Structure and evolution of the Amadeus, officer and Ngalia basins of central Australia

Kurt Lambeck

Research School of Earth Sciences, Australian National University, Canberra, ACT, Australia

Online Publication Date: 01 March 1984


To link to this article: DOI: 10.1080/08120098408729278
URL: http://dx.doi.org/10.1080/08120098408729278

PLEASE SCROLL DOWN FOR ARTICLE

Full terms and conditions of use: http://www.informaworld.com/terms-and-conditions-of-access.pdf

This article maybe used for research, teaching and private study purposes. Any substantial or systematic reproduction, re-distribution, re-selling, loan or sub-licensing, systematic supply or distribution in any form to anyone is expressly forbidden.

The publisher does not give any warranty express or implied or make any representation that the contents will be complete or accurate or up to date. The accuracy of any instructions, formulae and drug doses should be independently verified with primary sources. The publisher shall not be liable for any loss, actions, claims, proceedings, demand or costs or damages whatsoever or howsoever caused arising directly or indirectly in connection with or arising out of the use of this material.
Structure and evolution of the Amadeus, Officer and Ngalia Basins of central Australia

Kurt Lambeck

Research School of Earth Sciences, Australian National University, PO Box 4, Canberra, ACT 2600, Australia

The late Proterozoic to Palaeozoic evolution of the basin and basement terrains of central Australia cannot be ascribed to conventional basin-forming models based on thermal or stretching mechanisms. Foreland basin models also are inappropriate. The principal objection to the thermal and stretching models is that the basins are not in isostatic equilibrium, whilst the foreland basin model cannot explain the quasi-continuous evolution of the basins over periods of several hundreds of millions of years. A mechanical model is developed on the supposition that the crust has been in compression for long periods and that the evolution is determined by the balance of the compressive force, the buoyancy force, surface loading and erosion, and the elastic and viscous forces. The lithosphere is considered as a viscoelastic plate subject to a small irregular load in Late Proterozoic time. Some of the initial deflections caused by this load grow with time at a rate that is a function of the various forces and plate properties. Erosion of the uplifting areas, with the sediments deposited in the downwarps, much enhances the deformation. As bending stresses increase with time, the uplift rates increase and gravity sliding may become important. Failure by thrust faulting is also predicted, the basement being thrust over the basins. This would have occurred early in the Cambrian, corresponding to the Petermann Orogeny. The model predicts further significant deformation in the southern Arunta in Late Palaeozoic time, corresponding to the Alice Springs Orogeny and the final stages of the Ngalia Basin formation. The proposed model predicts a crustal structure that is in broad agreement with the available geophysical data and provides a framework for discussing the geological history of the basin sediments and basement metamorphics.

Key words: sedimentary basin formation, Amadeus Basin, Arunta Block, crustal compression, intracratonic basins

INTRODUCTION

The evolution of intracratonic basins remains poorly known, even though their structures are becoming better understood as a result of closer geophysical and geological scrutiny. To some, an understanding of basin evolution is important for evaluating the resources potential. To others, the study of basins is important for evaluating tectonic processes or for learning more about the rheology of the crust and upper mantle. The motives of this paper lie in the latter direction although, if an adequate evolutionary scheme can be devised, there are obvious consequences for evaluating the potential for mineral or hydrocarbon resources.

Of the Australian basins, the central structures forming the Officer, Amadeus and Ngalia Basins are of considerable interest because they cannot be described by the usual basin formation mechanisms: neither the thermal mechanisms as promulgated by Sleep (1971) and others, nor the stretching model of McKenzie (1978), nor the foreland basin models of Price (1973) and Beaumont (1981) appear to be reconcilable with the available, but limited, geological and geophysical evidence. Furthermore, the gravity anomalies observed in central Australia are an embarrassment. In general, the Australian gravity field is bland, in keeping with an old continent where the combination of stress-relaxation and erosion has resulted in a generally uninteresting and uninformative gravity field. The major exceptions to this are the quasi-linear E-W trending anomalies over central Australia (Fig. 1) where, in a general sense, the strong negative Bouguer anomalies correspond to the basins and the positive anomalies correspond to the intervening region of exposed basement. Anomalies such as these beg answers to three fundamental questions. What crustal and upper mantle structure do they imply? How did this
structure evolve into its present state? What are the rheological implications of this structure? Clearly, these three questions are intimately related. Yet most discussions still treat the three points separately. Thus Forman and Shaw (1973) and Mathur (1976) propose models for the existing structure without giving the physics of the mechanism that led to their proposed structure more than a fleeting glance. Dooley (1973) explores the stress state implied by these anomalies without really investigating how these stresses are supported for the requisite long time periods. One reason for this state of affairs is that unique answers cannot be found from gravity alone. Gravity anomalies indicate that there has been deformation (or lateral density structure), and that the stress state is non-hydrostatic. The anomalies do not, however, tell us uniquely what the deformation is. They tell us even less about the stress state.

As already emphasized, the questions of present structure, evolution and rheology are closely related and considerable progress can be made by introducing physical models based on thermal and mechanical principles and constrained by regional geological observations. One excellent example of this is Beaumont's (1981) Alberta foreland basin model formed by the Rocky Mountains fold-thrust belt. Additional geophysical observations, such as deep crustal reflection and refraction surveys, now largely serve the purpose of testing critical aspects of the model or of providing a means for testing between competing physical models. If the new measurements are coherent with the predictions, then we have not only confirmed our ability to apply physics to what are essentially geological problems, but we may also have established a basis for extrapolating to other basins and beyond. Even if the new measurements are inconsistent with the

---

**Fig. 1** Regional gravity field in central Australia based on the 1:5 000 000 gravity map of Australia published by the Bureau of Mineral Resources in 1975. Anomalies greater than $-20$ mgal occur over the exposed basement areas of the Arunta and Musgraves with maximum values reaching about $25$ mgal. Anomalies less than $-100$ mgal occur over the Ngalia, Amadeus and Officer Basins with the maxima occurring near the contacts of the basins with the blocks. A relative gravity high of $-60$ mgal separates the northern and southern parts of the Amadeus Basin. The profile on the right is along an approximately N-S line from the Officer Basin in the south to the Lander Trough of the Wiso Basin in the north.
model we still gain insight into the applications of physics to geology and provide a basis for building alternative models. It is in this spirit that the present paper is written: to provide a general framework for discussing the present structure and past evolution of the central Australian intracratonic basins and intervening 'arches' of uplifted and exposed Precambrian crust. A simple model will never explain all geological observations but such a model may provide a basis for discussing the geological data, for helping to distinguish between important and secondary issues, and for providing insight into possible refinements.

MECHANICAL MODELS FOR BASIN EVOLUTION

In its simplest form, the introduction of a model relating evolutionary processes to structure is a statement about isostasy. Mass excess created by elevated terrain or high-density, near-surface rocks is compensated by mass deficits deeper in the crust and vice versa. The compensation is either local, as the classical Pratt and Airy models, or it is regional, as in the Vening-Meinesz and Gunn models (Jeffreys 1959; Heiskanen & Vening-Meinesz 1958). In the event of a basin filled with low-density sediments, the isostatic models require a thinning of the crust below the basin or an increase in crustal density to offset the mass deficit in the basin. The Bouguer gravity anomaly over the basin will be negative by an amount that depends on (i) basin depth, (ii) density contrasts and (iii) the depth of compensation. If the compensation depth is very shallow the isostatic anomaly approaches the free air anomaly but if the compensating mass is deep the isostatic anomaly approaches the Bouguer anomaly. In regional isostatic models the compensating mass is distributed over an area that is more extensive than the load itself. This compensating mass is usually determined by the condition of a state of mechanical equilibrium being reached between the load, the buoyancy, and elastic or viscoelastic forces in the crust. Examples include the passive loading models in which the crust is loaded by ice, where a depression in a crust is filled with sediments or where the crust is loaded by a volcano: in all cases the crust is responding to the load not by failure, but by a regional deflection.

The local and regional states of isostasy are possible because the crust and lithosphere have a greater viscosity or long-term strength than the underlying asthenosphere or upper mantle. This is testified by the crustal rebound following glacial retreats, where typically mantle materials flow on time scales of $10^3$–$10^5$ years when subjected to stress differences of a few tens of bars or even bars (e.g., Peltier 1981; Nakiboglu & Lambeck 1982) $(1\text{ bar}=10^5\text{Pa})$. The much less viscous lithosphere or crust will respond more slowly to stress differences, typical time scales being of the order $10^6$–$10^7$ years (Lambeck 1981). In most loading problems it is possible to consider the lithosphere as an elastic or possibly viscoelastic layer overlying a fluid-like asthenosphere, provided that the decrease in viscosity with depth is at least two orders of magnitude across the lithosphere–asthenosphere boundary (Biot 1961).

When the crust or lithosphere is passively loaded by sediments, the isostatic response will, in the first instance, be a regional one whereby the load is partly supported by the strength of the crust. The regional compensation will approximate local compensation when the areal extent of the load is much larger than the crustal thickness, and it will often not be possible to distinguish between local and regional compensation. If the stresses associated with the loading exceed the strength of the crust, failure may occur and a state of near-local isostasy may be reached through vertical motions along the failure planes. Otherwise, with time, the stress differences existing in the crust may relax by creep, by very localized movements on pre-existing zones of weakness, or by localized mechanical failure. Compensation becomes increasingly local, with the stress differences migrating upwards into the stronger parts of the lithosphere. The rate at which the stress relaxation occurs is a function of the elastic and viscoelastic parameters, and also of the wavelength of the load, with the crust under long wavelength loads evolving faster than under short wavelength loads. Models for such evolutionary processes have been discussed by Beaumont (1978) and Lambeck and Nakiboglu (1981). However the state of local isostasy is not a natural end-state, since stress differences in an isostatically compensated crust are not necessarily small (Jeffreys 1959; Lambeck 1980). If the stress differences are sufficiently high, further relaxation may occur as the crust searches for a hydrostatic equilibrium state. This will be achieved by a creep of material such that the local isostatic state will, once attained, be maintained throughout.

Figure 2 illustrates schematically the possible evolution of a 1 km deep, water-filled depression, formed by some unspecified mechanisms and in a
state of local compensation. Sediment loading occurs until the depression is entirely filled. For a narrow basin the initial state will be one of regional compensation, but for a broad basin the state will be practically indistinguishable from the local state. The additional subsidence caused by the sediments replacing the water will reach \((p_s - p_w) / (p_m - p_s)\) of the original depression where \(p_m\), \(p_s\) and \(p_w\) are the densities of the mantle, sediments and water, respectively. With \(p_s = 2.5 \text{ g cm}^{-3}\) and \(p_m = 3.3 \text{ g cm}^{-3}\), the additional subsidence is approximately equal to the initial depression depth. This is a maximum basin depth for, if elastic forces within the crust are invoked as a means of supporting the extra load, the actual deflection may be less than this. To form a 10 km deep basin, the original depression must therefore have been some 5 km deep, and much of the sediments would have formed in a deep-water environment. In most basins the sediments have been deposited in shallow water or continental environments, and clearly this passive loading mechanism alone is unimportant.

If there is active coupling between the sediment source and the basin of deposition, considerable thickness of sediments can be accumulated in a shallow water or continental environment. Burke (1976) proposed such a model for the Chad Basin, in which the basin periphery, subject to ongoing uplift by some unspecified mechanism, provides a continuous sediment source. Perhaps the most successful model is that proposed by Price (1973) and others and quantified by Beaumont (1981), in which a foreland basin is mechanically coupled to an adjacent orogen. [See also Speed and Sleep (1982) who appeared to be unaware of Beaumont's work.] The supracrustal load, in the form of the fold-thrust belt, forms a depression in front of it and in which sediments collect. As the fold-belt grows and migrates basin-inward, moving ahead of an uplifting crystalline or metamorphic core zone, older sedimentary and basement materials are transported laterally over sedimentary sequences laid down at an earlier stage in the basin evolution (Fig. 3). The basin forming process is therefore very much associated with the active orogeny phase of geosyncline evolution. In the final phase, erosion unroofs the core and the fold-belt of the orogen and the basins are uplifted owing to isostatic rebound. Ultimately the contrasting topographies of mountains and basins will be removed. The Price–Beaumont model is primarily intended for basin formation along plate boundaries, where the source for the fold-thrust belt is provided by the subduction of the oceanic plate under the continent. In the Canadian model a fold-thrust belt zone, 100–150 km wide, moves ahead of an uplifting core zone several hundred kilometres wide. Typically, where the Rocky Mountains reach their greatest heights, a basin of some 2–3 km of sediments has formed in the 140 Ma since the Upper Jurassic.

Other basin-forming models emphasize the role of thermal processes in forming the basins. In their simplest form, the crust is heated from below or from within and thermal expansion occurs. This is followed by erosion of the uplifted area, by cooling and contraction of the lithosphere, and the
Fig. 3 Schematic evolution of a foreland basin. A load (1) is placed on the crust which deforms elastically and the surrounding depression is filled with sediments (1'). As the fold-thrust belt (the load) moves forwards a second load (2) is emplaced which deepens the basin (2'). The process continues until the orogeny ceases. Any elevated terrain will then erode and rebound of the basin occurs.

formation of a depression to be filled by sediments (Sleep 1971; Sleep & Snell 1976; Fig. 4). The dominant mechanism here is the thermal one, the role of gravity being one of passive loading. To produce deep basins, material must be removed on a large scale during the uplift and erosion phase. Uplift by thermal expansion alone appears to be limited, and it is unlikely that the final basin will exceed more than a few kilometres in depth. Subsidence due to thermal contraction will have a time constant comparable to the thermal conduction time constant, of the order $10^7$–$10^8$ years. Also the basin-forming process will decay with time, and if sedimentation keeps up with contraction, subsidence will decay with the square root of time. The thermal phase leaves a crust that is essentially in local isostatic equilibrium, whereby the mass deficit from the depression is compensated by an upward movement of the Moho. The subsequently deposited sediment load will tend to

be regionally compensated. Bouguer gravity anomalies, including corrections for sediment densities if required, will therefore be small. Alternatively, the subsidence may be attributed to the penetration of dense basic and ultrabasic intrusives into the lower crust (e.g. McGinnis 1970) or by metamorphism of the lower crust to denser granulite or eclogite facies (Falvey 1974) or by a combination of the two (Haxby et al. 1976). The intruded hot and initially lighter materials will

Fig. 4 The crust moves over a heat source resulting in thermal expansion (a). The uplifted area is eroded and the Moho moves upwards as a result of isostasy (b). When the crust cools a depression is formed (c) which is filled with sediments, enhancing the original depression and lowering the Moho towards its original position (d).
balance the heavier transformed crustal rock, but upon conductive cooling of the crust, subsidence will take place. These models too result in a depression that will be locally or partly regionally compensated. Time constants will be controlled predominantly by those of the heat conduction through the lithosphere.

Stress-based hypotheses include the graben models of Vening–Meinesz, elaborated further by Bott (1976) and Beaumont (1978), and the crustal stretching model proposed by McKenzie (1978). Graben models are characterized by normal faulting and subsidence of a crustal wedge when the crust is subjected to sufficient tension to cause failure in the upper crust and ductile flow at depth. The resulting rift valleys are generally narrow, less than 50–100 km, and the wider the valley the less the overall subsidence (Bott 1976); for a 20–30 km wide sediment filled graben, subsidence may attain 4–5 km, but for a graben twice as wide the subsidence is only about half this amount. Beaumont (1978) has explored graben-initiated basins in some detail. In these models the basin is initially controlled by faulting of an elastic lithosphere, but with time this initiating mechanism ceases and subsequent evolution is controlled by stress relaxation of the lithosphere (Fig. 2). Initially narrow grabens now evolve into quite wide basins in which a deep central graben is flanked by extensive but shallow basins. McKenzie (1978) produced basins by large-scale stretching of the crust. Hot mantle material wells up under the thinned crust, producing a thermal perturbation that gradually decays and leaves behind a thinned and subsiding area to be filled by passive sediment loading. Prior to this loading, local isostatic equilibrium is attained and the subsequent load is supported regionally. During stretching, brittle failure occurs in the upper crust, possibly producing graben-type structures, and the basin floor may resemble a sequence of parallel grabens and horsts. Deformation of the lower crust is likely to be by ductile flow, possibly with deeper crustal materials penetrating the upper crust, which further enhances the subsidence upon cooling (Fig. 5).

Fig. 5 Basin formation by crustal stretching. In (a) the crust is subjected to a tensional force and thinning occurs. In response to this, the Moho moves upwards (b). Sediments deposited into the basin magnify the depression (c). Any intrusion into the weakened lower crust will further amplify the basin depth.

Basin formation by crustal compression has not been given much attention in the recent literature, although such a mechanism appears to be relevant to intracratonic basin formation (Lambeck 1983). Possibly this neglect stems from the recognition that excessively large compressional stresses are required to induce buckling in the crust as a whole. But the buckling state is not required. Significant deformation can occur if (i) the crust is subjected to an initial load, (ii) viscous or viscoelastic elements are introduced into the crustal model and enough time is available for the folds to grow, and (iii) erosion is considered. In this model, the lithosphere is considered as a viscoelastic plate that has been subjected to a small normal load, distributed in an arbitrary manner. Under compression, initial deflections due to this load are magnified by amounts that depend on the plate’s physical properties, the magnitude of the compressive force, and on the spectrum of the initial perturbations, with some wavelengths being magnified by greater amounts than others. Because of the introduced viscosity, the deflections grow with time at a rate that is determined by the balance of forces on the plate and on the rheology of the lithosphere. Upwarped areas are eroded, providing the bulk of the sediments for deposition in the downwarps (Fig. 6). This not only results in a weakening of the plate if uplift becomes significant, but also modifies the balance between the forces acting on the plate.

The above-discussed basin forming mechanisms can be characterized according to whether the dominant process is gravity, thermal or stress based. In general, all three processes play important roles at various stages of a basin’s history; in initiating the tectonic upheavals, in shaping the initial depressions, in providing the source of sediments, and in their subsequent evolution with time. Thus the relative importance of the three
mechanisms lead to the classification into one of these three categories. One immediate consequence is that it is unlikely that any one model can explain all intracratonic basins within Australia. Perhaps what can result is a model that provides a framework for discussing a majority of intracratonic basins, specific basins differing from each other because of the different degrees of importance of the main processes involved.

THE CENTRAL AUSTRALIAN BASINS

The oldest geological formations in central Australia are the basement metamorphics of the Musgrave and Arunta Blocks with an Upper Archean to Lower Proterozoic sedimentation age. Igneous and mafic intrusions occurred at several times, most notably at about 1800–1700 Ma ago and again at 1200–1100 Ma. It appears that by about 1000 Ma cratonization of central Australia was largely complete (Plumb et al. 1981). The oldest extensive sediments overlying this basement are the Dean and Heavitree Quartzites of the Amadeus Basin, the Townsend Quartzites of the Officer Basin and the Vaughan Springs Quartzite in the Ngalia Basin. These sediments, which appear to have covered much of central Australia, formed from about 1050 Ma to 900 Ma ago (Compston & Nesbitt 1967; Majoribanks & Black 1974; Black et al. 1980). Their thickness appears to have been uniform over the area, pointing to a relatively long period of tectonic stability in which the sediments were deposited in shallow marine, littoral, neritic and lagoonal environments. This period of regional stability after the earlier periods of upheaval and igneous activity makes the time 1000–900 Ma a convenient starting point for modelling the Late Proterozoic and Palaeozoic evolution of the basins.

The subsequent sediment sequence within the basins is of considerable importance in that they contain, inter alia, the history of subsidence and post-depositional activity of the region from about 900 Ma to the present. Table 1 summarizes the very general stratigraphy of the Amadeus Basin, based on the work by Wells et al. (1970). For the present purpose, a simple stratigraphy, representing the principle depositional sequences, is adequate. More recent work in the basins has led to further subdivisions in which several members of the earlier stratigraphy have been elevated to formation status (e.g. Preiss & Forbes 1981; Wells & Moss 1983). Thus the Upper Areyonga of Wells et al. (1970) is now identified as the Aralka.
Formation, but in this paper the combined Aralka and Areyonga Formations go, for convenience mainly, under the latter name. Likewise, the Lower Pertatataka Formation of Wells et al. was referred to by Wells and Moss as the Olympic Formation, but the term Pertatataka is here taken to imply this, plus the older Olympic and Younger Julie Formations. The adopted ages (Table 1) are also open to some question, as there is not yet a general consensus for the chronology of the Precambrian and Palaeozoic sedimentary sequences.

Figure 7 illustrates the schematic N-S section across the basin, based on the isopach maps of Wells et al. (1970) for the Upper Proterozoic to Palaeozoic (see also Froelich & Krieg 1969; BMR 1976). This section corresponds to the gravity profile of Fig. 1 between, approximately, Mulga Park in the south and the MacDonnell Ranges near Gosse Bluff in the north. Parallel profiles to the west of this line are qualitatively similar, except that (i) the basin narrows somewhat, (ii) a thick sequence of synorogenic Mount Currie Conglomerate occurs along the southern margin, and (iii) the Bitter Springs and Inindia–Areyonga Formations are more commonly exposed near the basin centre. Parallel profiles to the east of this line differ mainly in that the southern part of the basin contains up to 1 km of the Palaeozoic Finke Group, but these sediments may not be part of the Amadeus story. The angular unconformities between the various units are indicative of regional movements; the Areyonga movement between the Bitter Springs and Areyonga Formations, the Precambrian Souths Range Movement, the Cambrian Petermann Ranges Orogeny, the Palaeozoic Rodingan and Pernjara Movements and the Alice Springs Orogeny. The movements are often restricted to readjustments of the sedimentary sequences within the basin and, in some instances, with uplift to both the north and south of the basin. The two orogenies are more significant. The Petermann Ranges Orogeny at the start of the Cambrian was confined predominantly to the SW margin of the Amadeus Basin, where it resulted in extensive folding, faulting and metamorphism of the sedimentary layers. The major tectonic event of the northern part of the basin is the Alice Springs Orogeny of Early Carboniferous time and is reflected in the formation of the Ormiston and Arltunga Nappe Complexes.

A review of the geology of the South Australian part of the Officer Basin has been given by Pitt et al. (1980) and of the West Australian part by Jackson and van de Graaff (1981). The stratigraphic relations between the various parts of the basin are in many instances questionable (e.g. Preiss & Forbes 1981) although some general

Table 1  Generalized stratigraphy of the central Australia basins. The ages are nominal only. The Officer Basin stratigraphy is adapted from Pitt et al. (1980), Krieg et al. (1976), Jackson and van der Graaff (1981) and Preiss and Forbes (1981). The Amadeus stratigraphy is adapted from Wells et al. (1970) and that for the Ngalla Basin from Wells et al. (1972) and Wells and Moss (1983). This stratigraphy ignores the many further subdivisions and the names adopted here reflect more convenience than precision.

<table>
<thead>
<tr>
<th>Approximate ages (Ma)</th>
<th>Officer Basin</th>
<th>Amadeus Basin</th>
<th>Ngalla Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Late Proterozoic</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1000–900</td>
<td>Townsend Qz</td>
<td>Dean Qz</td>
<td>Vaughan Springs Qz</td>
</tr>
<tr>
<td></td>
<td>Pindin Bd</td>
<td>Pinyinna Bd</td>
<td>Albinia Fm</td>
</tr>
<tr>
<td></td>
<td>Wright Hill Bd</td>
<td>Bitter Springs Fm</td>
<td>Naburula Fm</td>
</tr>
<tr>
<td>900–800</td>
<td>Rodda Bd</td>
<td>Inindia Bd</td>
<td>Rinkabeena Sh</td>
</tr>
<tr>
<td></td>
<td>Tapley Hill Fm</td>
<td>Areyonga Fm</td>
<td></td>
</tr>
<tr>
<td>800–700</td>
<td>Punktcr Bd</td>
<td>Souths Range Movement</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Observatory Hill Bd</td>
<td>Winnall Bd</td>
<td></td>
</tr>
<tr>
<td>700–600</td>
<td></td>
<td>Pertatataka Fm</td>
<td>Mt Doreen Fm</td>
</tr>
<tr>
<td><strong>Palaeozoic</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>580–500</td>
<td>Wanna Bd</td>
<td>Petermann Ranges Orogeny</td>
<td>Yuendumu Sd</td>
</tr>
<tr>
<td></td>
<td>Mount Chandler Sd</td>
<td>Pertaoorrra Gp</td>
<td></td>
</tr>
<tr>
<td>500–420</td>
<td>Carmichael Sd</td>
<td>Larapinta Gp</td>
<td>Dijagama Fm</td>
</tr>
<tr>
<td>420–370</td>
<td>Mereenie Sd</td>
<td>Mereenieie Sd</td>
<td>Kerridy Sd</td>
</tr>
<tr>
<td>370–320</td>
<td>Finke Gp</td>
<td>Pernjara Gp</td>
<td>Mt Eclipse Sd</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Alice Springs Orogeny</td>
<td></td>
</tr>
</tbody>
</table>
correlations with the southern Amadeus Basin have been established (Table 1). Some of the minor movements seen in the Officer basin may also correspond to those in the Amadeus. The only significant tectonic event occurred at the end of the Precambrian–Early Cambrian and is believed to be synchronous with the Petermann Ranges Orogeny (Jackson & van de Graaff 1981). For much of the basin this was a mild event, except that it became intense along the northern margin with folding and thrusting.

Basement within the Ngalia basin is overlain by the Vaughan Springs Quartzite which correlates with the Heavitree Quartzite to the south. Younger Proterozoic sequences, such as the Naburula Formation, Rinkabeena Shale and Mount Doreen Formation, are thin. Cambro–Ordovician sediments (the Yuendumu Sandstone and Djamarama Formation) are considerably thicker, totalling about 2500 m, and the Devonian–Carboniferous Kerridy and Mount Eclipse Sandstones also total about 2500 m in thickness. In a general way, the sediment sequence in the Ngalia Basin follows that within the northern Amadeus and most of the formations are separated by unconformities that correlate with those seen southward. The dominant tectonic activity was synchronous with the Alice Springs Orogeny and resulted in major thrusting and folding along the northern margin of the Ngalia basin. As for the other basins, Post Carboniferous sediments are patchy and thin.

The geological observations in the three basins can be characterized by two main cycles of tectonic activity since Late Proterozoic time. The first is a period of basin formation starting at about 900 Ma and culminating with the Petermann Ranges Orogeny in the Early Cambrian. Basin formation was most pronounced in the now-southern Amadeus and the Officer Basins, the evolution being similar in the two. Basin growth was initially slow, but increased with time up to the orogeny. Deposition appears to have been continuous and sediments formed predominantly in shallow water and continental environments. Basin growth was accompanied by significant uplift of the intervening Musgrave Block and, to a lesser degree, of the southern Arunta Block. These uplifted areas would have provided a major source for sediments. A shallow proto-Ngalia Basin may also have formed at this time. The second sequence of events, centred at the southern Arunta Block, started in the Cambrian and led to the further evolution of the northern Amadeus and Ngalia Basins. Early growth again appears to have been slow, but basin development increased rapidly leading up to the Alice Springs Orogeny. Significant uplift occurred within the southern Arunta in this time interval.

The ages and thicknesses and horizontal extent of the various sediment sequences in all three basins are only approximately known, but the trend of increasing rates of sedimentation appears to be valid throughout. For example, in the southern Amadeus, some 700 m of Pinyinna Beds were deposited from 900 to 800 Ma ago, nearly 200 m of Inindia Beds from 700 to 800 Ma, and at least a further 2000 m of Winnall Beds from 700 Ma to the Petermann Ranges Orogeny. The synorogenic Mount Currie Conglomerates average several kilometres in thickness. In the northern Amadeus the rates of sedimentation in the Precambrian were very slow in comparison, but in the Palaeozoic, reached nearly 2000 m in 70–80 Ma during the deposition of the Larapinta Group and 4000 m in the 100 Ma spanning the Mereenic and synorogenic Perntjara sequences.

The two major orogenies appear to be culminations of a sequence of subsidence and uplift events. Total uplift in both the Musgrave and Arunta may have exceeded 20 km (Goode & Moore 1975) but the time history of this uplift remains uncertain. If rates of uplift become excessive compared with rates of erosion, gravitational instabilities occur and extensive nappes may develop along the boundaries formed between the uplifting blocks and the subsiding basins. Failure of the crust as a whole in restricted zones, either by plastic flow or by brittle failure, may occur if or when the stresses associated with the deformations become excessive. The Petermann Ranges and Alice Springs Orogenies may therefore represent not so much the start as the end of the tectonic phase. The relatively widespread and mild movements such as the Areyonga and Souths Range Movements may be equally important as indicators of near-surface adjustments to increasing stress in the crust.

Geophysical observations relevant to the deep crustal structure of central Australia are sparse, being limited to gravity surveys and some deep crustal seismic reflection profiles (Brow 1970; Wells et al 1972). Observations of the shallow structure are more plentiful, for example, the seismic reflection surveys in the basins (Moss 1964; Froelich & Krieg 1969; Milton & Parker 1973) and aeromagnetic surveys (Milton & Parker 1973). The sediments in the basins alone can only account for about one-quarter to one-third of the gravity variations, and the anomalies illustrated in Fig. 1 point to further mass deficits below the basins (Lambeck 1983). The gravity observations
also point to a subdivision of the Amadeus Basin into two parts by the central E-W ridge of relative highs. The seismic reflection data constrains the sediment sequence in the basins and the deep reflection data (very limited in quantity) provide a constraint on the Moho depth in two localities. Systematic deep crustal seismic profiling has not yet been attempted at the time of writing.

A successful model that provides a basic framework for discussing the Late Proterozoic and Palaeozoic tectonics of central Australia must explain, by a quasi-continuous process, the formation of a series of parallel, linear basins and the coupled substantial uplifts of the intervening arches. The model must explain the two phases of tectonic evolution in each of which subsidence starts gradually but then accelerates until the actual orogeny results. Large negative gravity anomalies must be produced over the basins or at the basin margins, anomalies that cannot be attributed to the sediments alone. Finally, the model must maintain significant non-hydrostatic stresses for a very considerable time after the dominant basin-forming process has ceased to be significant.

In recent years a number of models for the intracratonic basins have been proposed (Anfiloff & Shaw 1973; Forman & Shaw 1973; Mathur 1976; Wellman 1978; Lambeck 1983). Anfiloff and Shaw proposed a model in which all lateral density anomalies lie in the upper 20 km of the crust, above the ‘Conrad’ discontinuity. The gravity highs correspond to uplift of granulite facies rocks whereas the lows are attributed to low-density granites and metasediments underlying the basins to depths of 20 km. The reason for restricting the density anomalies above a 20 km depth is not made clear. Possible they were motivated by rheological considerations, that the lower crust may not be able to support stress differences for extended periods, and that the Conrad and Moho discontinuities will, with time, follow surfaces of constant gravitational potential. Certainly the Conrad is not an ubiquitous feature of the crust and equally certain is that the Moho discontinuity varies in depth across many parts of Australia (e.g. Drummond 1979; Finlayson et al 1980) indicating that stress differences will persist for some time at these depths. The rheology of the lower crust remains essentially unknown, being dependent on chemistry, temperature, pressure, presence of water, and past tectonic history. What is known is that, if stress relaxation occurs, the first state acquired is one of isostasy, to be followed by a search for a hydrostatic state. Thus basins should be rebounding and the blocks should be subsiding with a concomitant undulation of the Conrad and Moho discontinuities. To prevent this from occurring, Anfiloff and Shaw (1973) postulated that the structure has been maintained through time by a large-scale compressive force. They made no attempt to estimate this force.

Forman and Shaw (1973) permitted a deformation of both the Conrad and Moho interfaces. The gravity highs were attributed to a Moho that is uplifted below the blocks such that lower crust granulites are exposed and flanked by amphibolite facies rocks. They postulated that thrusting occurred along the Woodroffe Thrust and the Redbank Zone on the two sides of the Amadeus Basin, but whether these ancient thrusts were reactivated during the Cambrian and Carboniferous Orogenies is not obvious. Large compressive forces again play a central role in creating the structure as well as in maintaining it throughout time. As pointed out by Mathur (1976), the degree of overthrusting in the model is perhaps excessive, since the resulting gravity anomalies are nearly twice those observed. This is readily modified and cannot be used as a serious argument against their model.

Mathur (1976) has proposed a section across central Australia that incorporates aspects of the Forman and Shaw model, in particular a Moho below the Musgrave and Arunta Blocks that is relatively shallow compared with the Moho depth beneath the basins. Thrust faulting has been introduced at several localities, including the northern margin of the Officer Basin, the north and south margins of the Amadeus Basin, and the northern margin of the Ngalia Basin. The mechanism that produced this structure is presumed to be one of the crustal folding and faulting, the basins forming in the downwarps and deep crustal rocks being uplifted to form the Musgrave and Arunta Blocks. As such, large-scale horizontal crustal compression is again implied.

All models discussed so far produce crustal sections that are out of isostatic equilibrium but which imply that some mechanical equilibrium is reached by horizontal compressive forces, surface and internal loads and buoyancy forces. Dooley (1973) attempted to evaluate the present stress state using an approximate formalism due to McKenzie (1967). This calculation is roughly comparable with the models discussed so far, except that the horizontal compressive force is not considered. In consequence, the stress differences are generally underestimated. Dooley concluded that, if the stress differences are confined to a 40 km thick lithosphere, the maximum stress reaches 500 bars.
For a 20 km thick lithosphere, as required by Anfiloff and Shaw (1973), the maximum stress will exceed 1.5 kbar (150 MPa) with this formalism. The rheology of the crust must be such as to be able to support these stresses since at least the time of the Petermann Ranges Orogeny in Early Cambrian time. Wellman (1978) departed from the above model by searching for solutions that are in isostatic equilibrium, either locally or regionally. In his model the positive anomalies occur over dense, high-grade metamorphic rock with a thicker-than-normal crust so that the mass excess is compensated by a mass deficit caused by the lowering of the Moho. The mass deficit is assumed to be at sufficient depth for the gravity anomaly to be dominated by the nearer-surface mass excess. Negative anomalies occur over the sedimentary basins which are assumed to be underlain by normal-density crust and by a relatively shallow Moho. Undulations of the Moho are in excess of 20 km, about 10 km deeper than average below the Arunta and Musgrave Blocks and about 10 km shallower than average below the basins. The stress state in this model is not insignificant, as supposed by Wellman (1978). If compensation is local, the stress differences at the Moho boundary are of the order $\Delta \rho g w$, where $\Delta \rho$ is the density contrast across the Moho, $g$ is gravity and $w$ is the depth difference of this layer. With $\Delta \rho = 0.3$ g cm$^{-3}$ and $w = 20$ km, the maximum stress differences are about 600 bars, of the same order as computed by Dooley (1973).

The stress state at the Moho can be reduced in the regional compensation model evaluated by Dooley by introducing models with depth-dependent rheology (Lambeck & Nakiboglu 1980, 1981). But this is not possible in Wellman’s model where the greatest stress differences are restricted to the lowermost part of the crust. Thus the stress state implied by his isostatic model may well place greater demands on crustal strength and crustal rheology than do the regional models supported by compressive forces. Figure 8 illustrates the main characteristics of these models. The differences are not so much in the density distributions as in the underlying physical implications of the models.

Rather than pursue these models further, it is more instructive to return to the basin-forming mechanisms discussed earlier and to compare them with geological observations. The passive gravitational loading mechanism is inadequate to form basins with depths in excess of a few kilometres if all sediments formed in shallow water and continental environments. This mechanism can therefore be seen only as one that amplifies the growth of basins initiated by thermal or stress mechanisms. As discussed by Beaumont (1981), the foreland basin model is very much associated with the active orogeny phase of the evolution of a geosyncline where the fold-thrust belt is driven by, for example, the continental tectonics associated with the subduction of an oceanic plate under the continent. These processes do not appear to be relevant to the intracratonic problem of central Australia. Instead, basin growth during both tectonic phases appears to take place on two sides of a rather narrow source region; for example, in the Late Proterozoic phase, two fold-thrust belts would have to migrate outwards in opposite directions and have been doing so for some $3 \times 10^8$ years. The Price-Beaumont model may be relevant to the late stages of basin formation, in the deposition of the synorogenic sediments — the Mount Currie Conglomerate of the Petermann Ranges Orogeny and the Brewer Conglomerate of the Alice Springs Orogeny. The mainly thermal based models also appear to offer unsatisfactory explanations for the central Australian basins. First, these models result in a nearly isostatic equilibrium state. Second, the time constants are problematical, the observed duration of the period of subsidence being very much greater than the conduction time constant. Third, the observed subsidence rates increase with time, whereas they decrease with time in the thermal models. Fourth, these thermal models do not address the question of uplifts of the areas adjacent to the basins. Finally, there does not appear to be any evidence for significant volcanic or igneous activity associated with the basin formation. Some basalts occur within the Bitter Springs Formation (Wells et al 1970; Bladon & Davies 1982) but their thickness does not appear to exceed 100 m. Nor is their extent known.

Stretching models are also inadequate. Conventional graben models as discussed by Bott (1976) cannot provide the broad and deep basins seen in central Australia. The Beaumont (1978) modification of this can produce broad basins, but the geometry of the basin floors do not match that predicted by this model. McKenzie’s (1978) stretching model is no better in this situation. In particular, the observations of gravity do not point to crustal thinning below the basins; rather the opposite. Nor is there evidence for graben and horst type observations in the basin floor. Instead, the limited seismic reflection data point to a very regular basin floor.
A model that matches the observations more successfully than the above ones is my compression model, since this was developed specifically for the central Australian region. It also incorporates some of the qualitative aspects of the models proposed by Forman and Shaw (1973) and by Mathur (1976). The application of this model to central Australia is discussed in the next section.

A SIMPLE MECHANICAL MODEL

The model is one in which the crust as a whole responds to compression. The crust or lithosphere is considered as a uniform plate overlying a fluid-like asthenosphere. This crust is loaded by a small but irregularly distributed normal load, which reflects past geological processes and topography. Under compression, any deflections \( w_i \) in the crust created by this load are magnified by an amount \( w_e \) that depends on the plate’s physical parameters, on the characteristics of the initial deflection, and on the magnitude of the compressive force. Wavelengths near the critical buckling value are considerably more magnified than others. Uplifted areas are eroded away with time, providing the sources for the sediments that fill the depressions and this ‘passive loading’ contribution magnifies the original deflection. More significant is the influence of a finite effective viscosity or relaxation time constant for the crust; with time the original elastic deflection \( w_e \) grows at a rate that depends on the magnitude of the compressive force, on the surface load, buoyancy forces and the elastic and viscous stresses. Provided enough time is available, deformation becomes significant, even if the crust’s viscosity is high. The model can be quantified mathematically be evaluating the deformation of a viscoelastic plate that is subjected to compression, to a surface or internal load, to a buoyancy force arising from the more dense but also much more ductile asthenosphere, and to the elastic and viscous forces within the plate.

The solution to the equations of deformation is given by (Lambeck 1983)

\[ w(t) = w_i e^{-\gamma t} e^{-\gamma x/l} \cos \left( \frac{\pi x}{2l} \right) \]

for times after \( t = 0 \) when the compressive force is applied. The deflection \( w \) is measured positively downwards. \( \gamma \) is positive and the deflection decreases with distance \( x \) away from the origin \( (x = 0) \) about which the deformation is assumed to be symmetric. \( l \) is the distance from the origin at which the deflection first reaches zero. Uplift occurs in the intervals \( 0 < x < l, 3l < x < 5l, \) etc. Subsidence occurs in the intervals \( l < x < 3l, 5l < x < 7l, \) etc. The relative importance of the successive zones of uplift (or subsidence) is determined by the attenuation parameters \( \gamma \). For \( \gamma = 0 \) there is no attenuation. Both \( \gamma \) and \( l \) are time-dependent, and their time evolution follows from

\[
\frac{1}{l} (\gamma, \pi/2) = \left[ \left( \frac{g(\rho_s - \rho_M)}{4D} \right)^{\frac{1}{2}} \left[ 1 + \frac{T_a}{\tau} - \frac{T_a}{\tau} e^{-\gamma t/\tau} \right]^{\frac{1}{2}} \right] \frac{w_i}{\tau} \left( \frac{w_i}{w_e} \right) e^{-\gamma t/\tau} \]
The various parameters are defined in Table 2 and discussed further below. With these equations the evolution of the basins can be computed throughout time for as long as the deformation remains in the viscoelastic regime; that is, up to the point where brittle failure or plastic flow occurs.

The parameters that characterize the forces and rheology of the crust or lithosphere have to be extracted from the geological observations themselves, both from the present morphology and from past records of uplift and subsidence. The principal unknowns are the effective flexural rigidity $D$, a measure of the elastic response of the crust to the applied forces, the effective viscosity $\eta$, and the effective compressive force $N$. I use the term ‘effective’ here in recognition that these parameters need not be constant with time, nor with depth, nor with horizontal location. Observations that provide some constraints on these quantities are the rates of uplift or subsidence, the wavelength of the basins, and the degree to which secondary basins form (Fig. 6). Independent estimates of these three parameters are not possible from the available data. For example, for an appropriate solution a comparable but not identical solution can also be found by increasing the effective flexural rigidity and decreasing the viscosity. Also important is a choice of the starting time. I have chosen the end of the Heavitree Quartzite deposition as the starting time on the grounds that these sediments and their equivalents appear to have formed a uniform blanket over much of central Australia. This is not to negate the importance of the earlier activity recorded in the exposed basement of the Musgrave and Arunta Blocks. Rather, these past events enter into the problem as the initial perturbations that influenced the subsequent evolution, with the advantage that, once initiated, the evolution is not strongly dependent on these conditions. Table 2 summarizes the model parameters used. Many of

Fig. 7  The geological cross section of the basin sediments across the Amadeus for the profile illustrated in Fig. 1 (based on isopach maps from Wells et al 1970, and from seismic data from Froelich & Krieg 1969).
these are not independent. The Maxwell relaxation time $\tau$ is defined by the ratio of viscosity to rigidity. The spatial attenuation factor $\gamma$ determines the ratio of the depths of the primary and secondary basins or of the uplifts of the primary and secondary intervening arches. For $\gamma = 0.75$ this ratio is about 5%, while for $\gamma = 0.50$ it is about 15%. Evolution of $\gamma$ will initially be rapid. The wavelength of the deformation is given by $4l$ and the width of the basin is $2l$. This quantity also evolves rapidly at first. For the parameters chosen, the final wavelength is about 50% of the initial value. $N$ is the compressive force. $w/w_e$ represents the ratio of the initial deformation to the elastic deformation when the compression is applied. The erosion parameter $\delta$ defines the amount of topography which is eroded away as uplift occurs. It also controls the basin depth. A value near unity is required (i.e. erosion keeps up with uplift) for much of the time since the basins cannot have been very deep for long periods. The reasons for the choice of these parameters is discussed in more detail in Lambeck (1983). In most instances the exact choice of individual parameters is not very important; this is because asymptotic solutions, which are not very sensitive to the original choice of $\gamma$ and $w/w_e$, are reached relatively quickly. Also the evolution is determined by the ensemble of the parameters, and different combinations of parameters can produce similar end results. One example is the trade off between viscosity and the magnitude of the compressive force.

Figure 9 illustrates the predicted sequence of events in the Late Proterozoic. The origin ($x = 0$) is initially taken at the centre of the present Musgrave Block. Early uplift of the Musgrave Block is slow, as is the evolution of the proto-Amadeus and Officer Basins over the 100 Ma following the deposition of the Dean–Heavitree Quartzites and their correlates to the north and south. The Pinyinna Beds and Bitter Springs Formation, deposited in the slowly subsiding basin, total perhaps 700 m in thickness (Wells et al 1970), although this amount is uncertain. Figure 10 summarizes the predicted deflections at several localities after $10^8$ years. During this time interval, the length parameter $l$ has been reduced and the basin margin has migrated by some 25 km towards the origin at $x = 0$. This migration can result in a complex margin and, in particular, the margins of the basin will be asymmetrical (Fig. 11). This complication has been ignored in Fig. 9.

If, at the end of the Pinyinna and Bitter Springs deposits, the compressive force is significantly reduced or ceased to operate, the basins will rebound and the arches subside. The previously deposited basin sediments erode away, resulting in a discontinuity in the sediment sequence. If the compression is assumed to have continued, the thickness of the Inindia Beds–Areyonga Formation deposited over the next 100 Ma can be predicted (Fig. 9). Uplift within the Musgrave Block becomes significant now, as does the uplift in the southern Arunta. In the absence of external sources, sediments deposited in the Amadeus Basin would come predominantly from the erosion of the slowly uplifting Musgrave block. Initially this would be the Dean Quartzite to be followed by the older underlying metamorphics. The Officer Basin will form a mirror image of this early stage of the Amadeus Basin with some 2 km of Inindia correlatives, possibly the Rodda Beds and Tapley Hill Formation, overlying the Wright Hill Beds. To the north of the Amadeus Basin a shallow basin would have formed with some 750 m of sediments. This could be the Naburula Formation.

The total stress state can be approximated by the super-positioning of the overall compression and the bending stresses. Quantitative estimates of these stresses are difficult to predict as this requires a detailed knowledge of the rheology. What can be predicted is where maximum stress differences occur. This is in the neighbourhood of $1.25l < x < 1.75l$ (Lambeck 1983), and if failure occurs this is in the form of shallow-angle thrust faulting near the surface (Fig. 12). The predicted fault plane steepens with depth.
Minor movements, reflecting readjustments of the basin sediments to the changing stress state, will occur throughout time but no significant region-wide break, involving the crust as a whole, appears to have occurred between the Inindia and Winnall sequences. Hence the computation is pursued for a further $10^8$ years up to Early Cambrian time. By this time, some 300 Ma after the onset of folding, the total predicted uplift along the Musgrave Block axis is such that rocks initially at depths of more than 17 km, 1000 Ma ago, are now at the surface. The southern part of the Amadeus Basin is now filled with 5–6 km of Late
Fig. 11 Schematic evolution of the margins of the proto basin from time \( t_1 \) to \( t_2 \). The basin margin moves inwards with time such that near the surface of the Inindia Beds are in direct contact with the Dean Quartzite. The Pinyinna Beds are exposed at the surface along the margin at \( x = 3l \). The subsequently deposited Winnall Bed may also be in direct contact with the Dean Quartzite if \( l \) continues to decrease with time.

Fig. 12 Schematic representation of the failure of the crust and thrust faulting during the early Cambrian Petermann Ranges Orogeny. The thrust fault will be inclined at between 30° and 45° to the horizontal with the Musgrave basement thrusting over the Proterozoic sediments. The synorogenic Mount Currie Conglomerates form at this time. The exact sequence of layers depends on where failure actually occurs. If it occurs at \( x < 1.5l \) then the Pinyinna Beds may be missing.

Proterozoic sediments. The Officer Basin has reached a similar depth with a comparable sedimentary sequence. Uplift of the southern Arunta is predicted to be about 1.8 km by this time. The shallow basin to the north has deepened a further 400 m with sediments that will be largely indistinguishable from the earlier sequence if no sediments originating from external sources have been deposited.

By Early Cambrian time, the model leads to stress differences in the crust that may well become excessive. Departures from viscoelastic deformation should therefore be anticipated. Thrust faulting could occur in the southern part of the Amadeus Basin with the basement moving over the basin sediments. If the crust is homogeneous this failure will occur near \( x \sim 1.5l \). But if the crust has a much older history of failure, this may occur at locations where the local stress differences first exceed the local strength of the crust. Thus movements along the Woodroffe and Davenport Thrust faults within the Musgrave Block may well have occurred at this time although the evidence for this is not overwhelming. Whether failure occurs in both the Officer and Amadeus Basins depends partly on the Proterozoic history of the crust. The geology does point to failure having occurred within both. The Petermann Ranges Orogeny at the northern edge of the Musgrave Block has been well documented (e.g. Forman 1966) and according to Jackson and Van der Graaff (1981), a comparable tectonic event occurred at about the same time at the southern margin of the Musgrave Block. This is confirmed by recent

Fig. 13 The predicted sedimentary sequences along the southern margin of the Amadeus Basin at some time after the Petermann Ranges Orogeny. Total displacement along the fault will, in this case, have been of the order of 5 km. If gravitational sliding occurred on the uplifted Block then remnants of Dean Quartzite, Pinyinna Bed or Inindia Bed may also be found to the north of the fault.
seismic reflection work carried out by the South Australian Department of Mines and Energy. Two sections across the southern margin indicate a thrust-fault down to at least 5 km depth. Milton and Parker (1973) had previously inferred such a structure from magnetic and gravity data. Prior to failure, any gravitational instabilities in the rapidly uplifting block may result in the development of nappes. It is the combination of the failure of the crust as a whole, along one or several faults, with the gravity sliding that makes up the Petermann Ranges Orogeny in this model. The thick sediments forming the Mount Currie conglomerate and the arkose of Ayers Rock would have been deposited at this time. It is at this stage that the foreland basin mechanism of Price and Beaumont may be appropriate. Figure 13 illustrates how the basin margin may have appeared some time after the orogeny. The actual deformation will depend on many factors. These include (i) the horizontal extent of the region of stress relaxation across the basin (ii) the amount of shortening of the scale parameter / that took place prior to failure, (iii) whether extensive nappes developed over the basin margin prior to, and at the time of, failure, in which case remnants of older material may be left on the younger formations such as the Winnall Beds, (iv) the exact distance x at which the failure plane developed, (v) the amount of movement that took place on this plane, and (vi) the amount of localized deformation that occurred within the basin prior to the time of the crustal failure. Along the margin the Inindia and Winnall Beds may lie unconformably on the Dean Quartzite or even on the older basement rocks. Bedding planes of the Winnall Beds may dip steeply on the basinward side of the fault, while on the other side any exposed Pinyinna or Dean deposits may also dip quite steeply. Within the centre of the block the bending stresses are tensional in the upper part of the crust, and the overall stress state may be one of tension if the magnitude of the localized bending stresses exceeds the overall compression. Hence, shallow graben-type features could form within the centre of the Musgrave Block. The predicted stresses correspond to those in the basement crust and upper mantle. The stress state within the basins themselves will be much modified by the

Fig. 14 Palaeozoic phase of tectonism computed on the assumption that no major faulting occurred previously. The upper figure corresponds to the Cambro–Ordovician sequences from time t4 to t5 (the Pertaaorria and Larapinta Groups) and the lower figure to the Siluro–Devonian sequences from time t6 to t7 (the Mereenie Sandstone and Pertnjara Group).
presence of weak or ductile layers, such as the evaporite sequences in the Bitter Springs. Failure within the basins will also occur at stress differences that are much less than those that cause failure of the crust as a whole. The sediments, deposited in a more or less continuous manner, may therefore be separated by minor discontinuities that reflect such failure and movements. They may also reflect a time dependent compressive force.

Once major failure is initiated, the analytical solution is no longer valid in any strict sense. Nevertheless once this has occurred, the stress differences are reduced in the vicinity of the failure zone, but not necessarily further away. If compression is maintained or reintroduced, deformation will occur predominantly where the stress differences are still significant. This turns out to be at a distance of about 4T from the original centre of uplift, with the consequence that deformation will now centre over the southern Arunta or at a similar distance to the south of the Officer Basin. The choice is again decided by the past geological history of the crust and the former is the obvious choice. Whether the second phase of activity started immediately upon the termination of the Petermann Ranges Orogeny, or whether there was first a period of basin rebound, remains uncertain. It is possible that part of the Winnall Bed and its correlations have eroded away but the evidence for this is again not obvious. What is clear is that at some time after this orogeny, significant uplift commenced in the southern Arunta, and basin evolution is now centred on the northern part of the Amadeus Basin and the Ngalia Basin. In comparing the computed deflections with the observed values in these two basins there is a suggestion that either the relaxation time \( T \) has increased since the Late Proterozoic, or that the magnitude of the compressive stress has decreased. If not, the subsequent rates of evolution become excessively large. To avoid this, \( T \) has been increased slightly in the calculations of the Palaeozoic phase of the evolution.

The Amadeus Basin formation continued with the deposition of the Pertaoorrtt Group over a 80–90 Ma time span into Upper Cambrian time, with the deposits being confined mainly to the northern part of the basin (Fig. 14). At the same time, a similar subsidence is predicted to the north, corresponding to the Yuendumu sequences of the Ngalia Basin. Uplift of the southern Arunta, slow in Precambrian time, may now be substantial (Fig. 10). Because the wavelength of folding decreases with time, the areas of secondary uplift will not necessarily coincide exactly with the regions of uplift during the early period of the Precambrian phase of the evolution. In particular, uplift of part of the southern or central Amadeus may occur. This results in partial erosion of the Winnall Bed and in the deposition of some Cambrian sediments on the Precambrian basement of the northern margin of the Musgrave Block.

Over the next 70–80 Ma, into the Ordovician, the northern Amadeus Basin will have evolved further, corresponding to the Larapinta sediments whose thickness is about 2 km. An equal amount of sediments is predicted for the Ngalia Basin and this could be the Walabri Dolomites, Djagamara Formation and related deposits. Uplift of the northern Arunta also becomes significant, reaching 1.5 km. Secondary basins may develop further northwards. To the south the central Amadeus continues to be uplifted and Ordovician deposits form near the margin with the Musgrave Block. These and the earlier Cambrian deposits may be largely indistinguishable from each other. More significant is that they mask the Early Cambrian fault zone along much of the Amadeus margin.

Predicted subsidence and uplift increases with time and by the Devonian the rates have become very significant (Fig. 10). This is what is observed in the Amadeus Basin, with some 2 km of Mereenie Sandstones deposited in about 40–50 Ma and a further 2 km of Pertnjara sediments deposited in the next 50 Ma up to Early Carboniferous time. Similar sediment sequences are seen in the Ngalia Basin. Predicted uplift of the southern Arunta also becomes rapid and the total uplift predicted by Late Devonian time is about 25 km. Figure 14 illustrates the sequence of events leading up to Late Devonian time. The northern Amadeus and Ngalia Basins are well developed and differ only in that the former contains more substantial thicknesses of late Proterozoic sediments. The secondary basin to the south may partly overlie the Musgrave while the secondary basin to the north corresponds approximately to the location of the Lander Trough in the Wiso Basin.

Stress differences by this time will have reached levels that are comparable to those found to the south in Early Cambrian time. Failure can therefore be expected, and the predicted sequence of events is comparable to that discussed for the Petermann Ranges Orogeny. To the south of the Arunta, failure is predicted within the northern Amadeus Basin, with the Arunta being thrust over the basin. As before, the location of the failure
zone may be influenced by the presence of older and reactivated fault zones, possibly including the Redbank Zone. But information on Palaeozoic movement in this region is incomplete. Movements along these failure zones, together with gravitational sliding of the rapidly uplifting southern Arunta, provide the elements for the Alice Springs Orogeny. During this event, much of the Pertnjara Group, particularly the Brewer Conglomerate, would have been deposited in the rapidly subsiding basin.

Stress differences will also have become excessive to the north; this time in the Ngalia Basin. If the stress state is wholly determined by the deformation during the Palaeozoic phase of tectonism, failure occurs preferentially in the southern half of the basin, with the fault plane dipping southwards. If, on the other hand, the residual stress from the Precambrian phase is included in the evaluation of the overall stress state, then the predicted failure plane is located in the northern half and dips northwards. The actual location is again influenced by any pre-existing zones of weakness in the crust. The geological evidence points to a thrusting of the northern Arunta over the basin (Wells et al. 1972) and sediments, such as the Mount Eclipse Sandstones, are probably synorogenic.

The predicted structures of the three basins by Carboniferous time are illustrated in Fig. 15. The Officer Basin will have been subjected to only minor adjustments since the Early Cambrian Orogeny, possibly with a shallow depression having formed in the northern part of the basin that is filled with Palaeozoic sediments. Unlike the Officer Basin, the Amadeus Basin has undergone considerable modification since the Petermann Ranges Orogeny (cf. Fig. 9). In the central part of the basin, significant uplift is predicted to have occurred, bringing the Late Proterozoic sediments to or near the surface. The northern part of the basin is now much deeper and the margin will be defined by the overthrusting of the southern Arunta. Little tectonic activity will have occurred.

Fig. 15 Predicted sediment sequences in the three basins with the predicted locations of major thrust faulting. The Amadeus Basin results can be compared with the observed section in Fig. 7. The shallow, post-Cambrian, secondary basin to the south of the Amadeus corresponds to the Finke and Carmichael Beds located mainly to the east of the section of Fig. 7. Note the vertical to horizontal exaggeration and that the fault planes will be inclined at about 45° to the horizontal.
in the Musgrave Block since the Petermann Ranges Orogeny. Rocks at a depth of 18 km \(10^9\) years ago should now be near the surface in the central regions of the block. This is illustrated schematically in Fig. 16 where the pre-Heavitree crust is idealized by stratified upper and lower crustal layers. The older tectonic activity is conveniently ignored. But this is not essential. If the older history can be independently developed it can be superimposed on these results of the later evolution. Uplift in the southern Arunta is more significant, totalling some 25 km in the central region. The metamorphics should therefore come from increasingly greater depths as one moves northwards from the contact with the Heavitree Quartzite to the region of maximum uplift. Again, what is actually exposed depends on the pre-Heavitree history.

That some post-Carboniferous evolution of the central Australian area has occurred is seen in the present-day morphology (Burek & Wells 1978) and in the limited Mesozoic and Cainozoic sedimentary history (Wells et al 1970). In the Amadeus Basin, scattered remnants of Permian sediments crop out towards the eastern and western parts of the basin and some Mesozoic sediments are found only in the SE. These point to a period of further subsidence followed by rebound, so that much of the post-Carboniferous sediment cover has been removed by later erosion. This implies a reduction in the magnitude of the in-plane force to the point where the folds in the crust can no longer be supported. Burek et al (1978) concluded, however, from a detailed analysis of field observations, morphology and air and satellite imagery, that the Cainozoic structural features are a consequence of N–S compression and that many major Carboniferous and older faults have been reactivated in the Cainozoic.

**CONCLUSION**

The model that has been developed is physically and mathematically simple and obviously can do little more than provide a general framework for discussing the evolution of central Australia since Late Proterozoic time. The mathematics and physics follow from the engineering literature (e.g. Timoshenko & Woinowsky-Krieger 1959; Nadai 1963; Flugge 1967; Kempner 1962; DeLeeuw & Mase 1962) and have been tested in many situations. One example is the stress analysis of a strapless evening gown (Siem 1963). If gowns occasionally fall down, this does not reflect on the inadequacy of the mathematical formulation but rather on the inappropriateness of the assumed boundary conditions.

A central assumption is that the continental crust has been in a state of compression for much of the time since the Late Proterozoic. The magnitude of this compressive force is model dependent and with the parameters given in Table 2, the force required is some \(10^{16}\) dyne cm\(^{-1}\). If the effective
STRUCTURE OF AMADEUS, OFFICER AND NGALIA BASINS

thickness of the crust is 30 km, then the compressive stress is about 3 kbar (300 MPa). For a 50 km thick layer, this stress is about 2 kbar. These are time-average values. Most old continents appear to be in a state of horizontal compression (e.g., Denham et al. 1979; Lambeck et al. 1984) but this need not be a universal condition. A test of the assumption may be possible from the examination of the sediment deposited in the basins. For example, unconformities may be indicative of periods of reduced compression. Models with time-dependent $N$ need to be developed. When matching the model predictions with the observed sediment thicknesses, there is a suggestion that the evolution becomes too rapid. Either, the viscosity of the plate must be increased with time or the average compression must be reduced. As the evolution proceeds, stresses increase, and if the rheology is non-linear the deformation will proceed faster than predicted by the model. But non-linear rheology models are not amenable to simple analytical solutions, and I have not attempted to explore this further. However, it is clear that, once started, the evolution will proceed with a reduced compressive force. Other arguments can also be marshalled to de-emphasize the importance of this rather fundamental assumption. One is that as erosion of the uplifted region proceeds, the crust becomes mechanically weaker because the effective flexural rigidity is proportional to the cube of plate thickness. Deformation is also faster than predicted by the above formulation, unless the compressive force is reduced. A second argument is that, as folding proceeds, isotherms in the crust and upper mantle are perturbed with warmer and more ductile crust occurring under the uplifts. This further accentuates the rate of uplift.

A second fundamental assumption is that the rheology of the crust can be characterized as a viscoelastic solid; that the overall response to deformation is first an elastic response, followed by time-dependent evolution. Stress relaxation will occur not only on the scale of grain sizes. It will also occur on a much larger scale by localized brittle failure. This would be reflected by microseismicity and very small earthquakes in the vicinity of the northern side of the Ngalia Basin which are occasionally recorded at the Australian National University’s seismic array to the north. More generally, the upper crust behaves as an inhomogeneous elastic medium when subjected to small stress differences and a more appropriate lithospheric model may be one of an elastic layer over a viscoelastic lower crust. In the absence of compressive forces the deformation (but not the stress state) of such a layered crust is not very different from that of an equivalent homogeneous viscoelastic plate (Lambeck & Nakiboglu 1981) and it would appear that the same assumption can be made here. This avoids the use of the more laborious models.

A third fundamental assumption is that the crust is a homogeneous layer. But, as this layer has been subjected to a surface or an internal load which initiates the process, an equivalent assumption is that the crust has been subjected to a complex thermal and tectonic history and that this is followed by a period of quiescence in which a state of local isostatic compensation is reached. If this early history is known, it can be superimposed on the results predicted by the present model to give the total present-day structure.

The boundary conditions imposed in the model have not been discussed here in any detail. These include the conditions away from and parallel to the structure. The choice of conditions is not very significant in computing the deformation of the crust as a whole, but is more important in determining deformations within the sedimentary basins. The stress state is more sensitive to the choice of boundary conditions, and within the basin some failure planes may be oriented at about $45^\circ$ to the present N–S axis.

Observations of the sedimentary sequences within the basins provide constraints on the evolution of the crust as a whole, something that may not always be appreciated by geophysicists. One would like to have more complete mapping of some of the principal sedimentary sequences and of their depositional chronology. Also it should be recalled that the deformation of the sediment layers in the basins may be considerably more complex than illustrated here because the stress state within the basin has not been modelled. This still needs to be done. Information on likely sediment sources and palaeocurrents will also be helpful, since the model predicts what these sources may be. But the evolution of central Australia did not occur in isolation of other events. In particular, the occasional marine intrusions in Palaeozoic time provide additional sources of sediments from outside the region.

Deep crustal seismic reflection profiles will provide a most important test of the model since the Moho is predicted to oscillate significantly in depth, being considerably deeper under the basins than under the Arunta and Musgraves. This is also characteristic of the models of present structure.
ACKNOWLEDGMENTS

Discussions with P. J. Cook, R. Shaw, R. Stephenson and A. T. Wells have been most helpful, as have their extensive comments on a previous version of this paper.

REFERENCES


*(Received 17 February 1983, accepted 12 May 1983)*