distinctive quartz rich sandstone sequences at the basin margins, e.g. the Algebuckina Sandstone in the southwest (Forbes, 1982; Moore, 1982, 1986) and the Ronlow Beds (Ranmoor and Jones Valley members) in the northeast (Burger and Senior, 1979; Burger 1982, 1986).

By the end of the Early Jurassic, the depositional environment in the Eromanga Basin sequence was connected by drainage systems to the Surat Basin to the east (Wiltshire, 1982b) and the Carpentaria Basin to the north (Smart and Senior, 1980; Smart et al., 1980). The Jurassic sedimentation in the Eromanga Basin is considered to be of a cyclic nature with 3 cycles of fining upwards clastics (Exon and Burger, 1981; Gravestock, 1982; Burger, 1982, 1986; John and Almond, 1987) in the central and northern Eromanga Basin region. The sandstone rich members of the three cycles are the Hutton, Adori/lower Namur and Hooray/upper Namur Sandstones and these represent a dominantly fluvial environment of deposition. The shale rich units are the Birkhead, Westbourne/middle Namur and Murta Formations and these are interpreted as either low energy fluvial or lacustrine. The temporal and geographical relationships of the sediment types within the cycles is complicated by facies variations, reflecting the changing depositional environments (e.g. Gravestock, 1982). For example, the Murta Member is very thin north of 26°S (Ambrose et al., 1982, 1986) where the Hooray Sandstone is relatively thick (Senior et al., 1978). Volcanogenic material is apparent in the finegrained sandstones within the shale sequences and also, to a lesser extent, in the upper two sandstone rich sequences (Senior et al., 1978; Gravestock, 1982; Watts, 1987; Duddy, 1987). The cyclicity is directly attributable to variations in the sediment source areas and the energy of the depositional environment. Veevers (1984, pg 265), P.S. Moore et al. (1986a), and Duddy (1987), noting the progressive increase in the proportion of volcanic material eastwards from the Eromanga Basin, across the Surat Basin to the Maryborough Basin (see fig. 3.1.1), attributed the lithological variations to fluctuations in the tectonic activity of a volcanic arc situated off the present day east coast of Queensland.
Certainly it is likely that the lithological variations are an immediate consequence of changes in the dominant direction of drainage of the major river systems. As the rivers drained eastwards out through the Surat Basin, cratonically derived quartzose detritus would have been deposited over the Eromanga Basin. Westerly flowing rivers initiated as a result of increased activity at the eastern margin (volcanics and probably an enhanced topography) would have ultimately drained into the Eromanga Basin region. The continued, albeit reduced, influence of easterly flowing rivers during these periods is attested to by the presence of over 500 m of the quartzose Algebuckina Sandstone, of Early to Late Jurassic age, in the western Eromanga Basin (Forbes 1982, Moore 1982, 1986).

By the time of deposition of the fluvial Hoornay Sandstone, the area of sedimentation had increased such that the present day outcrop limits of the Eromanga Basin were reached (fig. 3.1.2). By the Late Neocomian, the first marine influences were seen in the Eromanga Basin during the deposition of the marginal marine/paralic Cadna-Owie Formation or Transition Beds, a coarsening upwards sequence of clastics up to 100 m thick. During this time the connection to the open sea is thought to have been to the east across the Surat Basin where the laterally equivalent shallow marine/paralic Bungil Formation is about 150 m thick (Exon and Senior, 1976). As with earlier Jurassic, and later Cretaceous, sedimentation, the Cadna-Owie Formation reflects the proximity of the basement at the southern, western and northern margins where sandstones, often quartzose and of fluvial or shore line affinity, are developed (Exon and Senior, 1976; Forbes, 1982, 1986; Moore and Pitt, 1984; John and Almond, 1987).

Subsequently, during the Late Aptian, shallow marine conditions became widespread across the Eromanga and Surat Basins with the deposition of 150-200 m of a generally calcareous mudstone sequence, the Doncaster Member (Wallumbilla Formation) or its equivalent in the south west of the basin, the Bulldog Shale. In addition to the eastern connection to the open sea, marine influences came from the north through the Carpentaria Basin and this connection rapidly became dominant (Smart and Senior, 1980; Burger, 1982, 1986). The Coreena Member (Wallumbilla Formation), about 150 m thick, marks the
first regressive phase of the shallow marine environment and the upwards coarsening character of the unit reflects the increasingly paralic nature of the environment, especially in the eastern area of the Eromanga Basin. This is more evident in the Surat Basin where the equivalent of the Wallumbilla Formation sediments, the Griman Creek Formation, is of a paralic nature and slightly thinner, about 250 m (Exon and Senior, 1976). This supports the notion that the eastern connection to the sea was restricted. By this time there also seems to have been a distinct change in the provenance region as the detrital character of the sediments changed from quartzose/sub-labile to labile, being dominated by contemporaneous volcanics. Lack of evidence of a volcanic terrain to the south or west of the Eromanga Basin led Exon and Senior (1976) to conclude that a major arc system existed at this time off the present day eastern coast of Queensland similar, but more active, to that proposed in the Jurassic and the transport of large volumes of fresh (i.e. unweathered), often coarse volcanic detritus over distances of up to 2000 km was probably facilitated by a cool climate (Quilty, 1984). The presence of this arc system is supported by the observation of dacitic-andesitic pyroclastics of Albian age in the Whitsunday Island Group (Paine 1969) and in the Maryborough Basin area (Ellis, 1968; Ellis and Whitaker, 1976) on the eastern coast of Queensland.

Widespread shallow marine conditions returned in the Middle Albian, marked by the deposition of a thin (<60 m) but distinctive black carbonaceous bituminous shale/siltstone sequence, the Tooheebuc Formation in the central and northern Eromanga Basin and the laterally equivalent Woolridge Limestone in the south west and the Urisono Beds in the southeast (Ozimic, 1986; McMinn and Burger, 1986). During this time, the Eromanga Basin had a northerly dipping regional slope and the facies variations in the units mentioned above are attributable to a shallowing of the sea to the south. The marine conditions did not reach the Surat Basin and the Nebine Ridge, a persistent structure at the eastern margin of the Eromanga Basin during its history, began to act as a depositional divide (Senior, 1971). Sedimentation had effectively ceased in the Surat Basin after deposition of the paralic Griman Creek Formation, while in the Eromanga Basin, the shallow marine Allaru
Mudstone (Late Albian) was deposited. This unit is thin in New South Wales compared to over 200 m in the central Eromanga Basin. Its equivalent in the south west of the basin, the Oodnadatta Formation, similarly shows thinning and facies variations towards the basin margin (Moore and Pitt, 1982, 1984, 1985; Forbes, 1982). By the Latest Albian, the detrital character of the sediment was again markedly volcanioclastic and the Mackunda Formation, a series of fine-grained clastics generally 60-120 m thick, was deposited in a shallow marine/paralic environment. Facies variations in the Eromanga and Carpentaria Basins indicate that the sea was retreating to the north at this time (Exon and Senior, 1976; Smart and Senior, 1980; Smart et al., 1980; Burger, 1982, 1986). The final stage of the Cretaceous depositional history of the Eromanga Basin is represented by the fluvial Winton Formation (Cenomanian), a thick (up to 1200 m, Moore and Pitt, 1982) sequence of sandstones and siltstones, with minor shales and coals.

The maximum thickness of the Jurassic-Cretaceous sediments is observed over the southern Cooper Basin (e.g.Burley 2, 2300 m) and Simpson Desert Basin (e.g. Walkandi 1, 2700 m), and the depocentres are reflected in the present day isopachs and structure contours (fig. 3.4.1). These two areas are separated by the Birdsville Track Ridge across which the sediments are continuous but tend to show thinning (e.g. by a total of about 200 m for the Eromanga Basin sequence). This feature exerted a secondary influence on the subsidence mechanism of the basin and the main area of subsidence may be regarded as over the Cooper/Simpson Desert Basins. Outside this central area, the sediments tend to thin and show predictable facies variations towards the basin margins in the south west and north.

The depositional environments during subsidence and sedimentation in the Jurassic/Cretaceous can be summarized as follows:

The Jurassic (Basal Jurassic-Hooray Sandstone) was predominantly fluvial with some evidence of volcanically derived detritus coming from the east after the Middle Jurassic (from the Birkhead Formation on). A transition period before widespread marine
Fig. 3.4.1  Structure contours (metres below sea level) on the 'C' seismic horizon, near the top of the Cadna-Owie Formation (Early Cretaceous) in the Eromanga Basin. The contour interval is 500 m. The maximum sediment thicknesses occur in the southern Cooper Basin region and the Poolowanna Trough (adapted from Armstrong and Barr, 1986).
conditions is marked by the paralic Cadna-Owie Formation, although fluvial conditions persisted at the basin margins at this time (upper Hooray Sandstone). Initially the marine influence came from the east but a northern opening through the Carpentaria Basin quickly became the dominant control on marine conditions. The sea was at all times shallow and two apparent transgressive/regressive sequences are recognized. Volcanically derived detritus was prominent during the regressive phases. The first phase tends to correlate with a slowing up and eventual cessation of sedimentation in the Surat Basin. The final stage of fluvial sedimentation in the Cenomanian was rapid (over 1 km in, at most, 10 m.y.). The absence of contemporaneous sediment in the Surat Basin implies that the volcanogenic detritus bypassed the Surat Basin area and was effectively dumped in the Eromanga Basin.

3.5  ? Late Cretaceous-Tertiary sedimentation, post-Eromanga sediments

Post-Winton Formation, possibly Cretaceous, sediment, the Mt. Howie Sandstone, has been been observed in two localities in South Australia: near Innamincka (Wopfner, 1963, see well 4 on fig. 3.1.1) and near Maree (Forbes, 1972, about 320 km SW of Innamincka) although the latter is possibly Tertiary. However, the sandstone is at most 50 m thick, and developed on a very local (i.e. kilometre) scale and is not considered significant in the context of the basin evolution. Apart from the Mt. Howie Sandstone, the Upper Cretaceous to early Tertiary was marked by an lengthy period of gentle structuring and deep (100 m) chemical weathering of the labile (feldspar rich) Winton Formation (Senior et al., 1978). The Late Palaeocene/Eocene Eyre Formation (Wopfner et al., 1974) was deposited over much of the basin. This sequence is generally less than 70 m thick but thicknesses of up to 140 m are observed in synclines. This quartzose fluvial sandstone/conglomerate is thought to be derived from Jurassic and older sediments exposed at the northeastern margin of the Eromanga Basin and the dominant drainage direction was to the south west of the basin, i.e. Lake Eyre area (Wopfner et al., 1974).
3.18

Thin Quaternary sediments are also areally extensive in the Eromanga Basin but are generally less than 100 m thick (e.g. Packham, 1969) and these are attributed to reworking of the uppermost Cretaceous sediments (Senior et al., 1978). The thickest sequences of these sediments tend to be in areas of thickest Mesozoic sediments which were presumably topographic lows.

3.6 Structures in the Eromanga Basin

In detail the structural style of the Eromanga Basin is complex (e.g. Nelson, 1985). However consistent regional trends have been recognized from surface geology, remote sensing and seismic reflection data. The broad patterns are summarized below:

The dominant fault and fold axis directions lie between northeast and northwest. A predominantly N-NW trend is apparent in the northern Cooper Basin and Adavale Basin areas (fig. 3.3.3, and Senior et al., 1978; Senior and Habermehl, 1980; Pinchin and Senior, 1982; Moore and Pitt, 1984; Kuang, 1985; Finlayson et al., 1987a,b). In the southern Cooper Basin the major structural trend is NE and the northern and southern areas are separated by the WNW trending Pepita-Wackett-Nockatunga trend (fig. 3.3.3), which acted as a depositional hinge up to the end of the Permian (Stuart, 1976; Battersby, 1976; Moore and Pitt, 1984; Kuang, 1985). The major N-S trending structures seen in the Adavale Basin area were formed during the Carboniferous as a result of E-W compression (Pinchin and Senior, 1982; Wake–Dyster et al., 1983; Mathur, 1983, 1987; Finlayson et al., 1987a,b; Leven et al. 1987).

Subsidence in the early Permian was facilitated by active faulting and the dominant style is oblique reverse faulting, generally with a major dextral strike-slip component (Gravestock and Morton, 1984; Kuang, 1985; Nelson, 1985; Vincent et al., 1985; Delhi Petroleum, pers. comm., 1986). Mild thrust faulting occurred from a SE direction but the
overall oblique movement resulted in the thickest sequence of sediments forming at the northern end of the Nappamerri Trough where pronounced curvature of the structure allowed extension to occur locally (Kuang, 1985).

The influence of syn-depositional faulting declined through the Permian and by the Early Triassic, sedimentation was continuous between the northern and southern Cooper Basin areas. Kuang (1985) relates NW trending structures in the northern Cooper Basin and the erosional unconformity at the top of the Nappamerri Formation (Middle Triassic) to a NE-SW directed compressional episode. These structures are generally narrow, linear features but show some asymmetry in cross-section with the steeper reverse faulted side facing the NE. The development of the reverse faulting is interpreted as resulting from tectonic movement to the NE during uplift of the area to the SW of the Cooper Basin, and the narrow, linear nature of the structures as being consistent with a strike-slip or wrench component, although the sense of movement is unclear (Delhi Petroleum, pers. comm., 1986). As the Jurassic-Cretaceous sediments have similar dips as the underlying Permo-Triassic strata, the deformation related to this episode was small or most probably, given the narrow character of the structures, concentrated locally on faults. An erosional unconformity is recognized at the top of the Nappamerri Formation but it is difficult to estimate how much material was eroded. Estimates of the amount of material removed based on vitrinite reflectance studies are between zero and 500 m (Kantsler et al., 1978, 1983) and vary across the basin. As mentioned in section 3.3.c, the recognition of Late Triassic sediments by Wiltshire (1982a,b) suggest that the erosional period was not continuous through the Late Triassic.

During the Jurassic and Cretaceous subsidence phase, some syn-depositional movement of underlying faults occurred (e.g. Smith, 1983) but was very much a minor factor (Moore and Pitt, 1984; Kuang, 1985), and the next major phase of structuring occurred in the Tertiary after the cessation of widespread sedimentation in the Eromanga Basin. In fact, many of the structures observed in the Eromanga Basin today are thought to have developed in the Tertiary or at least were the result of major reactivation of older
structures at this time. However the lack of sediment makes timing of the structuring
difficult.

Northerly, NE and NW trending anticlines or monoclines are recognized at the
surface from geomorphology and outcrop patterns (Senior et al., 1978; Krieg, 1982) and
LANDSAT imagery (Senior and Habermehl, 1980; Senior, 1982). These are seen in
reflection profiles to pass down into faults at depth (Moore and Pitt, 1984; Kuang, 1985;
Nelson, 1985; Finlayson et al., 1987a,b). These structural trends are illustrated for the
central and southwestern area of the Eromanga Basin in fig. 3.5.1. The Tertiary
deformation is interpreted as resulting from compressional stresses. Kuang (1985) suggests
an E-W direction and Finlayson et al. (1987b) suggest NE-SW, presumably because they
are looking at reactivated structures in different areas of the basin, i.e. Cooper Basin and
Adavale Basin areas respectively. Examination of the surface geology has led others to the
general conclusion that more or less continuous mild activity was operating throughout the
Tertiary (Senior et al., 1978; Senior and Habermehl, 1980; Senior, 1982; Moore and Pitt,
1984) to at least the Oligocene and possibly to the present day (e.g. Krieg, 1982).

Moore and Pitt (1984) state that up to 800 m of the Winton Formation may have
been removed locally from the crests of anticlines in the central Eromanga Basin. Exon and
Senior (1976) recognize erosion of the upper Winton Formation and also suggest that uplift
of the northern and eastern areas of the basin altered the dominant drainage direction from
an overall northerly direction to the Carpentaria Basin to a southwesterly direction. As a
result, erosion of the basin margins must have occurred and this material may have been
subsequently redeposited in the Eromanga Basin region, e.g. the Tertiary Eyre Formation.
Fig. 3.5.1  Structural trend map of the central Eromanga Basin region showing the axes of major anticlines present at the 'C' seismic horizon, near the top of the Cadna-Owie Formation (Early Cretaceous). Many of these features pass up into monoclines and can be seen in LANDSAT imagery, e.g. Senior 1982. (from Moore and Pitt, 1984).
3.7 Concluding remarks

It is apparent that both the Adavale Basin and southern Cooper Basin areas were tectonically active between the Late Devonian and the Late Carboniferous. A major compressional force acted from the east, resulting in more severe deformation in the Adavale Basin area. The period of granite emplacement in the southern Cooper Basin region implies an above average geothermal gradient, and the intrusive process may have been enhanced by applied compression, as suggested by Neugebauer and Reuther (1987). During this time uplift and erosion occurred, and up to 8 km of section may have been removed in the Adavale Basin region (Passmore and Sexton, 1984) and probably at least 2-3 km in the southern Cooper Basin region.

Early Permian deposition in the southern Cooper Basin was dominantly fault controlled, but by the Late Permian sedimentation was more regular and widespread. The Triassic appears to mark the beginning of the expansion of the depositional area and it is probable that the Cooper and Simpson Desert Basins were linked at this time. The last stages of Triassic sedimentation were disrupted by tectonic activity, associated with the Hunter-Bowen orogeny (Middle-Late Triassic), which was most active to the east in the Bowen-Sydney Basin region. The subsequent Jurassic deposition covered an initially similar geographical area in the central Eromanga Basin region but by the end of the Jurassic, the present day basin limits were reached. Shallow marine conditions were apparent for the first time in the Early Cretaceous, but were short lived and the final stage of deposition is represented by over 1 km of fluvial sediments. The detrital character of the Jurassic-Cretaceous sediments varies between quartzose and labile, reflecting the fluctuating influence of a volcanic arc to the east of the Surat and Maryborough Basins. The influx of volcanioclastic detritus is most pronounced in the Cretaceous part of the sequence.

The Late Cretaceous-Tertiary was mainly a period of non-deposition, and the sediments that are possibly of this age are thought to be derived in part from the reworking of the Eromanga Basin sequence. The outcrop geology of the Eromanga Basin has a
distinctly bullseye pattern (fig. 3.1.2). The outermost sediments are the Upper Jurassic Hooray Sandstone or its marginal facies equivalents (Algebuckina Sandstone in the SW and the Ronlow Beds in the NE). The regional dip of the strata is towards the basin centre (e.g. Senior et al., 1978, fig. 3.4.1) and this, together with the outcrop pattern described above, implies that the area of deposition continued to expand during the Jurassic and possibly contracted during the later stages, in the Early Cretaceous. I discussed the different basin geometries predicted with the elastic and viscoelastic plate models in section 2.2.a, but the Eromanga Basin does not fit either of these two simple models which predict a progressively widening or narrowing basin respectively with time. It is probable that erosion of the basin margins has occurred over the last 90 m.y. and this may be the explanation for the present day outcrop pattern, rather than appealing to stress relaxation.

The present day fault and fold trends observed in the Eromanga Basin reveal a dominantly compressional influence directed in an approximately E-W (±45°) direction. The main periods of activity were pre-Permian (Late Devonian-Carboniferous), Early Permian, Middle/Late Triassic and Tertiary with the first and last episodes being the most influential on the present day structures. The Late Devonian-Carboniferous structures are related to convergent tectonics at the eastern margin of the continent. Outside the immediate area of interest, east-west compression was active in Lachlan Fold Belt (eastern NSW and Victoria) and was similarly associated with a convergent eastern margin (Veevers, 1984, pg 341). During the Middle Carboniferous time, N-S shortening occurred in the central Australian region (Lambeck, 1983 a,b; Shaw, 1987) and there is evidence of N-S compressional structures in the Lachlan Fold Belt region (Powell 1984). An extensional environment existed in the the north western part of the continent in the Canning Basin which developed as an asymmetric rift during the Devonian (e.g. Drummond et al., 1987).

As the effects of the pre-Tertiary compressional episodes are recognized across the Eromanga and Surat Basins it would seem reasonable to relate the origin of the force to the boundary conditions at the eastern margin of the continent, given the convergent character of this margin during the Permo-Triassic. The presence of volcanogenic detritus in the
Mesozoic sediments as described earlier is indicative of an active arc to the east and possibly a convergent margin during this time. Numerous authors have described the influence of oblique convergence or strike-slip movement had on eastern Australia during the late Palaeozoic-Mesozoic (e.g. Evans and Roberts, 1979; Henderson, 1980; Cawood, 1982; Jones and Veevers, 1983; Murray et al., 1987). It would seem likely however that a compressional component would be the most significant influence on deformation in the continental interior as the length scale of deformation associated with purely strike-slip motion is small, e.g. of the order of 5-10% of the wavelength of the strike-slip boundary (Sonder et al., 1986). The strike-slip nature of the Cooper Basin faulting may then be the result of compression acting obliquely on pre-existing structures. The lack of syndepositional deformation during the Jurassic and Cretaceous, although significant volcanic activity occurred at the continental margin, suggests that possibly the convergence rate was reduced during this time. The continental margin of Western Australia tended to be a zone of rifting through the Permian to Cretaceous (Plumb, 1979; Falvey and Mutter, 1981; Forman and Wales, 1981; Veevers, 1984). The southeastern margin of Australia was also undergoing active extension during the Cretaceous with the development of the Bass, Gippsland and Otway Basins (e.g. Etheridge et al., 1987).

The cause of the Tertiary deformation is not clear. Northern Australia is thought by Pigram and Panggabean (1984) to have been a passive margin from the Triassic to at least the late Jurassic although Veevers (1984, pg 113) suggests that the eastern end of the Papuan Peninsula was in a convergent environment related to the arc off eastern Australia. The major deformation of Papua New Guinea occurred in the Plio-Pleistocene (~3 Ma although the opening of the Coral Sea about 60 Ma (Falvey and Mutter, 1981) marks the movement of the Papuan Peninsular away from NE Australia and collision with a pre-existing volcanic arc occurred about 55 Ma (Veevers, 1984, pg 114-115). Denham (1983) relates the present day compressional stress state in Australia to the convergent northern boundary of Australia and Moore and Pitt (1984) tentatively suggest that the same collision may be responsible for the apparently pervasive Tertiary deformation in the Eromanga
Basin. Another possibly relevant tectonic event may be the change in direction of subduction in the New Hebrides trench. The present day easterly directed subduction is thought to have been initiated in the Late Miocene after a reversal from a more westerly directed subduction zone (Falvey, 1975, 1978; Carney and MacFarlane, 1982). On the southeastern margin of Australia, Hegarty and Houseman (1988) have suggested that two different structural trends in the Gippsland Basin may be the result of local and regional stress fields. Local extension, inferred to be associated with uplift, as a result of a mantle plume, during the Cretaceous to Early Eocene resulted in NW-SE trending normal faulting. Subsequently after the Eocene, when the vertical stress due to the plume was reduced, structures developed indicative of NW-SE compression as the local stress field was modified. Cloetingh and Wortel (1985) calculated the present day regional stress field in the Indo-Australian plate by using finite element methods, assuming a uniform elastic plate and plane stress (no vertical stresses). They incorporated the driving forces from the subducted slab (slab pull) and the spreading centres (ridge push). Resistive forces arise due to the buoyancy of the subducting slab and also shear stresses between the slab and the overriding plate and the slab and the underlying mantle. Given that a uniform elastic plate model is undoubtedly an over-simplification, Cloetingh and Wortel's (1985) results suggest that, at the present time, northeastern Australia is in a state of approximately E-W tension while central Australia is in a state of N-S compression. The variations in the stress state are attributed to lateral variations in the boundary conditions. Significant boundary features controlling compression in central Australia are the Himalayan collisional belt and the Banda Sea subduction zone. The tensional stresses in eastern Australia are attributable to the easterly directed subduction in the New Hebrides and Tonga-Kermadec trenches. The evolution of these large scale tectonic features, or those which existed prior to their formation, during the Tertiary may then be responsible for the development of structural trends in the Eromanga Basin as older structures were reactivated under changing regional stress regimes.
4. ANALYSIS OF SUBSIDENCE IN
SEDIMENTARY BASINS

4.1 Introduction

A common approach to sedimentary basin analysis is the reconstruction of subsidence, or strictly speaking, sediment accumulation as a function of time. The technique, known as geohistory analysis, ideally requires a knowledge, or inference, of the stratigraphic and absolute ages of the sediment, palaeowater depths and sea level variations (van Hinte, 1978; Falvey and Deighton, 1982; Sevier, 1983; Guidish et al., 1984). Thermal histories of the sediments may be modeled if heat flow is prescribed as a function of time and the relevant thermal parameters can be estimated or assumed. Also, because sediments can be expected to compact when progressively buried, i.e. porosity is reduced, it is important to incorporate this effect, especially if original sediment thickness or rates of sedimentation are desired.

Sleep (1971) suggested that subsidence of the United States Atlantic continental margin, as inferred from observed sediment thicknesses, has an exponential dependence on time such that the rate of subsidence decreases towards the present day. He proposed that this trend was the result of thermal contraction (section 2.3.a). Watts and Ryan (1976), and subsequently Steckler and Watts (1978), developed a method, similar to geohistory analysis, to quantify tectonic subsidence in sedimentary basins. Here tectonic subsidence is defined as that part of the observed subsidence being attributed to the (unknown) physical processes forming the sedimentary basin, for example the mechanisms discussed in chapter 2. Therefore, corrections are required to remove the loading effect of the sediment and water at the time of deposition. The method has become known as backstripping and has since been widely used to provide supporting evidence for particular basin forming mechanisms (e.g. Royden and Keen, 1980; Sclater and Christie, 1980;
4.2

Sclater et al., 1980; Feinstein, 1981; Brunet and Le Pichon, 1982; Sawyer et al., 1982; Yorath and Hyndman, 1983; Bond and Kominz, 1984; Barton and Wood, 1984; Sørenson, 1986; Chadwick, 1986; Heidlauf et al., 1986; Hegarty et al., 1987).

4.2. Backstripping: assumptions, method and errors

4.2.a Introduction

Despite their widespread use, the backstripping methods are based on a number of simplifying assumptions, the consequences of which appear to have not been fully explored. It is desirable, however, to know how sensitive the procedures are to these assumptions. In the following section I outline the method of backstripping, discuss the assumptions made therein and attempt to quantify some of the errors which may arise in the decompacted sediment thicknesses and basement subsidence as a result of an unrepresentative parameterization of the porosity/depth relationship. I define my notation here as follows and refer to fig. 4.2.1:

\[ H_s \] - sediment thickness (m)

\[ H_b \] - tectonic, or basement, subsidence (m), remaining after backstripping

\[ H_w \] - water depth during deposition, or palaeobathymetry (m)

\[ H_{sl} \] - height of sea level relative to the present day (m)

\[ z \] - depth (m)

\[ \rho \] - density (kg m\(^{-3}\)) : subscripts, s-bulk sediment, g-grain, w-water, i-infill, m-mantle

\[ \phi \] - fractional porosity (0-1)

\[ V \] - volume/unit cross-sectional area (m) : subscripts, t-total, s-solid, w-water
Fig. 4.2.1 Symbols used in the decompaction and backstripping procedures. $H_s$ is the present day observed thickness, $z_2 - z_1$, of a sedimentary unit whose upper surface is at a depth $z_1$. $H'_s$ is the decompacted sediment thickness, $z'_2 - z'_1$, with its upper surface at a depth $z'_1$, at time $t_1$. $V_s$ is the solid volume, or grain component, which does not change with decompaction. $V_w$ is the fluid component in the pore space which is increased on decompaction to $V'_w$. $H_w$ is the depth of water during deposition, or palaeobathymetry, and $H_{sl}$ the height of sea level at the time of deposition above the present day sea level. The depositional base level, or upper surface, during sedimentation is assumed to be the contemporary sea level. $H_b$ is the backstripped, or unloaded, basement depth in the absence of the sediment load at time $t_1$, relative to present day sea level. The unloaded basement depth is attributed to a tectonic or driving mechanism, and is thus referred to as the tectonic subsidence.
4.2.b Porosity and compaction in sediments

Most simply, a sedimentary rock can be considered to have two components. The first is the solid, grain or matrix (framework as opposed to cement) component, and the second is the void space which exists between the grains, known as the pore space or porosity. The latter component is measured as a fraction, or percentage, of the total rock volume, and can range from about 80% in an unconsolidated clay to effectively zero in fully lithified or cemented sediments. The permeability of a sediment is a measure of the ability of fluid to flow through the porous medium, or conversely, a measure of the resistance of the sediment to fluid flow. Permeability therefore depends on the size, shape and degree of interconnection of the pore spaces, and also on the tortuosity, a measure of the actual pore path length a fluid would travel through, relative to a length parallel to the overall flow direction. Physically, permeability is experimentally derived as a constant of proportionality relating the fluid flow velocity through the porous medium to the hydraulic, or applied pressure, gradient and the viscosity of the fluid. Typical permeabilities range from $10^{-7}$ m$^2$ in gravels to less than $10^{-20}$ m$^2$ in igneous and metamorphic rocks, which are then considered to be impervious. In practice, the commonly used unit for permeability is the millidarcy (md) where 1 md = 9.8697x$10^{-16}$ m$^2$.

In this section I discuss some processes involved in the reduction of porosity in sedimentary rocks. This is undoubtedly a complicated process (see, for example, Rieke and Chilingarian, 1974; Chilingarian and Wolf, 1976; Taylor, 1978 and papers discussed therein). However for regional basin studies, often with a limited data set, it is necessary to make simplifying assumptions regarding the processes involved, the implication being that the model will represent the dominant factor(s), or at least average conditions.

Mechanical compaction can be expected to exert a major control on porosity reduction in clastic sediments but factors such as cementation, leaching and high pore pressures may also be significant (Magara, 1980; Scherer, 1987). I briefly discuss these latter three factors before describing the approach taken to model mechanical compaction. Cementation will reduce porosity but not necessarily change an elemental thickness or volume - this
depends on whether or not the material ultimately precipitated as cement is derived from an external (or distant) source. Siliceous cements seen in sandstones often have been derived from within the sedimentary basin by mechanisms such as grain boundary dissolution and liberation of silica during mineral reactions, such as degradation of feldspar to clay minerals (Taylor, 1978; Leder and Park, 1986). In this situation, material is effectively removed from one place and redeposited in another (i.e. in the porespace). Although solid volumes before and after this process will not necessarily be equal, it is effectively only a local redistribution of material and except perhaps at the near surface, mechanical compaction should exert a more important control on porosity reduction (e.g. Scherer, 1987). The formation of authigenic, or in situ, clay minerals will have a more profound effect on permeability as a result of the constriction of pore throats. Leaching will increase porosity as a result of removal of unstable minerals and cements, resulting in what is termed secondary porosity and this is most commonly observed in carbonates. High pore pressures are often attributed to mechanical overloading (Shi and Wang, 1986) where rapid deposition of low permeability sediments (shales, mudstones) restricts the upward movement of pore fluid. This increases the pore pressure, provided that no significant horizontal flow occurs. The reduction of porosity is inhibited as the solid framework or matrix is partially supported by the excess pore pressure acting against the sediment overburden. In the absence of information regarding excess pore pressures, I do not pursue this aspect of compaction, but Sclater and Christie (1980) discuss this effect which is observed in some wells in the North Sea and show how it may be incorporated into decompaction models, if necessary.

Mechanical compaction is the result of a progressively increasing overburden load when a sediment becomes buried. The first order effects are the partial collapse of the grain framework, plastic deformation of soft fragments, fracturing and pressure solution (or grain boundary dissolution) (Rieke and Chilingarian, 1974; Chilingarian and Wolf, 1976; Nagtegaal, 1978; Scherer 1987). As might be expected, sedimentological factors exert major controls on mechanical compaction. It is observed that sorting, grain size and composition/mineralogy are important parameters (Beard and Weyl, 1973; Blanche and
Whitaker, 1978; Nagtegaal, 1978; Magara, 1980; Scherer, 1987). For example a well sorted, coarse grained sandstone would be expected to have a higher porosity than a poorly sorted, generally finer grained one. Sediments with a high proportion of mechanically unstable lithic fragments will tend to collapse more readily than quartz rich sandstones. It is important, then, to scrutinize lithological types when considering porosity as a function of depth.

Attempts to quantify porosity reduction with depth generally rely on the fitting of empirical curves to observed data (e.g. Magara, 1980; Baldwin and Butler, 1985; Scherer, 1987), without detailed consideration of the physical processes involved. Some studies have concentrated on specific aspects such as simple consolidation due to overburden load (Biot, 1941, 1955; Gibson, 1958; Parasnis, 1960; Korvin, 1984) or development of overgrowths, i.e. reprecipitation of minerals dissolved in pore fluids around pre-existing grains (e.g. Leder and Park, 1986). Korvin (1984) has shown that if simple compaction occurs (by vertical migration of pore fluid) then the porosity should be an exponential function of depth of the form,

\[ \phi(z) = \phi_0 \exp(-cz) \]  \hspace{1cm} (4.2.1)

and this is in fact observed in many studies of porosity in sediments (see Korvin, 1984 for references). Magara (1980) suggested that, although an exponential relationship is appropriate for shales, the influence of diagenetic processes such as those briefly described earlier, e.g. cementation, can result in a more linear porosity/depth relationship and Selley (1978) made a similar observation. Bond and Kominz (1984) comment on this and conclude the data used by Magara (1980) is unable to resolve such a difference as most of the observations are from deeper than 500-1000 m and below the depth where most rapid rates of compaction are expected. A similar comment may be made regarding the data used by Selley (1978) as his porosity data is obtained with samples from depths mainly greater than 500 m. Also in this context, it is worth noting that sandstones generally show a broad scatter of porosities, presumably as a result of the diversity of processes involved in the
porosity reduction. In such cases it is probable that an empirical curve of an exponential form could describe the data as well as a linear form. This would of course depend on the actual data set in question. A pseudo-linear form could be expressed, of course, as an exponential form depending on the value of c, the constant in the exponential, and the depth to which the relationship is considered relevant.

For the purposes of numerical decompaction it is desirable to define a simple porosity/depth function. The exponential form, eqn. 4.2.1, is the most commonly used (e.g. Steckler and Watts, 1978; Royden and Keen, 1980; Sclater and Christie, 1980; Sawyer et al., 1982; Beaumont et al., 1982; Sørensen 1986) but alternative relationships have been proposed. For example, Falvey and Deighton (1982) use a function of the form

$$\frac{1}{\phi(z)} = \frac{1}{\phi_0} + cz \quad (4.2.2)$$

which has similar properties to the exponential function, i.e. rapid early compaction approaching an asymptotic value with increasing depth. Issler and Beaumont (1987) favour a linear relationship of the form

$$\phi(z) = \phi_0 - cz \quad (4.2.3)$$

for sandstones of the Labrador margin. Bond and Kominz (1984) use a two-stage exponential porosity depth function for sediments from the Lower Palaeozoic miogeoclone of western North America. One stage is relevant to only the top 500 m of the evolving sedimentary column and has a more rapid rate of compaction than that used for the deeper section. Lucazeau and Le Douaran (1985) use a similar two-stage relationship for shales in the Viking Graben of the North Sea. Bond and Kominz (1984) also define maximum and minimum porosity functions for individual lithologies, simulating mechanical compaction (loss of porespace by framework collapse) and porespace reduction solely by externally
sourced cement infill respectively. These assumptions are required as the present day porosities in this area are effectively zero and petrographic studies reveal a significant component of cement, especially in the quartz rich sandstones and calcarenites. The textures are interpreted as being indicative of early cementation at shallow depths before major compaction occurred. In the absence of porosity observations in this area, Bond and Kominz (1984) use those of Schmoker and Halley (1982) from southern Florida which, although made on extensively cemented carbonates, show an exponential decreasing trend with depth. Fig. 4.2.2 shows some empirically derived porosity/depth curves for sandstone and shale.

In summary, it is assumed that generally porosity will decrease with depth primarily as a result of mechanical compaction. The commonly observed rapid rate of decrease in porosity near the surface where the overburden is smallest would appear to be at odds with this statement. This may be attributed to the likelihood that porosity reduction may occur more by grain settling. Also, cementation is often considered to occur early in the depositional history of a sediment presumably as ground water moves through the pore-space dissolving or precipitating minerals depending on the prevalent thermodynamic conditions.

4.2.c Numerical decompaction

The process of decompaction as described below involves attempting to restore the present day observed thickness of a given sedimentary unit to the value appropriate to it being at a shallower depth earlier in the subsidence history of the basin (fig. 4.2.1). The thickness change is implicitly attributed to some form of mechanical compaction. Cementation, where porosity reduction may occur with no thickness change, is treated by Bond and Kominz (1984) and will not be further discussed here.

The basic premise for simple compaction is that a given elemental volume, $V_t$, made up of a solid or matrix component, $V_s$, and a fluid component in the
Fig. 4.2.2  Empirical porosity/depth curves taken from published studies.


Upper panel - Shale: (a) Middleton (1978), (b) Sclater and Christie (1980), (c) Issler and Beaumont (1987), (d) Beaumont et al. (1982), (e) and (f) Royden and Keen (1980), (g) Lucazeau and Le Douaran (1985), (h) Falvey and Deighton (1982), (i) and (j), Baldwin and Butler (1985).
porespace, \( V_w \), changes in volume only as a result of changes in \( V_w \), i.e. expulsion of pore fluid (Perrier and Quiblier 1974). This of course implies that the matrix material is incompressible, and at all times its mass is conserved. The former assumption can be justified, as the low bulk compressibility of sedimentary rocks means that for an overburden of 5 km, and neglecting pore pressures, the solid volume at this depth would change by less than 0.2% relative to the surface value. Including pore pressures would reduce this estimate. The latter assumption regarding conservation of mass can be assumed to be valid if dissolution and precipitation occurs only on a local scale. Also pertinent is the fact that in practice not enough information is available to warrant a more complicated approach to the problem. Continuing with these assumptions we can express the total elemental volume, \( V_t \), as

\[
V_t = V_s + V_w \tag{4.2.4a}
\]

and

\[
V_s = V_t - V_w = \text{constant} \tag{4.2.4b}
\]

As volume changes due to compaction are assumed to be the result of the vertical expulsion of pore fluid, then

\[
V_s = V_t - \int_{z_1}^{z_2} \phi(z,t) \, dz \tag{4.2.5a}
\]

or

\[
V_w = \int_{z_1}^{z_2} \phi(z,t) \, dz \tag{4.2.5b}
\]

where \( \phi \) is the fractional porosity. In these formulations, compaction is assumed to take place instantaneously. The decompaction problem can be posed as follows - If we know the present (time = \( t_p \)) depth to the upper and lower boundaries of a particular unit, \( z_1 \) and \( z_2 \), and we wish to bring the unit up to a depth where its upper surface is at \( z_1' \) at a given time \( t_1 \), what is \( z_2' \), the depth to the base of the unit, or how thick was the sedimentary unit, at this time? From now on I shall refer to the elemental volume as a thickness, which is valid if a unit cross sectional area is considered.
Using 4.2.5a and 4.2.4b (i.e. \( V_s = \) constant) then

\[
z_2 - z_1 - \int_{z_1}^{z_2} \phi(z,t) \, dz = z_2' - z_1' - \int_{z_1'}^{z_2'} \phi(z,t) \, dz
\]  

(4.2.6)

Taking an exponential porosity/depth relationship as given by 4.2.1

\[
z_2 - z_1 + \frac{\phi_0}{c} \left[ \exp(-cz_2) - \exp(-cz_1) \right] = z_2' - z_1' + \frac{\phi_0}{c} \left[ \exp(-cz_2') - \exp(-cz_1') \right]
\]  

(4.2.7a)

which, as \( z_2, \ z_1 \) and \( z_1' \) are known, can be written in the form

\[
z_2' + \frac{\phi_0}{c} \exp(-cz_2') + \text{constant} = 0
\]  

(4.2.7b)

Similarly for the porosity/depth relationship as given by 4.2.2

\[
z_2 - z_1 - \frac{1}{c} \left[ \ln \left( \frac{1}{\phi_0} + cz_2 \right) - \ln \left( \frac{1}{\phi_0} + cz_1 \right) \right] = z_2' - z_1' - \frac{1}{c} \left[ \ln \left( \frac{1}{\phi_0} + cz_2' \right) - \ln \left( \frac{1}{\phi_0} + cz_1' \right) \right]
\]  

(4.2.8a)

or

\[
z_2' - \frac{1}{c} \ln \left( \frac{1}{\phi_0} + cz_2' \right) + \text{constant} = 0
\]  

(4.2.8b)

and for the linear porosity/depth relationship as given by 4.2.3

\[
(z_2-z_1)(1+\phi_0) - \frac{c}{2} (z_2^2 - z_1^2) = (z_2'-z_1')(1+\phi_0) - \frac{c}{2} (z_2'^2 - z_1'^2)
\]  

(4.2.9a)

or

\[
z_2'(1+\phi_0) - \frac{c}{2} z_2'^2 + \text{constant} = 0
\]  

(4.2.9b)

The solutions to 4.2.7b and 4.2.8b for \( z_2' \) may be found numerically and as 4.2.9b is a quadratic an analytical solution is possible, taking the positive root of course. Using
downhole stratigraphic data, the procedure is to bring a unit whose upper surface, or top, is of a given age \(t_1\) to the surface \(z_1=0\), find \(z_2\) and use this value as \(z_1\) for the underlying unit. Thus we obtain a decompacted section at the time \(t_1\).

4.2.d Backstripping and tectonic subsidence

Once sediments have been decompacted, we have an estimate of the total thickness of sediment at a given time for a given locality - this is one of the aims of geohistory analysis. Backstripping (Watts and Ryan, 1976; Steckler and Watts, 1978; Sclater and Christie, 1980) is a quantitative attempt to remove the displacement due to the sediment (and water) loads and reveal the tectonic, or basement, subsidence. The load due to a sediment thickness, \(H_s\), is simply \(H_s \bar{\rho}_s g\), where \(g\) is acceleration due to gravity and \(\bar{\rho}_s\) is the mean sediment density given by

\[
\bar{\rho}_s = \frac{1}{(z_2-z_1)} \int_{z_1}^{z_2} \rho_s(z) \, dz \tag{4.2.10a}
\]

with \(\rho_s(z) = \rho_g (1-\phi(z)) + \rho_w \phi(z)\). Therefore, if \(\rho_g\) is constant over the interval of integration, then

\[
\bar{\rho}_s = \rho_g - \frac{(\rho_g-\rho_w)}{(z_2-z_1)} \int_{z_1}^{z_2} \phi(z) \, dz \tag{4.2.10b}
\]

To remove the sediment load it is necessary to make some assumption about the response of the earth, or at least its outer part, to surface loading or what mode of isostasy is appropriate. For the purpose of the following discussion I assume Airy isostasy and will not consider flexural effects, except to say that if the lithosphere does have the ability to support shear stresses then backstripping under the assumption of Airy isostasy will over correct for the sediment load and underestimate the tectonic subsidence.

Assume at time \(t_1\) there is a thickness of sediment, \(H_s(t_1)\), of density \(\bar{\rho}_s(t_1)\) then the tectonic subsidence is given by
\[ H_b(t_1) = \frac{(\rho_m - \rho_s(t_1))}{(\rho_m - \rho_i)} H_s(t_1) \]  \hspace{1cm} (4.2.11a)

\( H_b(t_1) \) is interpreted as the depth of the sediment filled depression with sediment replaced by a fluid (water or air) of density \( \rho_i \) with an isostatic correction applied and is then interpreted as the amount of subsidence due to a tectonic driving mechanism (fig. 4.2.1). Eqn. 4.2.11a assumes that the sediment filled the depression or basin to depositional base level (usually defined as the contemporary sea level), but in marine basins this is not always so. If water depths during deposition can be inferred, for example from palaeontological studies (van Hinte, 1978) then a correction may be applied so that now

\[ H_b(t_1) = \frac{(\rho_m - \rho_s(t_1))}{(\rho_m - \rho_i)} H_s(t_1) + \frac{(\rho_m - \rho_w)}{(\rho_m - \rho_i)} H_w(t_1) \]  \hspace{1cm} (4.2.11b)

and if sea level variations are considered important then the difference in the reference level should be corrected for, giving

\[ H_b(t_1) = \frac{(\rho_m - \rho_s(t_1))}{(\rho_m - \rho_i)} H_s(t_1) + \frac{(\rho_m - \rho_w)}{(\rho_m - \rho_i)} H_w(t_1) - \frac{\rho_m}{(\rho_m - \rho_i)} H_{sl}(t_1) \]  \hspace{1cm} (4.2.11c)

where \( H_{sl}(t_1) \) is positive for a sea level rise relative to the present day.

Of course errors inherent in these three terms will be manifested in the estimate of the tectonic subsidence. Errors arising from uncertainties in \( H_w \) and \( H_{sl} \) will be linear or directly proportional to the errors in these terms and are given as

\[ \frac{(\rho_m - \rho_w)}{(\rho_m - \rho_i)} \Delta H_w \]  \hspace{1cm} (4.2.12a.)

and
\[
\frac{\rho_m}{(\rho_m - \rho_i)} \Delta H_{sl} \tag{4.2.12b}
\]

Taking \( \rho_i = \rho_{air} = 0 \) does not reduce the relative magnitude of the errors as the tectonic subsidence is correspondingly reduced. In this thesis I use \( \rho_i = \rho_w = 1030 \text{ kg m}^{-3} \) and \( \rho_m = 3300 \text{ kg m}^{-3} \) and then the error contributions to the calculated tectonic subsidence are simply \( \Delta H_w \) and \( 1.45 \Delta H_{sl} \).

4.2.e Errors associated with porosity in decompaction and backstripping procedures

Observed porosities as a function of depth can vary significantly even for the same lithology, as shown in fig. 4.2.2. This then raises the question of how important it is to constrain the porosity/depth function such that it is an adequate representation of the actual relationship for a particular locality. The implication at this stage is that, by fitting a curve to the observed data for a given lithology, the same mechanism(s) have been operative throughout the subsidence and compaction history in the basin as that lithology was progressively deposited and buried.

However, for the purpose of decompaction and backstripping calculations we assume porosity is an exponential function of depth and porosity reductions are caused solely by compaction. Even under these assumptions it is desirable to know how sensitive the results of the calculations are to the values chosen for \( \phi_0 \) and \( c \) and also the solid grain density \( \rho_g \), i.e. how uncertainties in these parameters are mapped into the estimates of rates of sedimentation and basement subsidence. The first term in 4.2.11c is a function of both the decompacted sediment thickness and the mean sediment density, which are themselves functions of the porosity and eqn. 4.2.11a can be rewritten in the following form:
\[ H_b(t_1) = \left( z_2' - z_1' \right) \frac{(\rho_m-\rho_w)}{(\rho_m-\rho_g)} - \frac{(\rho_g-\rho_w)}{(\rho_m-\rho_g)} \frac{\Phi_0}{c} \left[ \exp(-cz_2') - \exp(-cz_1') \right] \]

(4.2.13)

where the expression for \( H_b(t_1) \) has been replaced with \( z_2' - z_1' \), \( z_2' \) and \( z_1' \) being the depth of the top and base respectively of the particular sedimentary unit at time \( t_1 \), \( \bar{\rho}_s(t_1) \) has been evaluated with eqn. 4.10.b with an exponential porosity depth relation (eqn 4.2.1) and \( \rho_i = \rho_w \). In order to obtain error estimates for \( H_b(t_1) \), which may arise due to errors in \( \Phi_0 \), \( c \), and \( \rho_g \), it is possible to derive the partial derivatives of the solid grain thickness (\( V_s \)) and basement subsidence (\( H_b \)) with respect to these parameters. However, evaluation of the resulting expressions generally involves a linearizing approximation by truncation of a Taylor series. It will be seen that this is not valid for the exponential constant \( c \). Therefore the approach we adopt here is to simply use different values for the porosity constants, \( \Phi_0 \), and \( c \), and solid grain density, \( \rho_g \) for calculation and compare the results. Based on the spread of the empirical relationships shown in fig. 4.2.2 errors, \( \pm \Delta \Phi_0 \), \( \pm \Delta c \) and \( \pm \Delta \rho_g \), were chosen (table 4.2.1) for two models, one to simulate a sandstone and the other, a shale.

<table>
<thead>
<tr>
<th>( \Phi_0 )</th>
<th>( \Delta \Phi_0 )</th>
<th>( \frac{\Delta \Phi_0}{\Phi_0} )</th>
<th>( c )</th>
<th>( \Delta c )</th>
<th>( \frac{\Delta c}{c} )</th>
<th>( \rho_g )</th>
<th>( \Delta \rho_g )</th>
<th>( \frac{\Delta \rho_g}{\rho_g} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>%</td>
<td>%</td>
<td></td>
<td>km(^{-1})</td>
<td>km(^{-1})</td>
<td></td>
<td>kg m(^{-3})</td>
<td>kg m(^{-3})</td>
<td></td>
</tr>
<tr>
<td>Sandstone</td>
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<td>6.75</td>
<td>0.12</td>
<td>0.45</td>
<td>0.19</td>
<td>0.42</td>
<td>2650</td>
<td>133</td>
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<tr>
<td>Shale</td>
<td>59.0</td>
<td>11.0</td>
<td>0.19</td>
<td>0.525</td>
<td>0.125</td>
<td>0.24</td>
<td>2710</td>
<td>200</td>
</tr>
</tbody>
</table>

Table 4.2.1 Parameters used for estimating errors during compaction and backstripping

Porosity curves were calculated using five permutations of the porosity constants ; \((c,\Phi_0)\), \((c \pm \Delta c,\Phi_0 + \Delta \Phi_0)\) and \((c \pm \Delta c,\Phi_0 - \Delta \Phi_0)\) for both the shale and sand models and these are illustrated in fig. 4.2.3. The curves b and e in fig. 4.2.3 are approximately bounding curves on those, empirically derived from observed data, shown in fig. 4.2.2. I used compacted
Fig. 4.2.3 Model exponential porosity/depth curves for lower panel, sandstone, and upper panel, shale models. The porosity functions are defined as $\phi(z) = (a) \phi_0 \exp(-cz), (b) (\phi_0 - \Delta \phi_0) \exp(-(c+\Delta c)z), (c) (\phi_0 + \Delta \phi_0) \exp(-(c+\Delta c)z), (d) (\phi_0 - \Delta \phi_0) \exp(-(c-\Delta c)z), (e) (\phi_0 + \Delta \phi_0) \exp(-(c-\Delta c)z)$. The values of $\Delta c$ and $\Delta \phi_0$ are given in table 4.2.1. These are the curves used to estimate the errors in the decompaction and backstripping procedures.
4.14

thicknesses \((H_s = z_2-z_1)\) of between 100 and 1000 m with the upper surface, \(z_1\), varying between 250 and 4750 m at intervals of 250 m. This is then assumed to be the present day observed sedimentary unit. The solid thickness, \(V_s\), was evaluated with \(H_s = 100, 250, 500\) and 1000 m, for each of the 5 porosity curves for both the sandstone and shale models (table 4.2.1, fig. 4.2.3) using eqn. 4.2.6, to give,

\[
V_s = z_2 - z_1 + \frac{\phi_0}{c} \left( \exp(-cz_2) - \exp(-cz_1) \right)
\]

(4.2.14)

The unit was moved up to depths of \(z_1'\) given by 0, 250, 500, 1000, 1500, 2000, and 3000 m. This involved decompacting the unit along each porosity curve (fig. 4.2.3) by using eqn. 4.2.7b to estimate \(z_2'\) and the decompacted sediment thickness, \(H_s' = z_2' - z_1\). Finally, the loading effect of this decompacted sedimentary unit with removed eqn. 4.2.11a to estimate the basement subsidence, or the depth of a pre-existing depression which the sediment is assumed to have infilled to the same datum.

I discuss the results of these calculations in the text but they are also presented as plots in the following form:

(i) In figs. 4.2.4 a,b I show the value of solid thickness, \(V_s\) estimated with \(\Delta c = \Delta \phi_0 = 0\), and on the same diagram, the percentage error, \(100 \Delta V_s/V_s\), for the model parameters \(\pm \Delta \phi_0\) and \(\pm \Delta c\). The solid grain density does not affect \(V_s\). We will concentrate on the solid thickness rather than the total decompacted sediment thickness as later we use \(V_s\) to calculate rates of sedimentation.

(ii) In figs. 4.2.5 a-d, I show the calculated basement subsidence, \(H_b\) and the percentage errors, \(100 \Delta H_b/H_b\), for \(\pm \Delta \phi_0\), \(\pm \Delta c\) and \(\pm \Delta \rho_g\) with different values of \(z_1'\), the depth to which the present day unit \((z_2-z_1)\) is brought up to for decompaction (see fig. 4.2.1). The total mass of the unit will vary with the amount of decompaction as, although the solid volume is constant, the water volume, \(V_w\), increases due to the addition of water filled porosity as decompaction proceeds. For the basement subsidence, I only present the
results of calculations with $H_s = 100$ and 1000 m. These two can be considered bounding values on the errors as the results obtained with $H_s = 250$ and 500 m fall between these values. I discuss the effect of varying each parameter in turn.

(a) **Surface porosity term, $\phi_0$**

The solid thickness, $V_s$, is given by eqn. 4.2.14 and the error in $V_s$, $\Delta V_s(\phi_0) = V_s(\phi_0 \pm \Delta \phi_0) - V_s(\phi_0)$, due to an error $\pm \Delta \phi_0$ is then

$$\Delta V_s(\phi_0) = \frac{\pm \Delta \phi_0}{c} \left( \exp(-cz_2) - \exp(-cz_1) \right) \quad (4.2.15)$$

It can be seen from eqn. 4.2.15 that errors in $V_s$ are linear in $\phi_0$. As the term in the brackets is always negative, the sign of $\Delta V_s$ will be opposite to that of $\Delta \phi_0$. Figs. 4.2.4a,b (curve D) show that the errors decrease in magnitude for greater values of $z_1$, the equivalent of the present day observed depth in a well. This is because as the unit becomes more deeply buried, the porosity (and $V_w$) tends to zero, and the observed sediment thickness, $H_s$ then approaches the solid thickness, and the estimate of $V_s$ is less sensitive to the porosity function. The maximum error, $\Delta V_s$, for the sandstone model ($\phi_0 = 55.25 \pm 6.75\%$; $\Delta \phi_0/\phi_0 = 12.2\%$) is 11% for $H_s = 100$ and this decreases slightly to 8% for $H_s = 1000$ m. For the shale model ($\phi_0 = 60.0 \pm 12.2\%$; $\Delta \phi_0/\phi_0 = 18.7 \%$), the maximum error is 20% for $H_s = 100$ m (13%, $H_s = 1000$ m). However if the unit is at depths greater than 1000 m then the maximum errors are reduced to about 5 and 9% respectively.

The basement subsidence corrected for sediment loading is given by 4.2.11a, where the mean sediment density $\bar{\rho}_s$ is

$$\bar{\rho}_s = \frac{(\rho_g V_s + \rho_w V_w)}{V_t} \quad (4.2.16)$$

As there is direct tradeoff between overestimation of $V_s$ and underestimation of $V_w$ for a given $z_2$ and $z_1$, then
\[ \Delta \rho_s(\phi_0) = \frac{(\rho_g - \rho_w) \Delta V_s(\phi_0)}{V_t} \]  
(4.2.17a)

and therefore, from eqn. 4.2.11a,

\[ \Delta H_b(\phi_0) = - \Delta V_s(\phi_0) \frac{(\rho_g - \rho_w)}{(\rho_m - \rho_w)} \]  
(4.2.17b)

where \( V_t = H_s \). As \( \Delta V_s(\phi_0) \) is opposite in sign to \( \Delta \phi_0 \) then \( \Delta H_b(\phi_0) \) will be the same sign as \( \Delta \phi_0 \) and the actual magnitude of \( \Delta H_b(\phi_0) \) is linearly related to \( \Delta \phi_0 \). The errors follow the same trend as those for \( \Delta V_s(\phi_0) \), decreasing as \( z_1 \) increases (figs. 4.2.5a-d, panel (iii)). The maximum errors are less than 5% for the sandstone model and less than 9% for the shale model.

(b) Exponential constant term, c

The error in the solid thickness, \( V_s \), \( \Delta V_s(c) = V_s(c \pm \Delta c) - V_s(c) \), due to an error \( \pm \Delta c \) is, from eqn. 4.2.14,

\[ \Delta V_s(c) = \frac{\phi_0}{(c \pm \Delta c)} \left[ \exp(-(c \pm \Delta c)z_2) - \exp(-(c \pm \Delta c)z_1) \right] - \frac{\phi_0}{c} \left[ \exp(-cz_2) - \exp(-cz_1) \right] \]  
(4.2.18)

and this expression is non-linear as \( \Delta V_s(c) \) depends inversely and exponentially on \( c \pm \Delta c \).

Generally the error would be the same sign as \( \Delta c \), i.e. positive for \( +\Delta c \), but the magnitudes are different for \( +\Delta c \) and \( -\Delta c \), (curves marked B and C in fig. 4.2.4a,b), and tend to be greater for the latter. For a constant surface porosity \( \phi_0 \), the magnitude of \( \Delta V_s(c) \) will decrease with depth \( z_1 \) as the porosity approaches some lower limiting value (zero) for all values of \( c \). The magnitude of \( \Delta V_s(c) \) will also decrease towards the surface where the porosity approaches the same upper limiting value \( \phi_0 \) irrespective of the value of \( c \). These trends are apparent in figs. 4.2.4a,b, curves B and C. The maximum errors, \( \Delta V_s(c) \), for the sandstone model \( (c=0.45 \pm 0.19km^{-1}; \Delta c/c = 42\%) \) are \(<10\% (\pm c)\) and \(>14\% (-\Delta c)\) and for shale \( (c=0.525 \pm 0.126km^{-1}; \Delta c/c=24\%)\), \(<7\% (\pm c)\) and \(>8\% (-\Delta c)\).
The basement subsidence is also nonlinear in $c$ and so are the errors, $\Delta H_b(c)$. The individual errors in the decompacted thickness (eqn. 4.2.7) and the mean sediment density (eqn. 4.2.16) contribute to the total. The sign of the errors vary and depend on the depth of the unit and the depth to which it is moved for compaction. For $+\Delta c$ the basement subsidence tends to be underestimated as $z'_1$ increases but may overestimated as $z'_1$ approaches the surface (i.e. zero depth) and the opposite trend is apparent for $-\Delta c$ (figs. 4.2.5a-d, panel (ii)). For $+\Delta c$, the sandstone model gives errors are between $+6$ and $-14\%$ and the shale model between $+4$ and $-10\%$. The opposite trend is seen for $-\Delta c$ with errors of $+24$ to $-10\%$ for the sandstone model and $+15$ to $-6\%$ for the shale model.

(c) Grain density term $\rho_g$

The grain density only affects the basement subsidence and the error is

$$\Delta H_b(\rho_g) = -\frac{\Delta \rho_g V_s}{(\rho_m - \rho_w)}$$  \hspace{1cm} \text{(4.2.19)}

This error term is linear in $\rho_g$ and gives an error opposite in sign to, $\Delta \rho$ as the loading effect of the sediments is over (under) corrected for with $+\Delta \rho_g$ ($-\Delta \rho_g$). Because eqn. 4.2.19 depends on $V_s$ which is constant for a given $z_1$ and $z_2$ then $\Delta H_b(\rho_g)$ is independent of the depth to which a unit is brought up to for compaction ($z'_1$). However, as shown in the curves A-G in figs. 4.2.5a-d (panel (iv)) the percentage error for the grain density term decreases as the depth to which the unit is moved up to, $z'_1$, decreases (from 3000 m, curve G to 0, curve A). This is because, although $\Delta H_b(\rho_g)$ is constant, the total value of $H_b$ increases (panel (i)). The maximum errors for the sandstone model ($\rho_g = 2650 \pm 130$ kg m$^{-3}$; $\Delta \rho_g/\rho_g = 5\%$) are $< 14\%$ and for the shale model ($\rho_g = 2710 \pm 200$ kg m$^{-3}$; $\Delta \rho_g/\rho_g = 7.4\%$), $< 24\%$.

4.3 Discussion and concluding remarks

In a sedimentary basin, porosity is quantified as a function of depth in terms of one, or more, simple functions depending on the lithological variation. These functions are based, if possible, on observations made on core samples or interpreted from well logging
Fig. 4.2.4a Sandstone model: percentage errors (right side scale) associated with the solid thickness (left side scale), $V_S$ as estimated from eqn. 4.2.14, due to errors in the exponential porosity/depth function terms $c$ and $\phi_0$ ($\Delta c$ and $\Delta \phi_0$ respectively). The thick solid line (A) is the error free value of $V_S$, the thinner solid lines (B) and (C) are the percentage errors in $V_S$ for $\Delta c$ and $-\Delta c$ and the dashed line (D) is the percentage error for $-\Delta \phi_0$. The results are given for a thickness of $z_2 - z_1 = (i) 100$ m, (ii) $250$ m, (iii) $500$ m, (iv) $1000$ m with the upper surface $z_1$ at depths between 250 and 4750 m.
Fig. 4.2.4b  As Fig. 4.2.4a but for the shale model.
Fig. 4.2.5a  Sandstone model with a sediment thickness, $H_s = z_2 - z_1 = 100$ m.

Top left panel (i) is the backstripped basement subsidence, $H_b$, as a function of $z_1$, the depth to the top of the unit. The other three panels show the percentage error due to errors in the exponential porosity/depth function terms, $c$ and $\phi_0$ ($\Delta c$ and $\Delta \phi_0$) and the grain density, $\rho_g$ ($\Delta \rho_g$). Top right panel (ii) $\Delta H_b(c)$, the lower set of curves, marked with primes, for $+\Delta c$ and the upper set is for $-\Delta c$, bottom left panel (iii) $\Delta H_b(\phi_0)$ for $+\Delta \phi_0$ and bottom right panel (iv) $\Delta H_b(\rho_g)$ for $-\Delta \rho_g$. The results are given for 7 values of $z_1$, the depth to which the unit is moved for decompression, of (A) 0, (B) 250, (C) 500, (D) 1000, (E) 1500, (F) 2000, (G) 3000 m.
Fig. 4.2.5b  As Fig. 4.2.5a but with $H_b = z_2 - z_1 = 1000$ m.
Fig. 4.2.5c. Shale model with a sediment thickness, $H_s = z_2 - z_1 = 100$ m.

Top left panel (i) is the backstripped basement subsidence, $H_b$, as a function of $z_1$, the depth to the top of the unit. The other three panels show the percentage error due to errors in the exponential porosity/depth function terms, $c$ and $\phi_0$ ($\Delta c$ and $\Delta \phi_0$) and the grain density, $\rho_g$ ($\Delta \rho_g$). Top right panel (ii) $\Delta H_b(c)$, the lower set of curves, marked with primes, for $+\Delta c$ and the upper set is for $-\Delta c$, bottom left panel (iii) $\Delta H_b(\phi_0)$ for $+\Delta \phi_0$ and bottom right panel (iv) $\Delta H_b(\rho_g)$ for $-\Delta \rho_g$. The results are given for 7 values of $z_1$, the depth to which the unit is moved for decompression, of (A) 0, (B) 250, (C) 500, (D) 1000, (E) 1500, (F) 2000, (G) 3000 m.
Fig. 4.2.5d As Fig. 4.2.5c but with \( H_s = z_2 - z_1 = 1000 \) m.
(e.g. sonic, neutron logs). In the absence of detailed sedimentological studies and an understanding of the processes involved in porosity reduction in sediments, subsidence analysis generally assumes that compaction is the dominant mechanism controlling the changes in thickness of a sedimentary unit as it is buried. The empirically derived porosity/depth functions are then incorporated into the decompaction and backstripping procedures under this assumption to obtain estimates of the sedimentation rates and the tectonic, or unloaded basement, subsidence over time.

From the results of calculations described above and illustrated in figs. 4.2.4-5, it may be concluded that the errors associated with decompaction procedure are generally increased as a unit is brought up to progressively shallower depths where the porosity is greater, and so is the rate of change of porosity for an exponential function. These trends would be expected for other porosity/depth functions described such as those described earlier but the details would be slightly different. The percentage errors however are generally similar for different model thicknesses \((z_2-z_1)\) and as such the results in this form can be applied for any reasonable thickness of the sedimentary unit. From this analysis it would seem that errors of up to 20% may occur in the estimates of the solid thickness \(V_s\) and up to 40% in the basement subsidence \(H_b\), as a result of an inappropriate choice of a porosity function and sediment grain density. However it would be unrealistic to combine the individual errors from the \(\phi_0\) and \(c\) terms as the combined parameter range I adopt is over-cautious and gives a wide spread of possible values (fig. 4.2.3). In practice, the approach is to obtain a best fit curve through a set of observed data, giving a spread either side. In this situation the discrepancies between the observed and calculated porosities may be considered random and to destructively interfere, reducing the potential total error. If no data is available then it is generally necessary to obtain a relationship from an independent source. The range of curves shown in fig. 4.2.2 suggest it is possible that systematic errors may be introduced if the curve chosen is wholly unrepresentative of the true situation and these errors may be of the order of 10-20%. Similar comments may be made regarding the choice of the sediment grain density. It will be seen later that the overall shape of the subsidence curve is not sensitive to the porosity/depth function and errors of
greater magnitude are likely to arise from uncertainties concerning the palaeobathymetry and sea level variations, and the isostatic assumption.
5. DATA FOR SUBSIDENCE ANALYSIS OF THE EROMANGA AND COOPER BASINS

5.1 Introduction

In this chapter I give a description of the data sources and reduction for the subsidence analysis of the Eromanga and Cooper Basin sequences. Where relevant I extend the description to review some of the assumptions made in obtaining the data.

5.2 Data and Assumptions
5.2.a Stratigraphy and timescales

The stratigraphic nomenclature and absolute ages assigned to each sedimentary formation I use in this thesis are given in table 5.2.1. In recent years, considerable effort has been made trying to establish an absolute timescale and relate it to chronostratigraphic or palaeontologically based timescales (e.g van Hinte, 1976 a,b; Cohee et al., 1978; Harland et al., 1982; Odin, 1982; Snelling, 1985). The absolute ages are obtained from isotopic dating of volcanics and sediments and, more indirectly, from magnetostratigraphy. In order to subsequently relate these data to stratigraphic boundaries, it is necessary to have good biostratigraphic control. The references given above describe the methodology and data used to construct timescales and the subtleties of the approaches will not be discussed in detail here.

We are primarily interested in the Permian to Cretaceous stratigraphy for the Eromanga and Cooper Basin sequences. Unfortunately, the ages of the Jurassic and Early Cretaceous in the absolute timescales are poorly defined due to a lack of data. Comparison between the timescales proposed by Snelling (1985), Odin (1982) and Harland et al. (1982)
over this interval (table 5.2.2) reveal that the former two are most similar, although the Odin ages are of the order of 2-5 m.y. less than those of Snelling. The Harland et al. (1982) ages are between 4 and 11 m.y. greater than Snelling's. These differences can be resolved qualitatively as follows: the Odin timescale is based substantially on K-Ar ages from glauconite, a clay mineral found in marine sediments. Due to its clay structure, glauconite is susceptible to diffusive argon loss both with time and increasing temperature. As such, many regard glauconite K-Ar ages as minimum values and treat them accordingly (e.g. Carr et al., 1983). Harland et al. (1982) consider the ages they derive through the Jurassic and Early Cretaceous using glauconite K-Ar data as unreliable, i.e. too low. Instead they assume that from the top of the Anisian (Middle Triassic) to the base of the Albian (Early Cretaceous) each stage represents an equal time duration of 6-7 m.y. and, on this basis, assign absolute ages to the stage boundaries. They do admit, however, that not only would this approach be invalid for the Late Cretaceous where the individual stage durations vary by a factor of 15, but also that the K-Ar ages are, in part, reflecting real departures from this assumption of equal stage duration. Therefore the absolute ages in the Harland et al. (1982) timescale over the Jurassic and Early Cretaceous can probably be treated as maximum values.

The timescale of Snelling (1985) as outlined above falls between those of Odin (1982) and Harland et al. (1982) although is closer to the former. For this study I favour the Snelling (1985) timescale - because it represents the most recent comprehensive interpretation of the available data. However, the absolute ages given for chronostratigraphic boundaries are not regarded as precise but rather, as stated by Snelling (1985), to represent a subjective, being Snelling's, compromise between different data sets. Given also that the stratigraphic ages of the formations are often only cited to within a given stage name, I adopt an arbitrary error of ±5 m.y. on the absolute ages I use for plotting the subsidence curves to reinforce these uncertainties.
<table>
<thead>
<tr>
<th>Formation Name</th>
<th>Absolute age (Ma)</th>
<th>Epoch</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winton Formation</td>
<td>91</td>
<td></td>
</tr>
<tr>
<td>Mackunda Formation</td>
<td>95</td>
<td></td>
</tr>
<tr>
<td>Allaru Mudstone</td>
<td>96</td>
<td></td>
</tr>
<tr>
<td>Toolebuc Formation</td>
<td>100</td>
<td>Cretaceous</td>
</tr>
<tr>
<td>Wallumbilla Formation</td>
<td>101</td>
<td></td>
</tr>
<tr>
<td>Cadna-Owie Formation</td>
<td>113</td>
<td></td>
</tr>
<tr>
<td>Murta member (Mooga Fm)</td>
<td>127</td>
<td></td>
</tr>
<tr>
<td>Namur member (Mooga Fm)</td>
<td>133</td>
<td>Jurassic</td>
</tr>
<tr>
<td>Westbourne Formation</td>
<td>138</td>
<td></td>
</tr>
<tr>
<td>Adori Sandstone</td>
<td>144</td>
<td></td>
</tr>
<tr>
<td>Birkhead Formation</td>
<td>146</td>
<td></td>
</tr>
<tr>
<td>Hutton Sandstone</td>
<td>158</td>
<td></td>
</tr>
<tr>
<td>Basal Jurassic</td>
<td>175</td>
<td></td>
</tr>
<tr>
<td>Unconformity</td>
<td>2186 - 2205</td>
<td></td>
</tr>
<tr>
<td>Peera-Peera beds</td>
<td>205</td>
<td>Triassic</td>
</tr>
<tr>
<td>Walkandi Beds</td>
<td>235</td>
<td></td>
</tr>
<tr>
<td>Nappamerri Formation</td>
<td>205 - 230</td>
<td></td>
</tr>
<tr>
<td>Toolachee Formation</td>
<td>251</td>
<td></td>
</tr>
<tr>
<td>Daralingie Beds</td>
<td>259</td>
<td></td>
</tr>
<tr>
<td>Roseneath Shale</td>
<td>264</td>
<td></td>
</tr>
<tr>
<td>Epsilon Formation</td>
<td>269</td>
<td>Permian</td>
</tr>
<tr>
<td>Murteree Shale</td>
<td>272</td>
<td></td>
</tr>
<tr>
<td>Patchawarra Formation</td>
<td>277</td>
<td></td>
</tr>
<tr>
<td>Tirrawarra Sandstone</td>
<td>285</td>
<td></td>
</tr>
<tr>
<td>Merrimelia Formation</td>
<td>290</td>
<td></td>
</tr>
</tbody>
</table>

Table 5.2.1: Absolute ages assigned to the tops of sedimentary formations for the Eromanga (Jurassic-Cretaceous) and Cooper (Permian-Triassic) Basin sequences using the Snelling (1985) timescale. The stratigraphic ages are given in tables 3.3.1 and 3.4.1.
<table>
<thead>
<tr>
<th>Stratigraphic age</th>
<th>Absolute age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Lower Cretaceous</strong></td>
<td></td>
</tr>
<tr>
<td>Cenomanian</td>
<td>91</td>
</tr>
<tr>
<td>Albian</td>
<td>95</td>
</tr>
<tr>
<td>Aptian</td>
<td>107</td>
</tr>
<tr>
<td>Barremian</td>
<td>114</td>
</tr>
<tr>
<td>Hauterivian</td>
<td>116</td>
</tr>
<tr>
<td>Valanginian</td>
<td>120</td>
</tr>
<tr>
<td>Berriasian</td>
<td>128</td>
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<tr>
<td><strong>Jurassic</strong></td>
<td></td>
</tr>
<tr>
<td>Tithonian</td>
<td>135</td>
</tr>
<tr>
<td>Kimmeridgian</td>
<td>139</td>
</tr>
<tr>
<td>Oxfordian</td>
<td>144</td>
</tr>
<tr>
<td>Callovian</td>
<td>152</td>
</tr>
<tr>
<td>Bathonian</td>
<td>159</td>
</tr>
<tr>
<td>Bajocian</td>
<td>170</td>
</tr>
<tr>
<td>Aalenian</td>
<td>176</td>
</tr>
<tr>
<td>Toarcian</td>
<td>180</td>
</tr>
<tr>
<td>Pliensbachian</td>
<td>188</td>
</tr>
<tr>
<td>Sinemurian</td>
<td>195</td>
</tr>
<tr>
<td>Hettangian</td>
<td>201</td>
</tr>
<tr>
<td><strong>Triassic</strong></td>
<td></td>
</tr>
<tr>
<td>Rhaetian</td>
<td>205</td>
</tr>
<tr>
<td>Norian</td>
<td>210</td>
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<tr>
<td>Carnian</td>
<td>220</td>
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<tr>
<td>Ladinian</td>
<td>230</td>
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<tr>
<td>Anisian</td>
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<tr>
<td>Scythian</td>
<td>240</td>
</tr>
<tr>
<td><strong>Permian</strong></td>
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<tr>
<td>Tatarian</td>
<td>250</td>
</tr>
<tr>
<td>Kazanian</td>
<td>255</td>
</tr>
<tr>
<td>Kungurian</td>
<td>260</td>
</tr>
<tr>
<td>Artinskian</td>
<td>270</td>
</tr>
<tr>
<td>Asselian</td>
<td>280</td>
</tr>
</tbody>
</table>

**Table 5.2.2.** Comparison of the Snelling (1985), Odin (1982) and Harland et al. (1982) timescales for the Permian to Lower Cretaceous tops of stratigraphic stages.
5.2.b *Porosity Data*

I assume for the purpose of later calculations that porosity reduction is attributable to simple compaction and/or equal rates of dissolution and precipitation, such that solid mass is preserved during burial. Carbonate and silica cements and also development of authigenic clay minerals as a result of feldspar degradation are observed in the sediments, but on a local and irregular scale (Senior et al., 1978; Pecanek and Paton, 1984; Moore and Pitt, 1984; Williams et al., 1985; Moore, 1986; P.S. Moore et al., 1986b; Ambrose et al., 1986; Gravestock and Alexander, 1986). Moreover, the fact that the sandstone rich sequences in the Eromanga Basin constitute a major aquifer system, effectively continuous over distances greater than 1000 km (Habermehl, 1980, 1986), supports the statement above that cementation and clay minerals are local factors influencing porosity. The major factor controlling the degree of compaction would then be lithology and indeed it is observed that the finer grained, more argillaceous sediments have intrinsically lower porosities than the clean quartzose sandstones (Porter, 1976; Moore and Pitt, 1984; Moore, 1986; Gravestock and Alexander, 1986). I have used values of effective, or interconnected, porosity from the following sources:

(i) Helium injection measurements from well completion reports

(Delhi Petroleum Pty Ltd and Bureau of Mineral Resources).


(iii) My own measurements made by evacuating samples under water and measuring the dry and saturated masses ($M_{dry}$, $M_{wet}$) and the porosity is derived as follows:

\[ M_{dry} = \rho_g (1-\phi) + \rho_{air} \phi \]  \hspace{1cm} (5.2.1a)

\[ M_{wet} = \rho_g (1-\phi) + \rho_w \phi \]  \hspace{1cm} (5.2.1b)
and then

\[
\phi = \frac{\Delta M}{(\rho_w - \rho_{air})}
\]  

(5.2.1c)

The results of (iii) agree favourably with (ii) when the same samples were used (see Appendix 1, table 2b). The data was grouped in one of 4 lithologies - shale (including mudstone, claystone and coal), siltstone, fine grained sandstone and coarse grained sandstone. The use of downhole lithology logs precludes any finer division. I adopted an exponential porosity/depth function (eqn. 4.2.1) for all 4 lithologies. The parameters, \(\phi_0\) and c, were obtained by a direct grid search method in model \((\phi_0, c)\) space similar to the approach taken in the 4 parameter problem of earthquake hypocentre location by Sambridge and Kennett (1986). The model misfit statistic, \(\sigma(\phi_0, c)\), calculated was

\[
\sigma(\phi_0, c) = \prod_{i=1}^{n} \chi(d_i)
\]

where \(d_i\) is the i-th data point and the probability density function, \(\chi(d_i)\), is given by

\[
\chi(d_i) = \exp \left( -\frac{d_i^{\text{calc}} - d_i^{\text{obs}}}{\sigma} \right)
\]

To obtain a best fit the misfit statistic is maximized, because as the difference between the observed and calculated data decreases, so \(\chi(d_i)\) increases. The form of probability density function given above is more appropriate than, for example, a Gaussian distribution as it places less weight on outliers. Therefore, it is more suitable for the simplified physical situation we are interested in. I chose \(\sigma = 1.67\%\) (porosity units) so that the 95% confidence limit corresponds to an error of \(\pm 5\%\) (porosity units). Figs. 5.2.1 and 5.2.2 show the observed data for each lithology, the best fit and bounding curves and table 5.2.3 below gives the parameters derived for these curves. The bounding curves were obtained simply by eye.
Fig. 5.2.1 Observed porosity/depth relationships in the Eromanga/Cooper Basin sequences for (i) coarse sandstone, (ii) fine sandstone, (iii) siltstone (iv), shale with the best fit and bounding curves as defined by the exponential function, eqn. 4.2.1.
Fig. 5.2.2 Observed porosity in the Eromanga/Cooper Basin sequences for all lithologies with the best fit curves. Oval - coarse sandstone, diamond - fine sandstone, asterisk - siltstone, triangle - shale.
<table>
<thead>
<tr>
<th>Lithology</th>
<th>Best fit $\phi_0$ (%)</th>
<th>c (km$^{-1}$)</th>
<th>Upper bound $\phi_0$ (%)</th>
<th>c (km$^{-1}$)</th>
<th>Lower bound $\phi_0$ (%)</th>
<th>c (km$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>coarse sandstone</td>
<td>42.8</td>
<td>0.614</td>
<td>48.0</td>
<td>0.446</td>
<td>38.0</td>
<td>0.846</td>
</tr>
<tr>
<td>fine sandstone</td>
<td>43.3</td>
<td>0.822</td>
<td>48.0</td>
<td>0.688</td>
<td>38.0</td>
<td>1.128</td>
</tr>
<tr>
<td>siltstone</td>
<td>45.7</td>
<td>1.158</td>
<td>51.0</td>
<td>0.965</td>
<td>41.0</td>
<td>1.310</td>
</tr>
<tr>
<td>shale</td>
<td>50.4</td>
<td>1.616</td>
<td>55.0</td>
<td>1.300</td>
<td>45.0</td>
<td>1.760</td>
</tr>
</tbody>
</table>

Table 5.2.3: Best fit, upper and lower bound values derived for the parameters $\phi_0$ and c in the exponential porosity/depth function (eqn. 4.2.1) for coarse and fine grained sandstones, siltstones and shales. The curves and data are illustrated in figs. 5.2.1 and 5.2.2.

I make 2 comments regarding the best fit curves. Firstly, for all 4 groups, there is a lack of data in the interval between 500 and 1000 m. This is due to the fact that the lithology between these depths is commonly shale, mudstone or siltstone (Allaru Mudstone and Wallumbilla Formation) and, as such, is of little interest to exploration companies or hydrologists as reservoirs. Except for the shales, the problem is alleviated as near surface (<500 m) observations are available to constrain the exponential parameters. The shale curve has only 2 points in the top 1200 m and these are effectively both at the same depth. It is possible to make an estimate of porosity from interval travel times as recorded by sonic logs. The method uses the empirical time-average equation of Wyllie et al. (1958):

$$\phi = \frac{(\Delta t_{obs} - \Delta t_{ma})}{(\Delta t_{fluid} - \Delta t_{ma})}$$

Here $\Delta t_{obs}$ is the observed travel time, $\Delta t_{ma}$ the solid (matrix) travel time and $\Delta t_{fluid}$ the travel time of the fluid in the pore space. In practice it is the velocities ($V=1/\Delta t$) that are used. With $\Delta t_{ma} = 205$ $\mu$s/m for shale and $\Delta t_{fluid} = 560$ $\mu$s/m (Telford et al., 1976) the results obtained with this method were too high by 15-20% (porosity units) when
compared with measured values at depths between 1200 and 1600 m. It is well known that small amounts (<10%) of shaly matter in sandstones can result in an underestimate of the porosity when using this method (Schlumberger, 1972; Porter, 1976). Thus it may seem inconsistent that for a shale rich unit the values are too high. However the sonic log, as with most logging tools, has a depth of penetration on a scale of centimetres and only samples local velocities. For a shale, velocities may be reduced, and porosity overestimated, due to drilling disturbances (Hicks, 1959; Jones and Wang, 1981) or factors such as pore anisotropy, e.g. horizontally flattened as would be expected in shales (Toksöz et al., 1976). Therefore, the parameters for the shale curve cannot be regarded as particularly well constrained but, as was shown in section 4.2.e, this is not likely to be significant in the decompaction and backstripping procedures. The bounding curves will be used to calculate maximum and minimum bounds on the tectonic subsidence and sedimentation rate. Most importantly, it will be seen that the shape of the subsidence curves is not significantly affected by these constants.

5.2.c Density data

Grain densities were obtained for the 4 lithologies from the same data sources as the porosity data described in the previous section. Knowing the porosity allows an estimate of the grain density to be made from eqns. 5.2.1a-c. Histograms of these grain densities are shown in fig. 5.2.3. I have grouped together the two sandstones because the values are very similar. Based on the modal values of these data distributions, I adopt a grain density of 2670 kg m\(^{-3}\) for sandstones and 2680 kg m\(^{-3}\) for siltstones and shales. The shales, once again, suffer from a deficit of data and show the largest spread (±100 kg m\(^{-3}\)). However this is only half the error assumed for the calculations in section 4.2.e and, as I have taken the modal value of the observations, the maximum error due to this parameter would only about 25% of those calculated in section 4.2.e, approximately 5% of the calculated tectonic subsidence.
Fig. 5.2.3 Histograms of sediment grain density (gm cm\(^{-3}\)) for sandstone, siltstone and shale. The data were obtained from well completion reports and from measurements made on core samples, as described in section 5.2.c.
5.2.d Palaeobathymetry

Depth of water during deposition, or palaeobathymetry, is estimated from faunal correlations with similar present day distributions, but is difficult to constrain accurately (van Hinte, 1978; Wood, 1981; Hegarty et al., 1987). As most sedimentation in the Eromanga Basin is non-marine, water depths can, for the most part, be ignored. Marine influences first appeared in the Early Cretaceous (Neocomian) but were relatively short lived with fluvial conditions being re-established by the Middle Cretaceous (Cenomanian) although, as will be seen later, this stage of the basin’s history is significant. The fauna in these marine sediments are interpreted as being of shallow water (littoral) affinity with water depths of less than 50 m (Scheibnerová, 1976, 1980, 1986; Morgan, 1980; McMinn and Burger, 1986; Burger, pers. comm., 1986). In accordance with the water depths suggested by Scheibnerová (1986) and the relative variations of water depth proposed by Morgan (1980), I adopt a maximum water depth of 50 m during the time of deposition of the Toolebuc Formation, 30 m during the Wallumbilla Formation and Allaru Mudstone, and 10 m during the Mackunda Formation. The error in the basement subsidence associated with water depth is equal to the error in the water depth (see eqn. 4.2.12a). As the sediments are all shallow marine this is likely to be less than 50 m. Deep water sediments would have a greater uncertainty in their palaeobathymetry (e.g. van Hinte, 1978; Wood, 1981).

5.2.e Sea level variations

That mean global sea level has not been a constant datum over at least the last 500 Ma is currently a widely supported hypothesis. More unresolved are the mechanisms for, and the absolute magnitudes of, these variations. I briefly discuss the evidence for sea level
changes, the mechanisms and magnitudes proposed and the application of sea level curves to the Eromanga Basin.

It was recognized early this century (Suess, 1906) that the presence today of relatively undeformed marine sediments on continental platform areas implies flooding in the past. This could be accommodated by a sea level rise, continental subsidence or a combination of the two. The apparent global synchrony of these events favours a major contribution from sea level variations (Hancock and Kauffman, 1979; Harrison et al., 1981, 1983; Hallam, 1984). Sloss and Speed (1974) have suggested that the continents may have moved vertically in concert although the mechanism responsible for such movements is open to suggestion. In order to estimate the magnitude of sea level changes from sediment distributions it is necessary to know not only the water depth during deposition, but also the elevation of the continent relative to the contemporary sea level - the freeboard or hypsometry (Bond, 1976, 1978, 1979; Harrison et al., 1981, 1983; Wyatt, 1984, 1986; Cogley, 1984; Schubert and Reymer, 1985). The basic observation made today is that the larger a continent is in area, the higher its mean elevation. Thus for example, Australia has the lowest mean elevation of the major continents and Asia has the greatest. Taken at face value, the implication of this observation for past sea levels is this: the large continental area of Pangea (or eastern Gondwanaland), which existed up to the earliest Mesozoic (Audley-Charles, 1983), would be expected to have had a greater mean elevation than any of the present day continents which made up the super-continent. Therefore, the ancient land mass would have been less susceptible to marine incursions during sea level rises. Cogley (1984) makes the comments, however, that the mean height/area relationship is an artifact of the greater relative amounts of tectonism and orogenic belts in the larger continents and that modal heights are similar, about 250 m, for all continents currently above sea level. Therefore, inferences regarding Pangea's hypsometry should be regarded as equivocal. Wyatt (1986) proposed a hypsometric model to incorporate the constant modal height observation of Cogley (1984). He stated that if the high topography was contained in the interior of the continent, global sea level changes
should flood each continent to approximately the same depth. Bond (1978, 1979) shows that this probably is not always so as the amount of flooding for example, in the Cretaceous, is variable between continents and he attributes this to differential vertical movements, and hence hypsometries, of the continents.

Seismic stratigraphy, a technique developed within Exxon (Payton, 1977), uses the concept that seismically defined sequences are bounded by what are assumed to be chronostratigraphic, or constant time, boundaries. Progressive landward migration of these sequences and associated facies distributions is termed onlap and is taken as an indication of a sea level rise relative to the basement. Conversely, offlap or migration of the sequence away from the initial shoreline region is attributed to a relative sea level fall. Problems arise with the results of Vail et al. (1977a,b), not least in that the use of proprietary data effectively precludes an independent interpretation and assessment of the proposed sea level curves but more fundamentally, with the approach used to estimate vertical components of sea level changes (Bally, 1982; Watts, 1982; Hallam, 1984). Rates of sea level variation, subsidence and sedimentation are all important in controlling the stratigraphy of both continental margins and interior platforms (Sleep, 1976; Pitman, 1978; Watts et al., 1982; Watts, 1982; Pitman and Golovchenko, 1983; Turcotte and Willemann, 1983; Thorne and Watts, 1984; Schwarzacher and Schwarzacher, 1986). Additionally, sea level falls are complicated by removal of parts of the section. Thus, the curves of Vail et al. (1977a,b) show gradual rises in sea level with instantaneous falls, partly because they ignored facies differences between marine and continental environments (see Hallam 1984), although they eventually recognized the inconsistency of the original curves (Vail and Todd, 1981). Another limitation of the method is that it does not permit estimation of absolute magnitudes of sea level change and therefore requires calibration from other methods.

Watts and Steckler (1979) reformulated the backstripping equation (4.2.11c) to estimate the magnitude of sea level changes on the eastern coast of North America. They assume that the tectonic subsidence is thermally driven and, after accounting for the contribution of this, the subsidence due to the sediment loading and the variations in water
depth during deposition, are left with what they assume to be the global sea level term in eqn. 4.2.11c. Guidish et al. (1984) also use basement subsidence to provide a control on the eustatic component of subsidence. They propose that, by weighted stacking of subsidence curves from a world wide data set, the eustatic contribution will be enhanced as it is a common factor. This approach would also enhance any similar, if not related, tectonic subsidence features. Their proposed world-wide data set has nearly 50% of the wells from the North Sea and 23% from the northwest shelf of Australia. A similar criticism regarding the validity of a global interpretation of the results of Watts and Steckler (1979), as their data is limited to the eastern seaboard of the USA. Guidish et al. (1984) correlate their master rate of basement subsidence curve to the relative sea level curve of Vail et al. (1977a,b) but make no suggestions regarding the magnitude of the variations.

In spite of the diversity of techniques used to obtain curves of relative sea level changes, there are consistent features giving some credibility to the supposition of global, or eustatic, sea level variations over time. For example, most agree that there appears to have been a major highstand in the Late Cretaceous and a lowstand in the early Mesozoic (see fig. 5.2.4). These are the long term or first order cycles of Vail et al. (1977a,b) occurring over some 200-300 Ma. Second order (10-100 m.y.) and third order (1-10 m.y.) cycles are superimposed on the first order curve but many authors have reservations as to the global nature of these shorter term fluctuations. Thus, Watts (1982) has shown that onlap can occur at a passive margin as a natural consequence of the flexural response of a cooling lithosphere to sediment loading. He correlates ages of rift-drift transitions (i.e. change from fault controlled, locally compensated subsidence to regionally compensated thermal subsidence) observed at continental margins to some of the second order cycles of the Vail curves.

Pitman (1978) and Pitman and Golovchenko (1983) briefly discuss the effect of glacial episodes on on global sea level and conclude that recent (<15 Ma) variations of greater than 10 m/m.y. may well be due to successive periods of glaciation. Nakada and Lambeck (1987) present a comprehensive analysis of this effect and the application of
Fig. 5.2.4  Global sea level curve for the Mesozoic after Falvey and Deighton (1982). This curve is used for sea level corrections in the backstripping procedure described in section 4.2.d. The timescale is that of Snelling (1985)
observed sea level variations in estimating the viscosity of the upper mantle. They suggest that the earth responds to glacial unloading and meltwater loading on timescale of $10^{4-5}$ years. Guidish et al. (1984) suggest that in addition to major periods of glaciation, mountain glaciation may contribute a small (a few metres) amount of sea level rise or fall, being responsible for upward shoaling seen in sedimentary sequences. Melting of present day ice sheets, a volume of about $3 \times 10^{16}$ m$^3$ (Drewry, 1983), and distributing the meltwater uniformly over the present day oceans ($=3.63 \times 10^{14}$ m$^2$, including continental margins, Turcotte and Schubert, 1982) would give a sea level rise of about 60 m, after correcting for isostasy.

Cloetingh et al. (1985) proposed a simple mechanism to explain the regional variations apparent in many sea level curves. The mechanism involves the flexural interaction of a horizontal force and a pre-existing deflection of the lithosphere similar to that suggested by Lambeck (1983a,b) for the tectonic evolution of central Australia and discussed in section 2.3.c. Application of a horizontal compressional force will uplift the basin flanks and deepen the central area. The opposite effect is predicted under tension. The differential displacement relative to the zero datum is greatest at the basin margin, where seismic stratigraphy techniques would be applied and the magnitude of the additional deflection depends on the force, the pre-existing deflection and the strength of the lithosphere. Care should be taken when applying this method as the relative change depends on the position of sea level with respect to the zero datum of the flexural model and it possible to infer the opposite sense of applied force to that which may be responsible for the apparent sea level change. Lambeck et al. (1987) have applied the method to the North Sea Basin. They have shown that fluctuations in the regional stress field required to explain the apparent sea level variations coincide to some extent with episodes of, or changes in style of, tectonic activity in the area. Bally (1982) has also commented that many of the higher order sea level features in the Vail curves are most probably of a tectonic origin. It is clear that the site of apparent sea level variations should be examined in a regional tectonic context before one can attach any global significance to the observations.
The first order cycles appear to be related to major plate tectonic processes such as mid-ocean spreading rates and the opening of oceans. Increased spreading rates during the Cretaceous have been proposed to explain the observed highstand (Hallam, 1963; Hays and Pitman, 1973; Pitman, 1978). The mean depth of the ocean floor, relative to a constant datum, is reduced for an increased spreading rate as the long wavelength seafloor bathymetry is controlled by diffusive heat loss and thermal isostasy (Sclater and Francheteau, 1970). Thus a sea level rise is expected for a high spreading rate. A decrease in spreading rates since the Cretaceous is a possible explanation for the long term sea level fall to the present day. Pitman (1978) states that this fall could be as much as 350 m. Falvey and Deighton (1982) suggest that this is possibly overestimated by about 50% as the relevant magnetostratigraphy has been revised and also as Pitman (1978) had not considered the possibility of asymmetrical spreading. Harrison et al. (1981) also modify the Pitman estimate by applying a correction for the rapid ocean basin sedimentation which has occurred over the last 5 Ma and the presence of about 300 m of deep water carbonates deposited over the last 100 Ma, and their preferred estimate is 267 m. Kominz (1984) concluded that the variation in sea level due to changes in ridge volume between the Late Cretaceous and the present day was between 45 and 365 m, and her preferred value was 230 m.

Hager (1980) has argued that the relationship between sea level and spreading rates is not as simple as described above. An increase in spreading rates will be accompanied by an increase in the rate at which relatively cool lithosphere will be subducted and a local volume decrease in predicted in the sublithospheric mantle, due to the enhanced cooling. If subduction occurs under a continental area, the freeboard is reduced and an apparent sea level rise results. Conversely, if subduction occurs under a oceanic area, the sub-oceanic lithosphere subsides cools more rapidly and the volume of the ocean basins increases, giving an apparent sea level fall. Hager (1980) suggests that changes of the order of 10% in the distribution of subducted lithosphere beneath continents and oceans will have a more profound effect on global sea level variations than a doubling of the spreading rate.
Heller and Angevine (1985) incorporate both spreading and subduction (or lack of it), relating first order sea level changes primarily to the formation of Atlantic-type (passive margin) oceans. In principle, the model is similar to the spreading ridge hypothesis in that the average age of the ocean floor, or at least the age/area distribution is the major control on sea level. Subduction reduces the average age of the oceanic lithosphere, while the formation of an Atlantic type ocean basin will tend to progressively increase the average age as the newly formed basin is not being consumed at its margins. Therefore, when these basins form, sea level is expected first to rise due to the initially low average age and shallower bathymetry of the ocean floor. As the ocean continues to open the average age, and therefore mean depth of the ocean floor, increases and mean sea level will begin to fall. The simplified 2 ocean model of Heller and Angevine (1985) predicts a sea level peak of about 100 -150 m during the Cretaceous followed by a fall to the present day. Given current observations of plate processes it would seem reasonable to conclude that the ocean ridge and basin models described above are manifestations of the same plate tectonic activity, and that they might be expected to operate at the same time. These mechanisms can therefore explain sea level highs of between 100 and 300 m in the Late Cretaceous.

Sleep (1976) inferred a sea level high of 325 m from the observation of late Cretaceous marine sediments currently at elevations of 300-400 m above present day sea level. However, Bond (1978, 1979) concluded that the North American platform has been elevated by 120-160 m since this time, reducing Sleep's estimate to about 200 m. Hallam (1984) gives a value of 350 m derived from palaeogeographic maps illustrating inundation by marine conditions, or distribution of marine sediments. Wyatt (1986) suggests that the Late Cretaceous peak of Hallam's curve may be increased to over 400 m if the hypsometric assumptions are modified to include thinned continental margins and the inferred continent wide isostatic subsidence. Heller and Angevine (1985) also state that the effect of thinning continental margins may cause an apparent sea level rise of the order of 60-70 m. In discussion of other sea level curves, Hallam (1984) quotes, as supporting evidence for a 350 m high, 375 m from Sleep (1976) and 390 m from Bond (1976), although both of
these estimates have subsequently been revised. Watts and Steckler (1979), using the modified backstripping approach, arrive at a maximum value of 150 m which is similar to that of 150-200 m favoured by Bond (1978).

As stated earlier, the method used by Guidish et al. (1984) does not predict magnitudes of sea level variations. They conclude that the similarity of the long term components of their 'master rate of basement subsidence curve', the average subsidence calculated from 159 wells, to the Vail et al. (1977a,b) relative sea level curve implies a relationship between the rate of subsidence and sea level variations. This conclusion, although perhaps tenuous when taken in context, is noteworthy in that the sea level curves derived using different approaches do show essentially the same gross long term features although discrepancies arise in the estimates of the magnitudes of the sea level changes.

It is generally assumed that sea level represents the depositional base level during sedimentation. Therefore if this datum varies with time relative to the tectonic subsidence, it should be corrected for. For this study I use the sea level curve of Falvey and Deighton (1982) recalibrated to the Snelling (1985) time scale (fig. 5.2.4). This curve is based on the premise that the form of the first order relative sea level curve of Vail et al. (1977) is correct, but the magnitudes will be closer to the estimates of Watts and Steckler (1979) and Bond (1978). The Cretaceous highstand at 78 Ma is 197 m above present day sea level. However the role of sea level variations in sedimentation on continental platforms, or interiors, is not obvious, and will certainly depend on the elevation of the region relative to the contemporary sea level. For example, recent (Cainozoic) sediments in the Chad Basin (Burke, 1976) in Africa and the Amazon Basin (Gibbs, 1965) in South America are at elevations up to 300 and 200 m above sea level respectively.

As the sediments in the Eromanga Basin are all non-marine prior to about 120-113 Ma, the only influence sea level variations would probably have is on the drainage direction of the major river systems considered to be controlling deposition. Marine influences are first apparent during deposition of the paralic Cadna-Owie Formation, or aptly named
Transition Beds (Senior et al., 1978), between 127 and 113 Ma (table 5.2.1). For calculating the tectonic subsidence then, I assume that sea level at 113 Ma (+62 m) was the base level and any sea level fluctuations on the continent are relative to this datum. The implication is that the sea would have had to rise by 62 m before it began to encroach into the depositional area on the continent. Consequently the Late Cretaceous highstand at 78 Ma would have been 135 m of water in this area, if no deposition occurred. As sedimentation ceased at approximately 90 Ma (sea level = 130 m above present day) when sea level was still increasing up to the Late Cretaceous highstand, the maximum correction to the tectonic subsidence curve, from eqn. 4.2.12b, is \( \frac{\rho_m}{(\rho_m-\rho_w)} H_{sl} \) or \( 1.45 \times (130-62) = 99 \) m. Using the sea level value at 113 Ma as a zero datum also means that even if I was to use the Hallam (1984) or Wyatt (1987) sea level curves, the maximum correction to the tectonic subsidence would be less than 170 m, as the change in sea level between 113 and 90 Ma is less than 120 m for their curves, although the absolute values differ by some 200 m when compared to the Falvey and Deighton (1982) curve. The assumption made here that depositional base level during sedimentation is maintained at, or for marine sediments below, sea level will be discussed in sections 6.5 and 6.6 with particular reference to the calculated tectonic subsidence curves for the Eromanga Basin.

5.2.6 Rates of sedimentation

Although variable, rates of sedimentation in different tectonic settings show some consistent trends (e.g. Blatt et al., 1976; Schwab, 1976; Miall, 1978; Conybeare, 1979; Reading, 1982). Most simply the observation is that basins close to active plate margins (foreland basin, rift and wrench controlled basins) show higher rates (> 50 m/m.y.) and those distant from margins (intracratonic and deep ocean basins) have lower rates (< 50 m/m.y.).
It is important to consider what is meant by rates of sedimentation. It is well known that present day rates of sediment accumulation are commonly an order of magnitude higher than rates estimated from ancient basins (e.g. Miall, 1978). This discrepancy can arise as a result of the compaction of the ancient sedimentary sequences not being considered, as is the case in the values quoted in the preceding paragraph, but also because sedimentation is not as continuous as is supposed (e.g. flash flooding followed by a relatively long period of non-deposition), depocentres vary in time and space (e.g. rapid lateral migration of depocentres as in deltaic environments), poor control on the absolute ages of the sediments is likely and the fact that relatively large amounts of sediment have been deposited since the Pleistocene glaciation, suggesting that present day rates are likely to be abnormally high (e.g. Ager, 1973; Blatt et al., 1976; Miall, 1978; Conybeare, 1979; Reading, 1982; Tipper, 1987).

Where sedimentation is not continuous, the rate estimate may be regarded as a lower limit, although of course this would be invalid if erosion had occurred. Averaging sedimentation rates over a sedimentary basin may alleviate the problem of depocentre migration, although care should be exercised when summing sediment thicknesses from different areas as discrete episodes of sedimentation may be difficult to recognize, as in the wrench controlled Californian basins (Reading, 1982). The effect of compaction can at least be partially compensated for by applying the decompaction procedure described in section 4.2.c. However, using the sediment thickness derived with this approach is somewhat unsatisfactory as the lower part of the unit will already have been partially compacted by the weight of the overburden, and the effect will be greater for thicker units. Hutchinson (1985) has shown that, as a result of compaction, the rate of solid sediment accumulation at the surface is greater than the velocity of the solid material at depth and therefore the rate of subsidence is not directly (linearly) related to the changes in sediment thickness at depth.

Therefore we define here a surface sedimentation rate as the solid component, $V_s$, divided by the time interval over which the unit was deposited, $V_s/\Delta t$. Here $V_s$ is defined as a volume for a unit cross-sectional area, or equivalently, thickness. This sedimentation
rate is then taken to be a measure of the flux of solid detritus into the basin and is assumed to be linear and continuous for the duration of deposition. This approach allows a comparison of rates which is, in theory, independent of lithology. I neglect diagenetic processes such as cementation which are likely to be more common in sandstones than shales. If porosity is reduced early in the burial history due to cement infill then the rates may be overestimated. Errors in $V_s$ due to an unrepresentative porosity function were discussed in section 4.2.e. These are less than 20% and are directly applicable to the sedimentation rate estimates. In practice, however, greater uncertainties will be introduced as a result of incorrect assignment of the absolute ages. Van Hinte (1978) outlines methods to estimate the rate of sedimentation from well data and concludes that the approach taken above is the most suitable, although he rightly emphasizes 'the need to apply rate equations consistently with due regard for assumptions inherent in each' (pg 213).
6. BACKSTRIPPING AND SUBSIDENCE IN THE EROMANGA AND COOPER BASINS

6.1 Introduction

In this chapter I apply the backstripping method described in chapter 4 to well data from the Eromanga Basin region. The purpose of this study is to examine how the subsidence history may be used as a constraint on the mechanisms involved in the evolution of the basins in the area. For the Permian to Triassic time we concentrate mainly on the Cooper Basin, although some of the wells used are located in the Simpson Desert and Galilee Basins. As I mentioned in chapter 3 the Galilee Basin appears to have been related to processes to the east of the Eromanga Basin, and the Permo-Triassic sequence in the west of the Galilee Basin is relatively reduced compared to the Cooper Basin section. The Simpson Desert Basin only contains Triassic sediments, but the depositional area was probably connected with that in the Cooper Basin at the time. The later Triassic sediments may in fact be related to the Jurassic deposition in the Eromanga Basin although the presence of an unconformity during the Middle-Late Triassic makes the relationship obscure. Sedimentation was continuous in the Eromanga Basin from the Jurassic to the Middle-Late Cretaceous, and it is that part of the basin's history that we examine in most detail.

6.2.a Summary of methodology

Forty wells were used for the subsidence analysis (fig. 6.2.1a). The absolute ages for each sedimentary formation were discussed in section 5.2.a and an error of ± 5 m.y. was assigned to each value. Where available, downhole lithology logs were used to estimate the relative proportions of the four lithologies; shale, siltstone, fine grained and coarse grained
Fig. 6.2.1a  Location of wells used for backstripping.

1  Macumba 1  21  Orientos 1
2  Poonarunna 1  22  Wareena 1
3  Kuncherinna 1  23  Mt. Howitt 1
4  Walkandi 1  24  Ingella 1
5  Putamurdie 1  25  Galway 1
6  Pandieburra 1  26  Bodalla 1
7  Kalladeina 1  27  Chandos 1
8  Fly Lake 1  28  Canaway 1
9  Tirrawarra 1  29  Yongola 1
10  Gidgealpa 1  30  Cothalow 1
11  Moomba 1  31  Quilberry 1
12  Mcleod 1  32  Dartmouth 1
13  Burley 2  33  Gilmore 1
14  Innamincka 1  34  Log Creek 1
15  McKinlay 1  35  Stafford 1
16  Kidman 2  36  Bury 1
17  Toolachee 9  37  Gumbardo 1
18  Durham Downs 1  38  Eastwood 1
19  Naccowlah 1  39  Ban Ban 1
20  Naryilco 1  40  Fermoy 1
sandstone. Otherwise, where only the depth of the formation tops were available, what were considered to be representative values (Delhi Petroleum, pers. comm., 1985) were assigned to the lithology for each formation. This was the case for about 25 of the wells. Each sedimentary formation recognized in a well was decompacted under the assumption that it was composed of 100% of each individual lithology. Then the arithmetic mean of these four decompacted sediment thicknesses and the mean sediment densities were used to calculate the water loaded ($\rho_1 = \rho_w$) tectonic subsidence. The calculations were made using the best fit porosity/depth curves and an estimate of the uncertainty is provided for by using the upper and lower bounding curves to the porosity data (section 5.2.b). Corrections were made for the water depth during deposition (section 5.2.d), and sea level variations (section 5.2.e).

In the absence of satisfactory isopach maps for the Permian to Cretaceous formations, the backstripping calculations were made under the assumption of local (Airy) isostasy. As the Eromanga Basin is over 1000 km wide then it is probable that the central region of the basin will be in a state approximating local isostasy, although this would depend on the width of any loads relative to the flexural parameter of the lithosphere (see section 2.2.e). As nearly all the wells are over 200 km from the basin margins then flexural uplift of the margin should not have influenced the tectonic subsidence. The incorporation of regional isostatic models, as done by Nunn and Sleep (1984), would not change the form of the subsidence curves significantly, but will affect the magnitude of the tectonic subsidence term. As I have previously mentioned, if flexure of the lithosphere makes a significant contribution to supporting sediment infill during basin evolution then backstripping under the assumption of local isostasy will underestimate the tectonic subsidence. This is because, under regional isostasy, the sediment load itself will be responsible for a smaller part of the total deflection, or subsidence.

Sedimentation rates (m/m.y.) were estimated from the solid thickness, $V_s$ (section 5.2.f). The results are illustrated (tectonic subsidence and sedimentation rates) in fig. 6.2.1b as a function of time and the numerical values (tectonic subsidence, decompacted sediment thicknesses and sedimentation rates) are given in appendix 2. The tectonic
subidence obtained with the best fit porosity curve is shown by a diamond symbol and the limits in depth and time derived by using the bounding curves and the ± 5 m.y. are illustrated as rectangular boxes. Unconformities are marked by a box with no diamond symbol. An unconformity was modeled as a period of non-deposition rather than making any assumptions about uplift and erosion. Similarly with the sedimentation rates, the preferred value is shown by a diamond symbol and the uncertainty due to the decompaction process is illustrated by a vertical bar although these are generally small and appear as a dot. I did not use errors in the absolute ages in calculating the sedimentation rates but it should be borne in mind that individual rates may be erroneous. I will address this point in more detail later.

As an illustration of the overall lack of sensitivity of the shape of the tectonic subsidence curve to the values inferred for the porosity/depth function, fig. 6.2.2 shows the tectonic subsidence and sedimentation rates for the Mcleod 1 well obtained by the averaging method described above, and also the results for each individual lithology. As can be seen, although the actual values of each point differ, the shapes of the subsidence curves are very similar. The individual sedimentation rates show the most variation in the near surface units due to the greater differences in the porosities and estimated solid thicknesses for each of the four lithologies at shallower depths. However, as implied by figs. 4.2.4a and 4.2.4b, the uncertainties are less than 15 %. In the next sections I discuss some of the features seen in the subsidence curves and relate these to the mechanisms which may have been involved in the formation and evolution of the Eromanga Basin.

6.3 Results of subsidence analysis

6.3.a Permo-Triassic subsidence : 300 - 205 Ma

We firstly consider the subsidence during the Permian and Triassic in the Cooper Basin region (see figs. 3.1.1, 3.3.3). Permo-Triassic sediments here can be up to 1500 m in thickness although the sequence is variable and often less than about 700 m. In the
Fig. 6.2.1b. Water loaded tectonic subsidence (upper panels) and sedimentation rates (lower panels) from the Permian to the Cretaceous for the wells in fig. 6.2.1a. The preferred values are shown as a diamond symbol. The error boxes represent ±5 m.y. in the absolute age and the tectonic subsidence values obtained using the bounding porosity/depth relationships (section 5.2.b). The uncertainties in the sedimentation rates are also obtained using the bounding curves but no errors are assumed in the absolute ages. The absolute age scale is at the bottom of each page and the stratigraphic divisions are given at the top. The black arrow marks the time of the end of the Cadna-Owie Formation deposition (113 Ma).
Fig. 6.2.1b contd.
Fig. 6.2.1b contd.
Fig. 6.2.1b contd.
Fig. 6.2.2. Tectonic subsidence (upper panel) and sedimentation rates (lower panel) for the Mcleod 1 well calculated in turn with the 4 porosity/depth relationships given in section 5.2.b for coarse grained sandstone, fine grained sandstone, siltstone, and shale. The tectonic subsidence calculated with the values derived from the lithology log is also shown. The shape of the subsidence curve and the sedimentation rates are relatively insensitive to the assumed porosity/depth relationship.
southern Cooper Basin, the thickest and most complete sequences are in the Nappamerri Trough (fig. 3.3.3) e.g. Burley 2 (1382 m); Mcleod 1 (1521 m). These two wells give a tectonic subsidence for the Permo-Triassic of between 700-750 m. The Fly Lake 1 and Tirrawarra 1 wells in the Patchawarra Trough to the west have about 550-700 m of Permo-Triassic sediments and a tectonic subsidence of 380-450 m. In the northern Cooper Basin sedimentation commenced a bit later, and the Permian section is reduced compared to the southern area, although the Triassic section is of an equivalent, if not greater, thickness. The tectonic subsidence is generally 250-350 m in this area (e.g. Chandos 1, Ingella 1, Galway 1, Mt. Howitt 1). Further to the east in the Galilee/Adavale Basin area (figs. 3.1.1, 3.3.2), subsidence was also reduced with less than 300 m of sediment accumulating in the Permo-Triassic and a tectonic subsidence during this time of less than 200 m (e.g. Dartmouth 1, Log Creek 1, Eastwood 1, Cothalow 1). In the far west of the Eromanga Basin, sedimentation and subsidence during this time was again relatively reduced compared to the southern Cooper Basin area. The Pedirka Basin accumulated less than 500 m of Early Permian sediments and in the Simpson Desert Basin (figs. 3.1.1, 3.3.1), less than 300 m of sediment is observed and 200 m of tectonic subsidence is estimated during the Triassic (Macumba 1, Walkandi 1, Kuncherinna 1).

Sedimentation and subsidence during the Permo-Triassic occurred irregularly in the Eromanga Basin area of eastern Australia, with the most significant thickness of Permian sediment occurring in the southern Cooper Basin. Where reasonably complete sequences are observed (e.g Burley 2, Mcleod 1, Moomba 1, Toolachee 9, Kidman 2), the earlier Permian subsidence is clearly the most rapid phase and the subsidence curves have a pseudo-exponential form, decreasing with time and suggestive of a thermal influence. However, the early part of this subsidence was dominantly fault controlled and locally developed unconformities are common. The trend in the basin is towards increasingly widespread (regional) subsidence into the Triassic as evidenced by onlap onto basement highs and the greater number of wells intersecting the later units over a more extensive area. Early in the subsidence history of the Cooper Basin, sediments were mainly derived from local topographic highs, probably fault scarps. By the Triassic, the depositional area had
increased, implying that the basement highs had been eroded, and the sediment was now coming from a more distant source into a topographic low, spanning the southern and northern Cooper Basin and the Simpson Desert Basin areas, with generally lower energy sedimentary depositional environments. It is tempting then to characterize the evolution of the Cooper Basin by fault controlled subsidence initially followed by a more regional, thermally driven, subsidence phase as predicted, for example, by the extensional models described in section 2.3.b. However, evidence for crustal extension is not strong and compressional and oblique faulting appear to have been more significant influences on sedimentation and subsidence (see section 3.6.a). The question that also arises at this point is what is the relationship, if any, between the formation mechanisms of the Cooper Basin and the overlying Eromanga Basin?

6.3.b Pre-Jurassic unconformity

The presence of an unconformity between the Cooper and Eromanga Basin sequences is taken by Middleton (1978, 1980) to indicate separate formation mechanisms for the two basins. He therefore invokes two successive episodes of heating and deep crustal metamorphism to explain the subsidence and the presence of the unconformity. As I discussed in section 3.3.c, the duration of this erosional or non-depositional unconformity is not well constrained and more recent studies, with a more extensive database, tend to conclude that the original estimate of an age span from the Middle Triassic to the early Jurassic, about 50 m.y., is too large and sediments of both Middle and Late Triassic ages have been recognized (Wiltshire, 1982a,b; Moore, 1982, 1986).

Sleep (1976) has shown that sequence bounding unconformities, manifested as kinks in subsidence curves, may be the result of sea level fluctuations on a continuously subsiding continental platform. However all of the sediments, from the Permian to the Jurassic inclusive, are non-marine so this mechanism is not considered relevant here. The general conformity of the sediment dips across the unconformity indicates that no major deformation affected this area and so the unconformity may be the result of a reduction in
sediment influx, reworking of pre-existing sediments or a change in sediment provenance related for example to a change in the gross drainage patterns of the depositional and surrounding areas. The observation of a widespread Middle/Late Triassic unconformity across, and the terminal deformation of, the Bowen and Sydney Basins (Veevers, 1984) to the east of the Eromanga Basin area, suggests that the ultimate cause of the unconformity would have been tectonic activity on the eastern margin of Australia, rather than requiring a fundamental change in the subsidence mechanism in the Cooper/Eromanga Basin region, as proposed by Middleton (1978, 1980).

6.3.c Jurassic-Cretaceous subsidence : ? 186 - 91 Ma

Subsidence and sedimentation in the Eromanga Basin sequence became areally more widespread and continuous than during the Permo-Triassic in the Cooper and Simpson Desert Basins. The rate of subsidence is approximately linear or decreasing with time up to the Cadna-Owie Formation (127-113 Ma) in the Early Cretaceous. I mentioned in section 3.4.a how the area of deposition progressively widened up to this time, such that the present day outcrop limits were reached. The later stages of the Cadna-Owie Formation mark the first signs of a marine influence on sedimentation, and a contraction in the area of deposition which accounts for the overall lack of exposure of this formation (see fig. 3.1.2). The apparent reduction of the depositional area may be an artifact of facies changes whereby that the marginal areas of the basin remained non-marine and reflect local sediment sources. After the Cadna-Owie Formation was deposited, the subsidence rate shows a marked increase. In the last 20 m.y. or so approximately the same amount of tectonic subsidence appears to have occurred as in the 80 m.y. prior to this in the Jurassic. An examination of the subsidence curves (fig. 6.2.1b) clearly illustrates that all of the wells show this phase of rapid subsidence after the Cadna-Owie Formation, and this is therefore a basin wide feature. Some wells do, however, show an apparent uplift, or lack of subsidence, in the final stage of subsidence in the Winton Formation time, (95-91 Ma), e.g.
Bury 1, Ban Ban 1, Dartmouth 1, Eastwood 1, Gumbardo 1, Stafford 1, Quilberry 1, Wareena 1. This is attributable to erosion of the Winton Formation and isopach maps (Senior et al. 1978) support this conclusion, as the wells have been drilled on structural highs where the upper part of the sequence is reduced. As described in section 3.6, the majority of structuring in the Eromanga Basin is considered to be post-depositional and occurred during the Tertiary. The deformation at this time resulted in reactivation of pre-existing structures, uplift and erosion of the upper part of the Winton Formation.

Sedimentation rates are included on the figures showing the tectonic subsidence for each well (fig. 6.2.1b) and were calculated as described in section 5.2.f. Although there are likely to be errors in the absolute ages that I use, it can be seen that the general trend is for the sedimentation rate to increase very rapidly towards the end of the basin's subsidence history, by a factor of 5-10 relative to the earlier Jurassic sedimentation rates (190-120 Ma). Fig. 6.3.1 shows the cumulative percentage of wells with sedimentation rates below specified values for the Winton/Mackunda Formations (96-91 Ma), the Allaru Mudstone (100-96 Ma), the Wallumbilla Formation (113-101 Ma) and also the maximum rate calculated for any formation in the Jurassic. I have neglected the Toolebuc Formation (101-100 Ma) which quite probably has too long a time span assigned to it. However, if it was included with the Wallumbilla Formation the curve would barely change, but if it is included with the Allaru Mudstone then the calculated rates drop to values typical of the Wallumbilla Formation and the maximum Jurassic rates. It is clear that the Winton/Mackunda Formations have a consistently higher rate of sedimentation than any of the other formations, although show greater variability. The area of the Winton/Mackunda Formations curve with low rates is reflecting the erosion of the upper part of the sequence as mentioned earlier.

As rates of sedimentation are sensitive to the ages assigned, I also illustrate the distribution of rates over the total Jurassic and the total Cretaceous sequences (fig. 6.3.2) The Cadna-Owie Formation always has a low rate of sedimentation, generally less than 5 m/m.y.. This may be a result of an incorrect age assignment (127-113 Ma), and the result can be expected to bias the total rate to a lower value. Fig. 6.3.2 shows the total
Fig 6.3.1 Cumulative frequency, as a percentage of wells studied, showing sedimentation rates for (i) the Jurassic (maximum value for any formation), (ii) Wallumbilla Formation (113-101 Ma), (iii) Allaru Mudstone (100-96 Ma) (iv) Winton/Mackunda Formation (96-91 Ma).

Fig 6.3.2 Distributed sedimentation rates through the Jurassic and Cretaceous sequences.
(a) The Cadna-Owie Formation (127-113 Ma) is included with the Jurassic sediments (closed symbols)
(b) The Cadna-Owie Formation is included with Cretaceous sediments (open symbols)
sedimentation rates for the Cretaceous and the Jurassic sequences for two cases (a) where the Cadna-Owie Formation is included in the Jurassic and (b) where it is included in the Cretaceous. As can be seen, for case (b) the Cretaceous rates are reduced as is the spread, but they are still a factor of 2-3 greater than the Jurassic rates, which are not particularly affected by the presence or absence of the Cadna-Owie Formation. We return to this late stage increase in subsidence and sedimentation rate later and for the moment we consider primarily the earlier part of the subsidence history.

If, as implied earlier, the same subsidence mechanism may be responsible for both the Cooper and Eromanga Basins, why then is the latter areally much more extensive than the former? I have illustrated in section 2.2.a how the flexural response of the lithosphere during subsidence can influence the geometry of the subsiding area and the resulting stratigraphy. In the early stages of basin evolution subsidence is often accommodated by movement on faults, as is the case in the Cooper Basin, and the most appropriate isostatic model will usually be one of local compensation. If a thermal influence is involved then the lithosphere may be assumed to be hot, relatively weak and unable to support the stress differences generated by the driving mechanism and the sediment load. As the thermal perturbation decays, the lithosphere cools, strengthens and tends to support loads by regional compensation. Alternatively, fault controlled subsidence may simply reflect the more dominant influence of applied stresses initially, and the subsequent phase of subsidence is primarily the result of the increasing contribution made by a thermal driving load acting over a broader area. In either case, the deflection due to the driving mechanism is distributed over a progressively greater region and the area of deposition at the surface is widened. It seems, then, on a qualitative level that the two basins may be the result of the same formation mechanisms, predominantly of thermal origin and the present day observed areal distribution of the sediments is the result of the dominant mode of isostasy (local or regional) operative during subsidence and deposition.

Ignoring for the moment what processes are likely to be responsible for a thermal perturbation, we will assume that the tectonic subsidence follows a function of the form
\[ H_b(t) = H_f \left[ 1 - \exp\left(\frac{t - t_0}{\tau}\right)\right] \]  

(6.3.1a)

where \( t_0 = \) age of initiation of thermal subsidence, \( \tau = \) thermal time constant, \( H_f = \) final depth to which the basement subsides as \( t \to \infty \). This is a similar form to that given in eqn. 2.3.10b. The time, \( t - t_0 \), for a fraction, \( F(t) = H_b(t)/H_f \), of the total subsidence, \( H_f \), to have occurred is given from eqn. 6.3.1a as

\[ t - t_0 = -\tau \ln(1-F(t)) \]  

(6.3.1b)

The general form for the decay of a thermal perturbation in a slab is a sum of Fourier components, each of which has an exponential time dependence (Carslaw and Jaeger, 1959). The solutions are usually expressed as an infinite series with different time constants for each harmonic component of the temperature field. The thermal time constant \( \tau \) controls the rate of decay and the dominant contribution is from the first order harmonic (longest wavelength). Then, as stated in section 2.3.b, the time constant is given as

\[ \tau = \frac{a^2}{\kappa \pi^2} \]

where \( a \) is the slab thickness (m) and \( \kappa \) is the thermal diffusivity (m\(^2\)s\(^{-1}\)). In tectonic problems the thermal time constant is estimated to be about 50-70 m.y. (e.g. Sleep, 1971; Parsons and Sclater, 1977), consistent with a thermal lithosphere of 125-150 km for \( \kappa = 10^{-6} \) m\(^2\)s\(^{-1}\).

In table 6.3.1 I give the ratio \( F(t) \) for different \( t - t_0 \) and \( \tau \) to illustrate how the amount of the total subsidence achieved, after a given time, varies with the thermal time constant. The results in this table imply that if a sedimentary basin has a dominantly thermal influence on its evolution, then the duration of the basin's subsidence history, or at least the first 95% of the total predicted subsidence, will be characteristic of the thermal time constant of the lithosphere. For example, a basin forming as a result of a simple thermal
mechanism on a lithosphere with a thermal time constant of 50 m.y. should span only about 150 m.y. or so, whereas a basin forming on lithosphere with a time constant of 200 m.y. will only have achieved about half of its total subsidence after an equivalent time.

<table>
<thead>
<tr>
<th>t - t₀ (m.y.)</th>
<th>τ (m.y.) 50</th>
<th>100</th>
<th>200</th>
</tr>
</thead>
<tbody>
<tr>
<td>25</td>
<td>0.39</td>
<td>0.22</td>
<td>0.12</td>
</tr>
<tr>
<td>50</td>
<td>0.63</td>
<td>0.39</td>
<td>0.22</td>
</tr>
<tr>
<td>100</td>
<td>0.87</td>
<td>0.63</td>
<td>0.39</td>
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<tr>
<td>150</td>
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<td>0.53</td>
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<tr>
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<td>0.63</td>
</tr>
<tr>
<td>250</td>
<td>0.99</td>
<td>0.92</td>
<td>0.71</td>
</tr>
</tbody>
</table>

*Table 6.3.1. Normalized subsidence, F(t) = H₀(t)/ H₀, given by eqn. 4.4.1a as a function of time for τ = 50, 100 and 200 m.y.*

Fig. 6.3.3 shows the tectonic subsidence for Mcleod 1, one of the wells with the most complete Permian to Cretaceous sequence. The theoretical curves given by eqn. 6.3.1a for τ = 50, 100, and 200 m.y. and the initial time, t₀, was taken as 290, 260 and 230 Ma. The value of t₀ = 290 Ma corresponds approximately to the time when Permian sedimentation commenced, the value of 260 Ma to the time when the sedimentation and subsidence became less fault controlled, and the value of 230 Ma is the time (Early Triassic) when an expansion of the depositional area occurred and sediments from this age are seen in the northern Cooper Basin and the Simpson Desert Basin regions. As the value of H₀ is obviously not likely to correspond to the observed tectonic subsidence (1262 m) at 91 Ma due to the increased rate of subsidence, the theoretical curves were constrained to pass through the observed tectonic subsidence (971 m) at 113 Ma (end of the Cadna-Owie Formation).
Fig 6.3.3 Observed basement subsidence for Mcleod 1, a well with one of the most complete Permian-Cretaceous sequences, and theoretical exponential subsidence curves (eqn. 6.3.1a) constrained to pass through the observed value at 113 Ma (Cadna-Owie Formation) and $t_0$ taken as (i) 290 Ma, (ii) 260 Ma, and (iii) 230 Ma.
The curves for \( t_0 = 290 \text{ Ma} \) suggest that the Permo-Triassic subsidence is consistent with a thermal time constant of slightly less than 50 m.y., whereas the Jurassic-earliest Cretaceous subsidence implies a higher value of \( \tau \), about 100-200 m.y. The period of rapid subsidence after 110 Ma deviates significantly from all of the theoretical curves. Reducing the value of \( t_0 \) makes discrimination between different values of \( \tau \) less distinct, although the data (with the error bars) still tend to suggest a value of 100-200 m.y. to be more appropriate over the Triassic to earliest Cretaceous. This may be interpreted as a time dependent thermal constant, but as the Early Permian subsidence was predominantly fault controlled it is not clear what role thermal subsidence would have had during this period.

6.3.d An average subsidence curve for the Jurassic-Cretaceous

The total Jurassic-Cretaceous sequence is represented in nearly all the wells and it is therefore this part of the subsidence history we consider here. It obvious from a cursory examination of the subsidence curves that the total amount of subsidence is variable. In an attempt to obtain a representative subsidence curve we follow the approach taken by Sleep (1971). If we assume that the tectonic subsidence follows an exponential form as given by eqn 6.3.1a, then for a given datum of age \( t_i \) the tectonic subsidence is

\[
H_b(t_i, x) = H_f(x) \left[ 1 - \exp \left( \frac{-(t_i - t_0)}{\tau} \right) \right] \quad (6.3.2a)
\]

Defining a reference horizon of age \( t_\tau \)

\[
H_b(t_\tau, x) = H_f(x) \left[ 1 - \exp \left( \frac{-(t_\tau - t_0)}{\tau} \right) \right] \quad (6.3.2b)
\]

and normalizing by \( H_b(t_\tau, x) \) allows the elimination of the term \( H_f(x) \) gives