Late Cretaceous dysoxia in a southern high latitude siliciclastic succession, the Otway Basin, southeastern Australia

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Abstract

The warm greenhouse world of the Late Cretaceous created oceans that were poorly stratified latitudinally and vertically. Periodically these oceans experienced globally significant events where oxygen minimum zones enveloped the continental margins. Evidence of the effect of one of these Ocean Anoxic Events (OAEs) is preserved in the southern high latitude strata of the offshore Otway Basin in southeast Australia. During the Late Cretaceous, thick successions (up to 6 km) of mudstone-dominated deltaic to upper bathyal sediments (the Otway Delta) were deposited in an elongate inlet (ca. 500 km wide) between Antarctica and Australia located at least 70° S. The initial Turonian strata of this succession (the Waarre Formation) were deposited in upper delta plain to delta front conditions. The overlying late Turonian Flaxman Formation and basal Belfast Mudstone Formation preserve evidence of transgressive lower delta plain to prodelta conditions at inner to middle shelf depths. These units were subject to periodic dysoxia during deposition caused by intermittent freshwater input and deepening seas resulting in periods of thermohaline stratification and reduced bottom waters. Rapid subsidence from 89.3 Ma to 85.7 Ma created significant accommodation space leading to the seaward progradation of normal marine prodelta to upper bathyal mudstone-dominated facies at middle shelf to upper slope depths. After a period when the oxygen minimum zone contracted near the base of the Coniacian, upward-increasing dysoxia in the Belfast Mudstone Formation heralds the onset of Coniacian to Santonian dysoxic conditions. This event in the Otway Basin correlates to OAE 3, the last Ocean Anoxic Event of the warm Cretaceous before the onset of cooler conditions in the uppermost Cretaceous. The evidence suggests that, rather than confined to low latitude tropical areas, the effects of OAE 3 reached southern high latitude regions during the warm Late Cretaceous. The cessation of growth faulting after 85.7 Ma reduced accommodation space and delta front to prodelta facies prograded rapidly seaward. Hyposaline conditions and higher sedimentation rates due to delta front progradation and shallowing during this time
caused the local extinction and dissolution of many of the calcareous benthic foraminiferal taxa of the Belfast Mudstone Formation.
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Keywords: Late Cretaceous; Otway Basin; Deltaic facies; Foraminifera; Turonian; Coniacian–Santonian; OAE 3

1. Introduction

The Late Cretaceous ocean and climate is generally thought to have been much warmer than at present, with minimal latitudinal and vertical oceanic stratification (Huber et al., 1995; Huber, 1998; Barrera and Johnson, 1999; Haupt and Seidov, 2001; Bice and Norris, 2002). Turonian to early Campanian Southern Ocean sea surface temperatures were estimated by Huber (1998) to have fluctuated from 20 to >28 °C and deep water temperatures to have varied from 12 to 16 °C in the Antarctic. Periodic carbon burial occurred in this greenhouse world, depositing black shales during global Ocean Anoxic Events (OAEs) (Arthur et al., 1988; Leckie et al., 2002). Three major OAEs occurred, at the Aptian–Albian boundary (OAE 1), at the Cenomanian–Turonian boundary (OAE 2), and during the Coniacian–Santonian (OAE 3) (Jenkyns and Hallam, 1980; Arthur et al., 1988; Takjika, 1999). These were times when dysoxic and anoxic conditions (in oxygen minimum zones; OMZ) enveloped continental margins (Leckie et al., 2002). Two possible mechanisms for the formation of these OAEs were proposed: 1. oceanic stagnation in confined stratified basins such as the Black Sea; and 2. increased primary productivity causing the OMZ, allowing good preservation of organic matter. Another important factor that caused the accumulation of such large amounts of organic carbon in sediments was an increased input of terrestrially derived organic material.

During the warm Late Cretaceous the southern margin of Australia lay at 70° S (Norvick and Smith, 2001) in an oceanic inlet nearly 500 km wide open to the west but closed at its eastern end (Fig. 1). A thick succession (up to 5 km) of Turonian to Maastrichtian marine deltaic, shelfal to upper bathyal sediments were deposited in the region with two large delta systems, the Hammerhead Delta supersequence (Krassay and Totterdell, 2003) that devel-

![Santonian Palaeogeography ca. 85 ma](image-url)
oped in the Great Australian Bight, and the Sherbrook Group of the Otway Delta in the Otway region (Fig. 1). Over the last 25 years active hydrocarbon exploration in the Late Cretaceous Sherbrook Group in the Otway Basin has resulted in the discovery of economic gas reserves in the Turonian to Santonian intervals (Cliff et al., 2004). As a result, there exists a large amount of seismic and well data from the Late Cretaceous succession in the Otway Basin. This makes the Otway Basin one of the best Late Cretaceous marine sections along the southern margin of Australia.

In this paper we examine a >2 km thick subsurface Turonian to Santonian siliciclastic succession (Fig. 2) in the southern high latitude Otway Basin with the following aims: (1) to provide a detailed seismic and lithostratigraphic framework for the succession, this will then be; (2) combined with detailed foraminiferal, facies, rock eval pyrolysis analyses and preliminary geochemical data to interpret the palaeoenvironment of these strata; (3) these data will then be integrated to construct a facies model for the Otway Late Cretaceous succession, this is important since these strata are currently prospective for hydrocarbons (Cliff et al., 2004) and the mudstone-rich facies forms a regional seal on existing gas deposits (Krassay et al., 2004); and (4) to identify evidence for marine dysoxia related to the expansion of the oxygen minimum zone during the Coniacian–Santonian in southern high latitude strata, directly correlatable with OAE 3, the last OAE of the warm Late Cretaceous.

2. Methods

The data for this study come from a shelfal area in the offshore Otway Basin. Four wells are included in our study: Normanby-1, Discovery Bay-1, Voluta-1

Fig. 2. The Upper Cretaceous stratigraphy of the Otway Basin. The stratigraphy is modified from Boyd and Gallagher (2001), Partridge (2001) and Gallagher et al. (2004). The palynological zonal scheme is that of Schiøler et al. (2002) modified from Helby et al. (1987). The chronology of the stages and zones is that of Gradstein and Ogg (2004), Ogg (2004) and Gradstein et al., 2005. Note: the question marks in the Turonian interval indicate that the lower age limit of these units is unconstrained within this stage.
and Bridgewater Bay-1 (Figs. 3 and 4). The seismic lines that intersect these wells were obtained from Geoscience Victoria, Melbourne (Fig. 5) and Geoscience Australia, Canberra (Figs. 6 and 7). Wiltshire Geological Services supplied the gamma and velocity log data for Fig. 4. The primary palynological bio-

Fig. 3. The location of wells on Fig. 4 and the seismic lines on Figs. 5, 6 and 7: (A) An isopach map of the Sherbrook Group adapted from Megallaa (1986). (B) A Two Way Travel Time (TWT) thickness map of the Sherbrook Group modified from Constantine (2001).
stratigraphic data for each well were obtained from the STRATDAT Database of Geoscience Australia. The core, ditch cuttings samples and sidewall core samples were obtained from the sample library of Geoscience Victoria, Melbourne. The facies analyses of Voluta-1 were carried out by analyses of thin sections of cores, describing ditch cutting and sidewall core samples by texture, and analysing the microfossil samples, using a binocular microscope. Approximately 35 m of core, 10 sidewall cores and 88 ditch cuttings were logged (Figs. 8–10). The facies data are supplemented with biomarker geochemical data from twelve samples in Voluta-1 (Fig. 11A) and Rock-Eval pyrolysis data from twenty-four samples in Voluta-1 and from Constantine et al. (2001) where TOC (Total Organic Carbon), HI (hydrogen index) and OI (oxygen index) data were obtained for each well.

Macrofauna was obtained and identified from cores of the Belfast Mudstone Formation in Voluta-1 (Fig. 10, Table 1). In addition, the occurrence of macrofossils such as inoceramid bivalves was documented in the ditch cutting microfossil residues.

Eighty one ditch cutting (DC), twelve core and one sidewall core samples were processed for foraminifera in Voluta-1. Over 12,000 foraminifera were counted in the >63 μm fraction of these samples. It was not always possible to obtain >100 specimens per sample (especially in the Flaxman Formation, see below); nevertheless, the faunal data from such samples show useful trends. The taxonomy used in this study follows that outlined in Taylor (1964, 1965), Mello (1971), Bolli et al. (1994) and Revets (2001). A list of the taxa identified is included in Tables 2 and 3. The foraminiferal data are plotted quantitatively for Voluta-1 (Figs. 11B and 12) and semi-quantitatively for Bridgewater Bay-1 (Fig. 13A). The foraminiferal data in Voluta-1 are expressed as a percentage of the total foraminiferal fauna, the
Fig. 5. Uninterpreted and interpreted seismic line C–C’ through Discovery Bay-1 and Normanby-1; note: the depths are in milliseconds.
agglutinated fauna and the benthic rotaliid fauna. The diversity indices used include the Fisher alpha index (Murray, 1991) and the Shannon–Weaver index \( H(S) \) (Shannon and Weaver, 1949; Murray, 1991). This index takes into account the number of taxa in a sample and their equitability of distribution within

Fig. 6. Uninterpreted and interpreted seismic line A–A’ through Voluta-1; note: the depths are in milliseconds.
Fig. 7. Uninterpreted and interpreted seismic line B–B’ through Bridgewater Bay-1; note: the depths are in milliseconds.
that sample. The Shannon–Weaver index is calculated as follows:

$$H(S) = - \sum_{i=1}^{S} pi \ln pi$$

Where $H(S)$ = index of species diversity, $S$ = the number of species, $pi$ = proportion of total samples belonging to the $i$th species.

This study follows diversity classification of Holbourn et al. (2001) and assigns $H(S)$ values from 0 to $<1$ as low diversity assemblages, 1 to $\leq 2$ as moderate diversity assemblages and $>2$ as high diversity assemblages. Most samples in the Flaxman Formation yielded less than 50 specimens, the diversity calculations of these samples are included only for comparative purposes. The diversity analyses is most effective in the overlying Belfast Mudstone Formation where only 6 samples yielded less than 50 specimens (Fig. 11B). The oxia/dysoxia classification of Bernhard (1986) is followed in this work, where anoxic (anaerobic) conditions reflect dissolved oxygen contents ranged from 0 to 0.1 ml/l O$_2$, dysoxic (dyoxia or dysaerobic) conditions when dissolved oxygen values ranged from $>0.1$ to 0.5 ml/l O$_2$ and oxic (oxia or aerobic) conditions when values were greater than 0.5 ml/l O$_2$.

3. Stratigraphic framework

The stratigraphic correlation of the four wells (Fig. 4) is based primarily on wireline log attributes, and palynological data from well completion reports using the palynological zonal scheme of Fig. 2. The palynological zonal scheme is that of Schiøler et al. (2002) modified from Helby et al. (1987). The chronology of the stages and zones is that of Gradstein et al. (2005) adapted from Hardenbol et al. (1998). The palynological data are modified from Geoscience Australia’s STRATDAT database and from the Department of Natural Resources and Environment (1999) Victorian Petroleum Otway Biostratigraphic Reports CD and Partridge (2000). It is clear from the biostratigraphic data that there is poor age resolution in the Turonian Waarre/Flaxman Formations (Fig. 4). In addition, the palynological age assignment of the Belfast Formation in Voluta-1 is modified here, based on the first
occurrence of the inoceramid bivalve *Cremnoceramus bicorrugatus bicorrugatus*? (Marwick, 1926) in core 13 and 16 (see Fig. 10, Table 1). This is an index fossil for the New Zealand Cretaceous, confined to the uppermost Arowhanan to lowermost Teratan (late Turonian–early Coniacian) stages (see Crampton, 1996; Crampton et al., 2001). The genus *Cremnoceramus* apparently is restricted mostly to the lower and middle Coniacian in the Northern Hemisphere (Walaszczyk, 1992; Kauffman et al., 1993), but evidence does indicate an age extension into the uppermost Turonian in some regions (Kennedy et al., 1989; Zakharov et al., 1989).

The Waarre and Flaxman Formations have Turonian *P. infusorioides* zonal ages in Normanby-1 and Bridgewater Bay-1. The best studied section is in Normanby-1 where gas shows were recorded in three upward-fining sand to mudstone units (Lavin, 1998). The boundary between the Flaxman and Waarre Formations has been taken at a gamma and sonic log break in both wells, marking the transition to the slightly more mudstone-rich Flaxman Formation. The upper boundary of the Flaxman Formation is taken at a gamma and sonic log break denoting the uppermost occurrence of sand preceding the mudstone-dominated prodelta Belfast Mudstone Formation. The Turonian (*P. infusorioides*) to Santonian (*I. cretaceum*) Belfast Mudstone Formation is a relatively homogeneous mudstone-dominated unit that thickens to over 1 km in Voluta-1 and Bridgewater Bay-1. The chronology of the Belfast Mudstone Formation is shown on Fig. 4. The Turonian/Coniacian boundary (89.3 ± 1.0 Ma) is in the lower part (biofacies A). The *O. porifera* zone occurs in biofacies C. The base of
Fig. 10. Core logs of the Belfast Mudstone Formation showing the facies and macrofaunal distribution. For key to symbols see Fig. 8.
Fig. 11. (A) The biomarker distribution and Rock Eval pyrolysis data in Voluta-1. Note VR% data is >0.75% below 3150 m. (B) The distribution of foraminifera and microplankton diversity in Voluta-1. The % agglutinated, % rotaliids, % porcellaneous (porcell.) and % plankton are expressed as a percentage of the total fauna.
this zone has been dated as 85.7 Ma by Schiøler et al. (2002) lying close to the Coniacian/Santonian boundary. The I. cretaceum zone approximates the middle Santonian (Schiøler et al., 2002) and therefore the top of the Belfast Mudstone Formation biofacies D and the base of the sandstone/mudstone-dominated Paaratte Formation were deposited diachronously from 84.88 to 84.6 Ma. After 83.5 $F_0.7$ Ma the Paaratte Formation and Timboon Sandstone become increasingly sandstone-dominated up section with coarse sandstone and conglomerate in the uppermost strata. These Campanian to Maastrichtian units vary greatly in thickness, reflecting a complex of delta front to delta plain progradational depositional settings.

### 4. Depositional geometry of the Upper Cretaceous succession

The locations of the four wells studied and the seismic lines interpreted are shown on Fig. 3. The interpreted isopach thicknesses of the Sherbrook Group sediments of Megallaa (1986) and the TWT (two way time) isochron data of Constantine (2001) show that the principal depocentre with sediment thickness up to 6 km, $>2.75$ s occurs to the southeast of the studied area. Another depocentre can be identified southwest of Discovery Bay-1 in the Constantine (2001) data (Fig. 3). In general the Sherbrook Group thickness is $<3$ km ($<2$ s) in the study area, where northwest-southeast trending fault-controlled depocentres are best developed (Fig. 3). Normanby-1 and Discovery Bay-1 are situated on structural highs on the edge of elongate depocentres. Voluta-1 and Discovery Bay-1 are situated in northwest-southeast fault controlled depocentres where TWT thicknesses of up to 2 s are reached.

The locations of two dip seismic sections and one strike section that intersect the studied wells are shown on Fig. 3. The interpreted seismic lines are shown on Figs. 5, 6 and 7. Well logs were converted to TWT using sonic velocity data, and placed on the seismic lines to facilitate correlation. Complex normal faults dominate the northerly dip section (Fig. 5). The slightly less structurally complex strike section shows a series of growth faults and later inversion features around Voluta-1 (Fig. 6). These growth faults are also evident in the southerly dip section (Fig. 7). The Waarre and Flaxman Formations were mapped together since they are indistinguishable seismically. The units vary in thickness from 0.1 s to 0.5 s with at least four easily distinguished strong reflectors. The Belfast Mudstone Formation shows marked lateral thickness variations from 0.25 s to $>1$ s. The unit thickens into depocentre-bounding growth faults around Voluta-1 and Bridgewater Bay-1 (marked (i) and (ii) on Figs. 6 and 7). Transparent to poorly defined seismic reflectors characterise the Belfast Mudstone Formation; some downlapping is evident in the most southerly dip section (Fig. 6).

### 5. Turonian to Santonian facies variation

Facies analysis of the core, sidewall cores and ditch cuttings was carried out on Voluta-1. All units except the Flaxman Formation were sampled by core or sidewall core in this well. The Waarre Formation was not...
A list of the agglutinated foraminifers in the Turonian to Santonian of the Otway Basin

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<th>Palaeobathymetry</th>
<th>Textulariids</th>
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Key to symbols

- **Palaeodepth**
  - **MI** Marginal marine
  - **I** Inner shelf (0–50 m)
  - **M** Middle shelf (50–100 m)
  - **O** Outer shelf (100–200 m)
  - **ub** Upper bathyal (200–1000 m)
  - **mb** Middle bathyal (1000–2000 m)
  - **lb** Lower bathyal (>2000 m)
- **Abundance**
  - **1** Rare
  - **2** Common
  - **3** Abundant

The palaeobathymetric distribution is adapted from Murray (1991), van den Akker et al. (2001) and Nagy (1992) and Nagy et al. (1995).
reached in Voluta-1. It was however, intersected in Normanby-1 where interbeds of fine-medium quartz sandstone and carbonaceous siltstone are observed in the sidewall cores (Lavin, 1998). No facies analyses were carried out on Bridgewater Bay-1 or Normanby-1 wells, therefore a detailed examination of Voluta-1 is used as an example of the Turonian Flaxman Formation and Turonian to Santonian Belfast Mudstone Formation facies. These data will be combined with seismic stratigraphic, rock eval pyrolysis, organic geochemical and foraminiferal data to arrive at a depositional model for the succession (see later).

A significant sand fraction (from 3–15%) occurs in the dark grey mudstone-dominated facies of the Flaxman Formation, and is represented by very fine loose subangular to subrounded sand grains and as fine- to medium-grained white laminated and cross-laminated sandstone. Rare granule sized grains up to 3 mm in diameter are observed in the lower 100 m. The A to C biofacies subdivision of the Flaxman Formation is based principally on microfaunal data, wireline character and facies. Glauconite is present in most samples, but it is absent in biofacies B (Fig. 9). Inoceramid bivalve prisms and other molluscan debris occur in two intervals from 3917 m to 3871 m and in the upper 80 m of the unit. Occasional woody carbonaceous material occurs in the lower 100 m of the section. The upper 50 m is a dark grey mudstone-dominated unit.

The A to D biofacies subdivision of the Belfast Mudstone Formation is based principally on microfaunal data (see later), the relative abundance of macrofossils, and to a lesser extent, wireline character and facies. The top of biofacies B (i.e. the B/C boundary) in the Belfast Mudstone Formation at log level 2856 m is the disappearance of common inoceramid bivalve prisms in ditch cuttings. These reappear and last occur in biofacies C in core 13 (Fig. 10) in Voluta-1. Biofacies A to C of the Belfast Formation are similar lithologically and consist of dark grey mudstone with a minor (1–5%) fine white sandstone and fine sand component which include rare subrounded to rounded quartz grains from 0.5 mm to 4 mm sized granules. Glauconite is moderately common in biofacies A and B and decreases in abundance towards the top of biofacies C. Lamination is the main sedimentary structure present in the ditch cuttings, and bioturbation is more prevalent in biofacies A compared to biofacies B and C. In biofacies A and B intervals of inoceramid prisms are common with occasional serpulid worm and gastropod bioclasts in the ditch cuttings. Cores 21 to 16 represent biofacies A and B (Fig. 10). Homogeneous mudstone at the base (cores 21, 18 and 17) alternates with dark grey mudstone to siltstone with faint laminae and bioturbation (cores 20 and 16). Macrofossils are common in core 16 with rare irregular echinoderms and common inoceramids (see faunal list, Table 1). Macrofauna is sparse in the dark grey mudstone-dominated ditch cuttings of biofacies C in the Belfast Mudstone Formation. Occasional molluscs such as gastropods and ammonites occur in biofacies C of cores 15 and 13 (Fig. 10). Core 15 consists of faintly bioturbated dark grey siltstone to sandstone. In contrast to most of the ditch cuttings of biofacies C, core 13 is quite fossiliferous, with selactarian corals, nautiloids, ammonites, gastropods and bivalves (Cremnoceramus, see Table 1). The facies of core 13 is a dark grey siltstone to fine sandstone with occasional bioturbated and ripple horizons. The lower 50 m of biofacies D in the Belfast Mudstone Formation is lithologically very similar to biofacies C (the C/D boundary is based on microfaunal criteria). An increase in log gamma values associated with an upwards-increasing influx of 20–50% fine sand occurs in dark grey mudstone of the upper 30 m of biofacies D. The lithology of core 12 is characteristic of this upper unit, with laminated, cross-laminated and bioturbated siltstone with fine sandstone interbeds (Fig. 10). Rare carbonised woody material occurs in biofacies D.

6. Rock-Eval Pyrolysis variability

The Hydrogen Index (HI, mg HC/g TOC) versus Oxygen Index (OI, mg CO2/g TOC) data for Turonian–Santonian intervals in Normanby-1, Voluta-1 and Bridgewater Bay-1 (from Constantine et al., 2001) are illustrated on two van Krevelen-type diagrams (Fig. 14). The Turonian of Voluta-1 is characterised by Type III/IV kerogen and contains terrestrial and/or reworked organic material (Peters, 1986). However, the same interval in Normanby-1 shows a slight increase in HI that might relate to a slightly more algal-derived organic matter (i.e. a possible increase in the marine-derived organic input).
Table 3
A list of the rotaliid and porcellaneous foraminifers in the Turonian to Santonian of the Otway Basin

<table>
<thead>
<tr>
<th>Palaeobathymetry</th>
<th>Benthic rotaliids</th>
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<td>1 1 1 1</td>
<td>Hanzawaia</td>
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<td>1 1 1 1</td>
<td>Nodosaria</td>
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<td>1 1 1 1</td>
<td>Dentalina</td>
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<td>Frondicularia</td>
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<td>1 2 3 2</td>
<td>Alabamina</td>
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<td>1 3 3 2 1</td>
<td>Cibicides</td>
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<td>Stensioecina</td>
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<td>Gavelinella</td>
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<td>1 1 2 3 2 1</td>
<td>Hemirotubulinina (Marginulina)</td>
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<td>Marginulinopsis</td>
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<td>Marginulinopsis (Astacolus)</td>
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<td>Citharina</td>
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<td>Gyroidinoides</td>
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<td>Globulina</td>
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<td>3 2</td>
<td>Allomorphina</td>
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<td>Praebulimina</td>
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bulk of the organic matter in the Coniacian–Santonian interval also contains Type III/IV kerogen.

The TOC and HI data for the Sherbrook Group are tabulated for each well (Figs. 11A and 13), while the classification scheme of Wagner and Pletsch (1999) is used to describe organic richness:

- Low organic carbon content = 0 to <0.5% TOC
- Average organic carbon = 0.5 to <1% TOC
- High organic carbon = 1 to <2% TOC
- Very high organic carbon = ≥2% TOC

TOC values in the Waarre and Flaxman formations vary markedly from high to very high, leading to a strongly serrated TOC profile. Peaks of up to 60% TOC recorded the coals in the Waarre Formation, although on average the values are less than 14%. TOC values are also high in the Belfast Mudstone Formation (up to ~15%), although the values are more homogeneous than the underlying units. TOC values increase upwards to very high values in the Turonian interval in the Belfast Mudstone Formation in Normanby-1. An interval of average TOC occurs at the Turonian–Coniacian boundary in Normanby-1 and Voluta-1. However, some of the relatively low TOC values in Voluta-1 below log level 3150 m are possibly lowered due to hydrocarbon generation and expulsion, since all samples below this level have VR values greater than 0.75% (Fig. 11A). Occasional peaks in TOC values greater than 3% occur in the Belfast Mudstone Formation (Figs. 11A and 13), typically in biofacies B and C.

7. Preliminary biomarker analyses results

Gas chromatography (gc) and gas chromatography-mass spectrometry (gcms) were undertaken on bitumen extracts from 12 samples of core and
Fig. 12. (A) The distribution of agglutinated foraminifera in Voluta-1 expressed as percentage of the total agglutinated fauna. (B) The distribution of benthic rotaliids in Voluta-1 expressed as percentage of the benthic rotaliid fauna.
Fig. 13. (A) The distribution of foraminifera, microplankton diversity and Rock Eval pyrolysis data in Bridgewater Bay-1. (B) The distribution of microplankton diversity and Rock Eval pyrolysis data in Normanby-1.
cuttings from 2465 to 3962 m in Voluta-1 (Fig. 11A) in order to obtain an independent preliminary assessment of the organic facies (marine, terrestrial and bacterial inputs) and depositional environment. The \( n \)-alkane distribution is dominated by low molecular weight \( n \)-alkanes (\( n-C_{17}/n-C_{27}>1 \); Fig. 11A, GC ratio) consistent with a high input of marine organic matter to the saturated hydrocarbons. Further support for a strong marine influence comes from the low \( C_{19} \) tricyclic/\( C_{23} \) tricyclic ratio and the \( C_{24} \) tetracyclic/\( C_{23} \) tricyclic ratios (\( <1 \); Fig. 11A) and the dominance of \( C_{27} \) steranes over \( C_{29} \) steranes (\( >0.5 \); Fig. 11A), together with the occurrence of the marine-specific \( C_{30} \) desmethylsterane.

Fig. 14. A van Krevelen type diagram showing the Rock Eval pyrolysis data for the section. The fields of Types I, II and III–IV kerogen are indicated (cf. Peters, 1986).
(Peters et al., 1993). The intervals of elevated C19 tricyclic/C23 tricyclic ratio and the C24 tetracyclic/C23 tricyclic ratios may correspond to an increased terrestrial influx into the marine depositional environment or siliciclastics. The lack of a concomitant decrease in the n-C17/n-C27 ratio suggests that the latter process is the most likely cause. The pristane-to-phytane ratios (Pr/Ph; Fig. 11A, GC ratio) vary between 1.6 to 4.5, indicating a range of depositional conditions from dysoxic (1<Pr/Ph<3) to oxic (Pr/Ph>3). There are dramatic changes in the Pr/Ph ratio over the 1500 m sampling interval (Fig. 11A), which may indicate a fluctuating oxygenated water column throughout the Flaxman Formation and Belfast Mudstone Formation time. However, the Pr/Ph ratio is expected to show a response to maturity, since over the same interval the vitrinite reflectance increases from approximately 0.6–1.3% (i.e. it defines the oil window), Boudou (1984) has shown that with maturity the Pr/Ph ratio tends to maximise at the beginning of the oil window (77% TOC for coal (daf) or corresponding to 0.6 % VR; Fig. 8-38 in Hunt, 1996) decreasing thereafter, suggesting that maturity is not the dominant factor controlling the downhole variation in Pr/Ph. Furthermore, there is not a strong correlation between Pr/Ph and the source-dependent biomarkers in Fig. 11A (e.g. the low Pr/Ph at 2798.1 m is not accompanied by an increase in marine inputs while higher terrestrial inputs at 3285.8 m is not associated with high Pr/Ph), which suggests that redox conditions are the major control on Pr/Ph. Hence, stronger dysoxic conditions are evident during deposition of the Belfast Mudstone Formation biofacies C where Pr/Ph is at its lowest (2798.1 m; Fig. 11A). There is no evidence of anoxic conditions (Pr/Ph<1; Peters et al., 1993) in the 12 samples analysed for biomarkers in Voluta-1.

8. Variation in microplankton (dinoflagellate) distribution

The distribution of microplankton (dinoflagellates) in the Sherbrook Group has been tabulated for each well (Figs. 11B and 13). The purpose of this part of the study is to assess the relative marine influence in the strata in the absence of other diagnostic macrofossils and microfossils (such as foraminifera, see below). These data have been extracted from the Department of Natural Resources and Environment (1999) Victorian Petroleum Otway Biostratigraphic Reports CD. The relative abundance of microplankton in Voluta-1 (Dettmann, 1968) is here considered to be an underestimation since later work by Partridge (2000) has yielded >10 species of microplankton in several samples in the lower 1000 m of the well, whereas Dettmann (1968) recognised less than 3; nevertheless the Dettmann data yield useful trends above 2700 m. Microplankton is rare at the base of the Waarre Formation in Normanby-1 but the assemblage diversity increases upwards into the upper part of the Waarre and Flaxman Formations. The Belfast Mudstone Formation yields relatively diverse microplankton assemblages of up to twelve species, although the diversity fluctuates markedly, with some intervals yielding fewer than three species.

9. Turonian–Santonian foraminiferal distribution

Foraminiferal analyses were completed for the Sherbrook Group in Voluta-1 and Bridgewater Bay-1 (Figs. 11B, 12 and 13A). Several biofacies units are distinguished in the Belfast Mudstone Formation (A to D) and the Flaxman Formation (A to C). There are two main types of data represented on Figs. 11B, 12 and 13A: core and sidewall core data, and ditch cutting data. The yield of foraminifera is significantly less in the core samples, than in the ditch cutting samples in Voluta-1. This is directly related to the averaging effect of cutting collection during drilling, since each cutting represents at least a 3 m interval. However, the overall trends of diversity and faunal distribution between these sample types are very similar, and the following results are based primarily on the foraminiferal assemblages in ditch cuttings, since these represent all units in Voluta-1. A list of the foraminifera identified is given in Tables 2 and 3. The results of the analyses are presented in three parts: 1. Foraminiferal diversity and faunal trends; 2. Patterns in agglutinated foraminiferal distribution; and 3. Benthic rotaliid assemblage variability.
9.1. Foraminiferal diversity and faunal trends

9.1.1. Turonian Waarre and Flaxman Formation

Foraminifera are absent from the ditch cutting samples of the Turonian Waarre and Flaxman Formations and in the lower 200 m of the Belfast Mudstone Formation in Bridgewater Bay-1, where microplankton is relatively diverse. The Flaxman Formation in Voluta-1 yields a sparse (multiple tray picking yielded <50 specimens) moderate to highly diverse assemblage of foraminifera. This assemblage also characterises an interval with comparatively high HI. This variability yields a serrated profile of faunal diversity that is directly comparable and correlatable to the HI and TOC variability (Fig. 11). Rotaliids and miliolids are most common in biofacies A and C compared to biofacies B of the Flaxman Formation. Planktonic foraminifera are rare in biofacies A and B; they increase in percentage in biofacies C.

9.1.2. Turonian–Santonian Belfast Mudstone Formation

Although diverse microplankton/dinoflagellate assemblages are present in the Turonian Belfast Formation in Bridgewater Bay-1 foraminifera are absent. A diverse agglutinated-dominated foraminiferal assemblage occurs in the Turonian to Coniacian of the Belfast Mudstone Formation in Voluta-1. Intervals of moderate diversity occur within biofacies A, B and C, usually coinciding with a reduction in the percentage of rota
diids (typically 30–35% of the total fauna) to less than 20%. These intervals typically occur around 100 m below high-gamma spikes in the wireline log data. Foraminifer
diversity and the abundance of rota
diids decrease significantly into biofacies D of the Belfast Mudstone Formation (rota
diids are absent from this unit in Bridgewater Bay-1). Planktonic foraminifera are relatively uncommon, typically forming 1–10% of the fauna. However, abundance peaks (10%) occur at the A/B and B/C biofacies boundaries in Bridge
gate Bay-1 and Voluta-1, associated with relatively sparse foraminiferal yields (<50 specimens) with lower diversity assemblages. Other less prominent plankton peaks occur within units B and C. The planktonic fauna has a very low diversity, consisting of two taxa: Hedbergella spp. (specimens were poorly preserved and hard to speciate) and Hetero
helix reussi, with the addition of minor Whiteinella baltica in biofacies B in Bridgewater Bay-1. Typically, these peaks are associated with higher gamma values and high rota
diid abundance in the samples; however, planktonic foraminifera are rare or absent in the high gamma interval in biofacies D. Porcel
naceous foraminifera are a minor yet important component in several horizons yielding relatively low TOC values.

9.2. Patterns in agglutinated foraminiferal distribution

Haplophragmoides is the most common aggluti
nated genus in the succession. Two species dominate the assemblage: the keeled planispiral form Haplophragmoides excavatus Cushman and Waters and the rounded planispiral form Haplophragmoides topagor
tekensis Tappan.

9.2.1. Turonian Flaxman Formation

In Voluta-1 abundance peaks of calcareous agglu
tinated taxa such as Dorothia and Spiroplectammina alternate in this unit with peaks of Haplophrag
moides. The ratio of keeled to rounded Haplophrag
moides species also shows a marked variation with >60% rounded Haplophragmoides in the calcareous agglutinated-rich horizons (especially at the top of biofacies A and in biofacies B). Trochammina spp. dominate biofacies A and decrease in abundance up section. Peaks of Ammobaculites occur in biofacies B and C.

9.2.2. Turonian–Santonian Belfast Mudstone Formation

The proportion of Haplophragmoides spp. in the agglutinated assemblage is relatively constant in this unit in Voluta-1 (around 60%) although this proportion reaches >80% in Biofacies D. Rounded forms (H. topagorukensis) dominate in Voluta-1 and Bridge
gate Bay-1, but are replaced gradually upward by keeled forms in biofacies D. Some peaks of keeled forms occur near the base of biofacies A and in biofacies C and D. The relative abundance peaks in the calcareous agglutinated forms are similar to the rota
diid abundance. This group increases in abundance upward to the base of Biofacies C and thereafter
decreases. Ammobaculites and Hyperammina/Bathy-
siphon are relatively minor, though constant, elements
of the assemblage in Voluta-1 and Bridgewater Bay-1;
these decrease in abundance into biofacies D. Trom-
chammina spp. is relatively rare in the Belfast Mud-
stone Formation, decreasing upward in abundance in
each biofacies in Voluta-1.

9.3. Benthic rotaliid assemblage variability

9.3.1. Turonian Flaxman Formation
As described above the foraminiferal yield from
ditch cuttings in the Flaxman Formation in Voluta-1
was relatively low. The benthic rotaliid fauna is
sparse, yielding 2–8 species; this accounting in part
for the apparent large fluctuations in abundance of
various taxa on Fig. 12B. Nevertheless, trends in the
distribution of family groups are apparent in these
data. Gyroidinoides is rare in biofacies A and
increases in abundance into biofacies B and C. The
nodosariids (Pseudonodosaria, Nodosaria and Den-
talina) are relatively minor elements of biofacies A
and B, increasing in relative abundance in biofacies C.
Sporadic Cibicides spp. are present in biofacies B and
C. Vaginulinids such as Lenticulina, Saracenaria,
Marginulopsis (Astacolus), Hemirobulina (Margin-
ulina) are most common near the top of biofacies A
and C. Hanzawaia is relatively common in biofacies A
and B and dominates the benthic rotaliid assemblage at the Flaxman
Formation/Belfast Mudstone Formation boundary.

9.3.2. Turonian–Santonian Belfast Mudstone
Formation
A diverse benthic rotaliid assemblage is present in
the Belfast Mudstone Formation. The vaginulinids
(especially Lenticulina) dominate the benthic rotaliid
fauna, although this family is present in relatively low
abundance near the biofacies boundaries. Gyroidi-
noides and Cibicides are also relatively common in
the Belfast Mudstone Formation. These taxa increase
in abundance up section in biofacies C and D. Nodos-
sariids are common components of the rotaliid assem-
blage and are most common in biofacies C. This
biofacies is also characterised by peaks in the abun-
dance of Buliminella ovulum and Praebuliminula spp.

These taxa are present in sidewall core in Bridgewater
Bay-1 near the top of biofacies C. The gavelinellids
decrease in abundance up-section and disappear with-
in biofacies C. The distribution of Gavelinella and
Hanzawaia have an inverse relationship, peaks in
Hanzawaia coinciding with reduction in abundance of
Gavelinella especially at the base of the Belfast
Mudstone Formation and the biofacies A/B boundary.
This trend is less clear at the biofacies B/C boundary,
although the only two occurrences of these taxa in
Bridgewater Bay-1 are at this boundary. Biofacies D
has a lower diversity benthic rotaliid assemblage in
Voluta-1 and Bridgewater Bay-1 than in the underly-
ing units. Gavelinella and the lagenids increase in
abundance in this biofacies, where all other families
are less common or absent.

10. Discussion
The multidisciplinary analyses described in the
preceding sections are integrated to arrive at a depo-
sitional model and palaeoenvironmental history for
the Turonian–Santonian strata of the Otway Basin
(Fig. 15). The Turonian–Santonian interval of the
Sherbrook Group was deposited in a variety of envir-
onments ranging from flood plain, upper to lower
delta plain, delta front, prodelta and outer shelf to
upper slope environments (Fig. 15). Seismic data
and TWT isochron data (Constantine, 2001) suggest
that two principal depocentres received these sedi-
mants, to the southeast and southwest of the locations
studied. Although commonly disrupted by post Cre-
taceous faulting and inversion, the major stratigraphic
units show variable thicknesses related to delta depo-
sition. For example, the Belfast Mudstone Formation
thickens significantly seaward, downlapping onto the
Flaxman Formation south of Bridgewater Bay-1, sug-
jecting that the unit prograded to the south. Local
elongate sub-basins controlled by growth faulting
during Coniacian times from 89.3 ± 1.0 to 85.7 Ma
causedit deposition of thick (1 km) Belfast Mud-
stone Formation prodelta-outer shelf deposits around
Voluta-1 and Bridgewater Bay-1, leading to high sed-
imentation rates of up to 250 m per million years
(uncompacted linear sedimentation rates, Fig. 15B).
Such high sedimentation rates and thick mudstone
(shale) facies are typical of prodelta environments
Sedimentation rates were significantly lower landward of these growth faults from 89.3 to 84.88 Ma near Normanby-1 with values around 40 m/myrs (Fig. 15B). Similar large-scale growth faulting has been described by Winker and Edwards (1983) and Cohen and McClay (1996) close to the shelf edge in shale-dominated prodelta to outer shelf deposits. Sedimentation rates decreased to around 125 m/myrs after the cessation of the growth faulting during the lower Santonian from 85.7 to 84.88 Ma. Rapid progradation of the delta front to lower delta plain facies of Paarratte Formation and biofacies D after 84.88 Ma lead to the high sedimentation rates from 750 to 1200 m/myrs (Fig. 15).

The relatively high sedimentation rates estimated for the Belfast Mudstone Formation partly account for the low percentage of planktonic foraminifera; hence the low values recorded are not interpreted as directly related to palaeodepth (sensu Murray, 1991). The abundance of predominantly terrestrial-derived organic carbon in the Coniacian interval in Voluta-1 may be the result of the high sedimentation rates. The dominant foraminifera present in the succession is *Haplophragmoides* spp., agglutinated taxa typical of Mesozoic deltaic palaeoenvironments (Nagy, 1992), and found in marginal marine to upper bathyal depths today (Murray, 1991). Most of the agglutinated taxa in the succession such as *Spiroplectammina*, *Dorothyia* and *Ammobaculites*, are interpreted by van den Akker et al. (2001) and Nagy et al. (1995) as having broad environmental distributions from marginal marine to bathyal palaeodepths during the Mesozoic and Cenozoic.

### 10.1. Turonian transition to a prodelta environment

The absence of foraminifera in the Waarre Formation suggests a shallow marine, or even terrestrial environment. However, the presence of microplankton/dinoflagellates in this unit in Normanby-1 and Bridgewater Bay-1 suggests a marine influence. The variable mudstone/medium-grained sandstone facies, with carbonaceous woody material in this unit, suggests shallow marine deposition close to the shoreline. In the context of the seismic and facies evidence we concur with the interpretation of Buffin and Devine (1989) and Geary et al. (2001) that the Waarre Formation was deposited in a delta plain to delta-front environment (Fig. 15B). The lack of foraminifera is unexpected in this marine-influenced unit. Two factors may account for their absence: (1) a taphonomic effect that caused test dissolution; however, a sparse organic-walled agglutinated assemblage would be expected to remain after such dissolution; or more likely (2) the marine environment was too stressed to support a foraminiferal population. Such a stress may have been sea bed dysoxia in the delta front intervals.

The overlying mudstone-dominated facies of the Flaxman Formation preserves evidence of post-Waarre Formation deepening, with the first occurrence of foraminifera, common microplankton/dinoflagellates, glauconite and macrofauna such as the inoceramid bivalves. The relative rarity of marine microfauna in biofacies A and B, together with the presence of woody carbonaceous debris and a fine sandstone component, suggest that these units represent lower delta plain to shallow marine delta front deposition. This is further supported by the dominance of *Haplophragmoides excavatus* and *Trochammina* spp. in biofacies A. Further deepening to a lower energy delta front to prodelta environment is suggested for Biofacies C with the increase in plankton and decreasing sand component (Fig. 15B). The relatively sparse fauna in this biofacies as a whole makes environmental determinations quite difficult, although a few conclusions may be drawn from the fauna. For example the close correlation of relatively high foraminiferal diversity and TOC...
values and the marked variation of these values could be related to fluctuations in benthic organic flux and productivity related to dysoxic events. The environmental stress created by such variability may also account for the low relative foraminiferal yield in this unit, and the lack of foraminifera in Bridgewater Bay-1, even when the palynological evidence shows a strong marine influence. In addition, the limited pristane/phytane biomarker data in this interval suggest dysoxic conditions may have been prevalent during Flaxman Formation deposition in Voluta-1. Most of the benthic rotaliid taxa present in this unit, such as the nodosariids, are interpreted to have had a broad environment tolerance in the Cretaceous, from open shelf (Sikora et al., 1999), from 10 m to around 100 m palaeodepths in New Jersey (Olsson and Nyong, 1984), outer neritic (Carillo et al., 1996) and upper bathyal depths from 150 to 600 m (Kaiho et al., 1993). The Flaxman Formation is a lower delta plain to prodelta succession preserving inner to middle shelf deposition with the development of periodic dysoxia (Fig. 15B).

10.2. Macrofossil palaeoecology of the Coniacian environment

The Coniacian assemblage in cores 13 and 16 in the Belfast Mudstone Formation represent a low-energy hydrodynamic regime in a quiescent, outer shelf to upper slope environment, and the fragmented nature of some observed fossils may indicate a degree of post mortem disturbance by burrowing infauna rather than swift current activity. Fossils are generally scattered and sparse and of low species diversity. Of note, rare scleractinian corals in core 13 (biofacies C) attributed tentatively to the Cretaceous to recent genus, *Trochocyathus*, are significant in that they most likely flourished as ahermatypic, solitary forms in deeper waters in muddy or fine-grained substrates. Compared with most Australian Cretaceous macrofaunal assemblages, few apart from the Belfast Mudstone Formation and the Moonkinu Formation (mid-Cenomanian) of Bathurst Island (see Stilwell and Henderson, 2002) are as well preserved, with jewel-like preservation of original nacreous shells and little to no diagenetic alteration.

The most dominant macrofossil recorded in these two cores of the Belfast Mudstone Formation is the inoeceramid bivalve, *Cremnoceramus bicorrugatus bicorrugatus?* preserved in the cores as partial shells of adults or immature individuals and prismatic fragments, reflecting both natural burial processes associated with a range of ontogeny of the species, and also to the limited amount of available core. This bivalve was probably epifaunal, or even partial living in the soft muddy substrates of middle to outer shelf environments (see Crampton, 1996) and may have had a life habit as a mudsticker with the dorsal part of the shell buried in the sediment. Another possibility is that *C. bicorrugatus bicorrugatus?* may have been a free-resting recliner. The presence of this species in dark, organic-rich facies is consistent with other records of the group that suggest that they lived in dysoxic environments that contained depauperate assemblages of epibenthic organisms (see Crampton, 1996, and references therein). In some deposits, inoeceramid bivalves are the sole macrofossil group present. Indeed, few other taxa are recorded in the cores (Table 3) and all of the bivalves and gastropods are infaunal and/or semi-infaunal deposit feeders that preferred fine-grained substrates.

10.3. Turonian-Santonian prodelta to upper bathyal environment

The relatively organic-rich mudstone facies of the Belfast Mudstone Formation is typical of prodelta shale deposition (Fisher et al., 1969). The abundance of Type III/IV kerogen in this unit, suggests that the majority of this organic material is terrestrial; however the proportion of possible marine kerogen increases slightly in Bridgewater Bay-1, indicating deepening towards the main depocentre southeast of the study area (Figs. 3 and 14) this concurs with the seismic evidence which shows downlap progradation east of Voluta-1. While the organic yield is mainly land-derived, there is evidence that increased oceanic productivity and carbon flux occurred during Belfast Mudstone Formation deposition. In general, the benthic foraminiferal yield and abundance are relatively high in the Belfast Mudstone Formation, allowing a more detailed palaeoenvironmental interpretation of this unit compared with the underlying Flaxman Formation. However, the relatively high TOC values and low diversity foraminiferal assemblages typical of the Flaxman Formation continue into the Turonian part of
Belfast Mudstone Formation biofacies A. This suggests the persistence of shallow shelf oxygen-stressed palaeoenvironments in this interval. The change to lower TOC values and highly bioturbated fabric in the Coniacian mudstone facies of biofacies A indicates more oxic conditions after this time. This change is accompanied by an increased benthic rotalid diversity, and an increase in the abundance of *Gavelinella* spp. and *Stensioeina* spp., taxa typical of outer neritic to middle bathyal environments (Nyong and Olsson, 1984; Sikora and Olsson, 1991; Carrillo et al., 1996). Other associated taxa are the vaginulinids, such as *Lenticulina* spp., thought by Holbourn et al. (2001) to be typical of low-carbon-flux upper to middle bathyal environments during the middle Cretaceous. These diverse outer shelf to upper slope assemblages occur in biofacies A and biofacies B and persist into the lower part of biofacies C, where TOC values are high. The upward increase in TOC values and the reduction of bioturbation features in the mudstone from the upper part of biofacies A suggest that the environment became more oxygen deficient during the Coniacian (cf. Kauffman and Sageman, 1988), associated with a decrease in abundance of the low-carbon-flux indicator *Lenticulina* spp. This upward increase in dysoxia is also manifested by the change in pristane/phytane ratios from 5 (oxic) in the Turonian biofacies A to 1.6 (strongly dysoxic) in Coniacian biofacies C and the disappearance of common macrofaunal elements such as the inoceramids by biofacies C times. The plankton *Hedbergella* spp. and *Heterohelix reussi* that occur as peaks at the A/B and B/C biofacies boundaries and within biofacies C are typical high southern latitude Austral Realm taxa (Huber, 1992). Koutsoukos et al. (1990) suggested that such a poorly developed hedbergellid and heterohelicid plankton fauna typified Upper Cretaceous OMZ shelf-edge environments with a low to moderate degree of oxygen deficiency (i.e. “dysaerobic” conditions from 0.1 to 1.0 ml/l O2). In addition, the presence only of these two taxa in Upper Cretaceous assemblages may indicate eutrophism and upwelling (Premoli Silva and Sliter, 1999). Keller and Pardo (2004) suggest that the heterohelicids inhabited the OMZ in normal marine salinities during the Upper Cretaceous and classify the hedbergellids as normal marine salinity surface dwellers. Based on these data, these plankton events happened during a time when the OMZ expanded into the prodelta, outer shelf to upper bathyal realm in normal marine conditions. The strong association of these events with peaks in *Hanzawaia* spp. suggest that these taxa may be benthic dysoxic indicators. *Buliminella* spp. and *Praebuliminella* spp. are typical of Upper Cretaceous outer-shelf to upper bathyal organic-rich sediments (Holbourn et al., 1999) and are regarded by these authors as OMZ flux indicators. These taxa occur in biofacies C in the Belfast Mudstone Formation. By the time these taxa are most common in the Coniacian biofacies C coinciding with a plankton peak, *Gavelinella* and *Stensioeina* had disappeared. Oona et al. (1998) reported a similar transition from more oxic condition with common *Gavelinella* to dysoxia with *Neobuliminella* in the Cenomanian to Turonian strata of the western Interior Seaway, USA. These two taxa were replaced upwards by the outer shelf to upper slope *Gyroidinoideas* spp. (Sikora and Olsson, 1991; Holbourn et al., 2001) with significant peaks of nodosariids and *Cibicides* spp. The periodic influx of OMZ indicators in biofacies C suggests that occasional intensification of dysoxia alternated with more oxic conditions. This is also manifested by the occasional macrofaunal-rich intervals (with solitary corals) in the predominantly macrofossil-free biofacies C. The relative abundance of diagnostic OMZ foraminifera in biofacies C is not as great as in assemblages recorded from the Coniacian by Holbourn et al. (1999) from the tropical west African margin. This pattern is probably due to the southerly high latitude nature of the assemblages and the relatively high sedimentation rates during this interval.

The upward increase in sandstone and the presence of woody debris within the middle Santonian Belfast Mudstone Formation biofacies D heralds the onset of the delta front progradation (Fig. 15). The marked reduction in foraminiferal diversity and planktonic foraminiferal abundance in this interval is interpreted as related to shallowing of the environment, to predominantly shelfal conditions and to the dilution effects of high sedimentation rates during this progradation. It is also likely that increased fluvial input in this interval resulted in hyposaline conditions, limiting foraminiferal diversity. This shallowing with associated carbonate dissolution caused the local extinction of many of the calcareous foraminifera in the succession.
10.4. Turonian to Campanian subsidence, relative sea level change and dysoxia

The initial flooding during the Turonian on the southeast margin of Australia established the Waarre Formation delta plain to shallow marine delta front conditions. Subsequently, further deepening established the delta front to prodelta Flaxman Formation to Belfast Mudstone Formation biofacies A. While the relative chronology of these units cannot be solved by regional biostratigraphic analyses, they probably correlate to a Late Turonian regressive phase in relative global sea level (Miller et al., 2003, 2004; Fig. 15B). Significant subsidence on growth faults after 89.3 ± 1.0 Ma produced localised depocentres around Bridgewater Bay-1 and Voluta-1. This faulting created accommodation space that allowed the thick prodelta, outer shelf to bathyal mudstone-rich Belfast Mudstone Formation to prograde seaward onto the Flaxman Formation. Significant subsidence controlled deposition in this basin from 89.3 Ma-85.7 Ma when Belfast Mudstone Formation biofacies A to C were deposited. The cessation of active growth faulting in the middle Santonian (Belfast Mudstone biofacies D) reduced the accommodation space and led to rapid progradation seaward to inner shelf depths. The post-Coniacian Sherbrook Group strata were deposited during a transgressive global relative sea level (Miller et al., 2003, 2004; Fig. 15B). The transgressive event recorded by Miller et al. (2003, 2004) across the Santonian–Campanian boundary is associated with a regional mudstone-rich unit called the Skull Creek Mudstone and flooding surface in the Paaratte Formation (Krassay et al., 2004).

The shallow marine facies of the late Turonian Flaxman Formation preserve marked variations in foraminiferal diversity, TOC and HI values. This variation may reflect changes in marine dysoxia enhanced by intermittent freshwater and terrigenous influx that caused periods of thermohaline stratification and oxygen reduction in the bottom water. A period of less carbon-rich sedimentation occurred above the base of the Coniacian in the Otway Basin. The interval reflects an increase in oxia and increased sedimentation rates associated with growth faulting. Then followed a period of marked organic carbon enrichment and dysoxia when the oxygen minimum zone and its associated upwelling expanded across the shelf break from the upper bathyal to prodelta facies in the upper Coniacian to lower Santonian Belfast Mudstone Formation biofacies C. These Coniacian to Santonian strata are interpreted to preserve high latitude evidence for the occurrence OAE 3. This event is the last OAE of the warm Late Cretaceous ocean prior to the evolution of cooler modern oceanic conditions (Arthur et al., 1988). The published duration of OAE 3 (Fig. 15B) varies from: the entire Coniacian to Santonian stages (Jenkyns and Hallam, 1980), middle Coniacian to middle Santonian stage (Arthur et al., 1988), the lower Santonian strata of western tropical Atlantic (Hofmann et al., 2003) and the lower Santonian to lower Campanian strata of the La Luna Formation in Venezuela (Rey et al., 2004). OAE 3 is associated with δ13C enrichment in northwest Europe (Jenkyns et al., 1994) and represents upwelling conditions that developed in epicontinental marine settings (Arthur et al., 1988) similar to the Otway Basin. The evidence suggests that, rather than confined to low latitude tropical regions, the effects of OAE 3 reached southern high latitude regions where modelled sea surface temperatures varied from 6 to 20 °C (Haupt and Seidov, 2001).

11. Conclusions

This study integrates seismic, wireline log, fossil, facies and geochemical data from a Turonian–Campanian siliciclastic succession in the Otway Basin in southeast Australia to arrive at the following conclusions:

1. The Turonian–Campanian strata of the Otway Basin were deposited in a delta plain, delta front and shale-dominated prodelta to upper slope environment. Two principal depocentres to the southeast and southwest of the area (the Otway Delta) received up to 6 km of sediment in a 500 km marine inlet between Antarctica and Australia situated at least 70° S.
2. The late Turonian sandstone-dominated Waarre Formation was deposited in a marginal marine delta plain to marine delta front conditions.
3. The mudstone/sandstone facies of the overlying Turonian Flaxman Formation and basal Belfast Formation represent transgressive inner to mid-
dle shelf upper delta to prodelta facies. An influx of fresh water and/or terrigenous material caused intermittent thermohaline stratification and dysoxia during this time.

(4) From 89.3–85.7 Ma significant growth faulting created accommodation space allowing prodelta, outer shelf to bathyal mud-dominated facies to prograde seaward, depositing the Belfast Mudstone Formation.

(5) During the basal Coniacian a period of relatively carbon-poor oxic sedimentation occurred.

(6) Detailed foraminiferal, facies, macrofossil and preliminary geochemical evidence suggest that the transgressive conditions that occurred from the early Coniacian–Santonian were associated with an upward increase in dysoxia. This was caused by the migration of the oxygen minimum zone across the shelf break from the upper bathyal to prodelta environment.

(7) The dysoxic period during Belfast Mudstone Formation deposition correlates with OAE 3. This was the last OAE of the warm Cretaceous Ocean (Arthur et al., 1988) prior to the evolution of a more modern oceanic system. The evidence suggests that the effects of OAE 3 were not confined to low latitude tropical regions and that dysoxia spread to southern high latitude warm inland seas such as the Otway during the Coniacian–Santonian.

(8) The cessation of growth faulting after 85.7 Ma limited accommodation space and delta front to prodelta sandstone/mudstone facies of the uppermost Belfast Mudstone and lower Paaratte Formation prograded rapidly seaward. The high sedimentation rates that accompanied this shallowing diluted the foraminiferal assemblages and created hyposaline conditions causing the dissolution and local extinction of many of the calcareous foraminifera in the succession.

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References


Department of Natural Resources and Environment, 1999. Victorian Petroleum Otway Biostratigraphic Reports. CD Version 0.5c.


Walaszczyk, I., 1992. Turonian through Santonian deposits of the Central Polish Uplands; their facies development, inoceramid paleontology and stratigraphy. Acta Geologica Polonica 42 (112 pp.)
