

## Geology

### Search for evidence of impact at the Permian-Triassic boundary in Antarctica and Australia: Comment and Reply

John L. Isbell, Rosemary A. Askin and Gregory J. Retallack

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**Notes**

## Low-latitude sea-surface temperatures for the mid-Cretaceous and the evolution of planktic foraminifera: Comment and Reply

### COMMENT

G. D. Price

M. B. Hart

Department of Geological Sciences, University of Plymouth,  
Drake Circus, Plymouth, PL4 8AA, UK

In a recent article, Norris and Wilson (1998) presented an attractive and detailed isotopic study regarding evidence for low-latitude sea-surface temperatures and the depth stratification and evolution of mid-Cretaceous planktonic foraminifera from Ocean Drilling Program hole 1052E, Blake Plateau, North Atlantic. On the basis of isotopic data, they suggested that the foraminiferal species *Planomalina*, *Ticinella*, and *Biticinella* lived closer to the sea surface, compared with rotaliporids, which appear to have calcified several degrees cooler and by inference at a greater depth also. Further, Norris and Wilson (1998) asserted that hedbergellids generally record the coolest temperatures, and by implication occupied the greatest depth habitats. An understanding of foraminiferal depth stratification is of great importance in constraining sea-level changes, paleo-oceanographic events such as widespread anoxia, and the reconstruction of sea-surface temperatures by reference to derived oxygen isotope values. If the assertions of Norris and Wilson (1998) are correct, then existing models of foraminiferal depth habitats and the resulting interpretations of the sediments where they are found needs to be changed radically.

Hence, a number of important questions stemming from the work of Norris and Wilson (1998) need to be addressed, particularly with respect to the reconstruction and implications of depth stratification of mid-Cretaceous foraminifera. Depth preferences of Cretaceous foraminifera are generally thought to have been similar to present-day planktonic species (Savin, 1977) with globular forms inhabiting near-surface waters and the flattened, keeled morphotypes representing deeper habitats (Caron and Homewood, 1982; Leckie, 1987). This general trend has been supported through oxygen and carbon isotope data (e.g., Corfield et al., 1990; Huber et al., 1995; Price et al., 1998) and paleobiogeographic distributions (e.g., Hart and Bailey, 1979). If the distribution of planktonic foraminifera from the Albian to Maastrichtian onshore United Kingdom chalk sequences is plotted, for example, the smaller sized fractions are dominated by hedbergellids as is (largely) the 250–500  $\mu\text{m}$  fraction. Keeled planktonic foraminifera (including praeglobotruncanids, rotaliporids, and marginotruncanids) are only common in the mid-Cenomanian to mid-Turonian interval (Hart and Bailey, 1979), which is coincident with the climatic optimum and where water depths in shelf successions would be at a maximum. Offshore, throughout the North Sea Basin and Western Approaches, the keeled forms are found coexisting with the hedbergellids (Hart and Ball, 1986). Hence, in paleobiogeographical terms, this evidence tends to suggest that hedbergellids occupied the shallower parts of the water column, while keeled forms became abundant only when water depths increased. This theory is in contrast to the inferences of Norris and Wilson (1998) regarding depth stratification. Furthermore, in lineages such as *Contusotruncana fornicata*–*Contusotruncana contusa*, a hedbergellid growth stage in the early whorls is discernible (Robaszynski et al., 1984), implying a migration to shallower waters during ontogeny. This is, however, the reverse of inferences derived from modern fauna and relatively implausible in that a light, thin-shelled, relatively delicate hedbergellid form should migrate upward in the water column despite becoming large and very “heavy.” Although Norris and Wilson (1998) asserted that hedbergellids occupied the greatest depth habitats, the oxygen isotope data presented in their Figure 2A–D shows that both *Hedbergella*

*simplex* and *Hedbergella delrioensis* occur in the same range as the keeled forms *Praeglobotruncana delrioensis*, *Rotalipora brotzeni*, *Planomalina buxtorfi*, and *Rotalipora ticinensis*, suggesting essentially an overlap of depth-temperature-related habitats rather than significantly deeper and cooler habitats. Moreover, in Figure 2B, *H. simplex* appears consistently isotopically lighter (and therefore yielding warmer temperatures) than coexisting *Rotalipora appenninica* foraminifera. The case of Norris and Wilson (1998) may be somewhat overstated, as only *Biticinella breggiensis*, *Ticinella praeticinensis*, and *Ticinella raynaudi* (ancestral forms of the rotaliporids) consistently display lighter oxygen isotopic values, indicative of warmer temperatures and shallower depth habitats.

Norris and Wilson (1998) utilized the isotope data, by reference to the paleotemperature equation of Erez and Luz (1983), to assess sea-surface temperatures for the mid-Cretaceous. Estimates reach a maximum of 30–31°C. Norris and Wilson (1998) acknowledged the sensitivity of the  $\delta^{18}\text{O}$ -based paleotemperature calculations to the assumed oxygen isotopic composition of ambient waters ( $\delta^{18}\text{O}_{\text{seawater}}$ ). Although these estimates are likely to be accurate and compare favorably to recent estimates of low-latitude mid-Cretaceous temperatures by Barron et al. (1995) and Price et al. (1998), it must be emphasized that potential variability in the  $\delta^{18}\text{O}_{\text{seawater}}$ , particularly in response to variation in the extent of possible ice caps and freshwater influences, provides adequate leeway to argue for significantly cooler temperatures.

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## REPLY

R. D. Norris

MS-23, Woods Hole Oceanographic Institution, Woods Hole,  
Massachusetts 02543-1541, USA

P. A. Wilson

Department of Earth Sciences, University of Cambridge, Downing Street,  
Cambridge CB2 3EQ, UK

Price and Hart take issue with our isotopic investigation of depth habitats in Albian-Cenomanian planktic foraminifera. Their central contention is that the foraminifera were depth stratified, as they are today, with globular taxa inhabiting surface waters and compressed, keeled morphospecies growing largely within or below the thermocline. They argue that the hedbergellids, which had globular shells and large population sizes in epicontinental seas, must have grown largely in the surface ocean.

Considerable consensus exists between us. Our data clearly showed that the ticinellids and biticinellids, which have globular chambers without keels, were surface dwellers for much, if not all, their life spans, while keeled morphotypes such as the rotaliporids and praeglobotruncanids grew in cooler waters, probably within the thermocline. However, on the basis of  $\delta^{18}\text{O}$  similarity to rotaliporids in the same samples, we argued that the hedbergellids that we analyzed (*Hedbergella simplex* and *Hedbergella delrioensis*) grew within thermocline waters. In contrast, we demonstrated that a keeled species, *Planomalina buxtorfi*, which is classified as a deep-dweller on morphological grounds, has consistently lower  $\delta^{18}\text{O}$  (implying warmer calcification temperatures and shallower depth habitat) than rotaliporids. On balance, therefore, our results suggest that the basic relationship between morphology and depth habitat holds for many taxa, but apparently not for the hedbergellids (some of which may have grown deeper than expected) or the planomalinids (which apparently lived at shallower depths than expected).

We concede that our data for hedbergellids show some scatter because two  $\delta^{18}\text{O}$  measurements overlap those of the "surface dweller," *Planomalina buxtorfi*. However, the other eleven  $\delta^{18}\text{O}$  measurements of hedbergellids plot with the thermocline-dwelling rotaliporids and *Praeglobotruncana*. We believe that our data make a case for a thermocline habitat for *H. delrioensis* and *H. simplex*, but we agree that this conclusion needs to be tested in other localities. Indeed, data in Price et al. (1998) show that many hedbergellids in several Pacific cores have  $\delta^{18}\text{O}$ , and inferred depth habitats, like thermocline-dwelling rotaliporids. It is important to note that not all hedbergellids are the same from a geochemical or depth-habitat perspective. Our (unpublished) work has shown that a hedbergellid-like species, *Costellagerina libyca*, has much more negative  $\delta^{18}\text{O}$  than other planktic foraminifera and may be an excellent recorder of surface water temperatures in the Albian and Cenomanian. Hence, it is necessary to identify which hedbergellid species one is discussing in any analysis of depth habitats.

Price and Hart correctly note that hedbergellids are a common, if not dominant, component of epicontinental sea foraminifer assemblages. Unfortunately, this pattern does not provide a strong test of depth habitats in open ocean sites. Recent analysis of the ecologies of Cretaceous planktic foraminifera suggest that hedbergellids were extremely "opportunistic" (Premoli Silva and Sliter, 1995). The salinity stratification of some epicontinental seas may have permitted some thermocline species to reproduce in abundance because they were able to adapt to the low salinity and perhaps higher productivity of these shallow seas. The hedbergellids may have bloomed in areas of high primary production, such as exists along the modern continental margins, and followed the chlorophyll maximum into thermocline habitats in the deep ocean. Hence, the hedbergellids may have varied their depth habitat depending upon the depth of highest primary production. These are hypotheses requiring future testing, perhaps by comparing isotopic records of the same species in epicontinental seas and the open ocean.

Many modern planktic foraminifera grow to substantial size in the surface ocean and sink into deeper waters to complete reproduction, so it is entirely reasonable to suggest that extinct species did the same (e.g., Caron and Homewood, 1983). However, our data strongly suggest that Albian-Cenomanian planktic foraminifera did not migrate vertically during their growth. Similar isotopic studies of Paleocene and Eocene planktic foraminifera have also found that few thermocline species migrated vertically during ontogeny (Pearson et al., 1993; D'Hondt et al., 1994; Norris, 1996). Hence, it may be that the widespread evolution of vertical depth migration is a Neogene phenomenon and not an ecology that can be applied directly to extinct species on uniformitarian grounds.

Price and Hart apparently agree that our isotopic data suggest sea-surface temperatures (SSTs) of approximately 30–31 °C in the mid-Cretaceous tropical western North Atlantic, but they argue that uncertainties in the extent of possible ice caps and freshwater influences provide adequate leeway for significantly cooler temperatures. We agree that estimation of surface ocean salinity and  $\delta^{18}\text{O}_{\text{seawater}}$  is the greatest source of error in the estimation of SSTs. However, in the absence of any convincing evidence for the existence of volumetrically significant continental ice caps for Albian-Cenomanian time and in light of observations on modern geographic salinity variation and results from climate model simulations water (Bottomley et al., 1990; Bush and Philander, 1997), we believe that our calculated SSTs remain best estimates.

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## Search for evidence of impact at the Permian-Triassic boundary in Antarctica and Australia: Comment and Reply

### COMMENT

**John L. Isbell**

*Department of Geosciences, University of Wisconsin, Milwaukee,  
Wisconsin 53201, USA*

**Rosemary A. Askin**

*Byrd Polar Research Center, Ohio State University, Columbus,  
Ohio 43210, USA*

Retallack et al. (1998) examined events at the Permian-Triassic (P-T) boundary using terrestrial successions from Australia and Antarctica. Although the paper's data were inconclusive, we are concerned about the placement of the boundary at Mount Crean in southern Victoria Land, Antarctica. It is our contention that the P-T transition is not present at Mount Crean; therefore, we believe that Retallack et al.'s boundary requires further discussion and clarification.

Retallack et al.'s (1998) criteria for identifying the P-T boundary include: (1) a decrease in  $\delta^{13}\text{C}_{\text{org}}$ ; (2) extinction of the coal-forming *Glossopteris* flora and its distinctive *Vertebraria* roots; (3) disappearance of coal; (4) appearance of the *Dicroidium callipteroides* flora; and (5) appearance of the *Lystrosaurus* therapsid fauna. The criteria are ambiguous, and may be difficult to use at any one locality. For example, *Glossopteris* occurs in the Lower Triassic of India (Bose et al., 1990), the *Lystrosaurus* fauna is not ubiquitous, and although Lower Triassic rocks are devoid of coal, the vertical disappearance of coal is not a reliable time line, especially in fluvial coal measures like the Permian of Antarctica (Isbell and Cúneo, 1996).

At Mount Crean, Permian and Triassic strata consist of the Weller Coal Measures and the overlying Feather Conglomerate. Retallack et al. (1998) placed the P-T boundary within a "claystone breccia" located 2 m below the top of the Weller Coal Measures. Krull et al. (1996) reported (but did not illustrate) an isotopic excursion in  $\delta^{13}\text{C}_{\text{org}}$  within the bed. The "breccia" bed occurs several meters above the top of the highest coal at Retallack et al.'s (1997) site, which is in turn overlain by a *Vertebraria*-bearing paleosol.

We question Retallack et al.'s use of the term "breccia." The "claystone breccia" at Mount Crean is the same bed that Kyle (1976) described as a coal containing quartz grains 1 to 7 mm in diameter. Kyle's figure 2.4 clearly shows mainly rounded to subrounded quartz grains within the coal, which she interpreted as detrital in origin.

We believe that the boundary at Mount Crean cannot be identified using fossils and is not a paleobotanical boundary as stated. *Glossopteris* leaves are last recorded 15 m below the proposed boundary, though presumed *Vertebraria* roots do occur near (as noted above) Retallack et al.'s (1998) boundary bed (interpreted from Fig. 2 of Retallack et al., 1997; there are no fossils shown in Fig. 2 of Retallack et al., 1998). *Dicroidium* has not been described from either the uppermost Weller Coal Measures or the Feather Conglomerate (Askin et al., 1971; Retallack et al., 1997). In southern Victoria Land, *Dicroidium* does not occur below the overlying upper Lower to Upper Triassic Lashly Formation, and the *Lystrosaurus* fauna has never been described. The proposed boundary also conflicts with palynomorph data. Kyle (1976, 1977) and Kyle and Schopf (1982) used palynomorphs from the middle and upper Weller Coal Measures to define the *Protohaploxypinus* zone, which broadly correlates with the Lower Permian (Artinskian) Stage 4 palynomorph zone in eastern Australia (Kemp et al., 1977). Mount Crean palynomorphs are typical of the *Protohaploxypinus* zone (Askin, 1997). The only possibility for confusion is the presence of *Praecolpatites sinuosus* (Askin, 1997) within a sample collected approximately 2 m below Retallack et al.'s boundary bed. Although *P. sinuosus* is characteristic of Upper Permian rocks, it also occurs in Lower Permian Stage 3 and Stage 4 strata in Australia (Playford, 1990). No palynomorph restricted to only Upper Permian rocks has ever been recovered from the Weller Coal Measures (Askin, 1997) and no palynological evidence for a latest Permian age exists.

Previous Antarctic workers have acknowledged an Early Permian age for the Weller Coal Measures. However, considerable disagreement occurs as to the age of the overlying strata. Although Kyle (1976, 1977) reported Lower Triassic palynomorphs in the uppermost Feather Conglomerate, no micro- or megafossils have ever been recovered from the lower parts of the formation. At Mount Crean, an erosional surface separates the Weller Coal Measures from the Feather Conglomerate (Askin et al., 1971), which Isbell and Cúneo (1996) interpreted as a major disconformity, suggesting that the entire Upper Permian is missing. However, Barrett and Fitzgerald (1985) believed that the lower Feather Conglomerate is Upper Permian. Regardless of the correct scenario, palynomorph data indicate that the Weller Coal Measures at Mount Crean are Lower Permian to possibly lower Upper Permian, but not uppermost Permian.

We believe that Retallack et al.'s (1998) placement of the P-T boundary in Antarctica is oversimplified. Although we do not dispute the possibility that Antarctic strata may record the P-T transition, we believe that unequivocal evidence has yet to be collected and/or presented.

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## REPLY

Gregory J. Retallack

Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403-1272, USA

Isbell and Askin advance the “contention that the Permian-Triassic transition is not present at Mount Crean” with surprising confidence, considering they also state that in the absence of suitable paleontological data, a variety of hypotheses of age and disconformities are possible. The real problem is that a 95 m thickness of uppermost Weller Coal Measures and Feather Conglomerate at Mount Crean has not yet yielded age-diagnostic fossils. These strata are no older than early Late Permian megafossil plants and palynomorphs in the middle and lower Weller Coal Measures and no younger than early Middle Triassic fossil plants and palynomorphs in the upper Feather Conglomerate and overlying Lashly Formation (Retallack et al., 1997). What the uppermost unfossiliferous Weller Coal Measures do contain is a decrease in the carbon isotopic composition of organic matter (Krull, 1998), widely taken as an indicator of the Permian-Triassic boundary (Morante, 1996). Also remarkable are traces of iridium and shocked quartz found on this and not other horizons (Retallack et al., 1998a). Isbell and Askin’s claim that our proposed boundary “conflicts with palynomorph data” is not strictly true, because *Praecolatites sinuosis* (the principal age-diagnostic form at Mount Crean; Askin, 1997) ranges to the end of the Permian in other parts of Antarctica (Farabee et al., 1991). Furthermore, no one has yet found Triassic fossils below or Permian fossils above our proposed Permian-Triassic boundary beds.

Fortunately, there is additional fossil evidence available to test our chemostratigraphic hypothesis of age, and some of these fossils are currently under study by Askin and Retallack. Gabites (1985) has illustrated a thin section of permineralized peat from the Feather Conglomerate in the Allan Hills with fern leptosporangia so well preserved that other age-diagnostic fossils can be expected. The small, narrow, fine-meshed *Glossopteris* leaves and fructifications (*Senothea*, *Squamella*) in the upper Weller Coal Measures at Mount Crean are more like assemblages from the Late Permian Illawarra and Newcastle Coal Measures of Australia (Retallack, 1999) than the larger-leaved *Glossopteris* and reproductive structures (*Plumsteadia ovata*) of the middle and lower Weller Coal Measures of Antarctica.

There is also the prospect of a regional definition of the Permian-Triassic boundary using paleosols. This can be considered paleontological evidence in the sense that paleosols are trace fossils of ecosystems. The distinctive green, nodular paleosols of the Dolores pedotype have been found only in earliest Triassic (not latest Permian) in the richly fossiliferous

sequences of the central Transantarctic Mountains (Retallack et al., 1998b). Significantly, Dolores paleosols are common in the unfossiliferous uppermost Weller Coal Measures at stratigraphic levels above the last *Glossopteris* leaf or *Vertebraria* root (Retallack et al., 1997). After working in the central Transantarctic Mountains, our team also discovered that all very weakly developed Early Triassic paleosols had noncarbonaceous root traces (and so should be referred to the Shaun pedotype), whereas all comparable Permian paleosols have carbonaceous root traces (Susanne pedotype; Retallack et al., 1998b). On these grounds as well as from the presence of Dolores paleosols, the uppermost Weller Coal Measures can be considered earliest Triassic. This revision also means that there is not a single pedotype common to both Early Triassic and Permian rocks in Antarctica, as is the case also in Australia (Retallack, 1999).

Our claystone breccia at Mount Crean is very likely the same rock noted by Askin, because cursory examination in the field appears to show large (up to 7 mm) granules of vein quartz floating in black coaly material, and there is nothing like it elsewhere in the sequence. How such large grains could get out into a sediment-starved swamp is a sedimentological paradox that stimulated further laboratory study. Thin sections (Fig. 1), XRD traces (mainly illite, with traces of smectite), and chemical analysis (30.78 wt% SiO<sub>2</sub>, 13.32 wt% Al<sub>2</sub>O<sub>3</sub>) show that what in hand specimen appears to be a dark, coaly matrix is in fact deformed carbonaceous claystone clasts. The claystone breccia at Mount Crean is remarkably similar to the black claystone breccia at the Permian-Triassic boundary at Coalcliff and Wybung Head in the Sydney Basin of New South Wales, Australia (Retallack, 1999). Although this rock does not present a conundrum of hydrodynamic sorting, it remains intriguing and worthy of further isotopic, geochemical, and mineralogical study (Retallack et al., 1998a).

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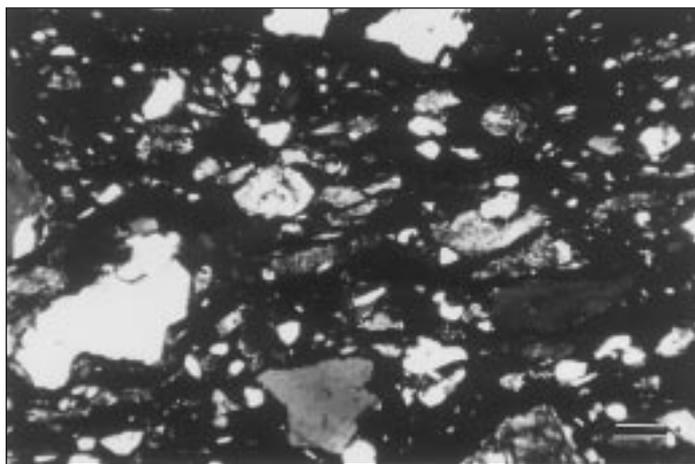


Figure 1. Petrographic thin section under crossed nicols of claystone breccia at Mount Crean, showing that large dark areas are clasts rather than matrix. Scale bar is 0.1 mm.

## Continental extensional setting for the Archean Belingwe Greenstone Belt: Comment and Reply

### COMMENT

**T. M. Kusky**

*Center for Remote Sensing, Boston University, Boston, Massachusetts 02215, USA*

**W. S. F. Kidd**

*Department of Geological Sciences, University at Albany, Albany, New York 12222, USA*

Hunter et al. (1998) used stratigraphic data from the basal Manjeri Formation of the Archean Belingwe Greenstone Belt to support an ensialic model for the entire structurally complex belt. They claimed their data refute claims by Kusky and Kidd (1992) that a several kilometers-thick pile of  $2692 \pm 9$  Ma (Chauvel et al., 1993) basalts and komatiites in the upper part of the greenstone belt is allochthonous.

Hunter et al. (1998) claimed that Kusky and Kidd (1992) stated the entire Manjeri Formation formed in a foreland basin, yet in other places they claimed we suggested the entire Upper Greenstone succession (Manjeri, Reliance, Zeederbergs, and Cheshire Formations) is allochthonous, and represents a single tectonic environment. We have not made such statements, but agree with Hunter et al. that rocks of the basal part of the Manjeri Formation were deposited in an extensional and/or passively subsiding shallow marine setting, with fluvial and transitional marine-fan delta influences higher in the autochthonous sequence. Kusky (1998) suggested the basal part of the Manjeri Formation is a remnant of a regionally extensive passive margin sequence deposited on the southeastern edge of the proto-Zimbabwe craton, in the interval of 2831–2700 Ma, prograding over older gneissic rocks and a 2904–2831 Ma rift sequence preserved in the Lower Greenstone sequences. We interpret the middle clastic part of the Manjeri Formation as a drowning sequence, reflecting the transition into a flysch basin. The upper part of the Manjeri Formation is consistently a strongly deformed banded iron formation. Only the middle part of the sequence was interpreted by Kusky and Kidd (1992) to be a foreland basin deposit, and data presented by Hunter et al. lend further support to this interpretation.

Kusky and Kidd (1992) and Kusky and Winsky (1995) suggested the Mberengwa allochthon (consisting of komatiites and basalts of the Reliance and Zeederbergs Formations) was emplaced over the Manjeri Formation using the banded iron formation at the top of the Manjeri Formation as the regional detachment at 2700–2650 Ma. The data presented by Hunter et al. (1998) are limited to the basal part of the 2831–2700 Ma Manjeri Formation. It is most unlikely that the basal Manjeri Formation has anything to do with a thrusting event that postdates deposition of these rocks by up to 130 m.y.

Hunter et al. presented misleading statements about local basement sources for the Manjeri Formation (implying the entire formation), when their chemical data is limited to the basal 20 m (Spring Valley Member) of section. These data have no significance for whether or not the structurally overlying Reliance and Zeederbergs Formations are allochthonous. In both interpretations, this basal part of the Manjeri Formation was deposited on the older gneissic terrane, and Hunter et al. produced geochemical confirmation of this relationship. It would have been informative if Hunter et al. had presented data on the upper part of the Manjeri Formation, which may, if the allochthonous model is correct, show a transition into a sequence containing a contribution from sources in the eroding allochthon. Interestingly, Hunter (1997) showed that the middle (Rubweruchena) Member of the Manjeri Formation is less fractionated than the underlying Spring Valley Member, consistent with the transition from a passively subsiding environment to a foreland basin as inferred by Kusky and Kidd (1992). These data were not reported by Hunter et al. (1998).

The facies analysis of the upper Manjeri Formation (Rubweruchena Member) presented by Hunter et al. is non-unique—they suggested graywackes, shales, sandstones, and conglomerates were deposited in a basin by “small-scale alluvial fans or fan-deltas.” This could be a continental exten-

sional basin as postulated by Hunter et al., or a foreland basin in which allochthonous plateau basalts were exposed, as suggested by Kusky and Kidd (1992). Foreland basins characteristically show sources, including the exposed allochthon and the underlying basement terrane exposed in normal fault scarps on the outer trench slope.

Hunter et al. previously denied, but now dismiss as insignificant, evidence of deformation along the contact between the Manjeri and Reliance Formations, despite remarkable similarities between this shear zone and stratigraphically controlled regional detachment zones of mountain belts worldwide (Kusky and Winsky, 1995). Hunter et al. attributed deformation along the Manjeri-Reliance contact to strain accommodation in a tightly folded syncline. However, their model did not explain why this particular contact is ubiquitously deformed, whereas other contacts with similar rheological contrasts and orientations are not deformed. It also did not account for the kinematics of the shear zone, which show subhorizontal lineations and movement directions in the low-grade chert-tectonites, with relative northward movement of the Mberengwa allochthon with regard to the Manjeri Formation and gneiss terrane, nor the presence of thrust ramps that cut out part of the upper Manjeri Formation, with consistent ramping up to the northwest.

Hunter et al. ended by pondering the paradox of how regionally similar volcanic belts might be deposited contemporaneously in isolated rifts across the craton. They simply were not. To support their model, Hunter et al. misquoted several previous works, including a report of alleged “basement zircon xenocrysts in more felsic units” (Wilson et al., 1995); Wilson et al. (1995) presented no data from the Upper Greenstones at Belingwe, but only from the underlying 2.9 Ga greenstone assemblage. Likewise, data of Chauvel et al. (1993) are overstated by Hunter et al. to support geochemical contamination of komatiites of the Reliance Formation by older continental crust. Chauvel et al. (1993) were more cautious in the interpretation of their Pb and Nd isotopic data. They simply noted that if there is contamination in the komatiites, it amounts to less than 1%, and they considered this as only one possibility along with others, including fractionation, alteration, and problems associated with sampling different, widely separated lava flows. Their samples were also considered to be contaminated by fluids that circulated through nearby late faults.

In summary, Hunter et al. (1998) presented data that confirm field relationships in the Belingwe Greenstone Belt but bear no relationship to the emplacement of the Mberengwa allochthon.

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## REPLY

M. A. Hunter\*

M. J. Bickle

Department of Earth Sciences, University of Cambridge,  
Cambridge CB2 3EQ, UK

E. G. Nisbet

Department of Geology, Royal Holloway University of London, Egham,  
Surrey TW20 0EX, UK

A. Martin

6, Autumn Close, Greendale, Harare, Zimbabwe

H. J. Chapman

Department of Earth Sciences, University of Cambridge,  
Cambridge CB2 3EQ, UK

The core arguments of Kusky and Kidd's Comment rest on whether there is evidence for a major allochthon in the Belingwe Greenstone belt. We see no evidence of an allochthon.

Kusky and Kidd (1992) suggested downwarping of the crust in front of a large accreting block and consequent deposition of the rocks we have described as the Rubweruchena Member of the Manjeri Formation as a drowning sequence. The Rubweruchena Member, where present, consists of 0–150 m of poorly sorted conglomerate, pebble beds, and sands. Erosion of an allochthon as suggested by Kusky and Kidd (1992) would create a basin with a thick sedimentary succession. It seems unlikely that the accretion of an allochthon many kilometers thick and ~45 000 km<sup>2</sup> would generate a basin fill only 0–150 m thick.

Kusky and Kidd's argument that the basal Manjeri Formation was deposited from 2831 Ma is unrealistic. The 2831 Ma age was recorded from a reworked dacitic clast in underlying rocks that were uplifted, deformed, and eroded prior to deposition of the Manjeri Formation. In contrast, we believe that the Manjeri Formation is broadly conformable with the rest of the Ngezi Group, and it is more likely that the sedimentary succession was laid down in a relatively short period around 2.7 Ga.

Material contributed to the Rubweruchena Member from a postulated advancing allochthon of Reliance and Zeederbergs Formations would

\*Present address: British Antarctic Survey, High Cross, Madingley Road, Cambridge CB30ET, UK.

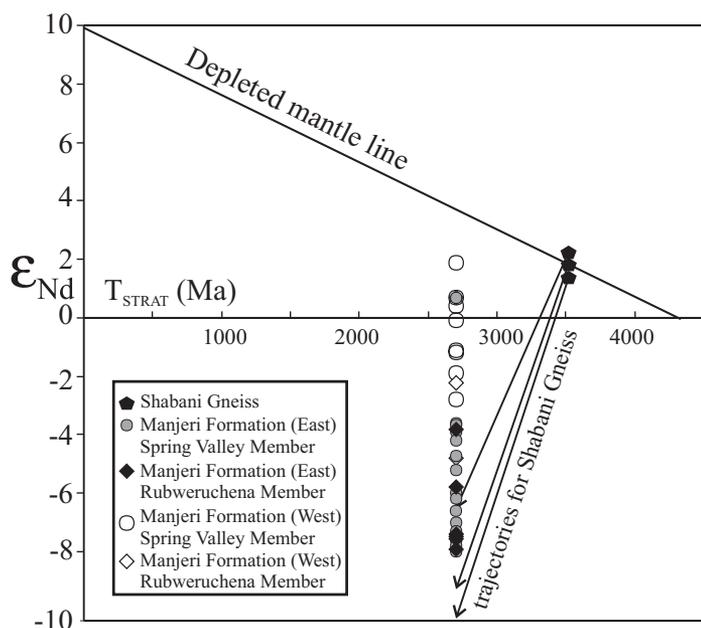


Figure 1: Plot of  $\epsilon_{Nd}$  against  $T_{STRAT}$  for Manjeri Formation.  $\epsilon_{Nd}$  calculated for  $T_{STRAT} = 2700$  Ma.

dominate the modal mineralogy of the Rubweruchena Member, reduce the apparent fractionation of rare earth element (REE) patterns, lower the model age of the source terrain, and raise the  $\epsilon_{Nd}$ . We observed none of these features in the sedimentary record of the Manjeri Formation; the mineralogy of the Rubweruchena Member is continental. REE patterns from the Rubweruchena Member indicate heterogeneous local basement sources throughout deposition (Hunter, 1997). This conclusion is reinforced by model ages from the Spring Valley and Rubweruchena Members (Hunter et al., 1998). Sm-Nd model ages from the eastern side of the belt (3234–3708 Ma) reflect the age range of the underlying granitoids, and younger model ages from the west (2969–3103 Ma) mirror the underlying greenstones. Most  $\epsilon_{Nd}$  are negative and show a distinct bimodality between east and west for a stratigraphic age of 2700 Ma (the age of the volcanic rocks and probably the stratigraphic age of the sediments; Fig. 1). As would be expected,  $\epsilon_{Nd}$  values from the Reliance Formation directly overlying the Manjeri Formation are positive at 2700 Ma (Chauvel et al., 1993).

The evidence of deformation at the contact between the Manjeri and Reliance Formations was described by us (Bickle et al., 1975) and has been discussed in several papers (Bickle et al., 1994 and references therein). The Jimmy Member is certainly a plane of strain accommodation, but it is not a mylonite developed on a large-scale detachment as suggested by Kusky and Kidd (1992). Grassineau et al. (1999) showed fine-scale, unambiguously biogenic, heterogeneity in S and C isotopic ratios in samples from this horizon ( $\delta^{34}S$  –15‰ to +17‰ in sulfide,  $\delta^{13}C$  –38‰ to –28‰ in kerogen); this heterogeneity is inconsistent with deformation of the gossan in a major shear zone.

Wilson has recognized the Manjeri unconformity widely across the Zimbabwe craton. This work, in several papers (summarized in Wilson et al., 1995), presents evidence for inherited (xenocrystic) zircons in the Upper Bulawayan. Several studies have confirmed continental contamination of the volcanic pile at Belingwe (e.g., Scholey, 1992). The conclusion is strong: the succession is ensialic. The arguments of Kusky and Kidd are contradicted by the sedimentary evidence in the Manjeri Formation, and the large scale thrusting they propose is precluded by preservation of fine-scale biogenic heterogeneity in the Jimmy Member.

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## Hydrothermal origin of Devonian conical mounds (kess-kess) of Hamar Lakhdad Ridge, Anti-Atlas, Morocco: Comment and Reply

### COMMENT

Michael M. Joachimski\*

Werner Buggisch

*Institute of Geology, University of Erlangen, Schlossgarten 5,  
91054 Erlangen, Germany*

Mounji et al. (1998) presented carbon and oxygen as well as strontium isotope data for microspars and diagenetic cements of Lower Devonian mud mounds from Morocco. On the basis of low  $\delta^{18}\text{O}$  values of the microspars, the authors proposed that the precipitation of the carbonate muds and the formation of the mud mounds was driven by hydrothermal venting. We acknowledge any attempt to explain the formation of the Lower Devonian carbonate mud mounds. However, the stable isotope data presented by Mounji et al. (1998) do not support the conclusion that hydrothermal venting initiated the formation of the Lower Devonian mud mounds.

Microspar samples collected from the intermound and mound limestones as well as from neptunian dyke infillings show  $\delta^{18}\text{O}$  values ranging from  $-7\%$  to  $-11.2\%$ . The light  $\delta^{18}\text{O}$  values are taken as evidence for precipitation of the microspars from hydrothermal fluids. We disagree with this interpretation. Microspar is considered not to be a primary precipitate but to be formed either as early diagenetic cement in meteoric (Steinen, 1978; Lasemi and Sandberg, 1984) or marine (Munnecke et al., 1997) pore waters or by aggrading neomorphism (Folk, 1965). We agree with Mounji et al. (1998) that there is no evidence for early meteoric alteration of the carbonate muds and argue that the micritic carbonate muds were transformed into microspar during burial of the sediments. Conodont color alteration indexes of conodonts from the Hamar Lakhdad (CAI = 4; Belka, 1991) indicate that maximum burial temperatures were beyond  $200\text{ }^\circ\text{C}$ . Assuming that the diagenetic fluid was mainly Devonian sea water with a presumed  $\delta^{18}\text{O}$  value of  $-1\%$  to  $-3\%$  (SMOW),  $\delta^{18}\text{O}$  values of the microspars translate into burial temperatures of  $62$  to  $75\text{ }^\circ\text{C}$ . This estimate is well below expected maximum burial temperature. Nonferroan radiaxial calcite cements (group Cl<sub>I</sub>) interpreted by Mounji et al. (1998) as early marine cements give  $\delta^{18}\text{O}$  values of  $-2\%$  to  $-6\%$  which are consistent with precipitation from Devonian sea water. Microspars sedimented between early marine cement crusts are again depleted in  $^{18}\text{O}$ . According to Mounji et al. (1998), the microspars of the mound and intermound facies precipitated from hydrothermal fluids, early marine cements crystallized in normal marine waters and microspars sedimented between these cement crusts precipitated again from hydrothermal fluids. We think that this scenario is highly unlikely and argue that the low  $\delta^{18}\text{O}$  values of the microspars simply reflect aggrading neomorphism of the carbonate muds under elevated temperatures.

The carbon isotope values of microspars ( $M_{II}$ ) as well as radiaxial (Cl<sub>II</sub>), nonferroan and ferroan granular calcite cements (C3<sub>II</sub>) from one locality described as "a dark limestone at the base of a Middle Devonian mud mound known as the Hollard mound and in the vicinity of a fault zone" are depleted with respect to  $^{13}\text{C}$ . The lowered  $\delta^{13}\text{C}$  values are interpreted by Mounji et al. (1998) as evidence (1) for primary precipitation of microspars and cements from methane-bearing fluids or (2) for diagenetic alteration of the calcites under the influence of methane.  $\delta^{13}\text{C}$  values as low as  $-17\%$  do not conclusively support the idea of a contribution of methane.  $\delta^{13}\text{C}$  of bicarbonate in pore waters may have been depleted in  $^{13}\text{C}$  due to oxidation of organic matter by bacterial sulfate reduction, especially since the negative  $\delta^{13}\text{C}$  values are only observed in samples taken from a dark and probably organic carbon-rich limestone bed. Bicarbonate derived exclusively from bacterial sulfate reduction would have a carbon isotopic composition of around  $-21\%$  to  $-22\%$  (assuming  $15$  to  $25\text{ }^\circ\text{C}$  and fractionation factors given by Romanek et al., 1992). As a consequence, a very high proportion

of carbon derived from the oxidation of organic carbon on total dissolved bicarbonate may explain the light  $\delta^{13}\text{C}$  values of cements and microspars. Nevertheless, even if venting occurred along the fault zone in the eastern part of the Hamar Lakhdad during the Middle Devonian, this does not prove that the formation of the Lower Devonian mud mounds is related to hydrothermal or methane venting, since microspars from the Lower Devonian mound and intermound facies (group M<sub>I</sub> and M<sub>II</sub>, average  $\delta^{13}\text{C}$  of  $-0.43\%$  and  $-7.84\%$ , respectively) do not reflect  $\delta^{13}\text{C}$  values characteristic for vent-related carbonates. Thus, although hydrothermal venting may represent a fashionable idea to explain the formation of the Lower Devonian mud mounds, the presented isotope data do not support this conclusion.

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### REPLY

Driss Mounji

P.-A. Bourque\*

*Département de géologie et de génie géologique, Université Laval,  
Québec G1K 7P4, Canada*

M. M. Savard

*Delta-lab, Geological Survey of Canada, Centre géoscientifique de  
Québec, Ste-Foy, Québec G1V 4C7, Canada*

We proposed that the finely crystalline material that forms the mounds and the intermound beds of the peculiar cone-shaped carbonate mounds of the Hamar Lakhdad, Morocco, is based on integration of sedimentological, architectural, and geochemical constraints (Mounji et al., 1998). Joachimski and Buggisch disagree with our interpretation of the oxygen isotope results we obtained on the finely crystalline material (microspar) that makes up the bulk of the mound and intermound limestone (the term microspar was used to indicate crystal size of  $4$  to  $20\text{ }\mu\text{m}$ , without any genetic connotation), and of the carbon isotope results we obtained on microspar and cements of one particular mound, the Hollard mound.

Among the three hypotheses Joachimski and Buggisch cite for microspar formation, the first two they discard, retaining only the aggrading metamorphism of Folk (1965). They argue that "the micritic carbonate muds were transformed into microspar during burial of the sediments," and that "the low  $\delta^{18}\text{O}$  values of the microspars simply reflect aggrading neomorphism of the carbonate muds under elevated temperatures." Our contention is that the microspar formed during early cementation of a primary finer-grained sediment, which at this stage was likely affected by a recrystalliza-

\*E-mail: joachimski@geol.uni-erlangen.de.

\*E-mail: bourque@ggl.ulaval.ca

tion that did not significantly change the primary isotopic signal. This assertion is based on the following.

Composition of the precursor sediment is difficult to assess. However, the very low Sr content (124–387 ppm,  $n = 11$ ; Mounji, 1999) and the scarceness of microdolomite inclusions into the microspar (scanning electronic microscope observations and microprobe analyses, work in progress), together with the near absence of available aragonitic skeletons (see Brachert et al., 1992, for an inventory of the biota), suggest that the primary material was mainly low magnesium calcite (LMC). We postulate that this material was cemented early. Early cementation on the sea floor or in the shallow burial environment is a well-known feature of most Paleozoic mud mounds. In the specific case of the Moroccan kess-kess, early cementation of the mound and intermound muds is attested by the presence of early marine cement in all kinds of cavities, particularly in the roof-unsupported stromatactid cavities, and the absence of sliding or slumping on mound steep flanks (Brachert et al., 1992; our own observations).

Since the pioneer work of Bathurst (1958, 1975) and Folk (1965), who considered that lime mud was lithified mainly during burial diagenesis by aggrading neomorphism that “digests” primary particules, more and more examples are described where muds were cemented early and where microspar is not the result of aggrading neomorphism (*sensu* Folk, 1965) but of cementation of the primary material (e.g., Lasemi and Sandberg, 1984; Lasemi et al., 1990; Munnecke et al., 1997). In our case, when viewed under cathodoluminescence, the kess-kess microspar is made up of nonluminescent particules, one quarter to half the size of the microspar crystals, embedded in a luminescent cement. The particule distribution with respect to the cement varies from closely to loosely packed. We interpret this situation as a representation of the primary mud cemented by early LMC cement (low Sr content and absence of microdolomites, as cited above). Moreover, scanning electron microscope observations on slightly etched polished surface show that the microspar crystals (4–20  $\mu\text{m}$  in size) contain smaller 1–2  $\mu\text{m}$  equant sub-crystals, more likely representing the primary mud particules. These particules should have been “digested,” and consequently not observable, if aggrading neomorphism was the acting process that produced microspar (see Bathurst, 1975).

It is possible that the early cemented mud that gave rise to microspar has been subsequently recrystallized (*sensu* Bathurst, 1975) during marine or shallow burial diagenesis, although this is difficult to demonstrate. The important point, however, is that it is likely that further burial of the limestone did not alter the stable isotope signature of the primary mixture of LMC particles and cement, a mixture we interpret as composed of hydrothermally precipitated mud and marine or very shallow burial cement. Of most examples we know where microspar was analyzed for C and O stable isotopes, particularly in mud-mound studies,  $\delta^{18}\text{O}$  values range in the early marine to shallow burial environment field. A particularly interesting case occurs in part of the Tafilalt-Maïder basin where mounds are not related to early sea-floor hydrothermalism. Microspar of the large mid-Devonian Aferdou el Mrakib mud mounds has  $\delta^{18}\text{O}$  values yielding in the mid-Devonian marine field, while the nearby Jbel el Otfal and Guelb el Maharch mounds that should have reached a similar burial depth as the Aferdou el Mrakib mound have yielded  $\delta^{18}\text{O}$  values 1.5‰ to 2.5‰ lower than the marine field (Kauffman, 1997; Mounji, 1999), pointing to cementation in the shallow burial environment. We doubt that these  $\delta^{18}\text{O}$  values of the microspar reflect burial temperatures in the 200 °C cited by Joachimski and

Buggish. To us, this indicates that the deep burial of the limestones does not significantly affect the isotopic signal of primary sediments and cements.

Therefore, we maintain that the most reasonable hypothesis to explain the low  $\delta^{18}\text{O}$  values of the kess-kess microspar is that the fine-crystalline carbonate was originally composed of particulate material precipitated under hydrothermal influence and marine cement, and that subsequent burial diagenesis more likely did not alter this primary mixed signal.

Concerning the low  $\delta^{13}\text{C}$  values of microspar ( $M_{II}$ ) and radiaxial non-ferroan ( $C_{1II}$ ) and ferroan ( $C_{3II}$ ) calcite cements of the Hollard mound, our work is in progress, as we wrote in the article. The Hollard mound harbors unique features including a fault zone cutting through a dark carbonate unit in which most of the early marine components cited above yield low  $\delta^{13}\text{C}$  values. This peculiar setting deserves particular attention, and therefore we plan to devote an entire article to it. For now, we are uncertain about the source or nature of the biogenic carbon. It is possible to envision that the light bicarbonate could be derived from a bacterially mediated oxidation of liquid or gas hydrocarbons (as we suggested), such as sulfate reduction or others, or from decarboxylation of hydrocarbons (Tissot and Welte, 1984). However, at this stage there is no indication that the light bicarbonates were seeping from a deeper source or produced *in situ*.

A compilation of our new data shows two populations trending from a marine, heavy  $\delta^{13}\text{C}$  (+2.5‰) to a light end-member (−18.5‰), one along a marine-like  $\delta^{18}\text{O}$  line (−3.5‰), and the other at around −11.0‰. These two populations strongly suggest that three geochemical systems mixed during the evolution of the Hollard mound. The two  $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$  trends clearly result from a mixture of marine-like bicarbonates, with biogenic bicarbonates of yet unknown origin, and with high-temperature waters; the biogenic bicarbonates could well be seeping with the hydrothermal waters in a marine setting.

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#### CORRECTION

Three-stage evolution of the Los Angeles basin, southern California

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The published version of Figure 1C was printed incorrectly. The abbreviation for Ridge basin (RB) was truncated along the northern margin of the map. Only the southern edge of RB is visible. All of the rotated Western Transverse Ranges should be moved to the north approximately 1 cm, so they are directly against the SG-CH fault. This revised position better aligns the SYCF with the northern end of the SAM. The Verdugo Hills block probably was northeast of the SG-CH fault in this revised reconstruction.