations at their particular localities. These diagrams are available on request to those who wish to tie their gravity survey to the Fundamental Gravity Network. In 1993 a new series of diagrams, including photographs and the latest gravity values, was developed using PC software. The new series of diagrams (Figure 3) will be introduced progressively as the Fundamental Gravity Network is refurbished.

Database structure

Gravity information contains basic data for investigations of the shape of the Earth and the structure and composition of its outer layer. Gravity anomaly data over the whole continent and its adjacent seas constitute a fundamental data set for systematic geological mapping, mineral and petroleum exploration and resource assessment. An efficient and versatile database management system is required to make full use of the data stored in the database, which currently contains data for more than 800 000 gravity observations. Before the advent of commercial database management systems in the 1980s, locally written special purpose systems were developed for each geophysical database, each having a different philosophy and structure, making interfacing between various databases very difficult.

The introduction of computers to process gravity data in the mid-1960s opened the possibility of digital storage of gravity data on magnetic tapes. Previously, data had been reduced to basic information on calculation sheets and hand-drawn graphs, which were then filed under survey number in a manual filing system. A card index was used to cross-reference all the survey information. Gravity maps were drawn by hand and each 1:250 000 map sheet required one or two months of drafting time. By the late 1960s a rudimentary suite of computer programs had been developed to process drift-controlled field observations for gravity and height, to list files of survey data stored on magnetic tape or disks at the CSIRO Division of Computing Research, to locate specified blocks of data, to edit data files and to produce station plot and crude contour maps (Bellamy et al. 1971, Townsend 1971).

By the early 1970s, the quantity of digital data and the

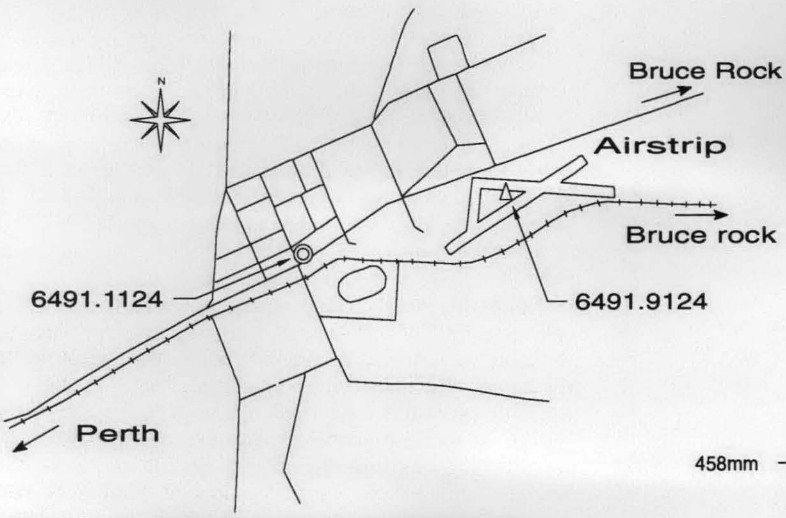
STATION NAME	STATION CODE
Quairading W.A.	6491.1124

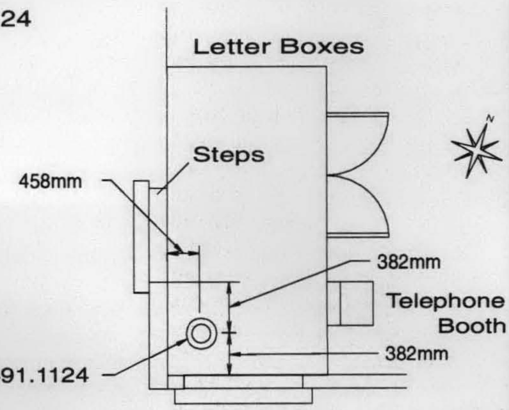
LOCATION	DESCRIBED / RECOVERED BY	OBSERVED GRAVITY (μms^{-2})	DATE
Post Office	A.G.S.O. *Drawn by P Wynne*	9794032.8	1980

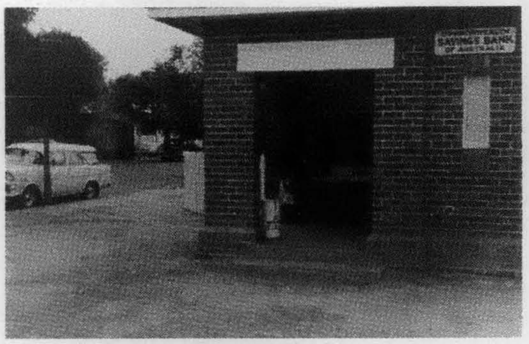
DESCRIPTION:

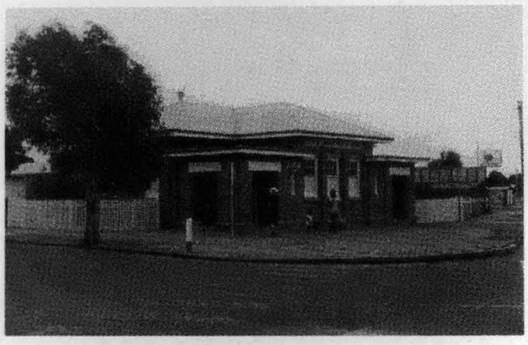
Station Class:

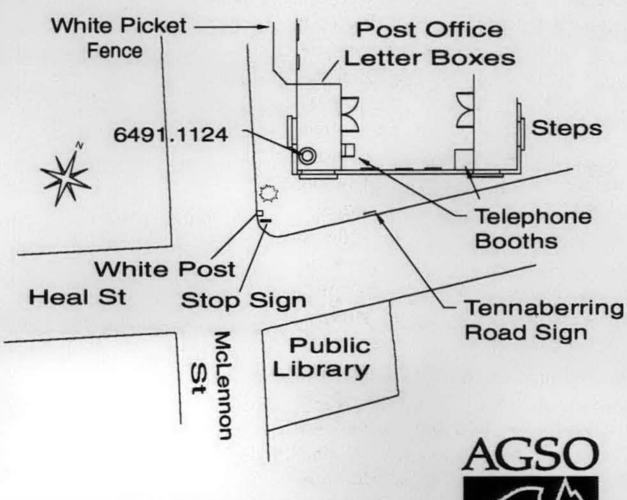
STATION TYPE: Base
POSITION: S 32 00' 36" E 117 23' 54"
Position Source: Report 261
ELEVATION: 247.35m
Elevation Source: Report 261











GRAVITY STATIONS
AUSTRALIAN NATIONAL GRAVITY NETWORK

AGSO
AUSTRALIAN GEOLOGICAL SURVEY ORGANISATION

Station No. 6491.1124 Quairading
Sheet 1 of 2

Figure 3. Example of a new Fundamental Gravity Network diagram.

Table 2. Number of gravity stations lying within each 1:1 000 000 map sheet area.

	49	*50*	*51*	*52*	*53*	*54*	*55*	*56*	*57*
C	142	199	2123	Melville I 6134	Cp Wessel 377	Torres Str 1714	Port Moresby 4572	Woodlark I 1774	266
D	250	1435	Brunswick 6315	Darwin 11245	Roper R 2850	Mitchell R 3150	Cooktown 4759	1540	159
E	1604	Rowley 14160	Broome 40214	Halls Ck 4648	Newcastle W 6838	Normanton 9922	Townsville T793	3490 3	1152
F	Cloates 4740	Hamersley 12074	Oakover 19948	L Mackay 16269	Alice Spr 12760	Cloncurry C6876	Clermont 15027	Rockhampton R600	4396
G	Carnarvon 6425	Meekatharra 11779	Wiluna 22567	Petermann 16740	Oodnadatta 22094	Cooper Ck 24456	Charleville 28407	Brisbane 316	2077
H	Houtman 983	Perth 16033	Kalgoorlie 13181	Nullarbor N8198	Tarcoola 83791	Broken Hill 5356	Bourke B6883	Armidale A875	Lord Howe 2572
I	4013	Albany 14557	Esperance 3173	Eyre 2965	Port Augusta 14383	Adelaide 42695	Canberra 3712	Sydney 4187	1831
J	731	1264	485	1047	Du Couedic 2040	Hamilton 44035	Melbourne 23894	Bodalla 1954	1075
K	360	593	357	534	1125	Stokes 2144	Tasmania 30925	928	913
L	154	391	1152	983	1024	569	2870	761	319

requirement to produce computer-drawn maps made it necessary to develop a comprehensive gravity database with integrated processing software. The basic foundation of the gravity database is the library of survey files. Each gravity survey is stored as one or more survey files, each file being the result of a single data-reduction process. Before the availability of large on-line disk storage, the database was held on magnetic tapes, each tape holding many survey files. The number of tapes which could be mounted simultaneously was limited, and access to large areas was severely restricted. The first suite of database programs (Murray 1974) was written to interface with the data stored on tapes. User input was needed to determine where the required data were held and to request that the necessary tapes be mounted.

In the late 1970s, on-line mass storage disks became available for mainframe computer systems. This enabled the complete database to be instantly accessible to the processing software. The change of storage medium allowed a new interactive automated accessing of the survey files, which are independent but referenced by a random access index which contains a summary of the data in each file. Any search for data automatically reads the index which references all the relevant files. The processing programs now have a front-end interactive menu and series of questions with a help facility. Most responses have a built-in default value which selects the most likely or sensible choice.

The survey files in the database consist of a series of interleaved blocked binary numerical and text records, each containing information on 50 gravity stations. The records have a header defining the survey block and a summary of the ranges of station number, coordinates and height in the block to aid in searching for data. The attributes of each gravity station are shown in Appendix 1.

Much more information can and should be added to the digital database, mainly in the form of supplementary information about surveys, detailed station descriptions, scanned photographs, benchmark descriptions, field sheets, processing and network details. The original field data for each survey must be used if the survey results are to be recomputed accurately to conform to new tie station and base network values, or to allow for an updated meter calibration factor. This additional information is currently held in hard copy files or as unreferenced ASCII digital files (see *Future developments*, below).

Scope and accuracy

Mainland continental Australia is covered by, at worst, an 11 kilometre grid of gravity stations except for three small holes (each about 2500 km^2) in the northern Canning Basin, Simpson Desert and Eromanga Basin. Much of the country is covered by more detailed work, as shown in Figure 1. Offshore, over the continental margins, the coverage is 50 km or closer spaced ship traverses. Not all of the measured marine gravity has yet been included into the National Gravity Database. The number of processed digital gravity observations or sampled values in the rectangle defined by 8–48°S and 108–162°E is currently slightly over 830 000 stations. The number of stations within each 1:1 000 000 map sheet area within this rectangle is shown in Table 2. Data have also been collected on Christmas and Cocos Islands in the Indian Ocean, Norfolk and Macquarie Islands, and the Australian Antarctic Territory.

The accuracy of the data varies from survey to survey and also depends on the control points used to define the reference datum of gravity and height. Checking of gravity station heights against other elevation sources as part of the Digital Elevation Model project (Bernardel et al. 1996) has shown a number of elevation errors greater than 10 m. Some 'levelled' surveys appear to be in error uniformly, while others show a sloping shift. Surveys which employed barometers to measure the height are known to have frequent errors up to 5 m and occasional errors up to 20 m, although new digital barometers give results to within 5 m accuracy for regional surveys and 1 m or better for detailed surveys. The elevations from the Digital Elevation Model will be used to control the heights of previously measured gravity stations where possible and should reduce the errors to less than 5 m, which is equivalent to 15 $\mu m.s^{-2}$ in gravity anomaly.

Gravity values from well-controlled surveys should have an accuracy of 1 $\mu m.s^{-2}$, which is certainly achievable using well-maintained and properly used gravity meters of the types in use since the mid-1950s. For reconnaissance surveys where control is less strict the accuracy should be better than 5 $\mu m.s^{-2}$ (the specified maximum misclosure for the contract helicopter reconnaissance surveys was 3 $\mu m.s^{-2}$). Of course, these degrees of accuracy can only be verified for the multiply read stations within a survey: individual singly measured stations may be in error due to misreading, misrecording or meter malfunction (tare, etc.). These single-station errors should be immediately obvious on the contour map or image of anomalies if they are large, but smaller subtle errors of about 5 $\mu m.s^{-2}$ may be suspected, but not provable unless the station spacing is less than 1 km. Modern gravity surveys are conducted with much stricter requirements and the introduction of GPS positioning will mean that accuracy of 0.5 $\mu m.s^{-2}$ or better can be achieved for regional surveying.

Data retrieval

Data from the database may be retrieved manually from the documents held in the hard copy survey files, or by computer using the gravity processing system. In both cases, the survey file is the basic unit of reference. Currently, the only data held in digital form are the gravity station basic data described above (see *Database structure*) and the index of gravity surveys, which contains a basic description of the survey and some statistical information. Diagrams of the Fundamental Gravity Network stations are being constructed on computer and all recent survey field readings and processing results are digitally stored.

The gravity processing system allows retrieval of gravity data from the digital database in the following forms:

- Survey index display — survey summary information may be displayed on screen or printed for one or more specified surveys or for all surveys. This information contains the survey name, survey datum descriptor, date of last amendment, the limits of latitude, longitude, elevation and station number, the survey numbers represented in this survey file and the map sheets which the data cover.
- Digital gravity station data — on magnetic tape or floppy disk in specified system of units. Data may be selected by survey number, station number range or coordinate limits, including map sheet specification.
- Gravity station data listing — printed hard copy list of the information held for each station, selected as for digital data.
- Station location map — displayed on screen or plotted as hard copy map. Stations may be assigned symbols and codes, differentiating surveys, and may be annotated with station number, Bouguer or free-air gravity anomaly and elevation. A choice of 7 map projections is available and the map limits, scale and graticule are defined by the user.
- Contour map — of Bouguer or free-air gravity anomalies or heights. This may also include station locations as above. Data are selected, gridded and contoured according to user-defined parameters.
- Pixel image — using ER Mapper software. An image may be generated from grids used to produce contour maps. An image of Australia based on a 2.5 km pixel size is shown as Figure 4. This image uses Geosat satellite coverage in offshore areas.

Gravity information is available to clients in the following forms:

- Fundamental Gravity Network station descriptions and

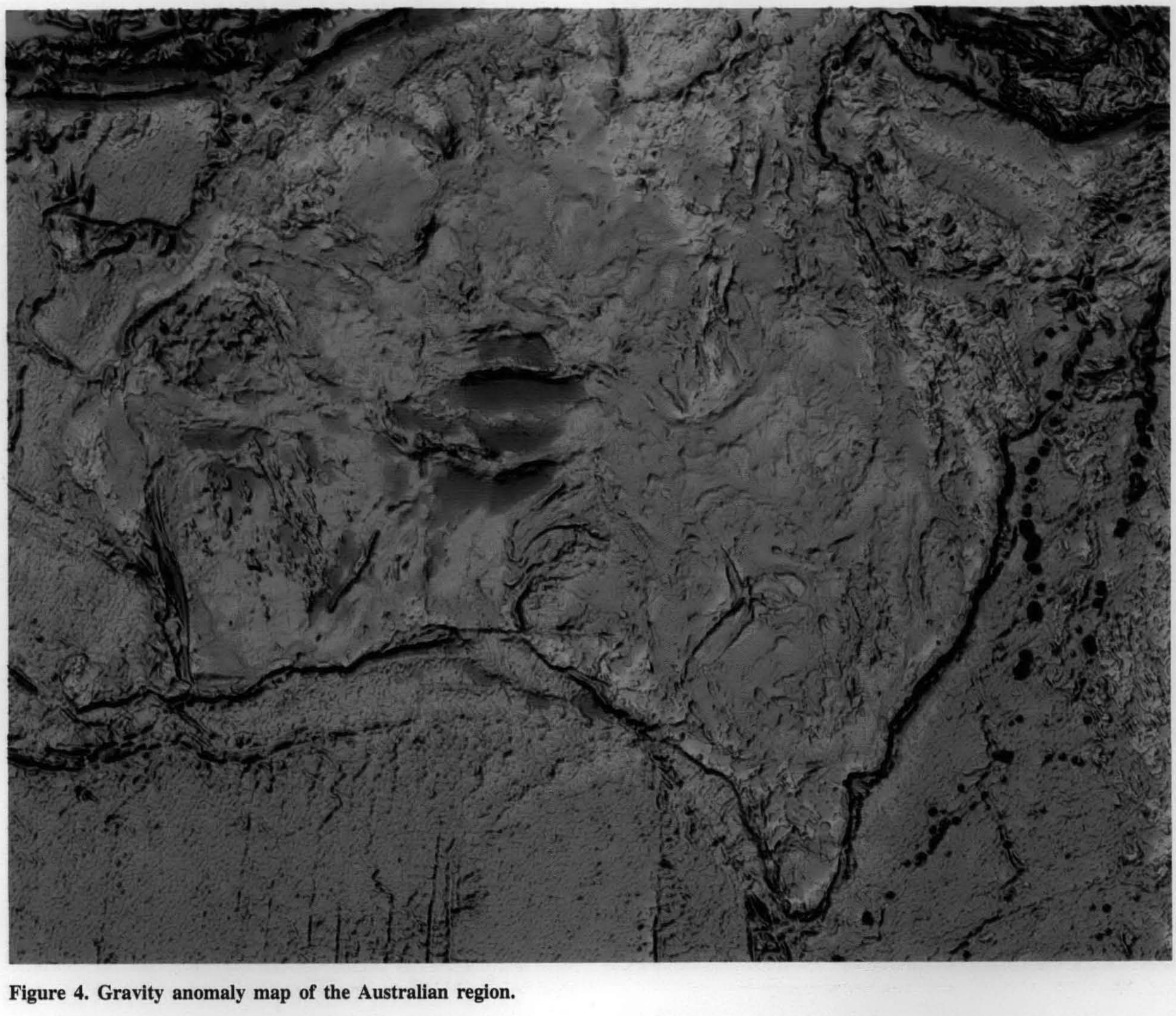


Figure 4. Gravity anomaly map of the Australian region.

values — the most accurate and well-documented set of gravity control stations throughout Australia. Refurbishment and augmentation of the network are in progress and a new series of diagrams, including photographs of the stations, is being introduced. An example of a new diagram is shown as Figure 3.
- Benchmark information — during the 1960s and 1970s AUSLIG, then part of the Department of the Interior, put in numerous third-order levelling traverses along roads at the request of BMR and other departments. A representative selection of these benchmarks was used as elevation control for gravity surveys and has a gravity measurement next to the benchmark. This information can be located for most specified areas and the strip map descriptions of the traverses can be used to locate the markers.
- Digital gravity station data — available by individual 1:1 000 000 map sheet area on floppy disks or as a complete national database on open reel or Exabyte magnetic tape. The national database does not include a limited number of confidential or recent surveys which may be available for sale elsewhere. Results of recent surveys conducted by the States or as part of the NGMA program are available for sale from the respective State; NGMA data are also available from AGSO in 1:250 000 map sheet area blocks on floppy disks. Availability of these data is advertised in press releases.
- Station location and contour maps — available in dyeline or transparency format from the AGSO Sales Centre in Canberra. Contour maps at 1:1 000 000 scale are available for the whole country. These maps show the station locations as dots and have Bouguer gravity anomaly contours at 20 μm.s^{-2} intervals. The original map series was produced using data gridded at a 3.0′ mesh size; this is being progressively replaced with a new series using a 2.5 km (1.5′) mesh size.
- Gravity survey reports — each gravity survey will be documented in an operational report covering survey method, statistics and processing, and will contain base-station diagrams.
- Pixel images — produced for NGMA areas where the coverage of data has been improved to 4 km station spacing or better. These images are at 1:500 000 or 1:1 000 000 scale with a pixel size of 400 m.

Future developments

The present database system is adequate for accessing basic data, but much more information is known about each survey and about particular groups of stations within each survey; for example, the date of the survey, the serial numbers of the instruments used, the observer's name, the accuracy of each parameter, the control points used and their values at the time of the survey, the processing statistics and the location of any documentation. This information, as well as graphical references to each survey, will be integrated into a comprehensive database system using such packages as Intrepid, Oracle and ARC-Info. Integration of the gravity database into the Intrepid system is now (1997) under way.

The Fundamental Gravity Network will be completely refurbished and augmented. Station sketch maps and photographs will be progressively integrated into the new set of digital base-station diagrams. These should be available to clients on a public access database, at a nominal charge, to encourage use of these stations to control all gravity surveys conducted in Australia. All measurements at stations forming part of the Fundamental Gravity Network will be digitally stored to allow automatic reprocessing of the network values when significant new measurements are introduced or as new absolute gravity determinations become available.

Images of 1:1 000 000 map sheets with a pixel size of 500 m will be progressively released for the whole country as data checking and upgrading are completed. New pixel maps of the whole continent at 1:5 000 000 and 1:10 000 000 scales will be produced as data quality and quantity increase. Coverage of the marine areas will be augmented by satellite measurements which are now publicly available in gridded form. The recently released Gravity Anomaly Map of the Australian Region (Murray et al. 1997; see Figure 4) is a pixel image formed by merging the onshore gravity grid with the offshore satellite coverage. The whole of the mapped area was gridded at a mesh size of 2.5 km.

A complete relevelling of all data held in the digital National Gravity Database is now overdue, as many surveys are still based on control points which have already been updated. Correlation of surveys can also be used to highlight datum errors. A systematic upgrading of survey data will be feasible when:
- the new Digital Elevation Model heights are compared with the gravity heights, allowing systematic survey height errors and spot errors to be detected in areas of relatively flat terrain;
- the World Geodetic System geocentric coordinates are formally adopted, giving an opportunity to adjust all previous surveys to this new system from the old AGD66 and Clarke 1858 grids;
- the Fundamental Gravity Network adjustment is completed and reliable values are known for all survey control points, enabling a recomputation or block shift of survey data to the new datum.

References

Arnautov, G.P., Boulanger, Yu.D., Karner, G.D. & Shcheglov, S.N., 1979. Absolute determinations of gravity in Australia and Papua New Guinea during 1979. BMR Journal of Australian Geology & Geophysics, 4, 383–393.

Barlow, B.C., 1965. Establishment of gravity meter calibration ranges in Australia, Bureau of Mineral Resources, Australia, Record 1965/19.

Barlow, B.C., 1967. Gravity meter calibration ranges in Australia. Bureau of Mineral Resources, Australia, Report 122.

Barlow, B.C., 1970. National report on gravity in Australia, July 1965 to June 1970. Bureau of Mineral Resources, Australia, Record 1970/62.

Bellamy, C.J., Lodwick, G.D. & Townsend, D.G., 1971. The gravity reductions storage and retrieval system used by BMR. Bureau of Mineral Resources, Australia, Record 1971/7.

Bernardel, G., Morse, M.P., Carroll, D. & Race, G., 1996. GEODATA 9 SECOND DEM. Australian Surveying and Land Information Group, Canberra.

Chamberlain, N.G., 1948. Preliminary report on the geophysical survey of the Collie Coal Basin. Bureau of Mineral Resources, Australia, Report 1.

Darby, F. & Vale, K.R., 1969. Progress of the reconnaissance gravity survey of Australia. Bureau of Mineral Resources, Australia, Record 1969/110.

Dooley, J.C. & Barlow, B.C., 1976. Gravimetry in Australia 1819–1976. BMR Journal of Australian Geology & Geophysics 1, 261–271.

Dooley, J.C., McCarthy, E., Keating, W.D., Maddern, C.A. & Williams, L.W., 1961. Pendulum measurements of gravity in Australia 1950–51. Bureau of Mineral Resources, Australia, Bulletin 46.

Fraser, A.R., Moss, F.J. & Turpie, A., 1976. Reconnaissance gravity survey of Australia. Geophysics, 41, 1337–1345.

Gusev, N., 1975. Determination of gravity acceleration at

Port Moresby PNG and Hobart with OVM pendulums. Bureau of Mineral Resources, Australia, Record 1975/106.

Hastie, L.M. & Walker, D.G., 1962. Two methods of gravity traversing with helicopters. Bureau of Mineral Resources, Australia, Record 1962/134.

Langron, W.J., 1966. Pendulum gravity ties between Tokyo and Melbourne. Bureau of Mineral Resources, Australia, Record 1966/109.

Morelli, C., Gantar, C., Honkasalo, T., McConnell, R.K., Tanner, J.G., Szabo, B., Uotila, U. & Whalen, C.T., 1973. The International Gravity Standardization Net 1971 (IGSN71). Special Publication No. 4, IUGG-IAG, Paris.

Murray, A.S., 1974. The Australian National Gravity Repository computer system. Bureau of Mineral Resources, Australia, Record 1974/68.

Murray, A.S. & Wynne, P.E., 1995. Fundamental gravity network in Victoria 1995 re-establishment survey. Australian Geological Survey Organisation, Record 1995/70.

Murray, A.S. & Wynne, P.E., 1996. Australian Fundamental Gravity Network 1994 Survey of Eyre Peninsula in South Australia. Australian Geological Survey Organisation, Record 1996/5.

Murray, A.S., Williams, J.W. & Wynne, P.E., 1995. Fundamental gravity network survey Mildura–Broken Hill–Mannahill Operations report. Australian Geological Survey Organisation, Record 1995/49.

Murray, A.S., Morse, M.P., Milligan, P.R. & Mackey, T.E., 1997. Gravity Anomaly Map of the Australian Region (Second Ed.), scale 1:5 000 000. Australian Geological Survey Organisation, Canberra.

Nakagawa, I. & Nakai, S., 1980. Precise calibration of scale constants of LaCoste and Romberg gravimeters and international gravimetric connection along the circum Pacific zone (preliminary investigation). Geophysical Institute, Kyoto University, Japan.

Nakagawa, I., Nakai, S., Shichi, R., Tajima, H., Izutuya, S., Kono, Y., Higashi, T., Fujimoto, H., Murakami, M., Tajima, K. & Funaki, M., 1983. Precise calibration of scale constants of LaCoste and Romberg gravimeters and international gravimetric connection along the circum Pacific zone (final report). Geophysical Institute, Kyoto University.

Shirley, J.E., 1966. Pendulum measurements in Australia. Bureau of Mineral Resources, Australia, Record 1966/162.

Townsend, D.G., 1971. Improvements to the BMR gravity storage and retrieval system. Bureau of Mineral Resources, Australia, Record 1971/8.

Wellman, P., 1975. Operational report on the Soviet gravity tie Moscow–Port Moresby–Hobart 1974. Bureau of Mineral Resources, Australia, Record 1975/48.

Wellman, P., 1976. Gravity ties to Australian Antarctic bases and Christmas Island. Bureau of Mineral Resources, Australia, Record 1976/11.

Wellman, P., Barlow, B.C. & Coutts, D.A., 1974a. Comparison of Western Pacific and Australian Calibration Line scales and an evaluation of secular variation. Bureau of Mineral Resources, Australia, Record 1974/121.

Wellman, P., Boulanger, Yu.D., Barlow, B.C., Shcheglov, S.N. & Coutts, D.A., 1974b. Australian and Soviet calibration surveys along the Australian Calibration Line. Bureau of Mineral Resources, Australia, Bulletin 161.

Wellman, P., McConnell, K. & Williams, J.W., 1985a. Australian gravity base-station — network: 1980 survey. BMR Journal of Australian Geology & Geophysics, 9, 225–231.

Wellman, P., Barlow, B.C. & Murray, A.S., 1985b. Gravity base-station network values, Australia. Bureau of Mineral Resources, Australia, Report 261.

Whalen, C.T., 1966. The Western Pacific Calibration Line survey, 1964-65. United States Air Force, ACCS Oflan 503, Phase Report No. 3.

Williams, J.W. & Murray, A.S., 1985. Gravity measurements at Sydney, Lord Howe and Norfolk Islands. Bureau of Mineral Resources, Australia, Record 1985/1.

Appendix 1. Attributes of gravity stations held in the National Gravity Database

Numerical

Station number — an eight or ten digit number (e.g. 9590.0117 or 7119.041250). The first four digits (9590) define the survey: the first two are the year marker (95) and the second two the type and serial number of the survey (90). The third digit is usually 9 for Fundamental Gravity Network control surveys. The final four or six digits (0117 or 041250) are the unique station number, which may be simply consecutive, related to a shot point meterage, or the day (04) and time (1250) of the observation, as commonly used in marine surveys. The choice of eight or ten digit numbers is a matter of convenience. Usually, onshore stations have 8 digit numbers and marine stations 10 digit numbers to accommodate the day and time information. This station number is often represented with a decimal point between the survey and station parts for convenience.

Latitude — expressed in decimal degrees to 6 decimal places (0.1 m) precision. Currently, coordinates are expressed in terms of the AGD66, but a shift to geocentric coordinates as measured by GPS systems is planned in the near future. The accuracy of coordinates varies widely from survey to survey, depending on the method of measurement. Many surveys over small areas were optically surveyed, surveys along roads at benchmarks were third-order levelled, and other surveys had station positions plotted manually on maps and the coordinates scaled from these maps. Most locations dating from before 1970 are only accurate to 0.1′ (166 m), many of the more recent data are accurate to 1 second (28 m), including single receiver, hand-held GPS readings. However, since the AGD66 and WGS84 coordinates differ by up to 200 m in the Australian region, absolute accuracy in position is hard to define; it may be estimated by using geoid to geoid conversion programs.

Longitude — expressed in decimal degrees. Comments under latitude also apply here.

Elevation of observation point (height of the gravity meter or pendulum) above mean sea level in metres to 3 decimal places. Heights are based on the Australian Height Datum (AHD). Strictly, this height should be measured to the centre of mass of the body being balanced in the gravity meter mechanism or the pendulum bob, but in all cases to date the height has been taken as the height of the surface on which the meter stands. Since relative gravity measurements do not themselves depend on an absolute gravity value any constant offset to the measurements will be eliminated when the gravity difference is computed. This height may be different from the ground surface height as described below.

Observed gravity value — expressed as $\mu m.s^{-2}$ to 2 decimal places. The value is based on the Australian Fundamental Gravity Network datum, which approximates the International Gravimetric Standardisation Net datum. The values of the Fundamental Gravity Network have and will continue to change as the network is refined and readjusted, but these changes should become progressively smaller each time. Not all the data currently in the National Gravity Database have been adjusted to the latest Fundamental Gravity Network datum, but in almost every case the correction will be less than the mean error intrinsic to the observed value.

Elevation of ground surface at the coordinates of the observation point — expressed as metres above AHD to 3 decimal places. For marine surveys this height will be the negative water depth. Other gravity stations which would have a meter height differing from the ground height are those in mines, on platforms or buildings, in aircraft, on lakes or glaciers, or those in submarines. In the case of solid concrete firmly seated on solid earth the level of the concrete is taken as the ground height.

Terrain correction — expressed in $\mu m.s^{-2}$ to 2 decimal places. This positive value is, by convention, based on a standard density of 2.0 $t.m^{-3}$ but may be scaled linearly to any desired density. Very few gravity stations have had accurate terrain corrections calculated, owing to the lack of surrounding height information. Generally, gravity observations are made in locally level environs to eliminate gross effects from close topographical changes. The new digital terrain model of Australia will permit automatic calculation of outer-zone terrain corrections for most stations, but near-station corrections will always need to be calculated individually.

Text

Station number — 8 or 10 ASCII characters, corresponding to the number described above, but no dot separates the survey from the station part. The database can accommodate alphanumeric station numbers, but this has been avoided so far. In some years the number of surveys is more than 100 and if all the data were or could be processed into the database an alphanumeric survey number may be necessary. Mines and Energy South Australia has used alphanumeric survey numbers for many years.

Informal station name — up to 30 ASCII characters. This optional station description may contain a benchmark number, alternative station number, shotpoint line and meterage, locality name, and brief description or pertinent information about the particular observation (noise, wind, etc.). This field is particularly useful to describe stations which may be used as control or tie points for a new survey. Not all stations in the database which should be labelled have been annotated. This information usually needs to be manually extracted from field notes, maps or data listings.

Datum definition — 10 characters, each of which is a letter code describing or defining an aspect of the individual gravity station. The characters, which must be chosen from a list of allowed letters, will control processing options, display formats, access to the data and reliability. With each station flagged in this way it is possible to display the data in different units and reference systems and to store data in their original form as required. With the introduction of geocentric coordinates the height datum may be redefined as a geopositional datum. Each of the 10 characters has the following significance:

1. Station format: '0' alphanumeric, '+' 10 digit, '=' 8 digit.
2. Coordinates: 'D' decimal degrees, 'M' minutes, 'S' seconds.
3. Height units: 'M' metres, 'F' feet.
4. Height datum: 'A' AHD, 'S' mean sea level (local), 'U' undefined.
5. Gravity units: 'M' milligals, 'U' micrometres per second squared.
6. Gravity datum:'O' Isogal65, 'P' provisional IGSN, 'I' FGN84.
7. Environment:'M' marine, 'I' ice, 'L' lake, 'G' ground surface, etc.
8. Completeness: 1 bit set for each numeric item present in data.
9. Reliability:'A' absolute, 'B' fundamental, 'C' control point, etc.
10. Classification:'C' confidential, 'R' restricted, 'U' unclassified, etc.

Date of last amendment — 10 characters in the format yyyy/mm/dd. The year month day format allows sorting or comparing of dates numerically without rearrangement. This is the date on which the station was processed into the database or when one of the station parameters was added, removed or edited.

AGSO

AUSTRALIAN GEOLOGICAL SURVEY ORGANISATION

AGSO Journal of Australian Geology & Geophysics

Volume 17, Number 1

AGSO JUBILEE 1946-96