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Did global tectonics drive early biosphere evolution? Carbon isotope record from 2.6 to 1.9 Ga carbonates of Western Australian basins

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Abstract

The $\delta^{13}C_{carb}$ record of well preserved carbonates in outcrop and core is here examined from the 2.6 to 1.9 Ga old basins of Western Australia. These data, which are constrained by a well defined stratigraphic and tectonic framework, and by U–Pb zircon ages, provide an insight into the variables coincident with the evolution of an oxidative atmosphere and the evolution of the early biosphere. In the latest Archaean (ca. 2.6 Ga) the secular $\delta^{13}C_{carb}$ curve is flat much like that seen in the later Palaeoproterozoic basins of Northern Australia (< 1.8 Ga). This implies that photosynthesis was a major component of the biosphere at that time and that the carbon mass balance was stable. In the early Palaeoproterozoic, beginning after 2.5 Ga and continuing until at least 1.9 Ga, the $\delta^{13}C_{carb}$ the curve is much more dynamic, with significant positive and negative excursions, including a major positive excursion (+9‰PDB) close to 2.2 Ga. These excursions can be correlated with the Lomagundi event identified in Africa, Europe and North America. Previously published studies of the overlying Meso- to Palaeoproterozoic Bangemall basin and of 1.8–1.5 Ga old basins in northern Australia suggest that the $\delta^{13}C_{carb}$ curve became relatively monotonic again after ca. 1.8 Ga and remained so for most of the following Mesoproterozoic. Comparisons with data from other ancient cratons, especially Africa, suggest that the secular carbon curve may be even more complex than presently understood and probably comparable to the major excursions seen in the Neoproterozoic. When the Western Australian data are placed in their stratigraphic and tectonic framework we find that the monotonic latest-Archaean curve coincides with a tectonically quiescent period in which carbonates formed in an basinal setting on a craton surrounded by passive margins. The data are consistent with an earth in which the carbon mass balance was in equilibrium. The $\delta^{13}C_{carb}$ curve began to oscillate following the onset of glaciation as the Pilbara and Yilgarn Cratons began to converge during the Capricorn Orogeny suggesting periods of rapid carbon burial during continental dispersal. However, the major positive excursion is preserved in carbonates from back-arc basins formed as the ocean closed and subduction began. Since similar tectonic processes can be recognised, not only in Northern Australia but also on other early cratons, it can be argued that the carbon excursions relate to supercontinent cycles and to major periods of mantle overturn and superplume development. We explain this coincidence of carbon isotopic excursions

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and tectonism by the sequestration of carbon during ocean closure with organic-rich passive margin sediments containing isotopically light carbon subducted into and stored in the lower crust and mantle. The global ocean thereby became enriched in isotopically heavy carbon, releasing oxygen to the atmosphere. A second stepwise increase in atmospheric oxygen in the Neoproterozoic may also have been connected with the assembly of Rodinia. This second event has been associated with the development of multicellular life and the evolutionary 'Big Bang'. Between the two events the carbon cycle, and to some extent the biosphere, appear to have entered a period of prolonged evolutionary stasis. This implies that the evolution of both the atmosphere, and the biosphere, may have been driven forward by planetary evolution, implying that biospheric evolution has largely been driven by the dynamo of earth's tectonism and its long term survival depends upon these endogenic (thermal) energy resources. If this is so it has fundamental implications, not only for life on earth, but for the more general problems surrounding the likelihood of life having evolved on other planetary bodies. Small planets with insufficient endogenic energy resources to sustain the crust/mantle interactions of plate tectonics (such as Mars) seem to us unlikely to have allowed evolution beyond single-celled life forms. Once a planet's energy resources are expended the biosphere would most likely enter a prolonged stasis and ultimately face extinction.

Keywords: Carbon isotopes; Archean; Palaeoproterozoic; Hamersley basin; Pilbara; Biosphere

1. Introduction

One of the most promising ways of tracking the evolution of earth's early biosphere involves analysis of biogeochemical signatures preserved in the sedimentary record (Knoll and Canfield, 1998). Useful insights into the carbon cycle and the growth of the crustal carbon reservoir may be gained, for example, through analysis of the stable isotopes of carbon (^{12}C and ^{13}C-expressed as $\delta^{13}C$ in ‰PDB) which are fractionated during autotrophic fixation of CO_2. The long-term secular curve for $\delta^{13}C_{carb}$ is bimodal with a major peak at approximately 2.0 Ga (see, Karhu and Holland, 1996; Bau et al., 1999—the so called 'Lomagundi Event'), and another well defined but oscillatory peak centred at approximately 0.65 Ga and associated with the appearance of multicellular life and the evolutionary 'Big Bang' in the Neoproterozoic–Cambrian interval. The complex Neoproterozoic carbon isotope record has been attributed to the rapid expansion of the metazoa, to nutrient flux and to the locking up of lighter ^{12}C during anoxic events (e.g. Schopf and Klein, 1992; Knoll and Walter, 1992; Derry et al., 1992, 1994; Berner and Canfield, 1989; Brasier and Lindsay, 1998; Brasier and Sukhov, 1998). In between ca. 1.9 and 1.0 Ga the secular carbon isotope curve is almost flat (Kaufman, 1997; Buick et al., 1995; Des Marais, 1997; Brasier and Lindsay, 1998; Lindsay and Brasier, 2000). It has been argued that the conspicuously bimodal nature of the secular carbon curve indicates that the global reduced carbon reservoir has grown episodically and this, in turn, may indicate that the atmosphere has become oxygenated in a stepwise fashion (Des Marais et al., 1992) as a result of episodic burial of carbon during large scale tectonic cycles (Des Marais, 1994). In between the episodes of oxygenation, it has been suggested that tectonic activity was low and that CO_2 in the ocean–atmosphere system reached a state of near equilibrium with respect to mass balance of the carbon cycle (Brasier and Lindsay, 1998; Lindsay and Brasier, 2000). Recently, Kump et al. (2001) have argued that this mechanism is not viable on the grounds that residence times for atmospheric O_2 are too short. Data from the present study support a tectonic mechanism as being important but do not rule out other mechanisms.

The Pilbara and Yilgarn Cratons of Western Australia, which are remnants of some of Earth's earliest crustal blocks, preserve a sedimentary record that offers one of the best opportunities to study the ancient isotopic record. The late Archaean and early Palaeoproterozoic basins that

rest upon these ancient cratonic blocks (Fig. 1) form a time series associated with the formation and ultimate disassembly of one of the Earth's earliest major continental masses. This was followed by the suturing of the Pilbara and Yilgarn Cratons along the Capricorn Orogen as a new supercontinent evolved in the Palaeoproterozoic. These basins contain an important and ancient sedimentary record of the early earth including some of earth's earliest carbonate platform deposits. The carbonates are well preserved, despite their age, and offer an ideal setting in which to investigate stable isotopes in a stratigraphic and structural framework that is well constrained by radiometric ages (Barley et al., 1997; Trendall et al., 1998; Arndt et al., 1991) at a critical time in Earth history when the atmosphere was first becoming oxygen rich. In this paper, we present the results of an extensive study of the stable isotopes of carbon and oxygen in these early carbonates. We also attempt to assess the factors controlling the evolution of the atmosphere and biosphere in relation to planetary dynamics.

2. The Western Australian basins

The Pilbara Block of Northwestern Australia, one of the earth's oldest crustal blocks, was established as a stable continental nucleus before 2770 Ma. A number of early basins rest, either on this ancient craton (Hamersley and Ashburton basins), or along its suture with the adjoining Yilgarn Block (i.e. Bryah, Yerrida, Padbury, Earaheedy and Bangemall basins), known as the Capricorn Orogen.

The Late Archaean–Early Palaeoproterozoic Hamersley basin, which lies close to the southern margin of the Pilbara Block (Trendall, 1983, 1990; Blake and Barley, 1992), overlies a normal crustal thickness (Drummond, 1981) and has a basin-fill architecture (cf. Krapez, 1996) very similar to that of the Neoproterozoic basins of central Australia (cf. Lindsay and Korsch, 1989; Lindsay and Leven, 1996). The basin began to subside at ca. 2.8 Ga following an episode of crustal extension. The basin fill is complex and polyphase with

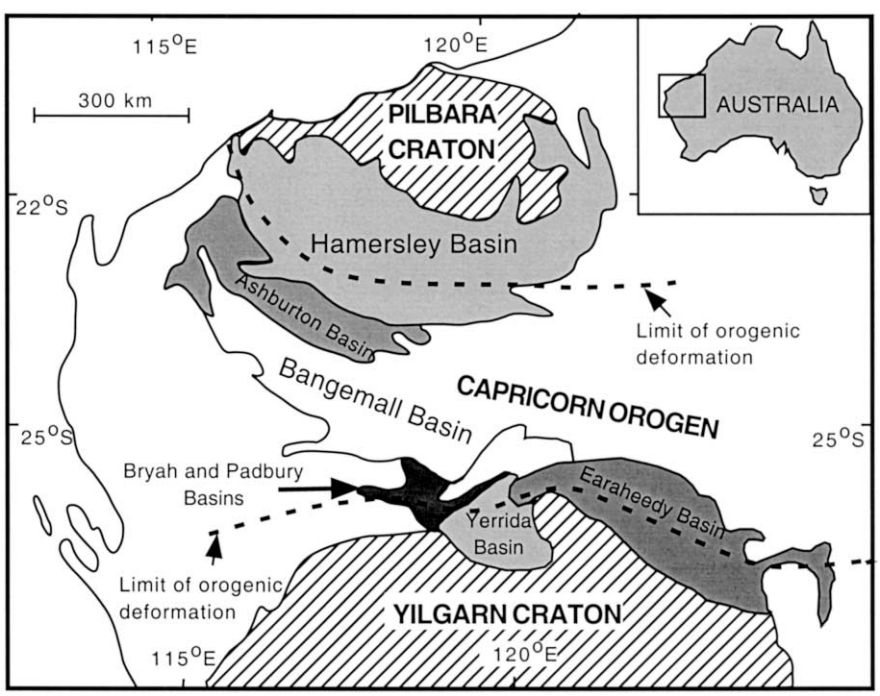

Fig. 1. The distribution of Western Australian basins sampled in this study in relation to the Pilbara and Yilgarn Cratons and the deformed zone of the Capricorn Orogen.

major erosional surfaces separating areally extensive megasequences/supersequences (Trendall, 1990; Krapez, 1996), much as seen in the Neoproterozoic and Palaeoproterozoic basins of central and northern Australia (Lindsay and Leven, 1996; Lindsay and Korsch, 1987; Lindsay and Brasier, 2000). During Fortescue and Hamersley Group time (Fig. 2), sedimentation began in a rift setting, in a siliciclastic-starved environment. Consequently, biogeochemical sediments and extrusive volcanic rocks dominate the early basin fill while clastic rocks are generally subordinate and fine grained. The succession is largely marine and includes major intervals of banded iron formation (BIF), for which the basin is well known (Trendall and Blockley, 1970), and some of the earth's earliest platform carbonates (ca. 2.7–2.5 Ga, Simonson et al., 1993a,b). The later stages of Hamersley basin sedimentation (Turee Creek Group times) record the early transition from a passive margin to a foreland basin setting and with it the establishment of a terrigenous sediment supply (Tyler and Thorne, 1990). The geometry and sequence stacking pattern of the succession are both very similar to younger intracratonic basins such as those encountered in central Australia where sequences are thin and extensive and dominated by highstand and, to a lesser extent, by transgressive systems tracts (see Lindsay et al., 1993). The Hamersley basin is thus the first clearly identifiable basinal setting preserved on the Australian craton in which marine sediments, and in particular platform carbonate rocks, have been preserved in a response to broad, regional, crustal subsidence.

The Ashburton basin (2.2–1.8 Ga) disconformably overlies the southern margin of the Hamersley basin along its the contact with the Capricorn Orogen (Thorne and Seymour, 1991). The basin forms the northern margin of the Capricorn Orogen, a deformed zone of low–high grade metamorphic rocks that occur along the contact between the Yilgarn and Pilbara Cratons. It was formed in a compressional environment during ocean closure. Loading of the ocean floor by the approaching Yilgarn Craton led to the development of a west–northwest oriented foreland basin (McGrath Trough, Horowitz, 1982) with an uplifted orogenic margin to the southwest (Blake and Groves, 1987; Blake and Barley, 1992) with the result that sedimentation was dominated by an abundant supply of terrigenous clastic material (Mount McGrath Formation). Continued loading of the crust ultimately led to the supply of terrigenous materials being disrupted producing a prograding carbonate shelf on which the Duck Creek Dolomite accumulated. Basin sedimentation was accompanied by active-margin mafic volcanism (Cheela Springs Basalt).

The Yerrida (ca. 2.2–1.9 Ga), Bryah (ca. 2.0 Ga), Padbury (ca. 2.0 Ga) and Earaheedy (ca. 1.9–1.65 Ga) basins lie to the southeast of the Hamersley and Ashburton basins and overlie the Capricorn Orogen. These now fragmentary basins formed along the northern margin of the Yilgarn Craton in a back-arc setting and record the convergence and collision of the Archaean Pilbara and Yilgarn Cratons (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Pirajno et al., 1998; Occhipinti et al., 1997, 1998). Due to their active margin settings, the fill of these small basins is dominated by clastic sediments. However, platform carbonate units are preserved in the Yerrida basin (Gee and Grey, 1993; Pirajno and Adamides, 2000) and a significant thickness of carbonate rocks also occurs in the Earaheedy basin, allowing for isotopic coverage of this critical time interval.

3. Analytical methods

Sampling was preferentially undertaken on carbonates from the less deformed parts of the basins and on lithologic intervals free from evidence of secondary alteration (Table 1). Where possible samples were collected from drill core at 5 or 10 m stratigraphic intervals through all major carbonate units. Where drill core was not available outcrop sections were sampled at a similar interval depending upon the availability of suitable exposures. Samples consisting of fine-grained, micritic, microbial or, less frequently, oolitic carbonates were selected and examined both macro- and microscopically for lithologic variation. Selected portions of carbonate were cleaned and analysed

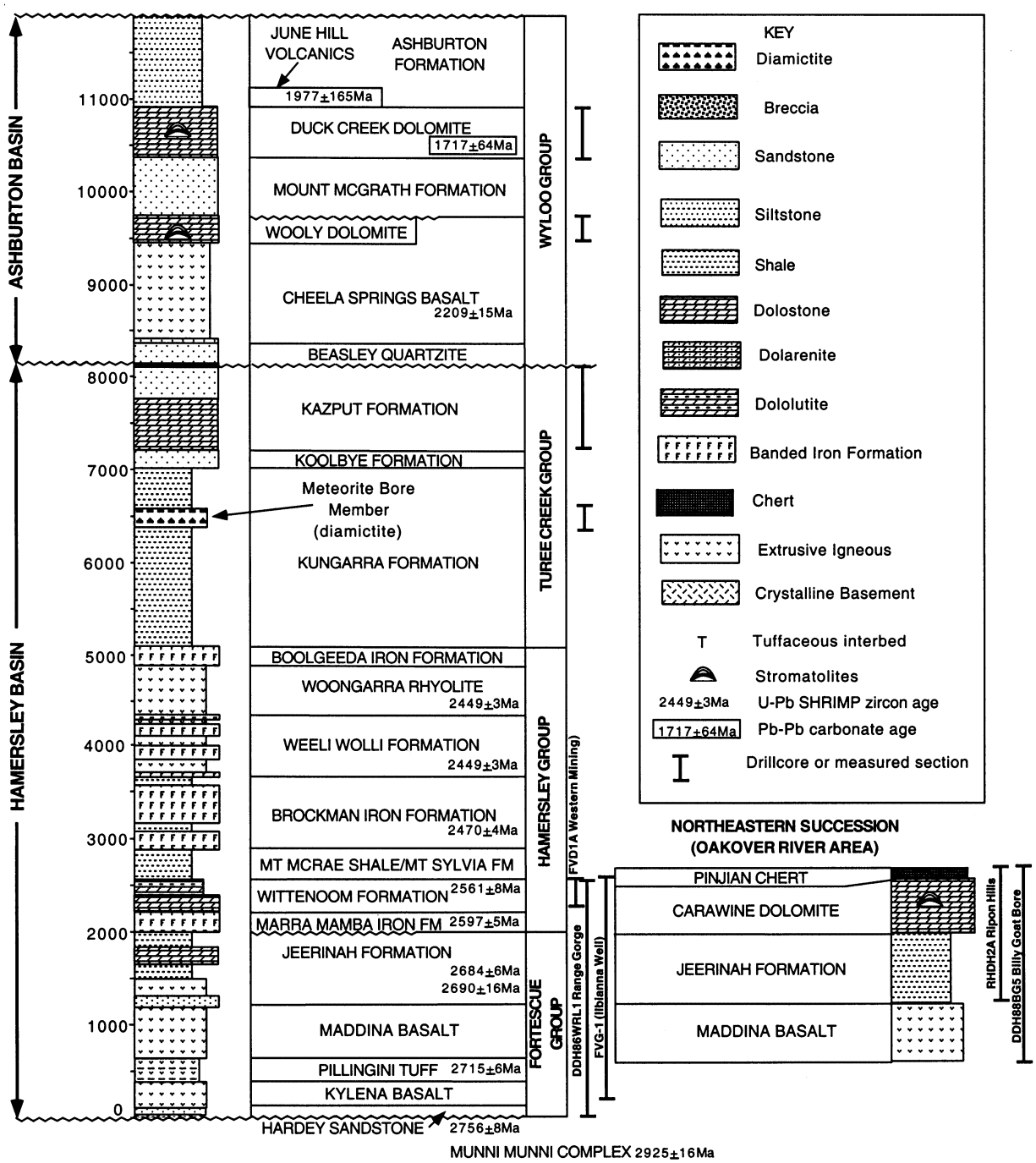

Fig. 2. Simplified stratigraphy for the Hamersley and Ashburton basins of Western Australia. Ages are mostly U–Pb SHRIMP zircon dates from several sources (Barley et al., 1997; Trendall et al., 1998; Arndt et al., 1991). For a comprehensive geochronological summary see Nelson et al. (1999).

Table 1
Sample statistics for data from the late Archaean and early Palaeoproterozoic basins of Western Australian basins

Drillhole/section locality	Location		Total depth (thickness) (m)	Formation	Formation thickness (m)	Carbonate carbon		Oxygen		Number of samples
	Latitude	Longitude				Mean	S.D.	Mean	S.D.	
Hamersley basin										
RHDH2A (Ripon Hills)	21°17′E	120°50′S	501.0	Carawine Dolomite	184.0	0.06	0.56	−6.66	1.38	20
DDH8BG5 (Billygoat Bore)	22°15′S	120°36′E	500.0	Carawine Dolomite	320.0	0.73	0.32	−7.43	1.03	62
FVG-1 (Ilbianna well)	22°33′S	119°30′E	1902.0	Wittenoom Formation	529.0	0.02	0.41	−8.80	1.27	53
DDH'86 WRL-1 (Range Gorge)	22°11.7′S	118°12.6′E	1963.0	Wittenoom Formation	258.0	0.22	0.18	−11.26	1.22	28
FVD 1a (Western Mining)	23°07′S	119°58′E	400.0	Wittenoom Formation	331.0	0.03	0.17	−8.15	0.89	55
Ashburton basin										
Duck Creek	22°29.8′S	119°19.0′E	517.0	Duck Creek Dolomite	517.0	0.49	2.31	−9.13	2.31	22
Horseshoe Creek	22°50.2′S	116°50.1′E	248.4	Kazput Formation	248.4	−2.40	2.53	−3.89	1.42	23
Yeera Bluff	21°42.4′S	116°08.3′E	–	Meteorite Bore Member	Outcrop	−0.52	n/a	−4.51	n/a	2
Mt de courcey area	22°47.3′S	116°27.5′E	59.0	Wooly Dolomite	59.0	−0.90	2.18	−8.26	2.89	16
Yerrida basin										
Peak hill sheet	25°38′S	118°43′E	–	Johnson Cairn Formation	Outcrop	−0.4	n/a	−7.6	n/a	1
QMW 83-1 (Quartermaine)	25°59.0′S	119°54.0′E	372.0	Juderina Formation	65.3	7.69	0.11	−6.40	0.48	5
PP-011 (ACM)	26°32.7′S	119°13.5′E	247.5	Juderina Formation	50.0	6.70	0.58	−7.77	1.66	6
Earaheedy basin										
Thurraguddy Bore	25°55.6′S	121°48.0′E	–	Kulele Limestone	Outcrop	0.46	n/a	−13.58	n/a	4
Tooloo bluff	26°24.6′S	122°14.3′E	–	Windidda Formation	Outcrop	−0.70	0.81	−9.20	2.60	8
Coonabildie trig point	25°46.3′S	122°41.5′E	–	Frere Formation	Outcrop	−0.7	n/a	−9.9	n/a	1

Table 1 (*Continued*)

Drillhole/section locality	Location		Total depth (thickness) (m)	Formation	Formation thickness (m)	Carbonate carbon		Oxygen		Number of samples
	Latitude	Longitude				Mean	S.D.	Mean	S.D.	
TDH-1 (RGC exploration)	25°41.1′S	120°39.8′E	451.0	Yelma Formation	146.8	0.63	0.28	−8.54	0.92	78
TDH-26 (RGC exploration)	25°35.9′S	120°39.9′E	517.0	Yelma Formation	127.9	1.16	0.72	−7.46	2.14	58
TDH-28 (RGC exploration)	25°38.9′S	120°39.1′E	394.0	Yelma Formation	80.0	1.76	0.56	−6.47	1.09	32

using a VG Isomass PRISM mass spectrometer attached to an on-line VG Isocarb preparation system in the Oxford laboratories (cf. Brasier et al., 1996; Lindsay and Brasier, 2000). Reproducibility of replicate standards was better than 0.1‰ for $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$. Calibration to PDB standard via NBS 19 and Cambridge Carrara marble was performed daily using the Oxford in-house standard (NOCZ, Brasier et al., 1994). Major and trace element analyses were carried out using XRF and ICP-MS.

The extreme age of the samples requires further careful analysis of possible diagenetic alteration even where thin section evaluation suggests diagenetic effects are minimal. Covariance in $\delta^{13}C_{carb}$/$\delta^{18}O_{carb}$ crossplots have been successfully used as an indicator of diagenetic alteration (Brand and Veizer, 1980; Derry et al., 1992; Lindsay and Brasier, 2000). Where correlation between the two variables is significant at the 95% level of confidence further evaluation was deemed necessary. In general, however, the crossplots show that most samples cluster tightly around mean $\delta^{13}C_{carb}$ values and cross correlation is not significant. The best indication of alteration in the $\delta^{13}C_{carb}$ data comes from the presence of anomalous outlying values. Where single $\delta^{13}C_{carb}$ values departed abruptly from the overall isotopic trend, especially when the deviation was negative, then the sample results were re-evaluated. Anomalous values, generally associated with proximity to brecciated zones or signs of faulting, suggest the influence of post-depositional fluids.

Elemental ratios such as Mn/Sr and Fe/Sr have been proposed elsewhere as indicators of diagenesis (e.g. Veizer, 1983). In Late Palaeoproterozoic sediments we have found considerable variation in these ratios suggesting that in older sediments they reflect local depositional conditions (Lindsay and Brasier, 2000). In late Archaean and early Palaeoproterozoic carbonates were deposited in oceanic environments where oxygen levels were variable (Bekker et al., 2001). It is also likely that the oceans we highly stratified with respect to oxygen concentrations as borne out by the present study. We have found that these elemental ratios vary considerably and often appear in phase with facies cycles (sequences and parasequences) suggesting that palaeo-water depth and thus oxygen levels were the main controlling factors. Fe in particular appears to be closely tied to the depositional sequences with BIF precipitation occurring during maximum flooding when water depths were greatest. Carbonates appear abruptly above the BIFs as a response to rising oxygen levels in the shallowing environment of the highstand. In short there are too many local variable affecting sediment geochemistry to make good use of such indicators.

Overall, we find that the primary fabrics of these ancient carbonates are well preserved, especially in the major platform carbonate units. We take this to suggest that early diagenesis, including dolomitization and silicification, was predominant (cf. Veizer et al., 1990, 1992; Buick et al., 1995; Lindsay and Brasier, 2000). Thin section analysis suggests that the carbonate rocks of the Western Australian basins were largely sealed against the passage of fluids during later diagenesis, thereby preserving their fabric and retaining the primary $\delta^{13}C_{carb}$ signatures.

4. Carbonate stratigraphy and stable isotopes

Carbonate-rich intervals are not uniformly distributed throughout the sedimentary succession in the Western Australian basins but instead occur in well defined settings, most typically in shallow marine facies that were starved of clastic sediment supply. Such conditions tend to occur during the late stages of sea level highstands, during the sag/extensional phases of basin evolution. As discussed by Grotzinger (1989) these conditions may have first appeared during the Archaean/Proterozoic transition, when cratonic stabilisation led to the formation of spatially extensive, stable cratonic masses; i.e. the first 'supercontinents'. Carbonate units, therefore, appear during the early evolution of the Hamersley basin (Hamersley Group) as highstand systems tracts developed on fully developed platforms and ramps (Simonson et al., 1993b) as the basin subsided during thermal recovery and while siliciclastic sedimentation was relatively subdued. In both the later stages of the Hamersley basin, and in the overlying Ashburton

basin, compressional tectonics played a major role so that clastic sediments dominated and carbonate platform units are neither extensive nor thick. Carbonate intervals are also much less abundant in the smaller basins that developed over the Capricorn Orogen in back-arc settings during collision of the Pilbara and Yilgarn Cratons. The smaller Padbury and Bryah basins that formed during this collision have basin fills that are entirely siliciclastic (Occhipinti et al., 1998). The Earaheedy basin, however, is a sag basin that formed in a back-arc setting (Tyler and Thorne, 1990) and contains carbonate units developed during the final stages of sea-level highstands. Thin carbonates also occur locally in the Yerrida basin, where clastic-sediment starvation occurred on maximum flooding surfaces in response to rising sea level.

4.1. Hamersley basin

The Hamersley basin was formed upon Archaean crust of granite and greenstone (Pilbara Craton). The basin, which presently covers 60 000 km^2, was initiated by crustal extension at approximately 2.8 Ga accumulating in excess of nine kilometer of sediment by 2.2 Ga (Fig. 2; Trendall, 1983). The basin fill has been subdivided into three groups, Fortescue; Hamersley and Turee Creek. The two lower groups were deposited in a siliciclastic-starved environment and are dominated either by chemical sediments (e.g. chert, carbonate, BIF) or by extrusive volcanic rocks, while clastic rocks are generally subordinate and fine grained. The dominant mechanism controlling basin subsidence at this time appears to have been thermal recovery in response to early extension, with the result that clastic sources were relatively subdued. Carbonate rocks are least abundant in the Fortescue Group, which was dominated by mafic volcanic and volcaniclastic sedimentation during the early stages of basin evolution, but are present to varying degrees in the two overlying groups. The minor carbonate units present in the Fortescue Group generally occur as thin turbidites, often within deeper water facies of shale and BIF, notably within the Jeerinah Formation.

Depositional patterns in the Turee Creek Group are significantly different from those of the earlier Fortescue and Hamersley Groups because the basin fill was responding to a compressive regime that led to the introduction of a much larger clastic component. The succession that accumulated in this foreland basin includes glaciogenic sediments (Meteorite Bore Member) that can be correlated with Huronian glacial events elsewhere (e.g. North America, South Africa, Fennoscandia) providing the earliest evidence for glaciation on the Australian craton (Trendall, 1976; Martin, 1999).

4.1.1. Hamersley Group

The Hamersley Group is well known for its BIFs which provide a significant proportion of the world's iron ore (Trendall and Blockley, 1970; Morris and Horowitz, 1983; Fig. 2). However, the Paraburdoo Member of the Wittenoom Formation and the Carawine Dolomite also form major carbonate intervals within the group (Figs. 3–7). These carbonates are exposed in two geographically distinct areas. The Wittenoom Formation is exposed across large areas of the main depocentre of the basin, where the Hamersley Group is thickest. The Wittenoom Formation is subdivided into three members, the discontinuous Angeles Shale at the base, the carbonate-dominated Paraburdoo Member and the uppermost Bee Gorge Member (a heterolithic but largely clastic unit). In contrast, exposures of the Carawine Dolomite are only found northeast of the main depocentre (in the Oakover River area, Fig. 2) where the Hamersley Group in considerably thinner. Detailed facies analysis of the two carbonate intervals suggests that they accumulated in distinct depositional environments. The sediments of the Carawine Dolomite are predominantly platform carbonates deposited in shallow marine environments involving stromatolite development and settings that were at times evaporitic and subareal. Carbonate turbidites are present in the Carawine Dolomite but only at its base. The Paraburdoo Member of the Wittenoom Formation was, in contrast, deposited in a much deeper-water ramp setting, at times involving carbonate turbidites (Simonson et al., 1993a,b).

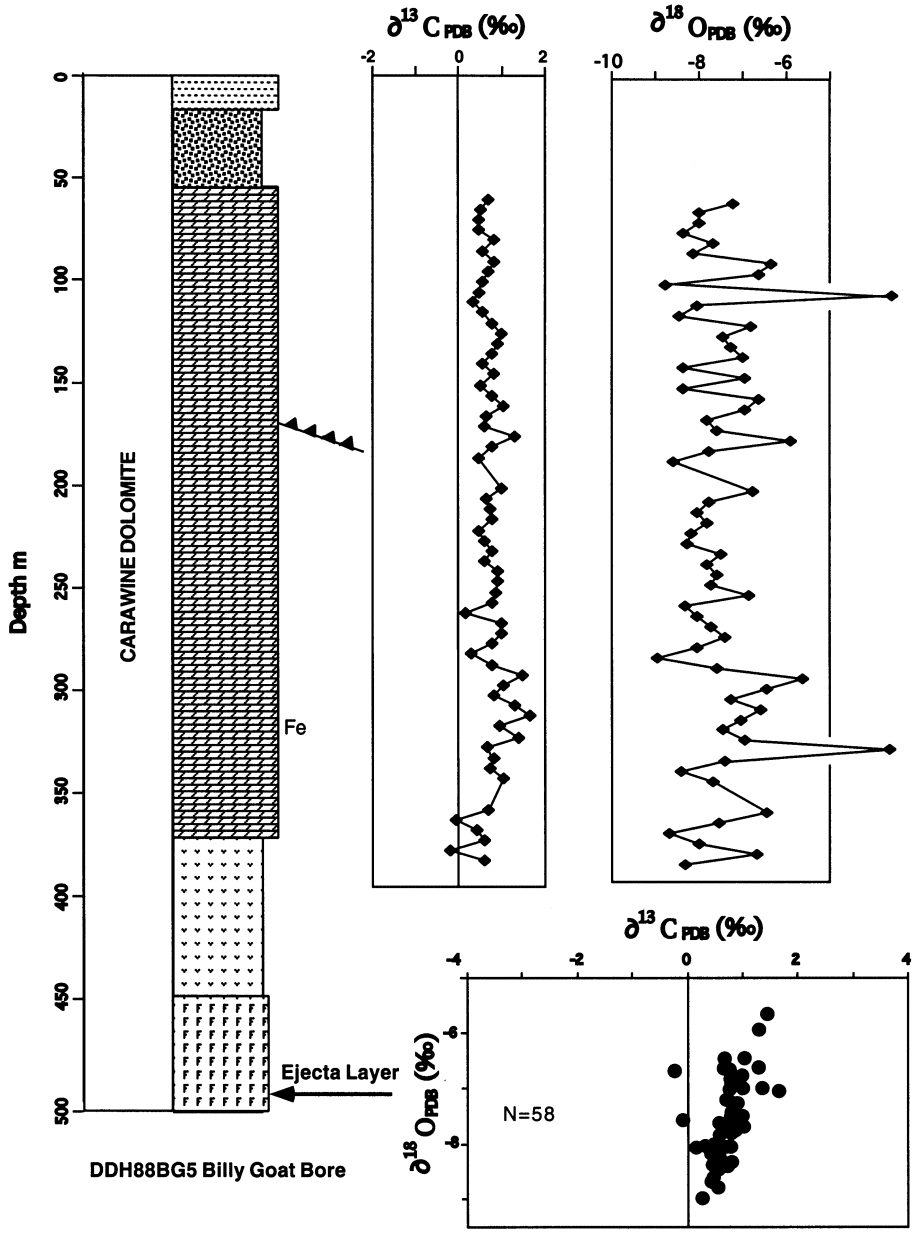

Fig. 3. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Billy Goat Bore drill core DDH88BG5 through the Carawine Dolomite, Hamersley basin, Western Australia (see Table 1). See Fig. 2 for Key.

The Paraburdoo Member and, in places, the Carawine Dolomite both conformably overlie the Mara Mamba Iron Formation which, in turn, conformably overlies the Jeerinah Formation. Locally, the Carawine Dolomite directly overlies the Jeerinah Formation suggesting that the Mara Mamba Iron Formation laps out or laterally interfingers with the upper part of the Jeerinah Formation towards the basin margin. The Jeerinah Formation consist largely of deeper water

shales with occasion carbonate turbidites. γ-Logs through both successions (Figs. 5 and 6) indicate continuity of sedimentation and suggest that the two carbonate units form the upper part of the highstand systems tract of laterally equivalent deposition sequences. The Jeerinah and Mara Mamba Iron Formations and the Angeles Shale Member of the Wittenoom Formation would then form the deeper-water transgressive systems tract and the basal parts of the highstand. The γ-logs

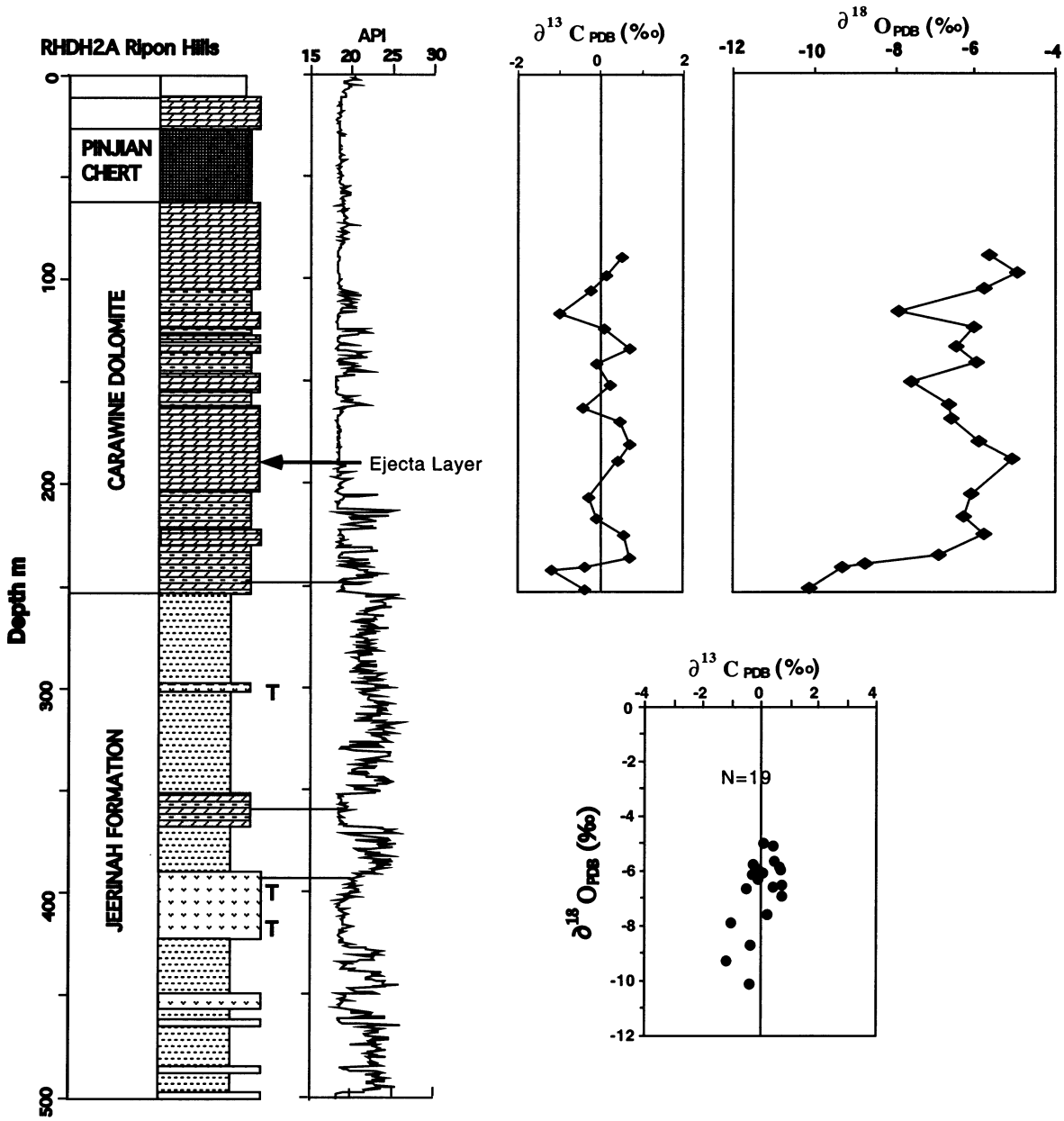

Fig. 4. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Ripon Hills drill hole RHDH2A through the Carawine Dolomite in the Hamersley basin, Western Australia (see Table 1). See Fig. 2 for Key.

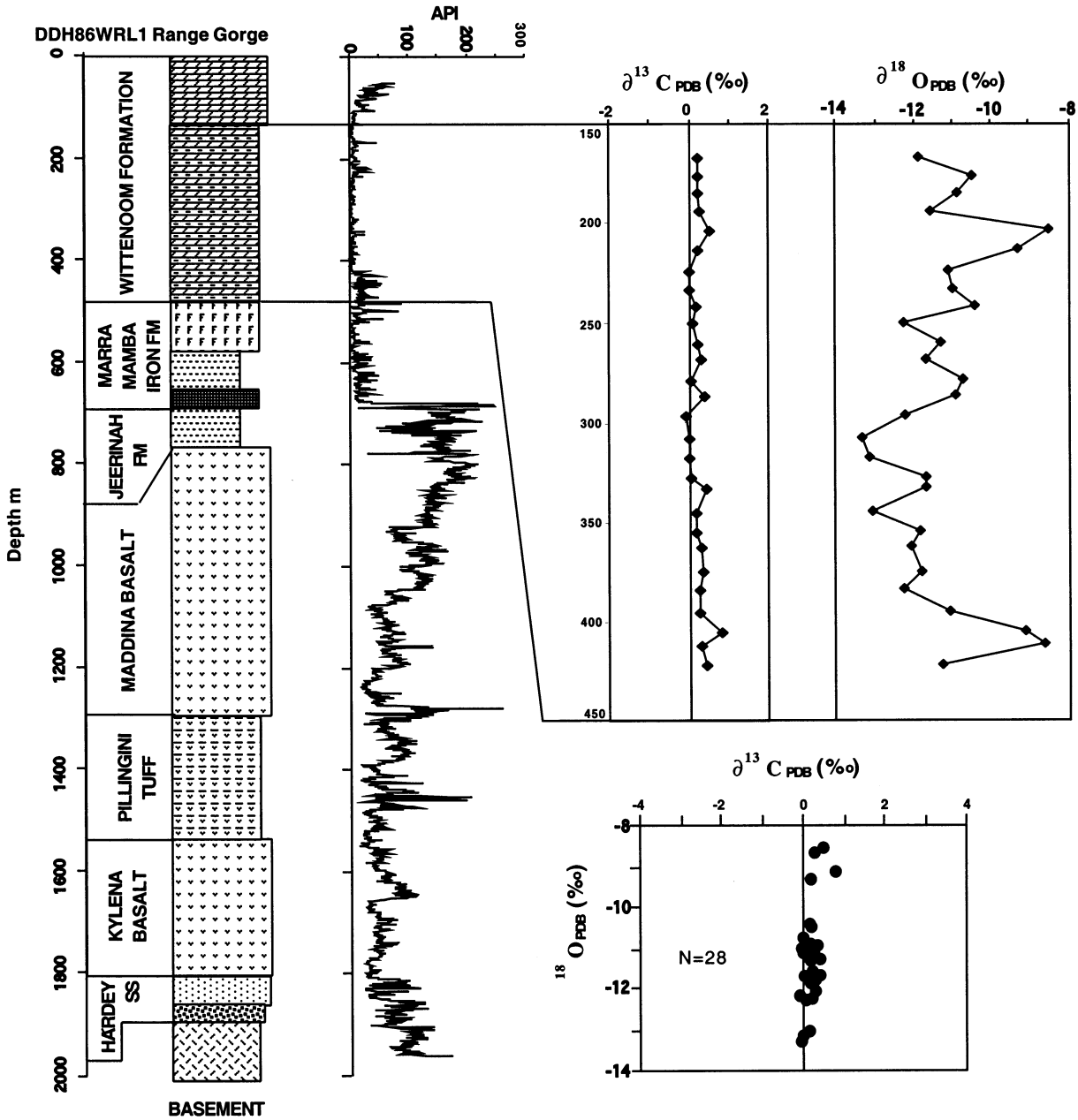

Fig. 5. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Range Gorge drill core DDH86WRL1 through the Wittenoom Formation, Hamersley basin, Western Australia (see Table 1). See Fig. 2 for Key.

indicate that both carbonate units contain thin clastic intervals near their bases and both rapidly become carbonate dominated higher in the formation. This implies that both formed part of a single deposition setting in which carbonates were generated in a proximal setting along the shallower northeastern margin of the basin (Carawine Platform) and dispersed to the southwest from the

platform towards a prograding ramp in the deeper-water, rapidly subsiding main depocentre of the basin (Paraburdoo Member). It also implies that deposition of BIF occurred in deeper water, at or near times of maximum flooding, as sea level rose and the basin became more anoxic. This is

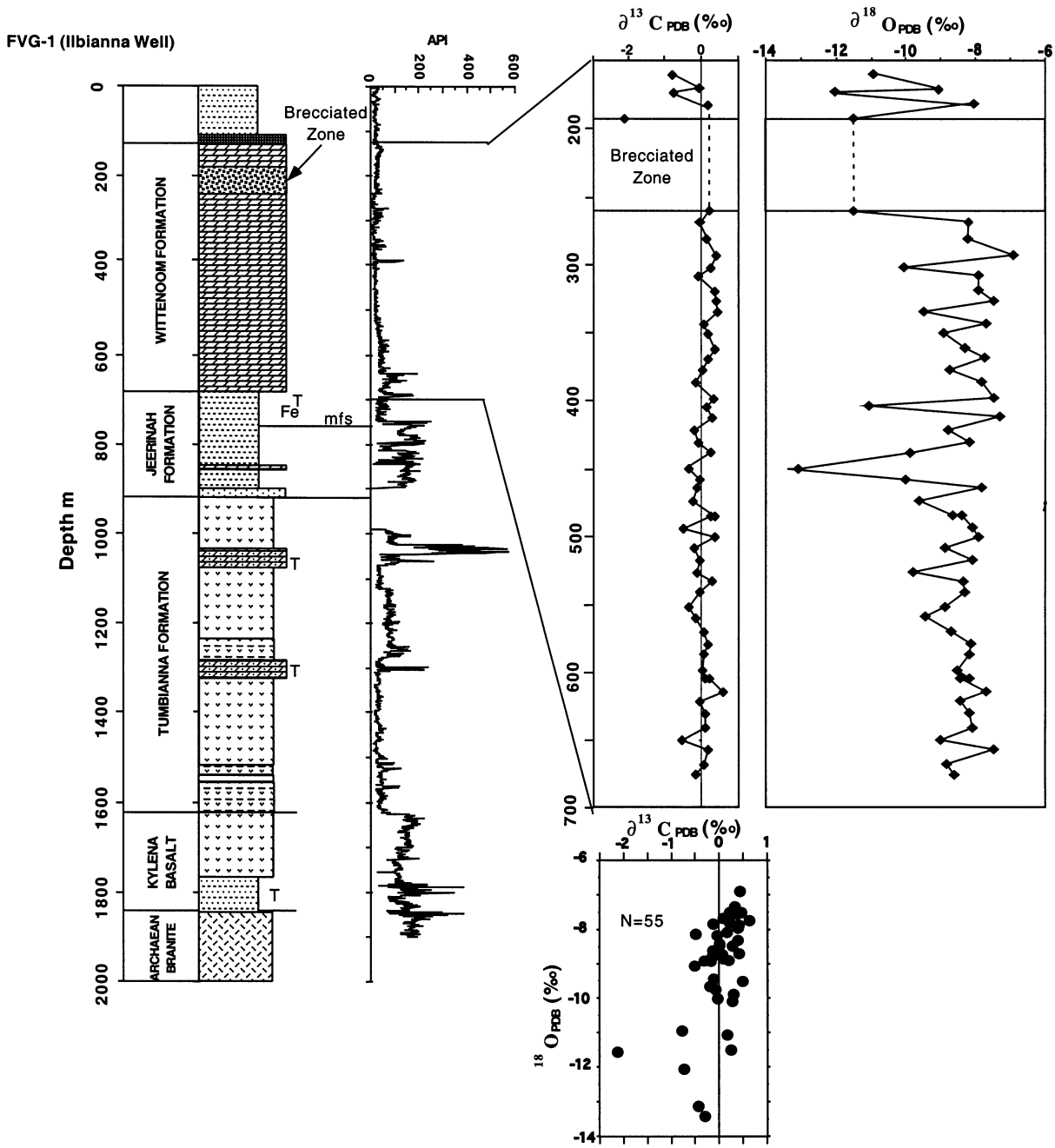

Fig. 6. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Ilbiana drill hole, FVG-1, through the Wittenoom Formation in the Hamersley basin, Western Australia (see Table 1). See Fig. 2 for Key.

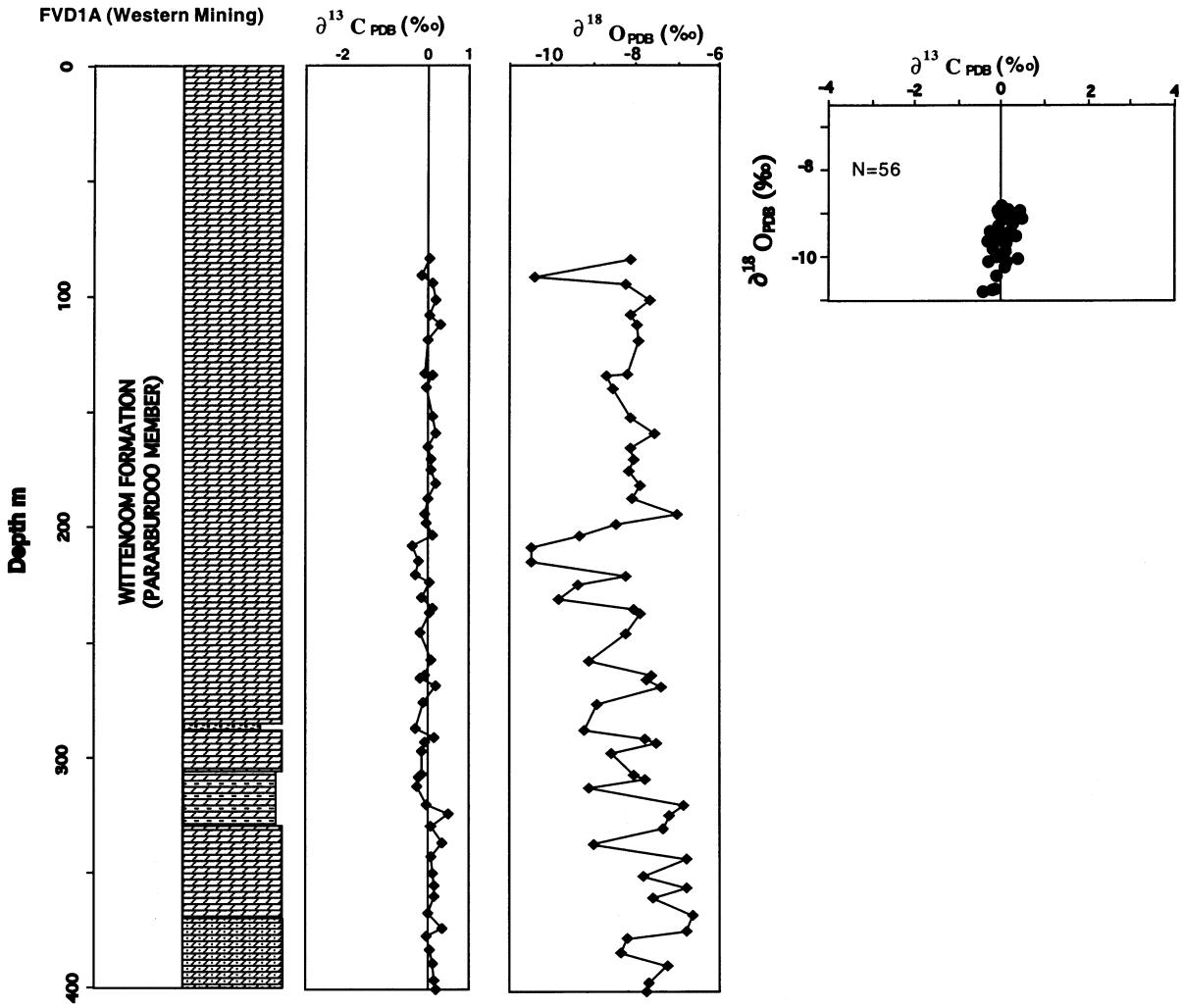

Fig. 7. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Western Mining drill hole FVD1A through the Paraburdoo Member of the Wittenoom Formation, Hamersley basin, Western Australia (see Table 1). See Fig. 2 for Key.

consistent with the model proposed by Simonson and Hassler (1996). The abrupt transition from BIF to carbonate deposition and the relatively low Fe content of the platform carbonates suggests that within the photic zone O_2 was abundant enough to eliminate Fe from the shallow platform setting.

The Bee Gorge Member of the Wittenoom Formation arguably forms the transgressive systems tract of a younger overlying transgressive systems tract that was fed sporadically by carbonate turbidites from the carbonate platform to the northwest. It is either time equivalent to the upper Carawine Dolomite or, more likely, related to a later carbonate interval that was removed through erosion around the basin margin.

Recently, Simonson and Hassler (1997) have carried out a detailed study of the Wittenoom Formation and the Carawine Dolomite and concluded, on the basis of impact-generated layers of glass spherules, that the two formations are not correlative and that the Carawine Dolomite lies stratigraphically above the Wittenoom Formation. Whilst this argument is appealing it is not

consistent with our understanding of regional facies relationships. On the basis of γ-log analysis (Figs. 5 and 6), the most parsimonious conclusion is that the two formations are correlative and form complimentary components of a single major depositional sequence. That is, the Carawine Dolomite is the product a major carbonate platform that fed detrital carbonates into a deeper water ramp setting to form the Wittenoom Formation. This interpretation implies that there may have been several distinct layers of glass spherules.

Five drill cores have been sampled to provide a stable isotope stratigraphy for the Hamersley Group. Two of the drill cores penetrate the Carawine Dolomite (Figs. 3 and 4), whilst three cores were sampled from the Wittenoom Formation (Figs. 5–7). The sampled Carawine Dolomite proved to be relatively uniform in composition with little siliciclastic material (typically less than 2.0%). The carbonates are dolomitic and compositionally very uniform (Table 2) with Mg/Ca being very consistent at a mean value of 0.53. The iron content of the carbonates is generally relatively low, less than 1%, except near the base of the formation, where it can be as high as 10%. The Wittenoom Formation is also dolomitic and very uniform in composition with a mean Mg/Ca ratio of 0.62 and a low silica content (typically less than 2.0%). The Fe content of these carbonates compared with the underlying sediments is relatively low at approximately 2%.

$\delta^{13}C_{carb}$ data from all three of the Wittenoom-Formation drill cores are very consistent, with mean values close to zero and small standard deviations (Table 1). $\delta^{18}O_{carb}$ values are considerably more erratic, although associated standard deviations (S.D.) are similar for all three cores. $\delta^{13}C_{carb}$ data from the Carawine Dolomite show slightly more variability than those of the Wittenoom Formation but there is no statistical difference (Table 1). In both formations the isotopic composition appears unrelated to Fe content. $\delta^{18}O_{carb}$ values are also similar to those of the Wittenoom Formation. Crossplots of $\delta^{13}C_{carb}$ versus $\delta^{18}O_{carb}$ indicate that there is no significant covariance for samples from either the Wittenoom Formation or the Carawine Dolomite (Figs. 3–7). Overall the data indicate that the $\delta^{13}C_{carb}$ curve was very flat during this time period and provide no evidence to support a difference in age or diagenetic history for the two formations.

Table 2
Summary statistics of relevant major and trace element geochemistry for the main platform carbonates sampled for the present study

Formation	Number of samples	Si (%)	Fe (%)	Mn (%)	Ca (%)	Mg (%)	Mg/Ca	Sr (ppm)
Hamersley basin								
Carawine Dolomite	83	1.6 ± 6.2	1.8 ± 2.5	0.7 ± 0.5	22.4 ± 5.3	11.8 ± 4.4	0.53 ± 0.19	18.5 ± 36.1
Wittenoom Formation	108	4.4 ± 13.2	1.6 ± 2.6	0.6 ± 0.3	20.5 ± 10.9	12.1 ± 6.5	0.62 ± 0.38	30.2 ± 14.5
Ashburton basin								
Duck Creek Dolomite	21	5.7 ± 7.6	1.2 ± 2.0	0.1 ± 0.1	20.2 ± 4.4	10.1 ± 2.6	0.50 ± 0.07	33.7 ± 14.4
Wooly Dolomite	16	2.5 ± 2.5	3.0 ± 1.7	0.1 ± 0.1	21.0 ± 2.2	11.2 ± 1.5	0.53 ± 0.03	31.5 ± 9.6
Kazput Formation	22	18.6 ± 10.6	4.3 ± 1.7	0.2 ± 0.2	9.5 ± 7.7	5.4 ± 2.9	0.57 ± 0.38	151.2 ± 105.8
Meteorite Bore Member	1	7.8	4	0.3	19.3	6.1	0.32	149.9
Earaheedy basin								
Yelma Formation	17	3.3 ± 4.9	0.4 ± 0.2	0.9 ± 0.1	20.1 ± 3.1	11.7 ± 1.7	0.58 ± 0.01	n/a

4.1.2. Turee Creek Group

The Turee Creek Group is more than 3000 m thick and consists of three formations, two of which, the Kungarra and Kazput Formations, contain carbonates (Thorne et al., 1995; Fig. 2). The Kungarra Formation is dominantly clastic in composition (approximately 2000 m of siltstone and fine sandstone) and includes the Meteorite Bore Member, a diamictite interval that has been correlated globally with the Huronian glaciation (Trendall, 1976; Martin, 1999) thus providing an important time line. The Meteorite Bore Member lies approximately 1800 m above the base of the formation and consists of approximately 270 m of diamictites, shales, sandstones with thin locally distributed interbedded carbonates. BIF is present locally and the member can be shown to be laterally equivalent to massive BIF and shale (Martin, 1999).

Thin (< 10 cm) dolomitic diamictites interbedded within siliciclastic diamictites of the Meteorite Bore Member were sampled at Yeera Bluff (near Deepdale on the Pannawonica Sheet, Table 1). The carbonate clasts are similar in composition to platform carbonates elsewhere in the succession (Table 2). Silica, iron and manganese contents are all low. $\delta^{18}O_{carb}$ and $\delta^{18}O_{carb}$ values lie within a narrow range comparable to those of the underlying Carawine Dolomite and Wittenoom Formation (Table 1).

The Kazput Formation, which conformably overlies the Koolbye and Kungarra Formations, forms the final stages of a major shallowing upward succession (upper highstand systems tract). The formation, which reaches 1100 m in thickness, consists of mudstones, siltstone, sandstones and conglomerates with rhythmic bands of silty dolomite and minor basalts and BIFs (Thorne and Tyler, 1997). The succession was deposited in a platform setting which was succeeded by a prograding deltaic environment. A section through a large part of the formation was sampled along Horseshoe Creek (Hardey Sheet) (Table 1, Fig. 8). With a mean silica content of 18.6%, the carbonates are significantly more silica-rich than other platform carbonates in the basin (Table 2). Sr values are similarly high, perhaps reflecting the onset of foreland tectonics or the after-effects of the earlier glaciation. Other major elements such as Fe, Mn, Ca and Mg are, however, little different from those of the other intervals. $\delta^{13}C_{carb}$ data from the Kazput Formation samples show greater variance than formations lower in the section (Table 1) while $\delta^{18}O_{carb}$ data are similar to those of earlier formations. A crossplot of the two variables indicates that there is no significant correlation suggesting that the high variance could well reflect the primary $\delta^{13}C_{carb}$ signal.

4.2. Ashburton basin

The Wyloo Group, which forms the fill of the Ashburton basin, lies unconformably over the Hamersley basin succession and onlaps the northern margin of the Capricorn Orogen covering an area of approximately 30 000 km^2 (Fig. 2). The group consists of 12 km of sedimentary and volcanic rocks deposited between ca. 2.2 and 1.8 Ga. The succession has been interpreted as the response of the southern margin of the Pilbara Craton as it shifted from an active margin to a foreland basin (Thorne and Seymour, 1991). The early Wyloo Group sediments were deposited during the final stages of ocean basin closure with the development of a foreland bulge as the ocean floor was loaded by the approaching and overriding Yilgarn Craton. Two major carbonate intervals, the Wooly and Duck Creek Dolomites, are present within this predominantly clastic foreland basin succession and have been analysed in this study.

The Wooly Dolomite (Seymour et al., 1988), which is areally restricted to the southern margin of the Wyloo Dome, reaches a maximum of 325 m in thickness but is only 60 m thick at the sample location some 20 km from Wyloo Station (Table 1). It consists of well stratified stromatolitic dolostone units which, towards the base of the formation, are also crossbedded. The top of the formation is defined by a prominent irregular surface which is locally ferruginous. The surface, a major sequence boundary, is abruptly overlain by the clastic sediments of the Mount McGrath Formation. The upward shallowing sequence was

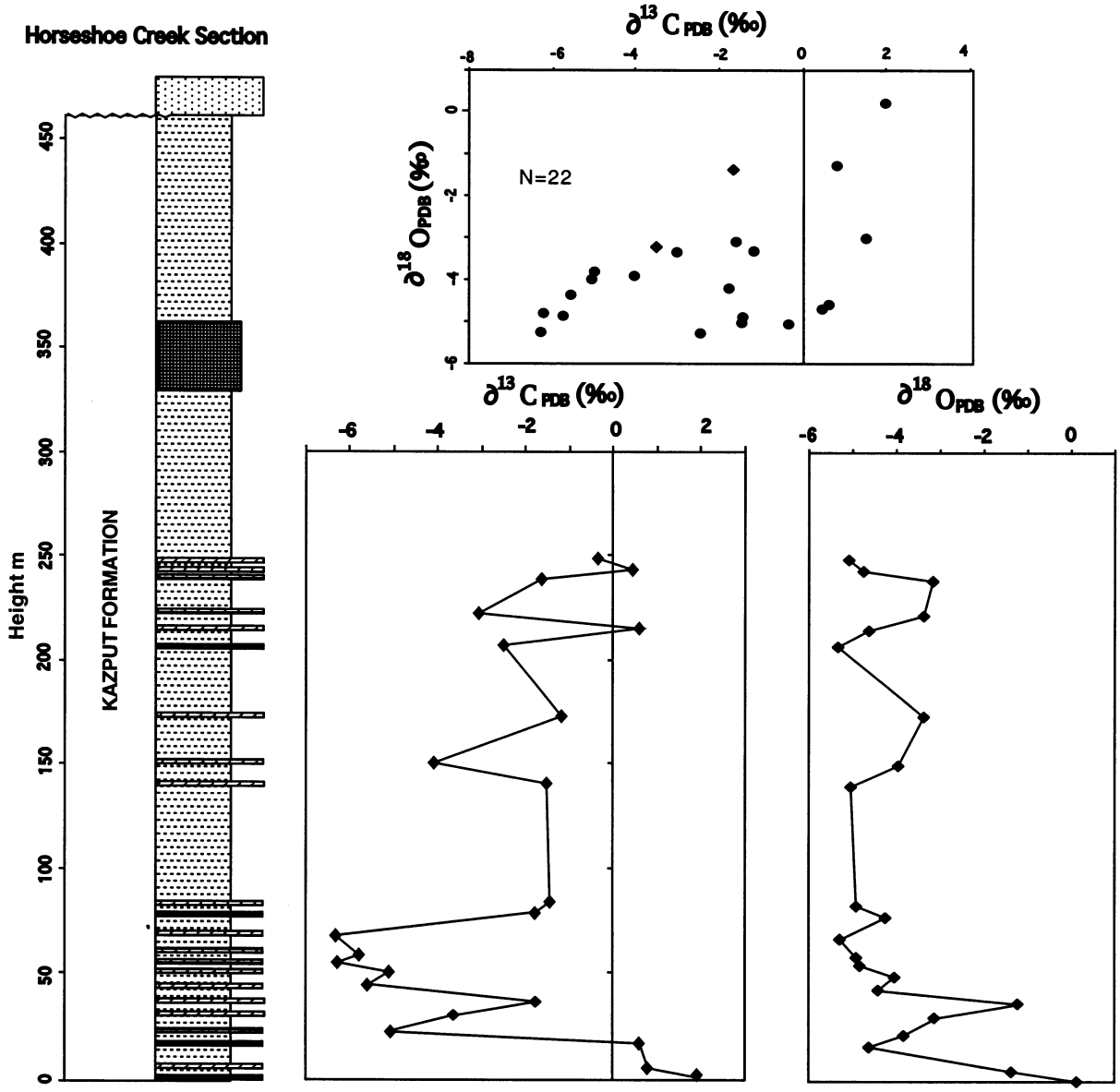

Fig. 8. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Kazput Formation sampled along Horseshoe Creek, Hamersley basin (see Table 1). See Fig. 2 for Key.

deposited in a platform or ramp/platform transitional setting with much of the sedimentation being tidal or shallow sub-tidal. It is a remnant of a major depositional sequence, the upper part of which may have been removed by erosion while the lower part was disrupted by volcanism. The restricted distribution of the formation appears to be the result of tectonism resulting in localised starvation of siliciclastic sediments.

Compositionally, the Wooly Dolomite is similar to the other platform carbonates and has a low silica content (Table 2). It also has a low Fe, Mn and Sr content and a comparable Mg/Ca ratio of between 0.5 and 0.6. $\delta^{13}C_{carb}$ data show a shift

from −4‰ at the base of the formation to a single sample with a value of close to +6‰ near the top (Fig. 9). Similar shifts are present in the $\delta^{18}O_{carb}$ data, however, there is no significant correlation between the two isotopic signatures, suggesting that the $\delta^{13}C_{carb}$ signal probably preserves primary values. It is notable that the $\delta^{13}C_{carb}$ values increase upward in parallel with inferred upward shallowing suggesting a possible connection. Ferruginous cements in the lower part of the section suggests that the depositional environment may have been oxygen-depleted deeper waters.

Siliciclastic-sediment starvation accompanying crustal downwarping resulted in widespread carbonate deposition on a distally steepened ramp in the Ashburton basin (Thorne and Seymour, 1991). The resulting Duck Creek Dolomite conformably overlies the clastic sediments of the Mount McGrath Formation and forms the upper part of a major shallowing-upward depositional sequence and the transgressive systems tract of an overlying sequence (Fig. 2). The basal and middle units form a shallowing-upward succession at the top of the sequence above the deeper water clastics of the Mount McGrath Formation. The depositional environment gradually shifted from a carbonate-ramp setting lower in the formation to a shallower platform setting. A sequence boundary between the upper two units marks the transition to the new sequence as sealevel rose and water depth increased, shifting the depositional environment back to a transgressive ramp setting before passing into deeper water clastics of the overlying Ashburton Formation.

The Duck Creek Dolomite, which is locally as much as 1000 m thick, consists of three discrete units (Seymour et al., 1988). The basal and upper units consist of thick-bedded dolorudite alternat-

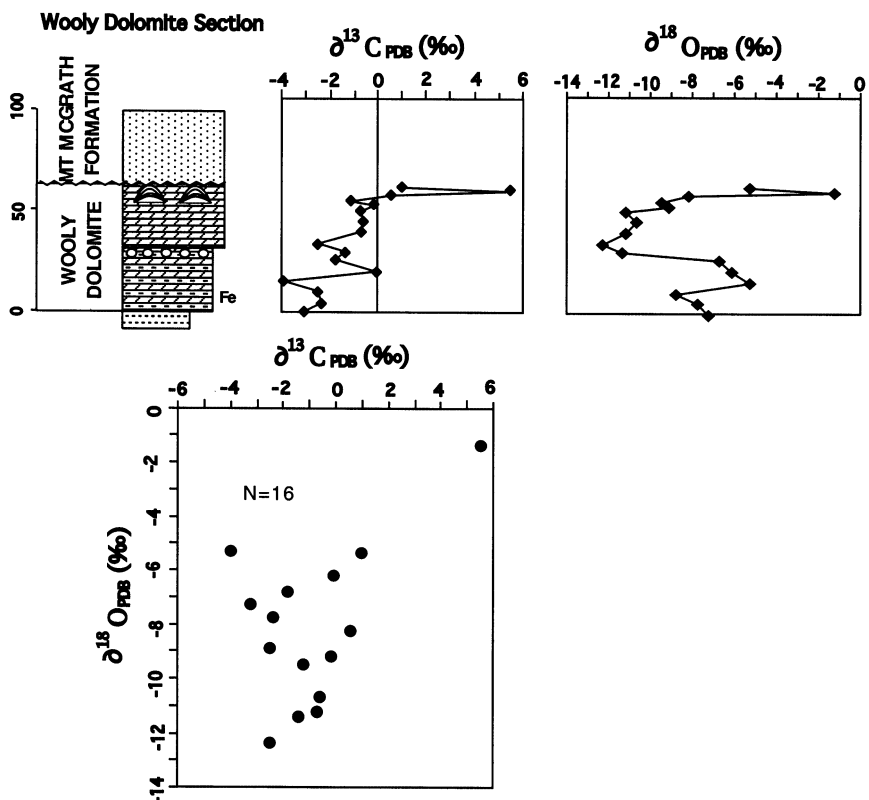

Fig. 9. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Wooly Dolomite sampled in the Mount De Courcey Area, Ashburton basin (see Table 1). See Fig. 2 for Key.

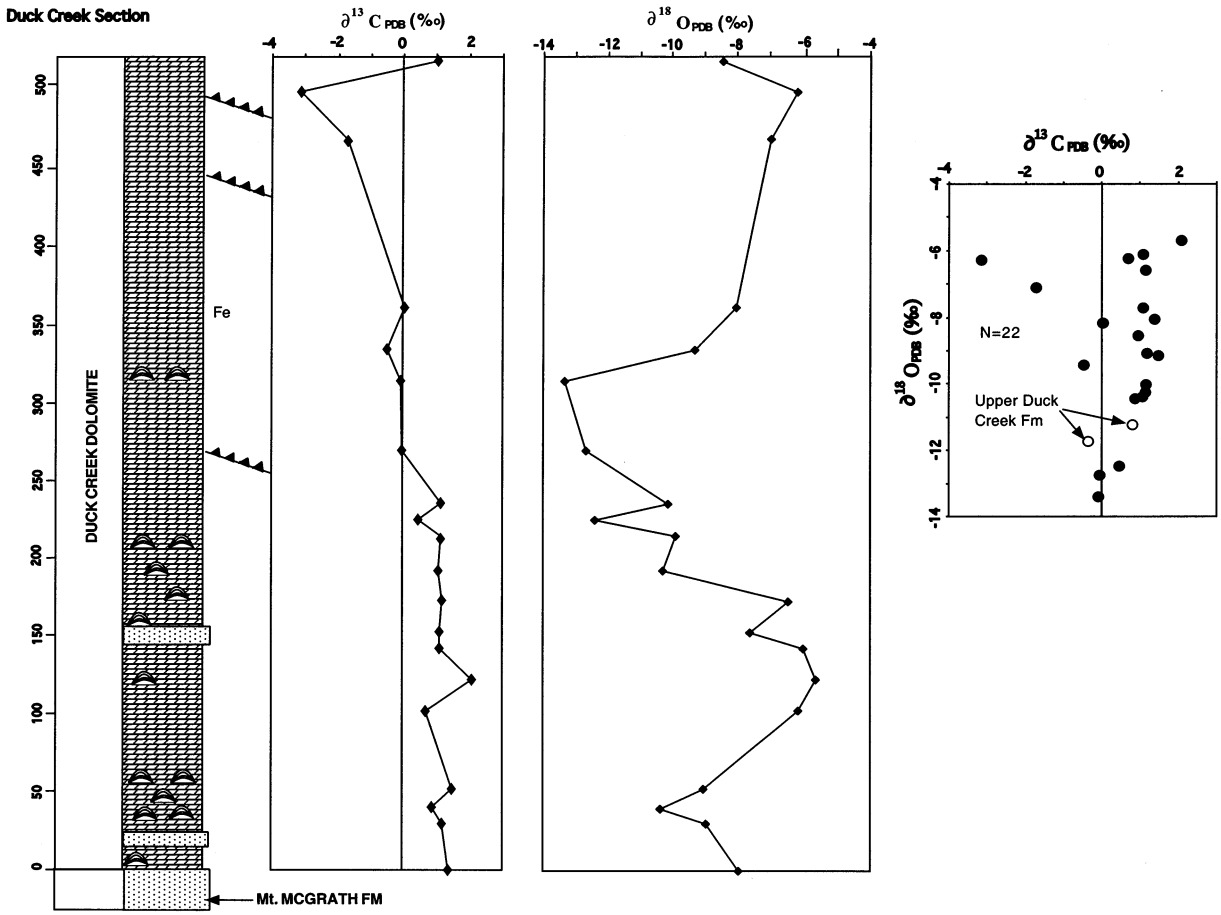

Fig. 10. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the Duck Creek Dolomite sampled along the type section of the formation at Duck Creek, Ashburton basin (see Table 1). See Fig. 2 for Key.

ing with interbeds of thinly bedded dolostone with nodular chert layers. Mudstones containing synsedimentary folds occur interbedded with the dolostone. The middle unit of the Duck Creek Dolomite contains stromatolites that typically occur at the tops of regular shallowing upward parasequences (Thorne, 1985a,b; Grey and Thorne, 1985). Domal stromatolites of at least three types (Grey, 1985) alternate with planar stromatolitic dolostone and dolomitic grainstones which may contain stromatolitic fragments. Locally the formation is heavily silicified, although, as discussed below, higher silica content is usually associated with the transgressive systems tracts.

More than 500 m of the lower to middle Duck Creek Formation was sampled at the type locality around Duck Creek (Table 1, Fig. 10; see Thorne, 1985a,b). The Duck Creek Dolomite samples are relatively pure carbonates with a low silica content (Table 2). Fe and Mn are also low and, along with Mg/Ca values, are similar to other platform carbonates. Silica values were found to be elevated within thin transgressive systems tracts at the base of inferred shallowing upward cycles. Crossplots of $\delta^{13}C_{carb}$ versus $\delta^{18}O_{carb}$ show a considerable spread of $\delta^{18}O_{carb}$ values but relatively tightly clustered $\delta^{13}C_{carb}$ values, with little indication of cross correlation. $\delta^{13}C_{carb}$ varies from + 2‰ at the base of the section with a gradual upward decline to − 3‰ near the top. Two additional samples were collected from the upper

Duck Creek Dolomite approximately 250 m above the main sampling interval. The samples have $\delta^{13}C_{carb}$ values of +0.8 and −0.3 with $\delta^{18}O_{carb}$ values of −11.2 and −11.7, respectively. They are thus well within the range of values obtained from the lower Duck Creek Dolomite samples.

4.3. Yerrida basin

The Yerrida basin covers the relatively small area of 10 000 km^2 (Fig. 1). The basin formed in a back-arc setting, initially by sagging, and subsequently by rifting, of continental crust along the margin of the Yilgarn Craton just prior to the collision of the cratons (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Pirajno et al., 1998; Occhipinti et al., 1998). The basin fill, the Yerrida Group, consists of seven formations that are largely siliciclastic or volcanogenic. However, the two lower formations, the Juderina and Johnson Cairn Formations, contain carbonate intervals (Fig. 11). The Juderina Formation, which averages 160 m in thickness, is a mixed clastic and carbonate succession the lower part of which consists of sandstones and siltstones while the upper part consists of microbial laminites and stromatolitic dolostones. Locally the carbonates include silicified evaporitic facies (El Tabakh et al., 1999). The interval forms an overall upward-shallowing depositional sequence. The Johnson Cairn Formation is predominantly laminated, iron-rich shale but includes thin graded silty units and thin carbonate interbeds, some oolitic. Russell (1992) carried out a Pb–Pb analysis of the carbonates of the Juderina Formation and established an isochron that provided an age of 2258 ± 180 Ma. A later Pb–Pb analysis by Woodhead and Hergt (1997) refined the age to 2173 ± 64 Ma.

Two drill cores (QMW 83-1 and PP-011, Figs. 12 and 13) that penetrate the Juderina Formation were sampled and analysed for carbon and oxygen isotopes. Both cores provide high positive $\delta^{13}C_{carb}$ values averaging 7.69 ± 0.11 and 6.57 ± 0.56‰, respectively. Russell (1992) also measured a high positive $\delta^{13}C_{carb}$ value on one outcrop sample from the Juderina Formation. $\delta^{18}O_{carb}$ values are not significantly different from those encountered in other formations whether older or younger (Table 1). Outcrop samples of the Juderina Formation were also analysed and produced similar values. $\delta^{18}O_{carb}$ values for the interval fall within a very narrow range and cluster tightly when plotted against the $\delta^{13}C_{carb}$ data suggesting little diagenetic alteration. The Johnson Cairn Formation is not cored but was sampled in outcrop. A single sample from a thin carbonate unit gave a $\delta^{13}C_{carb}$ value of −0.4‰ and a $\delta^{18}O_{carb}$ value of −7.6‰.

4.4. Earaheedy basin

The Earaheedy basin covers approximately 40 000 km^2 (Fig. 1) and contains a single lithostratigraphic group, the Earaheedy Group consisting of six formations (Fig. 11). The Earaheedy basin postdates the Yerrida basin and appears to have resulted from sagging of the continental crust in a back-arc setting during the collision of the Pilbara and Yilgarn Cratons (Tyler and Thorne, 1990). The basin fill is largely siliciclastic but includes some granular iron formation and carbonate rocks in the Frere Formation and carbonate rocks in the Yelma and Windidda Formations and the Kulele Limestone. Pb–Pb analysis of whole rock carbonate samples from the Yelma Formation at the base of the succession indicate that the basin was initiated at approximately 1.9 Ga (Russell, 1992). There is, however, a large error bar associated with this age.

Drill core is available for the Yelma Formation but only isolated outcrop samples were available for the Frere and Windidda Formations and the Kulele Limestone (Table 1, Fig. 11). Three closely spaced drill cores that penetrate part of the Yelma Formation were sampled (TDH-01, TDH-26, TDH-28, Table 1, Figs. 14–16). TDH-01 penetrates the lower Yelma while TDH-26 and 28 provide overlapping information on the upper part of the formation. Small numbers of outcrop samples were also collected from the Frere and Windidda Formations and the Kulele Limestone (Table 1).

Closely spaced sampling (1–5 m) of the three cores from the Yelma Formation provide a well defined curve for this interval. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ crossplots for all three cores provide no evidence of correlation suggesting that the signatures are primary. This is further supported by the close match of duplicate data provided by TDH-26 and TDH-28, both show a well defined positive excursion of +2.0‰. By contrast the lower part of the formation, as defined by drill-core THD-01, has a relatively flat curve that averages +0.63‰ (Table 1). The excursion at the top of the formation closely matches a similar excursion measured in the lower and middle Duck Creek Dolomite in the Ashburton basin (Fig. 10).

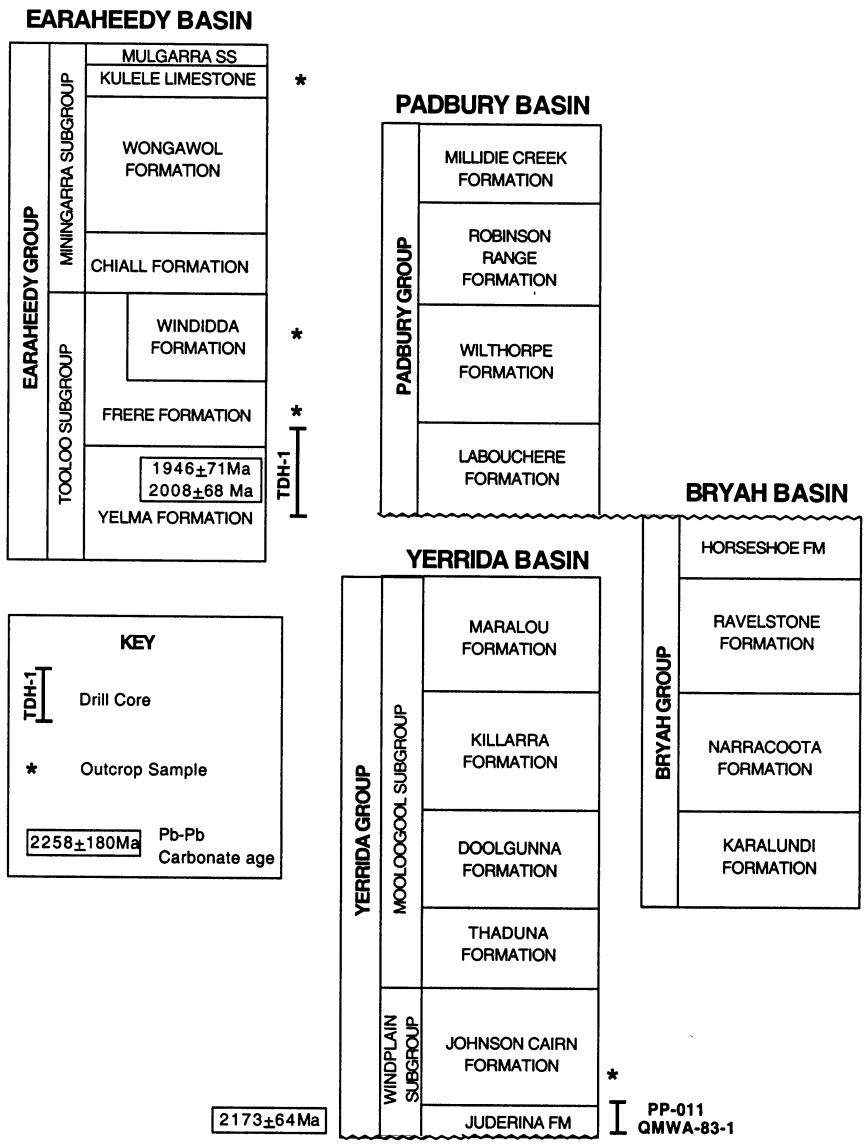

Fig. 11. Simplified stratigraphy for the Yerrida, Bryah, Padbury and Earaheedy basins. Adapted from Occhipinti et al. (1997), Pirajno et al. (1998). Pb–Pb ages from Russell (1992), Woodhead and Hergt (1997).

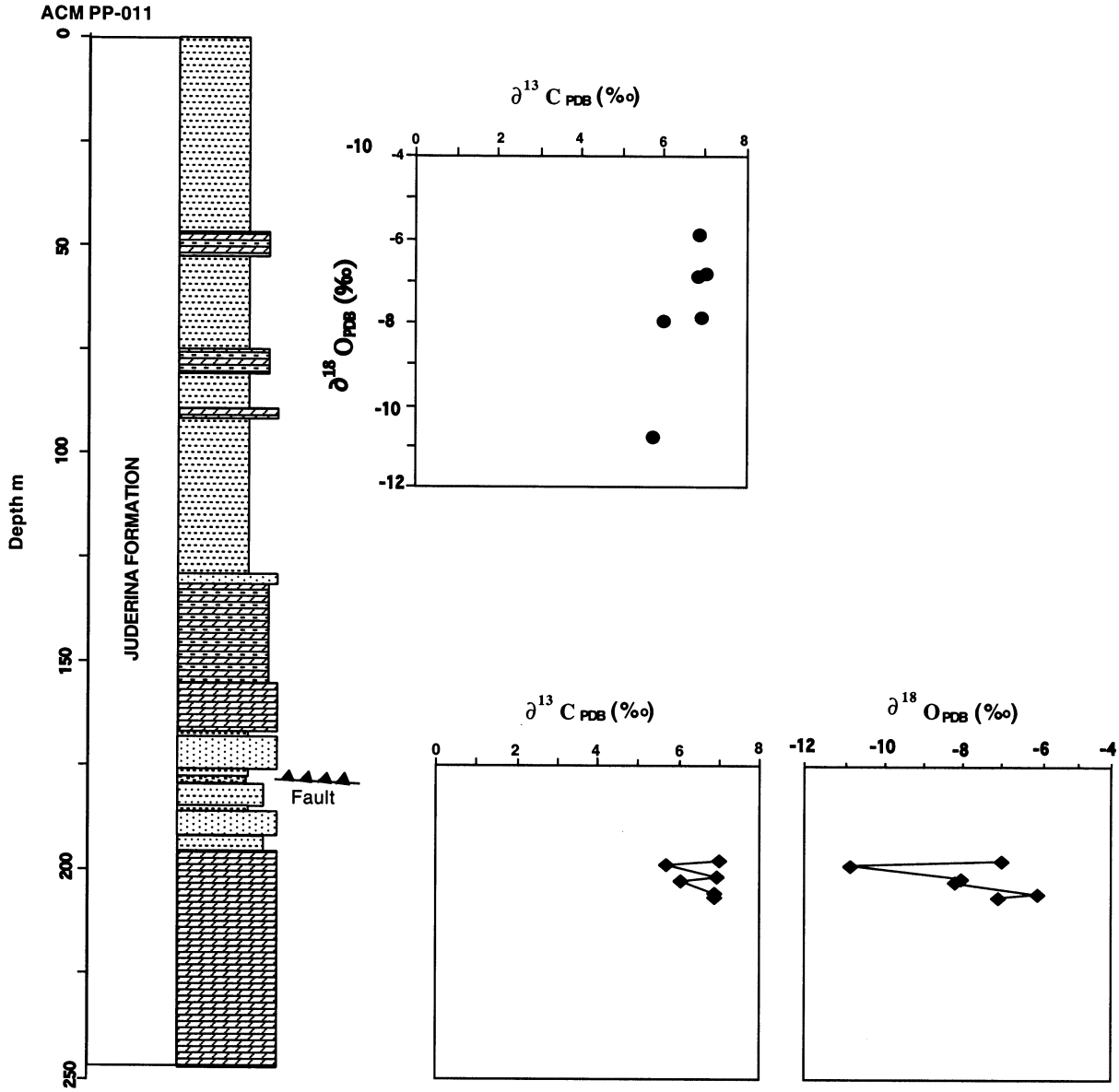

Fig. 12. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of Australian Consolidated Mineral's drill core PP-011 through the Juderina Formation, Yerrida basin, Western Australia (see Table 1). Note the very heavy δ^{13}Ccarb values.). See Fig. 2 for Key.

Analyses of outcrop samples from the Frere and Windidda Formations and the Kulele Limestone provide $\delta^{13}C_{carb}$ values of -0.7 and $+0.5$‰, respectively (Table 1) suggesting that the curve has again become monotonic. $\delta^{18}O_{carb}$ values are within the range encountered elsewhere in the back-arc basins (Table 1).

5. Discussion

Although the succession of carbonate sediments in the Western Australian basins is discontinuous, it provides a valuable sample of what we take to be the global oceanic carbon isotope signature during a critical span of time in Earth history (ca.

2.7–1.8 Ga) when the oxygen content of the atmosphere is interpreted to have been increasing (Des Marais et al., 1992). U–PB dates through the Hamersley and Ashburton basins provide good time constraints (Fig. 2). Dating is less secure for sediments from the smaller back-arc basins. However, their tectono-stratigraphic relations are well established and in combination with a small number of Pb–Pb carbonate ages it is possible to establish a reasoned timescale (Fig. 11). The relationship between the stratigraphy of the back-basins to that of the Ashburton foreland is strengthened by the correlation of the positive $\delta^{13}C_{carb}$ excursion measured in the lower and middle Duck Creek Dolomite (Fig. 10) and the upper Yelma Formation (Figs. 15 and 16).

Our data begin in the lower Hamersley basin succession, with a very flat $\delta^{13}C_{carb}$ curve consisting of values close to 0‰ that are reminiscent of data obtained from late Palaeoproterozoic to Mesoproterozoic rocks in the Mount Isa and McArthur basins of northern Australia (1.7–1.5 Ga; Fig. 17). The northern Australian data were interpreted as reflecting a period when the global ocean-atmosphere system reached a prolonged state of equilibrium with respect to the carbon

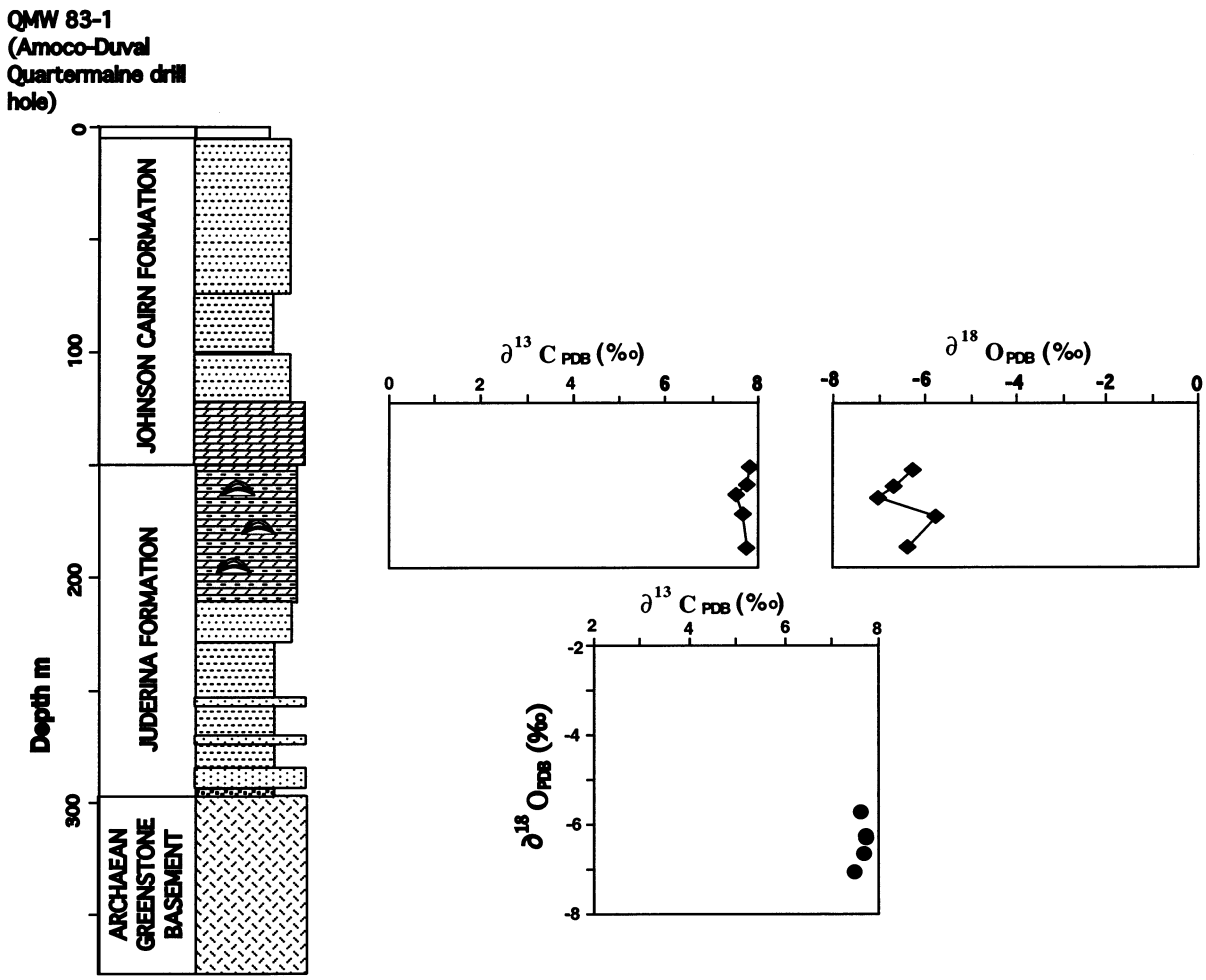

Fig. 13. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of Amoco-Duval's Quartermaine drill core QMW 83-1 through the Juderina Formation, Yerrida basin, Western Australia. Note the exceptionally heavy $\delta^{13}C_{carb}$ values (see Table 1). See Fig. 2 for Key.

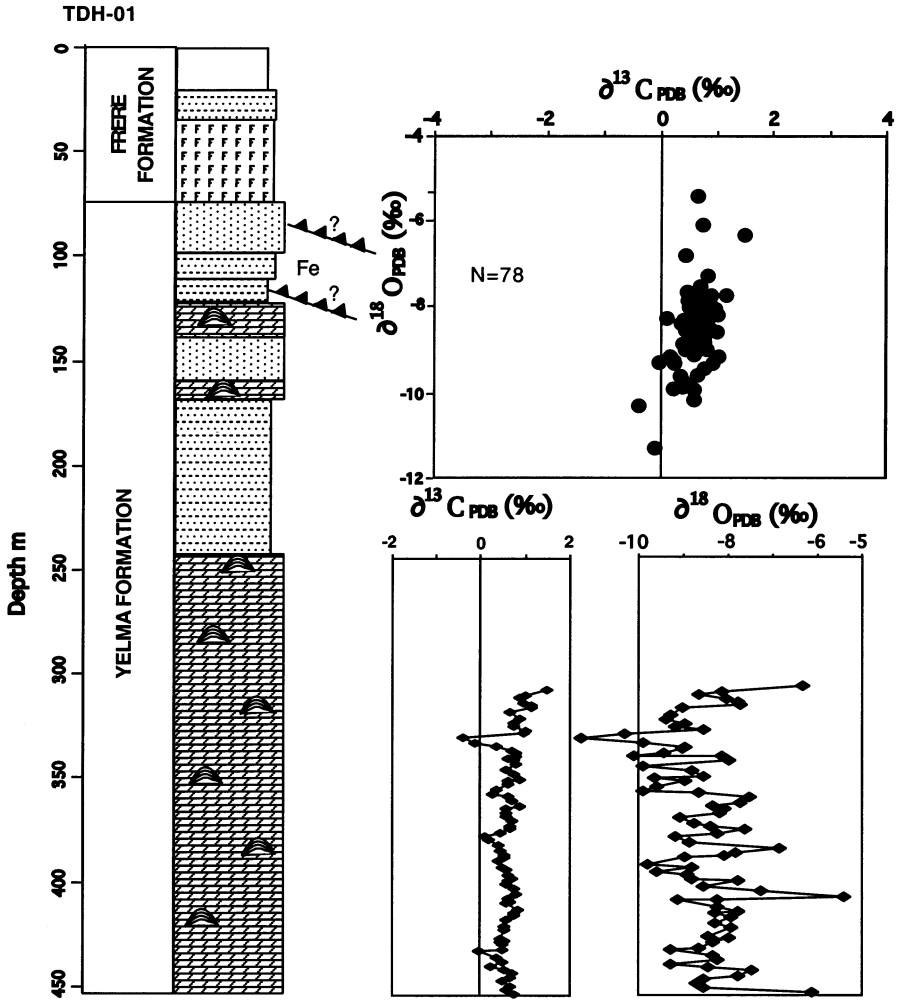

Fig. 14. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the TDH-01 drill hole through the lower Yelma Formation in the Earaheedy basin, Western Australia (see Table 1). See Fig. 2 for Key.

cycle (Brasier and Lindsay, 1998; Lindsay and Brasier, 2000). To maintain the carbon mass balance in such a stable state arguably requires relatively low levels of tectonic activity. The 2.7–2.2 Ga old Hamersley basin also appears to have developed during a period of continental stability (see Tyler and Thorne, 1990 for a synthesis). This is the first clearly identifiable basinal setting preserved on the Australian craton in which marine sediments have been preserved in a response to broad, regional, crustal subsidence. Comparison of the architecture of the Pilbara region (and the Hamersley basin) with the Kaapvaal region of southern Africa (e.g. Cheney, 1996) has been used to suggest that, by late Archaean time, Australia formed part of a large crustal block, perhaps the first crustal aggregation that could be referred to as a continent or supercontinent. More recent geochronological studies by Nelson et al. (1999) suggest, however, that the two cratonic blocks evolved at about the same time but independently. Either way the data suggest that by Hamersley basin time, major continental blocks had evolved. The flat secular $\delta^{13}C_{carb}$ curve encountered in carbonate rocks of the lower Hamersley basin successions could, therefore, be taken to reflect a

tectonically quiescent period associated with development of the first stable continental blocks and associated cratonic basin sedimentation and an indication that the role of photosynthesis in the biosphere was well established by that time. This interval (2.7–2.2 Ga) was a time when oceanic geochemistry is likely to have been dominated by the mantle flux rather than the continental flux (Veizer et al., 1982) and sedimentation was largely chemical or biogeochemical. Transgressive systems tracts at this time are dominated by cherts and BIFs whilst highstand deposits grade upward into platform carbonates (Fig. 2). The geometry of the sequences and their stacking patterns are comparable with those encountered in younger intracratonic settings (e.g. the Neoproterozoic Amadeus basin, Lindsay et al., 1993).

The isotopic record that follows the early Hamersley basin data is much more variable, with swings from as low as −5.0 to a high of +9.0‰, suggesting that rapid changes were occurring in the partitioning of carbon between major reservoirs and/or the biosphere. These changes occurred as the Pilbara and Yilgarn Craton began to converge and prior to the development of an active margin. By ca. 2.3 Ga, closure of the ocean was underway and the Pilbara and Yilgarn cratonic margins became active, with oceanic crust being subducted beneath the Yilgarn Craton.

The first evidence of complexity in the curve appears in association with, and immediately fol-

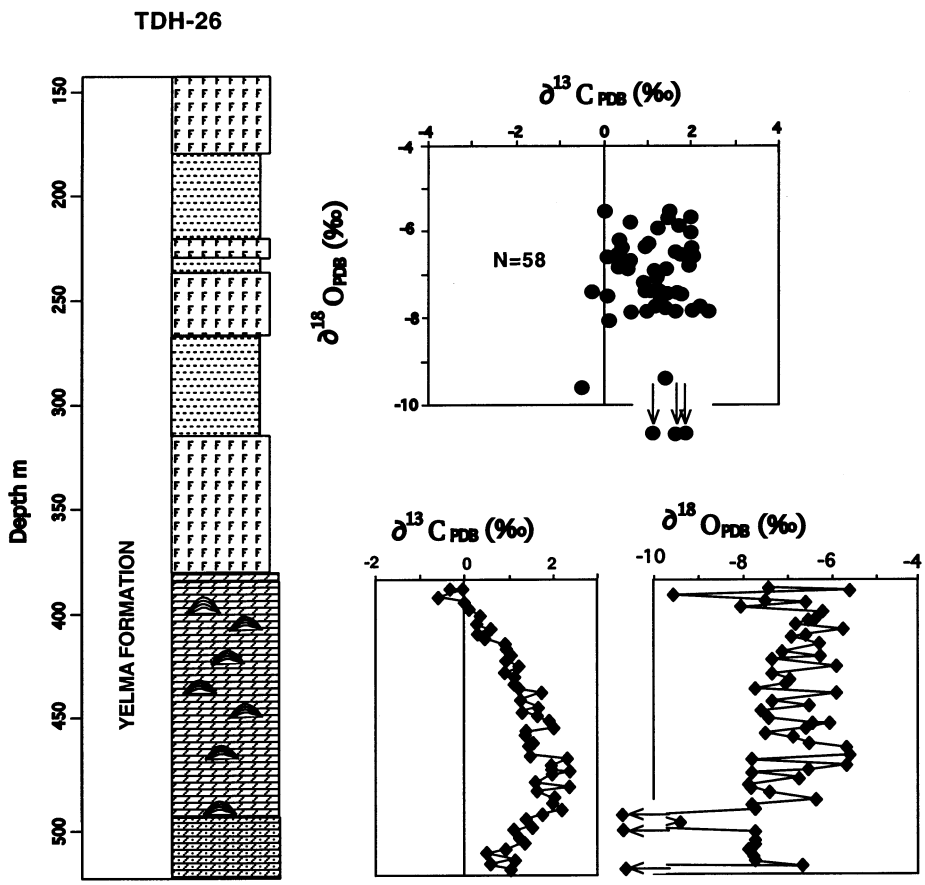

Fig. 15. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the TDH-26 drill hole through the upper Yelma Formation in the Earaheedy basin, Western Australia (see Table 1). See Fig. 2 for Key.

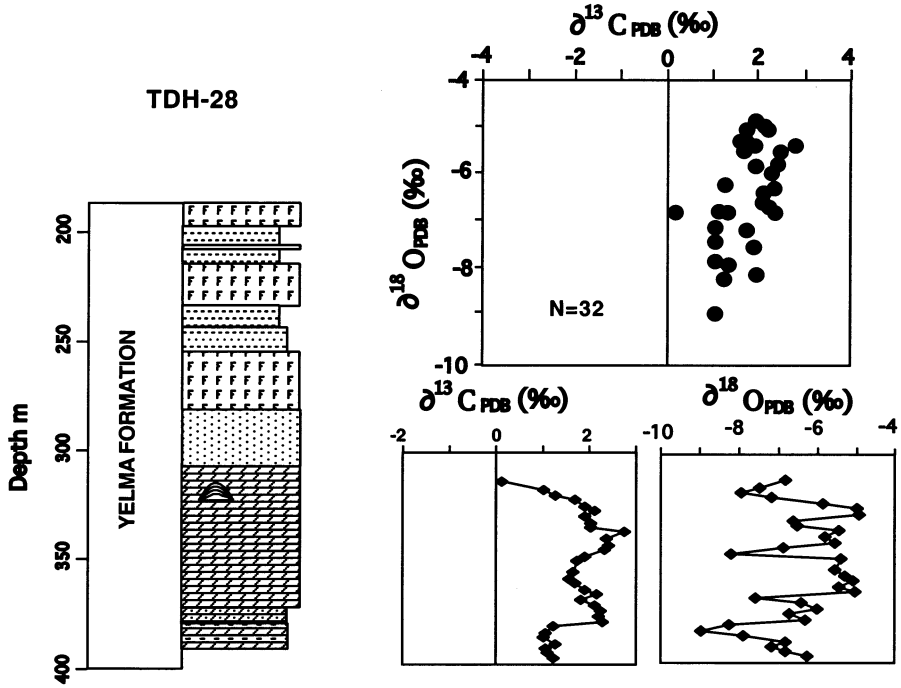

Fig. 16. $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ stratigraphy of the TDH-28 drill hole through the upper Yelma Formation in the Earaheedy basin, Western Australia (see Table 1). Note consistency with Fig. 15. See Fig. 2 for Key.

lowing, the glacial sediments of the Turee Group. A small poorly documented negative excursion is associated with thin carbonates in the Turee Creek Group which is in turn followed by a well defined negative excursion of up to −5‰. Only a single glacial episode has been recognised in the Hamersley basin (Martin, 1999) whereas three discrete events have been documented on the South African and North American cratons (Eriksson and Clendenin, 1990; Young, 1973, 1988). Available zircon U–Pb ages constrain the Hamersley basin glaciation to 2.45–2.20 Ga which is within the age constraints so far established for the North American, Fennoscandian and South African successions (Martin, 1999). It is thus conceivable that the second negative excursion that follows the Turee Creek glaciation (Kazput Formation) may relate to a second glaciation not documented sedimentologically in the Australian succession (Fig. 17).

A major positive isotopic excursion (+9‰) is recorded in the Juderinah Formation. Unfortunately, the back-arc basins containing this record are dominated by clastic rocks such that the isotopic record is fragmentary and age constraints limited. Carbonate Pb–Pb dates (Russell, 1992; Woodhead and Hergt, 1997) place this event at ca. 2.17 Ga, which is consistent with the regional tectonostratigraphic model. The event thus appears to lie stratigraphically well above the glacials of the Turee Group forming a consistent pattern with data from the other ancient cratons (cf. Bekker et al., 2001). The implications are that this is a complex interval during which several positive excursions with intervening negative excursions (associated with glacial events much as is seen in the Neoproterozoic record) and then a major positive excursion following the Huronian glaciations.

The major positive excursion and associated period of isotopic complexity recorded in the Western Australian $\delta^{13}C_{carb}$ succession coincides with the 'Lomagundi Event', which includes a major positive early Palaeoproterozoic excursion

in the secular $\delta^{13}C_{carb}$ curve. This event was first noted by Schidlowski et al. (1976) in samples from the Lomagundi Formation of Zimbabwe its global significance was only recognised when similar values were measured on rock samples from Norway, Scotland and North America (Fallick and Hamilton, 1989; Melezhik and Fallick, 1994). Recently, Karhu and Holland (1996), Bau et al. (1999) have compiled available carbon isotope data for the earliest Palaeoproterozoic and found evidence of a major $\delta^{13}C_{carb}$ positive excursion of ca. +12.0‰ in the interval from 2.22 to 2.06 Ga. They associated this distinctive excursion with a rise in atmospheric oxygen as carbon was biogenically fixed by photosynthesis and then sequestered.

It is interesting to note that, as subduction was initiated (Tyler and Thorne, 1990) the major positive $\delta^{13}C_{carb}$ excursion is recorded in back-arc settings in Western Australia. We here draw attention to the fact that this first major isotopic excursion followed the first appearance of basins on the Australian continent (Hamersley basin) and the beginning of ocean closure and subduction between the Pilbara and Yilgarn Cratons. This raises the possibility of a causal connection between the Lomagundi excursion and the plate tectonic evolution of the Earth's crust. A plausible explanation, for example, could be the subduction of biogenic carbon reservoirs into the mantle and

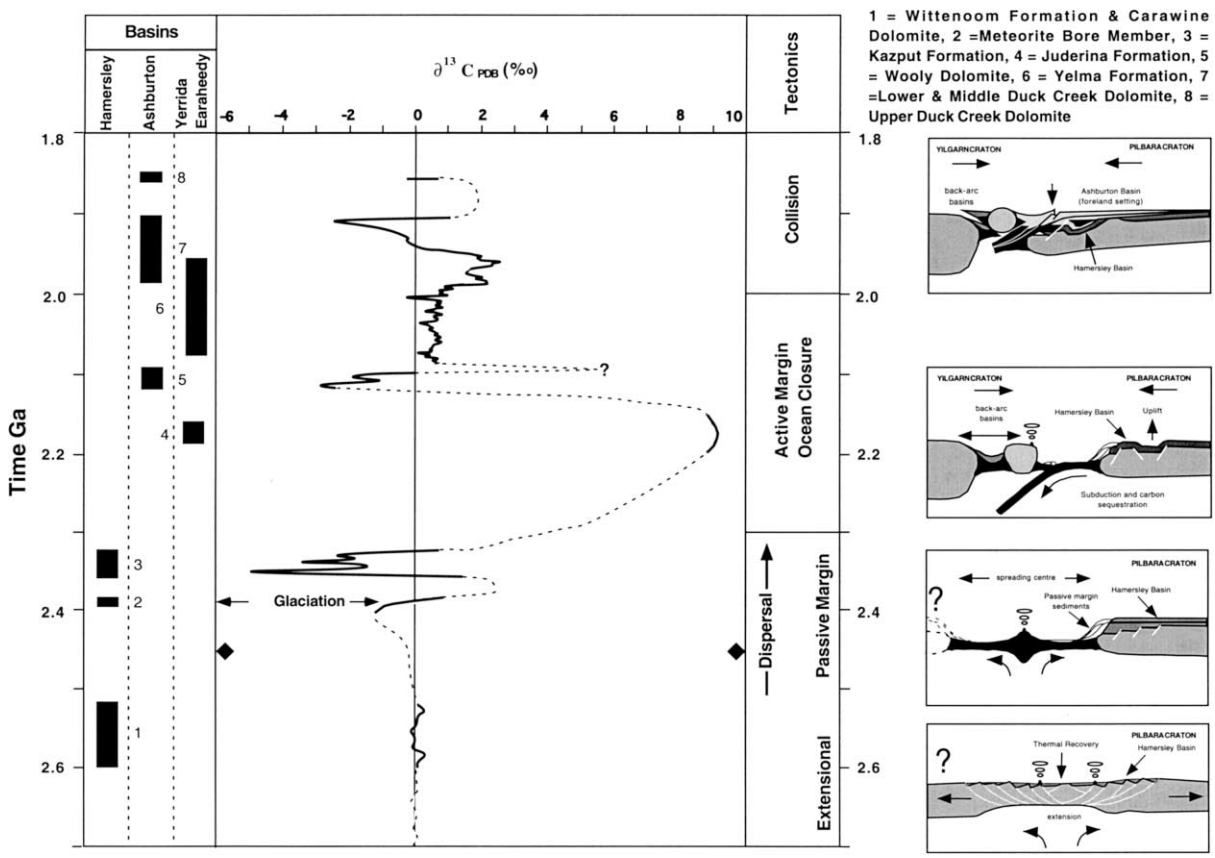

Fig. 17. Composite secular carbon isotope curve for the time interval 2.6–1.8 Ga based on data from Western Australian basins presented in this paper. Note the abrupt increase in the $\delta^{13}C_{carb}$ values at ca. 2.3 Ga as the Pilbara and Yilgarn Block converged marking the transition from a passive to an active margin setting and the subduction of intervening ocean floor sediments beneath the Yilgarn Craton. Tectonic outline based on Tyler and Thorne (1990), Occhipinti et al. (1998), Pirajno et al. (1998). Black diamonds (2.47–2.45 Ga) indicate the abrupt period of mantle overturn proposed by Kump et al. (2001).

lower crust, preferentially removing lighter carbon from the biosphere and leaving the global ocean enriched in heavy carbon and the atmosphere richer in oxygen. The extreme rate of ^{13}C removal, compared with most later time periods (except the Cryogenian and terminal Neoproterozoic) could be explained if the oceans were becoming strongly stratified, with large anoxic layers beneath a thin oxic layer—a model consistent with observed facies patterns in the transition from BIF to platform carbonates in the Hamersley basin. An incremental increase in atmospheric oxygen at this time has been proposed by Des Marais et al. (1992), Des Marais (1994) which they also suggested was associated with major tectonic cycles. Our data thus imply that organic-rich passive margin sediments were being rapidly subducted into the lower crust and mantle thus distorting the carbon mass balance for a significant time period. In contrast the large volumes of carbonates produced during the same time period remain preserved and relatively unaltered on the craton. Kump et al. (2001) have proposed a model involving mantle overturn and the development of superplumes near the Archaean–Proterozoic transition (2.47–2.45 Ga). Such a model is not necessarily inconsistent with the model proposed here and elsewhere in that it could well imply the involvement of supercontinent assembly and dispersal in the sequestration of carbon. Large volumes of isotopically light organic carbon are temporarily sequestered in passive margin sediments during periods of dispersal when continental margins are of greatest extent only to be subducted and sequestered in the lower crust and mantle as the new supercontinent forms. The timing of mantle overturn proposed by Kump et al. (2001) agrees well with our data from the Pilbara Craton. The proposed overturn event occurred as continental dispersal began and just before the first of the excursions defined by our $\delta^{13}C_{carb}$ curve (Fig. 17). Thus, whether we appeal directly to tectonic burial or a change in the redox state of the mantle it seems likely that the process is driven by planetary dynamics.

There is a growing body of evidence to suggest that there is a periodic cycle of supercontinent coalescence and dispersal, the so called 'Wilson Cycle' (Duncan and Turcotte, 1994; Murphy and Nance, 1992; Worsley et al., 1984), driven by large scale mantle convection (Gurnis, 1988; Anderson, 1982; Kominz and Bond, 1991). It is generally agreed that supercontinents assemble over geoid lows, mantle downwellings, and disperse over the geoid highs at mantle upwellings. This cycle is likely to be continuous because the same forces that fragment the old supercontinent over the geoid high are effectively assembling the next supercontinent over the associated low (Condie, 1998). It thus seems reasonable to argue that the growth of the carbon reservoir has been stepwise and related to the supercontinent cycle.

In the period following the major positive excursion the Western Australian $\delta^{13}C_{carb}$ curve indicates continuing instability in the longer term mass balance of carbon with isotopic swings of several ‰, including low values of -4.0‰ and a single value as high as $+6.0$‰. As more detailed information becomes available from this time interval, it seems likely that the carbon isotope record will prove to be more complex than is presently visualised, perhaps with events comparable to the major excursion in the Neoproterozoic carbon isotope record as suggested by Bekker et al. (2001). As in the Western Australian record, limited data from the Transvaal Supergroup in South Africa suggests that the Lomagundi event is indeed complex with at least three events ranging from 0 to $+10$‰ (Buick et al., 1998; Bekker et al., 2001). The $\delta^{13}C_{carb}$ record did not stabilise again until the long period following 1.8 Ga, as can be seen in the younger Bangemall basin (Buick et al., 1995) and in the northern Australian basins (Brasier and Lindsay, 1998; Lindsay and Brasier, 2000).

The events leading to and following the convergence of the Pilbara and Yilgarn Cratons were repeated across large areas of northern Australia during the Barramundi Orogeny (1890 ± 8 Ma, Page and Williams, 1988). Hoffman (1989a) noted that most continental collisional and arc-accretionary orogens in North America were active between 1.98 and 1.65 Ga. Crustal accretion in West Africa began at approximately 2.1 Ga and culminated in a major series of tectonic episodes at 1.9 Ga (Boher et al., 1992; Abouchami et al.,

1990; Davies, 1995). Tectonic events documented in Laurentia and Baltica (Gower, 1985; Gaal and Gorbatschev, 1987; Hoffman, 1988; Hoffman and Bowring, 1984; Rogers, 1995) suggest that these terrains were either fused with Australia or lay close by. Palaeomagnetic data support these conclusions (Idnurm and Giddings, 1988, 1995). It seems likely that Australia and western Canada were fused as early as 2.0 Ga (cf. O'Dea et al., 1997; Hoffman, 1991). Hoffman (1989b) has argued that the large intracratonic basins of Canada (Athabasca, Thelon and Sioux basins) also relate to supercontinent aggregation. Overall, there is now good evidence to suggest a Palaeoproterozoic supercontinent that included much of Gondwana (including Australia), Baltica, Laurentia and Siberia (Worsley et al., 1984; Hoffman, 1991; Moores, 1991). Condie et al. (2000) have provided considerable supporting evidence to suggest a mantle superplume at ca. 1.9 Ga. The Palaeoproterozoic supercontinent thus appears to have assembled over a considerable time span coming together by 2.0 Ga and then dispersing again at approximately 1.8 Ga. The data presented here suggest that carbon sequestration occurred during the early stages of assembly.

Beginning at approximately 1.8 Ga, large areas of the northern Australian craton began to subside, possibly as a response to mantle instability and the intrusion of anorogenic granites at a time of the breakup of a putative Palaeoproterozoic supercontinent (cf. Gurnis, 1988; Pysklywec and Mitrovica, 1998; Wyborn, 1988; Idnurm and Giddings, 1988; Lindsay, 1998). As a consequence, Palaeo- to Mesoproterozoic sedimentary rocks of similar age to the Bangemall basin (Muhling and Brakel, 1985) are widely distributed across the northwestern part of the Australian craton (i.e. Kimberley, McArthur, Mount Isa and Victoria River basins) containing up to 10 km of sedimentary rocks (Plumb et al., 1990). These widely distributed basins exhibit many similarities in their stratigraphic successions, suggesting common tectonic controls which were perhaps associated with supercontinent assembly and breakup (Lindsay, 1998; Lindsay and Brasier, 2000).

The secular $\delta^{13}C_{carb}$ curve remained relatively stable from the time of assembly of this putative Palaeoproterozoic continent, at around 1.8 Ga, for most of the following billion years, arguably until ocean closure and subduction began again as the supercontinent Rodinia assembled (e.g. Bond et al., 1984; Lindsay et al., 1987; Li et al., 1986; Dalziel, 1991, 1992). Current models of the oceanic carbon cycle suggest that the maintenance of an equilibrium in the mass balance of carbon over this long time period would have required relatively low or stable levels of tectonic activity, which in turn suggests that availability of land-derived biolimiting nutrients, such as phosphorus, is likely to have been stable. Under such conditions optimal utilisation by primary producers is likely to have led to extreme nutrient depletion across large stretches of the photic zone (e.g. Brasier, 1995). Prolonged nutrient stability may, therefore, have exerted a major influence upon the evolution of the biosphere over this time interval, most notably upon the evolution of photoautotrophy in eukaryotes (Brasier and Lindsay, 1998). At the end of this period the $\delta^{13}C_{carb}$ curve rose rapidly and increased in complexity, as critical activity increased and large swings took place in the climate. The biosphere also entered a period of escalating complexity with the appearance of multicellular life followed by the so called 'Big Bang' of metazoan evolution in the Late Neoproterozoic and Cambrian. The close connection between the carbon isotope record and both the climatic record and biospheric evolutionary history suggests a scenario in which both atmospheric and biospheric evolution have been driven forward by the tectonic evolution and dynamics of the planet.

6. Conclusions

The $\delta^{13}C_{carb}$ record preserved in the late Archaean and early Palaeoproterozoic basins of Western Australia provides evidence indicating a stable carbon mass balance in the late Archaean and early Palaeoproterozoic suggesting tectonic quiescence and a well establish role for photosynthesis in the early biosphere. However, some time after ca. 2.5 Ga the $\delta^{13}C_{carb}$ record became much more complex with positive to negative excursions

occurring after glaciation younger than 2449 ± 3 Ma and with a major positive excursion of Lomagundi type occurring at ca. 2173 ± 64 Ma. It seems likely that this early period of instability in the $\delta^{13}C_{carb}$ record is similar to and as complex as that encountered in the Neoproterozoic. When placed in the larger stratigraphic and tectonic framework of the region, the data suggest that carbon burial and concomitant oxygen enrichment of the atmosphere could well have been associated with the supercontinent cycle possibly with periods of continental dispersal and then of ocean closure. We suggest that sediments rich in isotopically light organic carbon that had accumulated in passive margin settings when continental fragments were dispersed were later subducted into the lower crust and mantle during the early stages of supercontinent assembly thereby enriching the global ocean in isotopically heavy carbon and potentially releasing oxygen into the atmosphere.

These observations potentially have profound implications for our understanding of the evolution of the biosphere and atmosphere. It has long been assumed that life evolved opportunistically on earth in a simple interactive relationship with its environment, sustained for the most part by an exogenic energy source, the sun. This data and a growing body of evidence elsewhere (see Knoll and Canfield, 1998 for a summary) suggests that the biosphere could, in a very general way, have been driven forward to greater complexity by the earth's endogenic energy resources (as expressed in plate tectonics). Conceivably, the long term survival of the biosphere depends upon those energy resources. Without those energy resources and a hydrosphere to facilitate plate tectonics (cf. Campbell and Taylor, 1983) it would not be possible to sustain the necessary mass of carbon in the lower crust and mantle to drive the biosphere. Without the driving energy provided by the evolving planet the biosphere would have entered a prolonged stasis and ultimately faced extinction. If this is so, it has fundamental implications, not only for life on earth, but for the more general problems surrounding the likelihood of life having evolved on other planetary bodies. Smaller planets with limited endogenic energy resources and short-lived atmospheres and hydrospheres are unlikely to develop a biosphere and, if they did, their evolutionary development is unlikely to extend beyond simple single-celled life forms.

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