A syn-depositional age for Earth’s deepest $\delta^{13}C$ excursion required by isotope conglomerate tests

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ABSTRACT

The most negative carbon isotope excursion in Earth history is found in carbonate rocks of the Ediacaran Period (635–542 Ma). Workers have interpreted the event as the oxidation of the Ediacaran oceans [Rothman et al., Proc. Natl. Acad. Sci. USA 100 (2003) 8124; Fike et al., Nature 444 (2006) 744; McFadden et al., Proc. Natl. Acad. Sci. USA 105 (2008) 3197], or as diagenetic alteration of the $\delta^{13}C$ of carbonates ($\delta^{13}C_{\text{carb}}$) [Knauth and Kennedy, Nature 460 (2009) 728; Derry, Earth Planet. Sci. Lett. 294 (2010) 152]. Here, we present chemo-stratigraphic data from the Ediacaran-aged Wonoka Formation (Fm.) of South Australia that require a syn-depositional age for the extraordinary range of $\delta^{13}C_{\text{carb}}$ values (−12 to +4‰) observed in the formation. In some locations, the Wonoka Fm. is 700 metres (m) of mixed shelf limestones and siliclastics that record the full 16$^{18}O_{\text{carb}}$–$\delta^{13}C_{\text{carb}}$ excursion. In other places, the Wonoka Fm. is host to deeply (~1 km) palaeocanyons, which are partly filled by tabular-clast carbonate breccias that are sourced from eroded Wonoka canyon-shoulders. By measuring the isotopic values of 485 carbonate clasts (an isotope conglomerate test), we show that canyon-shoulder carbonates acquired their $\delta^{13}C_{\text{carb}}$–$\delta^{18}O_{\text{carb}}$ values before brecciation and redeposition in the palaeocanyons.

Introduction

The Ediacaran Period (Knoll et al., 2006) is the bridge between the Proterozoic world and the animal-abundant Phanerozoic. Sponges (Love et al., 2009; Maloof et al., 2010; Sperling et al., 2010; Brain et al., 2012) appear before the ~635 Ma (Hoffmann et al., 2004; Condon et al., 2005) terminal–Cryogenian ice age, but decimetre-scale organisms (animals, giant protists and macro-algae known as the Ediacaran Biota (Xiao and Laflamme, 2009)) do not appear until ~579 Ma (Bowring et al., 2003). Broadly synchronous with these first appearances, the deepest carbon isotope excursion in Earth history is recorded in carbonates from at least four continents – most famously from Oman (Burns and Matter, 1993), South Australia (Calver, 2000), South China (McFadden et al., 2008) and southwestern USA (Corsetti and Kaufman, 2003). The extreme isotopic depletion seen in these Ediacaran basins is colloquially referred to as the ‘Shuram’ anomaly, although global synchronicity has not been established independently (Grotzinger et al., 2011). Chemostratigraphic (Prave et al., 2009) and sparse geochronological data (Condon et al., 2005; Bowring et al., 2007) data suggest that the globally observed anomalies, synchronous or not, are hosted in sediments younger than the ~580 Ma Gaskiers glaciation (Bowring et al., 2003) and older than ~551 Ma (Condon et al., 2005).

The dominant paradigm among chemostratigraphers is that $\delta^{13}C_{\text{carb}}$ in carbonate rocks reflects the $\delta^{13}C$ of dissolved inorganic carbon (DIC) in contemporaneous sea water, which is virtually uniform around the world due to the long residence time of DIC and the short mixing time of the oceans (Kump and Arthur, 1999). Similarities in shape and magnitude, and the broadly Ediacaran age, have been used to argue that the ‘Shuram’ anomaly fits this model, and is therefore globally synchronous and can be used for global stratigraphic correlation (Halverson et al., 2005). A primary DIC origin for the ‘Shuram’ is also the foundation for models involving stepwise oxidation of the terminal Neoproterozoic ocean (Rothman et al., 2003; Fike et al., 2006; McFadden et al., 2008) that resulted in pulsed inputs of light carbon to the ocean-atmosphere system (Rothman et al., 2003), and references therein.

Absolute time constraints, although not available, are necessary to quantify the light carbon fluxes and oxidant budget required, and thus test the model’s viability.

Alternatives to the global-DIC hypothesis have sought to explain the ‘Shuram’ anomaly, and include (1) meteoric and (2) burial diagenesis models. Under the meteoric model, the depleted $\delta^{13}C_{\text{carb}}$ values and co-varying $\delta^{13}C_{\text{carb}}$–$\delta^{18}O_{\text{carb}}$ features of most, but not all, Shuram-like anomaly profiles are interpreted as a record of remineralising fluids charged with DIC issued from organic matter respiration in soils (Knauth and Kennedy, 2009; Swart and Kennedy, 2012). The burial diagenesis model invokes post-burial fluid-rock interactions at depth, wherein a high $p$CO$_2$, low $\delta^{13}C$ fluid, developed from buried organic matter, mixes with an $^{18}O$-rich basinal brine (Derry, 2010). Both diagenesis models therefore argue for acquisition of negative $\delta^{13}C_{\text{carb}}$ after the carbonate sediments had been deposited (although in the meteoric case, alteration can happen immediately after deposition). If the ‘Shuram’ anomalies are a record of diagenesis, one must ask (1) what type of process would lead to diagenetic alteration of $\delta^{13}C_{\text{carb}}$, synchronously or not, in Ediacaran basins around the world, and (2) why might Ediacaran sediments be uniquely predisposed to such intense alteration (Grotzinger et al., 2011).

We present data from six measured stratigraphic sections (Fig. 1B,C) from the Ediacaran Wonoka Fm. of the Adelaide Rift Complex (ARC) of South Australia that (1) demand syn-depositional acquisition of its low $\delta^{13}C_{\text{carb}}$ values (down to ~12‰) and the observed covariance between $\delta^{13}C_{\text{carb}}$ and $\delta^{18}O_{\text{carb}}$. (2)
burial diagenesis model, and (3) constrain the styles of meteoric diagenesis that could be invoked to explain the observations from Australia.

**Geological setting**

The ARC (Fig. 1A) was part of a continental margin formed to the present-day east of the Stuart Shelf, and the Wonoka Fm. is part of the Ediacaran Wilpena Group (Preiss, 2000; Preiss and Robertson, 2002).
The Nuccaleena Formation, the distinctive cap dolostone to the glacial Elatina Formation (Plummer, 1979; Williams, 1979; Rose and Maloof, 2010), forms the base of the Wilpena Gp., and frequently is correlated with the younger Cryogenian glacial cap units found around the world (Halverson et al., 2005) (dated to be /C24/ ~635 Ma in Namibia (Hoffmann et al., 2004) and South China (Condon et al., 2005)). The Wonoka Fm. varies between 400 and 1500 metres (m) thick, and coarsens and shallows upwards into the siliclastic, Ediacara Biota-bearing Pound Subgroup. At its base in the central Flinders Ranges (e.g., 19, 67, and 84 in Figs 1B and 2D), the Wonoka Fm. is a deep-shelf sequence of turbidites with climbing-rippled sands and allodapic (i.e., transported down-slope) carbonate beds. The Wonoka Fm. transitions upwards into a shallower, storm-dominated mid-shelf sequence containing abundant hummocky and swaley cross-stratified sands (Fig. 2E), and wavy limestone laminites and grainstones. At the top of these units, the Wonoka Fm. coarsens abruptly into thick (~50–100 m), trough cross-bedded fine-to-medium grained sandstones (unit 10 under the terminology of Haines (1988); see Fig. 1 caption).

In contrast, the northern Flinders (5, 6, 9, 10 in Fig. 1C) record deeper outermost-shelf and slope settings, with ~1000-m deep palaeocanyons that cut into the underlying Bunyeroo and Brachina formations (Fig. 2A,B) (Haines, 1988; Christie-Blick et al., 1990; DiBona and von der Borch, 1993; Giddings et al., 2010) and are filled with mixed carbonate and siliciclastic turbidites and tabular-clast limestone breccias (Fig. 2C). Henceforth, we refer to outer shelf sections as ‘canyon-shoulders’, and sections with palaeocanyons as ‘canyon-fill.’ The uppermost Wonoka Fm. is characterised by ~80–100 m of microbialite and stromatolite bioherms (unit 11 of Haines (1988)), which blanket the canyon-shoulders and some canyon-fill sections (Fig. 3).

Methods

Carbonates were sampled at 1.0 m resolution whilst measuring six stratigraphic sections from across the Adelaide Rift Complex (ARC; Fig. 1 A). Clean limestones and dolostones with minimal siliciclastic components were targeted. A total of 1049 samples were slabbed and polished perpendicular to bedding and 5 mg of powder were micro-drilled from individual laminations for isotopic analysis. Sections 5–6, 9, 10, and 19 (Fig. 1B,C) were measured at the University of Michigan Stable Isotope Laboratory on a Finnigan MAT Kiel I preparation device coupled directly to the inlet of a Finnigan-

Fig. 2 (A) Google Earth image (acquired 2/26/2010) of a ~1000-m-deep Wonoka palaeocanyon, with palaeocurrent directions from Eickhoff et al. (1988) – photos B and C are located on the north side of the palaeocanyon near the ‘W’ in the Wonoka Fm. label in (A); (B) angular and erosive contact (black line) between Brachina Formation canyon wall and Wonoka Formation canyon-fill (white dashed lines depict change in bedding attitude) at Mount Thomas (sections 5–6); (C) a 2-m-thick debris flow breccia, dominated by tabular clasts (1–75-cm long) of micritic marine limestone; (D) canyon-shoulder outcrop near Parachilna Gorge (19); (E) small-scale hummocky cross-stratified (HCS) carbonate grainstone typical of canyon-shoulder outer shelf facies from Saint Ronan (67); (F) 20–40-cm thick grainstone (red in colour)-to-microbialite (yellow) couplet typical of upper Wonoka Fm. stratigraphy at Parachilna Gorge (19).
MAT 251 triple collector isotope ratio mass spectrometer. All other samples were measured at Princeton University on a GasBench II preparation device coupled directly to the inlet of a Thermo DeltaPlus continuous flow isotope ratio mass spectrometer. 

$d^{13}C$ and $d^{18}O$ data are reported in the standard delta notation as the difference from the VPDB standard (Vienna Pee Dee Belemnite), and measured precision is 0.1 (1σ) for both values. For a more thorough discussion of these methods, see Rose et al. (2012).

**Results**

The canyon-shoulder sections of Parachilna Gorge, Saint Ronan and Black Range Spring (19, 67, and 84 in Fig. 1B) exhibit very low $d^{13}C_{\text{carb}}$ values and $d^{13}C_{\text{carb}}-d^{18}O_{\text{carb}}$ covariation (see Fig. 4A). Isotopic values are most negative at the base of unit 3 (down to −12‰), and the profile recovers smoothly and gradually towards positive $d^{13}C_{\text{carb}}$ values. At the top of unit 8, the curve crosses 0‰, and $d^{13}C_{\text{carb}}$ remains positive in units 9–11. The isotopic profiles of canyon-shoulder units 1–11 are remarkably consistent between sections over

![Fig. 3 Bing Maps image (acquired 11/17/2004) of the Saint Ronan canyon-fill and canyon-shoulder complex (map area marked with black outline in Fig. 1A). Black lines indicate conformable contacts between units or formations, while white lines indicate sequence boundaries. Field mapping shows that the microbialite reef facies of unit 11 caps both the canyon-fill and canyon-shoulder sequences, thus indicating that canyon cutting and filling occurred before final Wonoka deposition.](image1)

![Fig. 4 (A) $d^{13}C_{\text{carb}}-d^{18}O_{\text{carb}}$ cross plot showing data from all stratigraphic sections, coded by section number, lithofacies, and lithology (‘l’ stands for limestone; ‘dl’ for dolomite). All canyon-shoulder sections show $d^{13}C_{\text{carb}}-d^{18}O_{\text{carb}}$ correlation, especially for $d^{13}C_{\text{carb}}$ values less than −5‰ (r² = 0.49, 0.49 and 0.44 for 19, 67 and 84 respectively; for all sections, P < 0.001). Such correlation is not observed in the fine-grained, allodapic carbonates of the canyon-fill sections (r² = 0.003 and P = 0.285 for all three sections combined). In the figure legend, lithofacies are organised in order of increasing permeability (i.e., ranging from fine-grained micritic wavy laminates to coarse grainstones and sandy carbonates). No pattern of dependence between isotopic values and lithofacies is observed. (B) Photomicrograph of a cm-scale fining–upward carbonate turbidite from the base of the Saint Ronan canyon-shoulder section (67; see Fig. 1B) that shows the lack of coarse recrystallisation observed in Wonoka Fm. carbonates. Even the fine-sand fraction of carbonate grains, which are most susceptible to recrystallisation are preserved in primary sedimentary textures (e.g., the dashed white line marks the erosive beginning of another turbiditic sequence). These fabrics contrast with rocks from unit 11 (C), where the first evidence of recrystallisation and growth of dolomite rhomboids is observed (one such crystal is outlined (C), with dashed lines depicting orientation of cleavage). Dolomites are only found in the upper 150–200 m of the Wonoka Fm., where carbon isotopic values are at their most positive (+2 to +8‰). The isotopic values of (B) and (C) are depicted in panel (A) with lettered symbols; both samples are ~85% carbonate by weight.](image2)
50 kilometres distant (Fig. 1B). The lower half of most canyon-fill units is carbonate poor, except for the frequent tabular-clast carbonate breccias (Fig. 2C). Fine-grained allogenic carbonate beds are common above 300–400 m of basal canyon-fill, and δ¹³C_carb values from Mount Thomas, Mount Goddard and Oodnapanicken (5–6, 9, 10 on Fig. 1C) vary between −8.5 and −5‰. Canyon-fill sections therefore do not record (1) the nadir seen in the canyon-shoulders, (2) δ¹³C_carb–δ¹⁸O_carb covariation ($r^2 = 0.003$ and $P = 0.285$ for all three datasets combined), or (3) the recovery to positive δ¹³C_carb values.

In palaeomagnetics, a conglomerate test is used to determine whether clasts in a conglomerate were magnetised prior to transport and deposition (preserving random magnetic directions) or after deposition (preserving uniform magnetic directions, despite random clast orientations). Analogously, we performed an isotope conglomerate test on the breccia units of Mount Thomas, Oodnapanicken, and Saint Ronan canyon-fill (5–6, 9, and 80, respectively, on Fig. 5A,B) to assess the provenance and relative timing of acquisition of δ¹³C_carb and δ¹⁸O_carb by measuring the isotopic values of carbonate clasts. The breccia units range from 0.2 to 11 m in thickness, are clast-supported in a matrix of fine sand (Fig. 2C), and are most common at the base of the canyon-fill stratigraphy. At Mount Thomas and Oodnapanicken, the tabular, 1–100-cm long carbonate clasts consist of two dominant lithologies: grey, micritic limestone and brown dolostone. At Saint Ronan, the clasts are distinctly different from Mount Thomas and Oodnapanicken breccia clasts, and identical to the green, wavy laminites and coarse red limestone grainstones from the adjacent canyon-shoulder strata (Fig. 4). Individual breccia units record δ¹³C_carb variability of 6 and 16‰, and significant δ¹³C_carb–δ¹⁸O_carb covariation (Fig. 5A,B). The δ¹³C_carb range of canyon-shoulder units 1–9 (−12 to +4‰; 19, 67 and 84; Fig. 1B) is seen in the breccia units of the more distal canyon-fill of Mount Thomas and Oodnapanicken (5–6 and 9 on Fig. 5A), while the range (−9 to −3‰) of the more proximal canyon-fill of Saint Ronan (Fig. 5B) matches that of the locally eroded canyon-shoulder (the first ~550 m (units 1–7) of J67 on Fig. 1B). More distal sourcing cannot be ruled out, however, because the canyon-shoulder δ¹³C_carb profile is remarkably reproducible across 20 000 km² of basin map-area (Fig. 1A,B).

**Discussion**

We interpret the breccia clasts to be carbonates sourced from units 1–9 of canyon-shoulder localities (e.g., 19, 67, and 84 on Fig. 1B). In certain palaeocanyons, 0.05 to 5 m-thick grey-to-yellow micritic limestone layers are found along the edge of the canyon wall (Fig. 2A,B). This carbonate is of debated origin (Eickhoff et al., 1988; Giddings et al., 2010), and it may represent an additional source of clasts for the basalt breccia beds of Mount Thomas and Oodnapanicken, although the observed δ¹³C_carb range of this in situ carbonate (−6.5 to −9.3‰) (Giddings et al., 2010) covers less than one-fourth of the full range observed in the carbonate breccias. These limestone beds are not present at Saint Ronan, where the lithofacies, colour and isotopic composition of the clasts are identical to units 1–7 (Fig. 1B) of the canyon-

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**Fig. 5** Results of isotope conglomerate tests from Mount Thomas and Oodnapanicken canyon-fill [labelled 5–6/blue and 9/red respectively, in (A)] and Saint Ronan canyon-fill (labelled 80/green in (B)). If the clasts acquired their δ¹³C_carb values and δ¹⁸O_carb correlation in situ on canyon-shoulders, then they should exhibit a random collection of values representing the full isotopic range present on the canyon-shoulder at the time of canyon filling. In contrast, if the extremely negative δ¹³C_carb (down to −12‰) and δ¹³C_carb–δ¹⁸O_carb correlation in canyon-shoulders are a result of post-depositional diagenesis, then clasts from individual breccia beds should either (1) reflect the original pre-diagenic isotopic values in the shoulder (i.e., not extremely depleted in δ¹³C_carb), or (2) a consistent diagenetic value homogenous within breccia units. The clasts of the more distal, Oodnapanicken and Mount Thomas canyon-fill exhibit the full range of δ¹³C_carb values observed in units 1–9 in canyon-shoulder sections throughout the basin (19 and 84; Fig. 1B). The more proximal Saint Ronan canyon-fill appears to have sampled a smaller canyon-shoulder range; breccias exhibit a smaller δ¹³C_carb range, but the clasts match the colour, lithologies and isotopic range observed in units 1–7 of the immediately adjacent canyon-shoulder (67; Fig. 1B; Fig. 3). More distal sourcing, however, is still possible because the canyon-shoulder δ¹³C_carb profile is remarkably consistent across the ARC (Fig. 1A,B).
shoulder section 6.5 km distant. Thus, these observations require that both the negative δ13C values and the δ13C–δ18O covariation were acquired in Wonoka Fm. canyon-shoulder carbonates before those carbonates were brecciated and redeposited in the palaeocanyons.

While the mechanism for the formation of the Wonoka canyons remains controversial [subaerial (Christie-Blick et al., 1995) vs. submarine (Giddings et al., 2010)], there has been broad agreement that canyon incision occurred early during Wonoka Fm. deposition (placed variously at a cryptic unconformity with Haines’ canyon-shoulder units 1–5 (Christie-Blick et al., 1995; Giddings et al., 2010); see Fig. 1B). Given the presence of carbonate clasts with positive δ13C values in canyon-fill breccia units, our results require a higher surface (i.e., above unit 8) to be responsible for at least some of the canyon cutting at Mount Thomas and Oodnapanicken. We do not claim, however, that this proposed surface represents all canyon cut-fill sequences within the Wonoka Fm., as workers have documented numerous intervals of canyon incision throughout Wonoka Fm. deposition (DiBona and von der Borch, 1993). We therefore propose placing a sequence boundary at the abrupt appearance of unit 10 sandstones, where units 8 and 9 are variably absent (Fig. 3) and the δ13C profile first recovers to positive δ13C values. Thus, the canyon-fill sampled units 1–9 of intact canyon-shoulder stratigraphy, whose carbon isotopic range (~12 to +4‰) matches that of the tabular-clast breccias. The most positive δ13C values observed from the Saint Ronan breccias is ~3‰, corresponding isotopically to the top of unit 7 of the intact canyon-shoulder stratigraphy. Thus, the canyon cutting surface could be placed lower at Saint Ronan, so long as the canyon remained open to sediment infill throughout the deposition of unit 7 on the canyon-shoulder.

As the canyons filled completely, the fine-grained, allogenic carbonate beds of the upper canyon-fill (upper 700 m of Fig 1C) represent a recycled and homogenised canyon-shoulder section, producing δ13C profiles (5–6, 9, and 10 in Fig. 1C) that are stratigraphic averages of a canyon-shoulder profile (Fig. 6). To test this hypothesis, linearly interpolated datasets were created for representative canyon-shoulder (units 1–9; 19 – Parachilna; see Fig. 6A) and fine-grained canyon-fill δ13C curves (10 – Mount Goddard; Fig. 6B). Linear interpolation serves to weight δ13C values by the stratigraphic thickness over which they occur, thus accounting for the stratigraphic weighting that occurs during brecciation (i.e., thicker units will contribute more breccia clasts). The distributions of the Parachilna and Mount Goddard interpolated datasets overlap, with the variance of Mount Goddard being much smaller. This result is expected, as mixing and homogenisation of a canyon-shoulder profile at a scale below that sampled for isotopic measurement should decrease the variance, but produce a similar mean; where canyon-fill is not well mixed at the scale of isotopic measurement (i.e., breccia clasts), the variance of the distribution is much larger (Fig. 6C).

These stratigraphic and isotopic observations preclude burial diagenesis models for the negative isotopic values (Derry, 2010) in South Australia. The low δ13C values and δ13C–δ18O covariation of the Wonoka Fm. must either be primary or a relatively early meteoric diagenetic signal (i.e., before the Wonoka

Fig. 6 The distribution of stratigraphically weighted δ13C values for representative (A) canyon-shoulder (units 1–9; 19 – Parachilna; see Fig. 1B) and (B) fine-grained canyon-fill (10 – Mount Goddard; Fig. 1C). The resulting distributions (C) overlap with a similar mean, and are thus consistent with the fine-grained canyon-fill consisting of recycled, homogenised and redeposited canyon-shoulder carbonates. Also displayed is the observed distribution of breccia clast δ13C (Fig. 5A), which represents discrete sampling of a canyon-shoulder isotopic profile. Its histogram distribution looks very similar to the canyon-shoulder profile for δ13C values less than ~7.5‰, but the breccia clasts have a long tail into positive values that is not as well developed in the canyon-shoulder. We interpret this observation as a sampling bias; the rarer dolostone clasts were over-sampled to gain clast diversity in our sampled population, and it is dolomite that carries the most positive δ13C values. A second contributing factor may be that unit 9 of the canyon-shoulder is truncated in places by unit 10 sandstones (Fig. 4). Thus, unit 9, and its associated positive δ13C values, may be under represented in the interpolated canyon-shoulder profile.
canyons began to fill, and certainly prior to burial). In classic examples of meteoric diagenesis, negatively altered δ13C_carb values are associated with exposure surfaces, indicative of top-down diffusion of the altering fluid (e.g., 18O-depleted rainwater charged with isotopically light DIC originating from remineralised organic carbon (Allan and Matthews, 1982; Swart and Kennedy, 2012)). In the Wonoka Fm., the only physical evidence for subaerial exposure is fenestral carbonate fabrics in the microbialite reef facies of the uppermost unit 11. At Saint Ronan, however, we see this unit developed at the top of both canyon-shoulder and canyon-fill sections, thus indicating that the canyons were cut and filled before unit 11 deposition (Fig. 3). This observation requires that canyon formation occurred during ongoing Wonoka sedimentation, and therefore demands a syn-depositional age for the low δ13C_carb values of the canyon breccia clasts and associated shoulder stratigraphy. If subaerial exposure of unit 11 resulted in meteoric diagenesis, it cannot be invoked to explain the isotopic values of underlying units 1–9 of the Wonoka Fm.

Although our proposed sequence boundary at the base of unit 10 could be submarine, with canyon cutting and filling being a subaqueous process of mass wasting (Giddings et al. (2010), and references therein), some have argued that the Wonoka canyons formed subaerially, with the base level fall accomplished by basin isolation and Messinian-style evaporitic drawdown (Christie-Blck et al., 1990). This scenario would leave the canyon-shoulders as much as 1.5 km above local sea level and susceptible to meteoric diagenesis. However, based on the isotope conglomerate test (Fig. 5A, B), in this scenario, diagenesis would need to occur after the basin was exposed, but before any substantial canyon cutting and redeposition of shoulder rocks into canyon-fill had occurred. Although our data do not rule out this possibility, there is no visible evidence of such alteration in the form of widespread recrystallisation (primary sedimentary fabrics (Haines, 1988; Giddings et al., 2010) are exceptionally well preserved in the Wonoka Fm.; see Figs 2E, F and 4B). Recrystallisation is only pervasively observed in unit 11 dolomites, where δ13C_carb values are at their most positive (+2 to +8‰; Fig. 4C). Also, as the meteoric model hinges upon fluid–rock interactions, the lightest isotopic values should be found in horizons most amenable to fluid flow. Fluid flux would be controlled by primary porosity and permeability, which is a function of grain size, shape and packing, and thus directly related to lithofacies. Contrary to these predictions, we observe no pattern of permeability-dependent isotope modification (Fig. 4A).

Furthermore, the δ13C_carb profile of the canyon-shoulders (most negative δ13C_carb at the base, and increasing in value towards unit 9 and the top) is opposite of that expected from top-down buffering with meteoric water charged with DIC issued from organic matter remineralisation (Allan and Matthews, 1982). Finally, whether Shuram-style δ13C_carb anomalies are globally synchronous or not, Messinian-style diagenesis would be a local event, restricted to the Adelaide Rift Complex and the Wonoka Fm., and would not explain the negative carbon isotope signatures of other Ediacaran basins.

Conclusions

Based on the isotope conglomerate tests, acquisition of δ13C_carb values (−12 to +4‰) in South Australian carbonate sediments was synchronous with deposition of Wonoka Fm. canyon-shoulder sediments. The filling of palaeocanyons occurred during the latter stages of Wonoka Fm. deposition, and the canyons were cut and filled before development of upper Wonoka Fm. microbialite reefs (unit 11 on Fig. 1B). These findings are inconsistent with a burial diagenesis origin, and the expected first-order stratigraphic and microtextural patterns predicted by meteoric diagenesis are not observed. Therefore, the balance of evidence supports a syn-sedimentary origin for the extraordinary range of δ13C_carb values seen in the Wonoka Formation of South Australia.

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Author contributions

Field work was conducted by J.M.H. and A.C.M. (2 field seasons) and B.S. (1 field season), following initial project planning by J.M.H. and A.C.M; J.M.H. analysed the data, wrote the manuscript and drafted figures, all with input from A.C.M. and B.S.

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Isotope conglomerate test • J. M. Husson et al.


