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# Paleoenvironment of the Western Interior Seaway inferred from $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of molluscs from the Cretaceous Bearpaw marine cyclothem

S. He<sup>a,1</sup>, T.K. Kyser<sup>b,\*</sup>, W.G.E. Caldwell<sup>c</sup>

<sup>a</sup>Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, Canada S7N 5E2

<sup>b</sup>Department of Geological Sciences and Geological Engineering, Queen's University, Kingston, Ontario, Canada K7L 3N6

<sup>c</sup>Department of Earth Sciences, The University of Western Ontario, London, Ontario, Canada N6A 5B7

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## Abstract

*What are these?*

$\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values of ammonites and inoceramids provide information on the paleoenvironmental conditions in the Western Interior Seaway (WIS) during the Campanian–Maastrichtian Bearpaw marine cycle, the last major Transgression–Regression (T–R) cycle to affect the seaway during the Cretaceous Period. Ammonites and inoceramids exhibit distinct stable-isotope values: inoceramids have higher  $\delta^{13}\text{C}$  values (0.6‰ to 5.0‰) and lower  $\delta^{18}\text{O}$  values (−4.0‰ to −2.8‰), whereas coeval baculitid ammonites have lower  $\delta^{13}\text{C}$  values (−4.8‰ to −0.3‰) and higher  $\delta^{18}\text{O}$  values (−2.3‰ to 0.0‰), with other ammonites having stable-isotope values ranging between the baculitids and inoceramids. The isotopic composition of the inoceramids and ammonites are unlikely due to kinetic or vital effects because species from open ocean environments have values expected from marine water and there is no correlation between the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values of the inoceramids as has been reported for modern foraminifers and calcareous algae suspected of exerting vital effects. The heterogeneity in stable-isotope values of coeval molluscs implies that the Bearpaw Sea was isotopically stratified. Isotopic exchange between the water and sediments, and the formation of  $^{18}\text{O}$ -depleted deepwater mass by mixing of Tethyan and Boreal waters in the WIS, may have caused such isotopic stratification.

A clear relationship exists between the  $\delta^{18}\text{O}$  values of Bearpaw zonal baculitids and their biostratigraphic sequence. Baculitids from zones during peak transgression have the lowest average  $\delta^{18}\text{O}$  values (−2.3‰ to −0.7‰), whereas those from the underlying and overlying zones have higher  $\delta^{18}\text{O}$  values (−0.8‰ to 0.2‰). This pattern of  $\delta^{18}\text{O}$  values can be explained by fluctuations in temperature rather than variations in freshwater influx, this influx probably having been reduced by lower precipitation and run-off under drier, warmer, climatic conditions. The Bearpaw Sea was not brackish and

\* Corresponding author. Tel.: +1 613 533 6179; fax: +1 613 533 6592.

E-mail address: [kyser@geol.queensu.ca](mailto:kyser@geol.queensu.ca) (T.K. Kyser).

<sup>1</sup> Present address: Department of Earth Sciences, The University of Western Ontario, London, Ontario, Canada N6A 5B7.

other paleoenvironmental factors likely account for the character of the Bearpaw fauna that is dominated by ammonites and inoceramids and lacks many of the taxa prevalent in the contemporaneous open oceans and seas.

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## 1. Introduction

During the Cretaceous Period, the Western Interior of North America was the site of a vast foreland basin,

extending from the Gulf of Mexico to the Arctic Ocean and from eastern British Columbia to western Ontario (Fig. 1). Both Tethyan waters from the Gulf of Mexico and Boreal waters from the Arctic Ocean



Fig. 1. Map showing the extent of the Cretaceous Western Interior Seaway of North America during the Bearpaw cyclothem at *Baculites baculus* time (after Williams and Stelck, 1975).

invaded the basin. They became confluent in Late Albian time to form the Western Interior Seaway, which persisted, variably expanded and shrunken, until Maastrichtian times. Of Kauffman's (1977) ten eustatically controlled, major T–R marine cycles recognized during the Cretaceous Period, four (the Greenhorn, Niobrara, Claggett, and Bearpaw cycles) occupied most of the Late Cretaceous Epoch and are widely recorded in the sedimentary rocks in the Western Interior Basin (WIB).

Knowledge of the paleoenvironments of the seaway is critically important to understanding the climatic development of North America in later Mesozoic time. Consequently, the seaway has been the subject of numerous geochemical investigations, including those by Tourtelot (1962, 1964), Tourtelot and Rye (1969), Forester et al. (1977), Pratt (1983), Wright (1987), Whittaker et al. (1987), Whittaker and Kyser (1993), Kyser et al. (1993), Pratt et al. (1993), Cadrin et al. (1996), He et al. (1996), and Fisher and Arthur (2002). However, interpretation of the data has always been problematic in one way or another. Difficulty in determining the paleoenvironmental conditions of the seaway has arisen partly from the fauna being dominated by ammonites and bivalves and lacking many of the taxa that typically prevailed in the coeval open oceans and seas, including most corals, bryozoans, brachiopods, echinoderms, and skeletonized sponges (Gill and Cobban, 1966; Caldwell, 1968; Kauffman and Caldwell, 1993). This characteristic of the fauna is usually attributed to the prevalence of somewhat brackish-water conditions in both surface and bottom waters of the seaway during the Late Cretaceous Epoch (Kauffman and Caldwell, 1993). Moreover, geochemical studies of Late Cretaceous sedimentary rocks and fossils from the basin (Wright, 1987; Cadrin, 1992; Kyser et al., 1993; Pratt et al., 1993; Cadrin et al., 1996; Fisher and Arthur, 2002) also indicate somewhat more negative  $\delta^{18}\text{O}$  values for the seaway than those of normal open-marine water, which suggests the existence of brackish waters in the seaway. More negative  $\delta^{18}\text{O}$  values of these rocks and fossils, however, could also imply higher average temperatures in the seaway than in contemporaneous open oceans and seas. The variations in  $\delta^{18}\text{O}$  values of molluscs from the seaway could, therefore, be due to more than one process.

The purpose of this study was to analyze the paleoenvironmental conditions of the seaway during the latest part of the Late Cretaceous Epoch through a detailed investigation of the stable-isotope compositions of the molluscan fossils from the Late Campanian–Early Maastrichtian Bearpaw cyclothem and its equivalents. The fossils used are mainly from localities in the broad, mid-latitudinal portion of the WIB, where the fossils record the meeting and mixing of waters from circum-polar and subtropical seas.

## 2. Bearpaw cyclothem

The Bearpaw marine cyclothem, the last major T–R cycle to affect the WIB, is represented primarily by the Bearpaw Formation (Fig. 2), which outcrops in southwestern Saskatchewan and southern Alberta (Fraser et al., 1935; Russell and Landes, 1940; Furnival, 1946; Lines, 1963; Caldwell, 1968; Rask, 1969), and by the Bearpaw Shale in Montana and north-central Wyoming (Tourtelot, 1962; Gill and Cobban, 1973). It is coeval with middle to upper part of the Pierre Shale in eastern Saskatchewan, Manitoba, and the northeastern part of the basin in the United States (Caldwell, 1968; Gill and Cobban, 1966, 1973; McNeil and Caldwell, 1981). The Bearpaw Formation and its equivalents form a westward-thinning cyclothem wedge of predominately marine silty clays and sands, underlain by a complementary eastward-thinning wedge of predominately non-marine sands, silts, and clays of the Judith River Formation and overlain by a similar wedge formed by the Eastend and Whitemud formations in Canada and the Fox Hills Sandstone and Hell Creek Formation in the United States (Caldwell, 1968; Gill and Cobban, 1973; North and Caldwell, 1975; Caldwell et al., 1993).

Like the older Greenhorn, Niobrara, and Claggett formations, the Bearpaw Formation and its equivalents contain a rich fauna of marine organisms, dominated by foraminifers, bivalves and cephalopods, but one which is poor in other normal marine animals, such as corals, brachiopods, bryozoans, and echinoderms and sponges (Gill and Cobban, 1966; Caldwell, 1968; Kauffman and Caldwell, 1993). The Bearpaw fauna, however, differs somewhat from the related faunas of the older formations by containing some distinct non-molluscan organisms. For example, cheilostomate

So either brackish water  
or higher temperatures (or both)

Stage	Reference Section Western Interior			Section in Plains of Western Saskatchewan	Section in Plains of Eastern Saskatchewan and Manitoba		
	Fort Union			Ravenscrag	Ravenscrag	Turtle Mountain	
MAASTRICHTIAN	Hell Creek			Frenchman	Frenchman		
	Fox Hills			Battle Whitemud	Whitemud		
	MONTANA	Pierre Shale	Elk Butte	Eastend			Eastend
			Mobridge				Unnamed
			Virgin Creek				Odanah
			Verendrye				Millwood
			De Gray				
			Grow Creek				
			Gregory				
			Sharon Springs				
CAMPANIAN	Eagle		Judith River (Belly River)	Pierre Shale	Pembina		
			Claggett ( Pakowki)			Lea Park	
			Eagle (Milk River)				

Fig. 2. Correlation chart of the Upper Cretaceous formations in the study area (after North and Caldwell, 1975; McNeil and Caldwell, 1981).

After it  
so cool!

STAGES		AGES (m.y.)	AMMONITE ZONES	FORAMINIFERAL ZONES
MAASTRICHTIAN	lower	69.9	<i>B. grandis</i>	<i>HAPLOPHRAGMOIDES</i> <i>EXCAVATA</i>
		70.4	<i>B. baculus</i>	
		70.9	<i>B. eliasi</i>	
		71.4		
CAMPANIAN	upper	72.9	<i>B. jenseni</i>	<i>ANOMALINOIDES</i> SP.
		72.4	<i>B. reesidei</i>	
		72.9	<i>B. cuneatus</i>	<i>PRAEBULIMINA</i> <i>KICKAPOOENSIS</i>
		73.4	<i>B. compressus</i>	
		74.1	<i>D. cheyennense</i>	<i>GAUDRYINA</i> <i>BEARPAWENSIS</i>
		74.8	<i>E. jenneyi</i>	<i>DOROTHIA</i> SP.
		75.3	<i>D. stevensoni</i>	<i>EOEPONIDELLA</i> <i>LINKI</i>
		75.9	<i>D. nebrascense</i>	

Fig. 3. Biostratigraphy of the Bearpaw Formation and its equivalents in the Cretaceous Western Interior Basin of North America. For the ammonite genera, *B.*=*Baculites*, *D.*=*Didymoceras*, *E.*=*Exiteloceras* (after Caldwell et al., 1978, 1993; Kauffman et al., 1993).

Table 1  
Sample descriptions of molluscan fossils from the Bearpaw cyclothem<sup>a</sup>

Sample ID	Ammonite zone	Taxon	Sample locality
BL(a)	<i>B. clinolobatus</i>	<i>Discoscaphites constrictus</i>	Dewey, South Dakota
BL(b)	<i>B. clinolobatus</i>	<i>B. clinolobatus</i>	Dewey, South Dakota
BL(c)	<i>B. clinolobatus</i>	<i>Pholadomya</i> sp.	Carson, South Dakota
BL(d)	<i>B. clinolobatus</i>	<i>B. clinolobatus</i>	Carson, South Dakota
BL(e)	<i>B. clinolobatus</i>	<i>B. clinolobatus</i>	Haakon, South Dakota
BG(a)	<i>B. grandis</i>	<i>B. grandis</i>	Pennington, South Dakota
BB(a)	<i>B. baculus</i>	<i>B. baculus</i>	Frenchman River valley, Sask.
BB(b)	<i>B. baculus</i>	<i>B. baculus</i>	Haakon, South Dakota
BB(c)	<i>B. baculus</i>	<i>B. baculus</i>	Niobrara, Wyoming
BE(a)	<i>B. eliasi</i>	<i>S. plenus</i>	South Sask. River valley, Sask.
BE(b)	<i>B. eliasi</i>	<i>B. eliasi</i>	Crook, Wyoming
BE(c)	<i>B. eliasi</i>	<i>B. eliasi</i>	Crook, Wyoming
BJ(a)	<i>B. jenseni</i>	<i>B. jenseni</i>	Huerfano, Colorado
BJ(b)	<i>B. jenseni</i>	<i>B. jenseni</i>	Huerfano, Colorado
BR(a)	<i>B. reesidei</i>	<i>B. reesidei</i>	South Sask. River valley, Sask.
BR(b)	<i>B. reesidei</i>	<i>B. reesidei</i>	South Sask. River valley, Sask.
BR(c)	<i>B. reesidei</i>	<i>B. reesidei</i>	South Sask. River valley, Sask.
BR(d)	<i>B. reesidei</i>	<i>B. reesidei</i>	South Sask. River valley, Sask.
BR(e)	<i>B. reesidei</i>	<i>B. reesidei</i>	Wyoming
BCN(a)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Haakon, South Dakota
BCN(b)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Haakon, South Dakota
BCN(c)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Haakon, South Dakota
BCN(d)	<i>B. cuneatus</i>	<i>Inoceramus</i>	South Dakota
BCN(e)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Cypress Hills, Sask.
BCN(f)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Cypress Hills, Sask.
BCN(g)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Cypress Hills, Sask.
BCN(h)	<i>B. cuneatus</i>	<i>B. cuneatus</i>	Cypress Hills, Sask.
BC(a)	<i>B. compressus</i>	<i>B. compressus</i>	Pennington, South Dakota
BC(b)	<i>B. compressus</i>	<i>B. compressus</i>	Pennington, South Dakota
BC(c)	<i>B. compressus</i>	<i>Acanthoscaphites</i> sp.	Pennington, South Dakota
BC(d)	<i>B. compressus</i>	<i>Baculites</i> sp.	Pennington, South Dakota
BC(e)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(f)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(g)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(h)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(i)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(j)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
BC(k)	<i>B. compressus</i>	<i>B. compressus robinsoni</i>	South Sask. River valley, Sask.
DC(a)	<i>D. cheyennense</i>	<i>Inoceramus</i> sp.	Pennington, South Dakota
DC(b)	<i>D. cheyennense</i>	<i>D. cheyennense</i>	Pennington, South Dakota
DC(c)	<i>D. cheyennense</i>	<i>Baculites</i> sp.	Pennington, South Dakota
EJ(a)	<i>E. jenneyi</i>	<i>Inoceramus</i> sp.	Carter, Montana
EJ(b)	<i>E. jenneyi</i>	<i>scaphite</i>	Carter, Montana
EJ(c)	<i>E. jenneyi</i>	<i>scaphite</i>	Carter, Montana
EJ(d)	<i>E. jenneyi</i>	<i>E. jenneyi</i>	South Sask. River valley, Sask.
DS(a)	<i>D. stensonsi</i>	<i>Baculites</i> sp.	Albany, Wyoming
DS(b)	<i>D. stensonsi</i>	<i>Inoceramus</i> sp.	Albany, Wyoming
DN(a)	<i>D. nebrascense</i>	<i>D. nebrascense</i>	Fall River, South Dakota
DN(b)	<i>D. nebrascense</i>	<i>D. nebrascense</i>	Fall River, South Dakota
DN(c)	<i>D. nebrascense</i>	<i>Baculites</i> sp.	Fall River, South Dakota
DN(d)	<i>D. nebrascense</i>	<i>Baculites</i> sp.	Fall River, South Dakota
BS(a)	<i>B. scotti</i>	<i>Eutrophoceras</i> sp.	Fall River, South Dakota

(continued on next page)

Table 1 (continued)

Sample ID	Ammonite zone	Taxon	Sample locality
BS(b)	<i>B. scotti</i>	<i>Menuites</i>	Fall River, South Dakota
BS(c)	<i>B. scotti</i>	<i>B. scotti</i>	Fall River, South Dakota
BS(d)	<i>B. scotti</i>	<i>Inoceramus sublaevis</i>	Fall River, South Dakota
BS(e)	<i>B. scotti</i>	<i>Menuites</i> sp.	Fall River, South Dakota
BS(f)	<i>B. scotti</i>	<i>B. scotti</i>	Millwood, Manitoba
BS(g)	<i>B. scotti</i>	<i>B. scotti</i>	Millwood, Manitoba
BS(h)	<i>B. scotti</i>	<i>B. scotti</i>	Millwood, Manitoba
BS(i)	<i>B. scotti</i>	<i>B. scotti</i>	Millwood, Manitoba

<sup>a</sup> *B.*=*Baculites*, *D.*=*Didymoceras*, *E.*=*Exiteloceras*, *S.*=*Scaphite*, *Sask.*=*Saskatchewan*.

bryozoans exist at many levels of the Pierre Shale in the United States and in the Bearpaw Formation in the South Saskatchewan River valley (Gill and Cobban, 1966; Caldwell, 1968), and boring sponges, corals, calcareous worms of various kinds, and shrimp and lobster parts are found at several localities in the basin, both in the United States and Canada. Caldwell and Evans (1963) also reported a rudistid bivalve from the South Saskatchewan River valley.

Mollusca, rich in number and variety throughout the Cretaceous marine sequences of the Western Interior Basin, have been used for the principal biostratigraphic zonal schemes. The molluscan biostratigraphic system, applicable throughout the greater part of the basin, is among the most temporally refined in the world. The zonation has an average of 0.45–0.85 m.y./biozone for all Cretaceous stages (range 0.04–2.42 m.y./biozone), and 0.28–0.68 m.y./biozone for the Late Cretaceous stages (Kauffman et al., 1993), based on the time-scale of Obradovich (1993). This precision has been achieved largely by the integration of well-studied evolutionary lineages of molluscs, particularly ammonites and various groups of bivalves. Fifty-nine ammonoid zones, many based on species of *Baculites*, have been identified for the Late Cretaceous sequences. The Bearpaw marine cyclothem spans up to eleven ammonite range zones, from the *Didymoceras nebrascense* zone at the base to the *Baculites grandis* zone at the top (Fig. 3).

### 3. Samples and analytical procedures

Samples used were exclusively molluscan, taken from ammonites and bivalves housed in the collections of the United States Geological Survey

(USGS) in Denver, the Geological Survey of Canada (GSC) in Ottawa, and the Department of Geological Sciences at the University of Saskatchewan. These collections were from outcrops located in the areas described in Table 1. If possible, both ammonites and bivalves were sampled from each ammonite zone spanned by the Bearpaw Formation and its equivalents (Fig. 3). The ammonite specimens included baculitids, scaphitids, didymoceratids, and *Menuites*. Information on the sampled fossils is also provided in Table 1.

The post-embryonic shells of the ammonites consist of two prismatic layers of calcite and one layer of aragonite between them (Birkelund, 1981), whereas inoceramids have a shell with an inner aragonite layer and an outer calcite layer. Few fossil samples collected have preserved the calcite layer of their shells. As a result, only the aragonite layer was sampled for stable-isotope analyses. Prior to sampling, fossil shells were washed in acetone to remove any surface organic material and rinsed several times in distilled water. Samples for stable-isotope analysis were taken from the aragonite layers of shells using a dental drill. As aragonite may be altered into calcite during diagenesis, only well-preserved aragonite samples (>98% aragonite) were selected for the analyses. The state of preservation of these samples was determined using XRD (Davies and Hooper, 1963) and a Scanning Electron Microscope (SEM). The accuracy of this quantitative XRD method is  $\pm 1\%$ .

Stable-isotope analyses were conducted in the laboratory at the Department of Geological sciences of University of Saskatchewan. The stable-isotope compositions of aragonite were determined using 100% phosphoric acid (McCrea, 1950). The carbon- and oxygen-isotope results are reported in the usual  $\delta$ -



Table 2

Stable isotope compositions of the molluscan fossils

Sample ID <sup>a</sup>	Taxon <sup>b</sup>	$\delta^{13}\text{C}$ (‰, PDB)	$\delta^{18}\text{O}$ (‰, PDB)	Sample ID <sup>a</sup>	Taxon <sup>b</sup>	$\delta^{13}\text{C}$ (‰, PDB)	$\delta^{18}\text{O}$ (‰, PDB)
BL(a)–1	<i>Discoscaphites constrictus</i>	–1.4	0.1	BR(c)–1	<i>B. reesidei</i>	–1.8	–1.0
BL(a)–2	<i>Discoscaphites constrictus</i>	–2.1	0.1	BR(c)–2	<i>B. reesidei</i>	–1.2	–1.4
BL(a)–3	<i>Discoscaphites constrictus</i>	–1.9	0.2	BR(d)–1	<i>B. reesidei</i>	–2.3	–1.5
BL(a)–4	<i>Discoscaphites constrictus</i>	–1.0	–0.1	BR(d)–2	<i>B. reesidei</i>	–1.3	–1.3
BL(b)–1	<i>B. clinolobatus</i>	–1.0	–0.1	BR(e)–1	<i>B. reesidei</i>	–1.3	–2.1
BL(b)–2	<i>B. clinolobatus</i>	–0.6	0.2	BR(e)–2	<i>B. reesidei</i>	–1.1	–2.1
BL(c)–1	<i>Pholadomya</i> sp.	0.4	0.1	BR(e)–3	<i>B. reesidei</i>	–1.1	–1.5
BL(c)–2	<i>Pholadomya</i> sp.	0.3	0.2	BR(e)–4	<i>B. reesidei</i>	–1.0	–1.7
BL(d)–1	<i>B. clinolobatus</i>	–1.2	–0.4	BCN(a)–1	<i>B. cuneatus</i>	–2.3	–0.9
BL(d)–2	<i>B. clinolobatus</i>	–1.9	–0.4	BCN(a)–2	<i>B. cuneatus</i>	–2.4	–0.7
BL(e)	<i>B. clinolobatus</i>	–1.7	–0.5	BCN(a)–3	<i>B. cuneatus</i>	–2.7	–1.2
BG(a)–1	<i>B. grandis</i>	–1.3	–0.6	BCN(a)–4	<i>B. cuneatus</i>	–2.4	–0.7
BG(a)–2	<i>B. grandis</i>	–2.0	–0.6	BCN(a)–5	<i>B. cuneatus</i>	–2.7	–1.0
BG(a)–3	<i>B. grandis</i>	–1.6	–0.3	BCN(a)–6	<i>B. cuneatus</i>	–3.0	–1.4
BG(a)–4	<i>B. grandis</i>	–1.6	–0.2	BCN(a)–7	<i>B. cuneatus</i>	–3.5	–1.2
BB(a)–1	<i>B. baculus</i>	–1.5	–1.0	BCN(a)–8	<i>B. cuneatus</i>	–3.2	–1.7
BB(a)–2	<i>B. baculus</i>	–1.9	–0.8	BCN(a)–9	<i>B. cuneatus</i>	–2.8	–1.7
BB(a)–3	<i>B. baculus</i>	–1.1	0.1	BCN(a)–10	<i>B. cuneatus</i>	–2.2	–1.4
BB(b)	<i>B. baculus</i>	–2.0	–0.1	BCN(a)–11	<i>B. cuneatus</i>	–1.8	–1.6
BB(c)	<i>B. baculus</i>	–0.5	–2.6	BCN(a)SF–12	<i>B. cuneatus</i>	–2.3	–0.5
BE(a)–1	<i>S. plenus</i>	1.9	–1.5	BCN(a)SF–13	<i>B. cuneatus</i>	–3.4	–0.8
BE(a)–2	<i>S. plenus</i>	2.0	–1.7	BCN(a)SF–14	<i>B. cuneatus</i>	–1.9	–1.6
BE(a)–3	<i>S. plenus</i>	1.3	–1.5	BCN(a)SF–15	<i>B. cuneatus</i>	–1.2	–1.8
BE(b)–1	<i>B. eliasi</i>	–4.1	–0.2	BCN(b)–1	<i>B. cuneatus</i>	0.2	–0.9
BE(b)–2	<i>B. eliasi</i>	–3.9	–0.5	BCN(b)–2	<i>B. cuneatus</i>	0.0	–0.9
BE(c)	<i>B. eliasi</i>	–3.4	–0.6	BCN(b)–3	<i>B. cuneatus</i>	0.6	–1.0
BJ(a)–1	<i>B. jenseni</i>	–0.7	–0.9	BCN(b)–4	<i>B. cuneatus</i>	–1.2	–1.1
BJ(a)–2	<i>B. jenseni</i>	–0.6	–0.9	BCN(c)–1	<i>B. cuneatus</i>	–4.4	–0.9
BJ(a)–3	<i>B. jenseni</i>	–0.5	–0.9	BCN(c)–2	<i>B. cuneatus</i>	–4.5	–0.9
BJ(a)–4	<i>B. jenseni</i>	–0.3	–1.1	BCN(c)–3	<i>B. cuneatus</i>	–4.4	–0.9
BJ(a)–5	<i>B. jenseni</i>	–0.5	–1.4	BCN(c)–4	<i>B. cuneatus</i>	–4.5	–0.9
BJ(b)–1	<i>B. jenseni</i>	–0.7	–0.5	BCN(c)–5	<i>B. cuneatus</i>	–4.0	–1.0
BJ(b)–2	<i>B. jenseni</i>	–0.5	–0.4	BCN(c)–6	<i>B. cuneatus</i>	–4.4	–1.2
BJ(b)–3	<i>B. jenseni</i>	–0.6	–0.5	BCN(d)	<i>Inoceramus</i>	4.5	–2.8
BJ(b)–4	<i>B. jenseni</i>	–0.5	–0.3	BCN(e)	<i>B. cuneatus</i>	–1.1	–2.0
BJ(b)–5	<i>B. jenseni</i>	–0.5	–0.6	BCN(f)	<i>B. cuneatus</i>	–0.7	–0.5
BJ(b)–6	<i>B. jenseni</i>	–0.6	–0.6	BCN(g)	<i>B. cuneatus</i>	–1.7	–0.4
BJ(b)–7	<i>B. jenseni</i>	–0.4	–0.7	BCN(h)	<i>B. cuneatus</i>	–1.5	–2.2
BJ(b)–8	<i>B. jenseni</i>	–0.3	–0.5	BC(a)	<i>B. compressus</i>	–3.9	–2.1
BJ(b)–9	<i>B. jenseni</i>	–0.5	–0.4	BC(b)–1	<i>B. compressus</i>	–3.0	–1.3
BJ(b)–10	<i>B. jenseni</i>	–1.0	–0.8	BC(b)–2	<i>B. compressus</i>	–0.5	–1.7
BR(a)–1	<i>B. reesidei</i>	–1.3	–1.2	BC(b)–3	<i>B. compressus</i>	–2.5	–1.8
BR(a)–2	<i>B. reesidei</i>	–1.9	–1.2	BC(b)–4	<i>B. compressus</i>	–2.5	–2.0
BR(a)–3	<i>B. reesidei</i>	–3.1	–1.3	BC(c)	<i>Acanthoscaphites</i> sp.	–2.9	–3.1
BR(b)	<i>B. reesidei</i>	–2.1	–1.2	BC(d)	<i>Baculites</i> sp.	–1.6	–1.7
BC(e)	<i>B. compressus robinsoni</i>	–1.3	–1.4	DN(a)–6	<i>D. nebrascense</i>	1.8	–2.0
BC(f)–1	<i>B. compressus robinsoni</i>	–2.0	–2.0	DN(a)–7	<i>D. nebrascense</i>	1.6	–2.3
BC(f)–2	<i>B. compressus robinsoni</i>	–2.6	–1.8	DN(a)–8	<i>D. nebrascense</i>	1.8	–2.0
BC(f)–3	<i>B. compressus robinsoni</i>	–2.9	–1.8	DN(a)–9	<i>D. nebrascense</i>	1.2	–2.1
BC(f)–4	<i>B. compressus robinsoni</i>	–2.4	–1.7	DN(a)–10	<i>D. nebrascense</i>	1.6	–1.9
BC(f)–5	<i>B. compressus robinsoni</i>	–2.8	–1.8	DN(a)–11	<i>D. nebrascense</i>	1.4	–2.2

(continued on next page)



Table 2 (continued)

Sample ID <sup>a</sup>	Taxon <sup>b</sup>	$\delta^{13}\text{C}$ (‰, PDB)	$\delta^{18}\text{O}$ (‰, PDB)	Sample ID <sup>a</sup>	Taxon <sup>b</sup>	$\delta^{13}\text{C}$ (‰, PDB)	$\delta^{18}\text{O}$ (‰, PDB)
BC(g)	<i>B. compressus robinsoni</i>	−2.1	−2.2	DN(a)–12	<i>D. nebrascense</i>	1.4	−2.0
BC(h)–1	<i>B. compressus robinsoni</i>	−2.2	−1.0	DN(a)–13	<i>D. nebrascense</i>	1.6	−1.9
BC(h)–2	<i>B. compressus robinsoni</i>	−2.5	−0.7	DN(a)SF–1	<i>D. nebrascense</i>	1.2	−2.1
BC(h)–3	<i>B. compressus robinsoni</i>	−2.8	−1.7	DN(a)SF–2	<i>D. nebrascense</i>	1.4	−2.2
BC(h)–4	<i>B. compressus robinsoni</i>	−1.8	−1.3	DN(a)SF–3	<i>D. nebrascense</i>	1.2	−2.4
BC(i)–1	<i>B. compressus robinsoni</i>	−0.5	−1.4	DN(a)SF–4	<i>D. nebrascense</i>	1.2	−2.1
BC(i)–2	<i>B. compressus robinsoni</i>	−0.7	−1.4	DN(a)SF–5	<i>D. nebrascense</i>	1.8	−2.1
BC(i)–3	<i>B. compressus robinsoni</i>	−1.3	−1.6	DN(a)SF–6	<i>D. nebrascense</i>	1.8	−2.0
BC(i)–4	<i>B. compressus robinsoni</i>	−0.9	−1.5	DN(a)SF–7	<i>D. nebrascense</i>	1.9	−2.0
BC(j)	<i>B. compressus robinsoni</i>	−3.5	−1.4	DN(a)SF–8	<i>D. nebrascense</i>	1.5	−2.3
BC(k)	<i>B. compressus robinsoni</i>	−2.0	−1.2	DN(b)	<i>D. nebrascense</i>	1.6	−1.9
DC(a)–1	<i>Inoceramus sp.</i>	4.7	−3.6	DN(c)–1	<i>Baculites sp.</i>	−2.6	−0.3
DC(a)–2	<i>Inoceramus sp.</i>	4.9	−3.7	DN(c)–2	<i>Baculites sp.</i>	−2.2	−0.4
DC(a)–3	<i>Inoceramus sp.</i>	4.8	−3.7	DN(d)–1	<i>Baculites sp.</i>	−2.0	−0.3
DC(a)–4	<i>Inoceramus sp.</i>	5.0	−3.6	DN(d)–2	<i>Baculites sp.</i>	−2.1	−0.7
DC(a)–5	<i>Inoceramus sp.</i>	5.0	−3.5	DN(d)–3	<i>Baculites sp.</i>	−1.7	−0.1
DC(a)–6	<i>Inoceramus sp.</i>	4.8	−3.5	BS(a)	<i>Eutrophoceras sp.</i>	−0.4	−0.2
DC(a)–7	<i>Inoceramus sp.</i>	5.0	−3.4	BS(b)	<i>Menuites</i>	−0.3	0.4
DC(a)–8	<i>Inoceramus sp.</i>	5.0	−3.5	BS(c)–1	<i>B. scotti</i>	−1.3	−0.5
DC(b)–1	<i>D. cheyennense</i>	1.9	−1.6	BS(c)–2	<i>B. scotti</i>	−1.3	0.0
DC(b)–2	<i>D. cheyennense</i>	1.4	−1.8	BS(c)–3	<i>B. scotti</i>	−1.5	−0.3
DC(b)–3	<i>D. cheyennense</i>	2.2	−1.6	BS(c)SF–1	<i>B. scotti</i>	−1.4	−0.2
DC(b)–4	<i>D. cheyennense</i>	1.6	−1.9	BS(c)SF–2	<i>B. scotti</i>	−1.6	−0.5
DC(c)–1	<i>Baculites sp.</i>	−0.6	−0.1	BS(c)SF–3	<i>B. scotti</i>	−1.7	−0.3
DC(c)–2	<i>Baculites sp.</i>	−1.1	−0.3	BS(c)SF–4	<i>B. scotti</i>	−1.5	−0.7
EJ(a)–1	<i>Inoceramus sp.</i>	1.9	−2.6	BS(c)SF–5	<i>B. scotti</i>	−0.9	−0.6
EJ(a)–2	<i>Inoceramus sp.</i>	3.8	−4.7	BS(c)SF–6	<i>B. scotti</i>	−0.8	−0.8
EJ(b)–1	<i>scaphite</i>	−0.5	−1.0	BS(d)–1	<i>Inoceramus sublaevis</i>	3.6	−3.0
EJ(b)–2	<i>scaphite</i>	−1.2	−1.3	BS(d)–2	<i>Inoceramus sublaevis</i>	3.3	−3.2
EJ(c)	<i>E. jenneyi</i>	0.9	−3.3	BS(d)–3	<i>Inoceramus sublaevis</i>	3.3	−3.2
EJ(d)	<i>Inoceramus sp.</i>	0.6	−3.8	BS(e)–1	<i>Menuites sp.</i>	−3.2	−1.8
DS(a)–1	<i>Baculites sp.</i>	−0.6	−2.3	BS(e)–2	<i>Menuites sp.</i>	−2.6	−1.5
DS(a)–2	<i>Baculites sp.</i>	−0.1	−2.9	BS(e)–3	<i>Menuites sp.</i>	−2.7	−1.2
DS(a)–3	<i>Baculites sp.</i>	−0.8	−2.6	BS(e)–4	<i>Menuites sp.</i>	−3.0	−1.4
DS(b)	<i>Inoceramus sp.</i>	3.9	−3.8	BS(e)–5	<i>Menuites sp.</i>	−3.4	−0.6
DN(a)–1	<i>D. nebrascense</i>	1.5	−1.9	BS(e)–6	<i>Menuites sp.</i>	−4.6	−0.8
DN(a)–2	<i>D. nebrascense</i>	1.0	−2.1	BS(f)	<i>B. scotti</i>	−0.9	0.0
DN(a)–3	<i>D. nebrascense</i>	0.9	−2.3	BS(g)	<i>B. scotti</i>	−0.7	0.3
DN(a)–4	<i>D. nebrascense</i>	1.1	−2.1	BS(h)	<i>B. scotti</i>	−1.4	0.4
DN(a)–5	<i>D. nebrascense</i>	1.3	−2.1	BS(i)	<i>B. scotti</i>	−0.7	0.5

<sup>a</sup> A series number related to one individual indicates samples taken along the length of the specimen.

<sup>b</sup> B.=Baculites, D.=Didymoceras, E.=Exilloceras, S.=Scaphite.

notation relative to the reference standard of the Peedee belemnite (V-PDB; Craig, 1957),

$$\delta = [(R_{\text{sample}}/R_{\text{standard}}) - 1] \times 1000\text{‰} \quad (1)$$

where  $R$  represents  $^{13}\text{C}/^{12}\text{C}$  or  $^{18}\text{O}/^{16}\text{O}$ . Replicate isotopic analyses indicated a reproducibility of  $\pm 0.05\text{‰}$  for both.

#### 4. Results

As reported in previous studies (Tourtelot and Rye, 1969; Forester et al., 1977; Wright, 1987; Kyser et al., 1993), marked heterogeneity is evident in the stable-isotope compositions of coexisting molluscan fossils throughout the central portion of the WIB (Table 2; Fig. 4). Generally, inoceramid

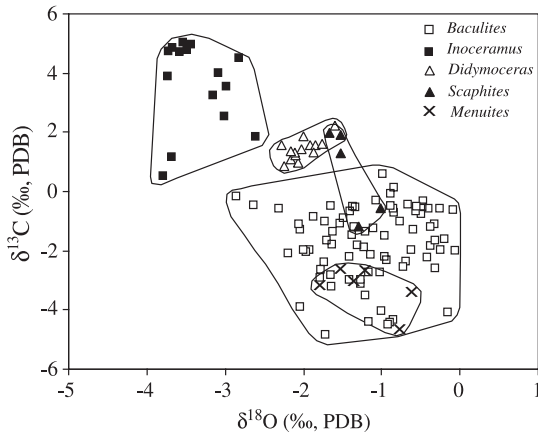


Fig. 4.  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values of molluscan shells from the sedimentary rocks of the Bearpaw cyclothem in the central portion of the Western Interior Basin of North America.

shells have the highest  $\delta^{13}\text{C}$  values (0.6‰ to 5.0‰) and the lowest  $\delta^{18}\text{O}$  values (−4.7‰ to −2.8‰), whereas baculitid shells have the highest  $\delta^{18}\text{O}$  values (−2.8‰ to 0.0‰) and the lowest  $\delta^{13}\text{C}$  values (−4.8‰ to −0.3‰). The isotopic compositions of other molluscan shells are generally between the

ranges of baculitids and inoceramids, with *Didymoceras* and *Exiteloceras* having lower  $\delta^{18}\text{O}$  values (−2.3‰ to −1.9‰) and higher  $\delta^{13}\text{C}$  values (1.0‰ to 1.8‰) than baculitids, and *Menuites* having the same range of stable-isotope compositions as baculitids. Some scaphitids have stable-isotope compositions similar to those of *Didymoceras*, whereas others fall within the range of the baculitids.

Despite the exact stratigraphic positions of some of the fossils within the biostratigraphic zones being unknown,  $\delta^{18}\text{O}$  values of the baculitids from places in the central portion of the basin vary with biostratigraphic sequence (Fig. 5). For example, the baculitids from the *Baculites compressus*, *Baculites cuneatus*, and *Baculites reesidei* zones, denoting the period of most widespread transgression, have the lowest average  $\delta^{18}\text{O}$  values (−2.3‰ to −0.7‰), whereas those from the underlying and overlying zones have higher  $\delta^{18}\text{O}$  values (−1.4‰ to 0.0‰). The  $\delta^{13}\text{C}$  values also change with the zonal sequence, but not necessarily in relation to the  $\delta^{18}\text{O}$  values. Generally, baculitids with higher  $\delta^{18}\text{O}$  values have lower  $\delta^{13}\text{C}$  values. Differences also exist in the stable-isotope compositions of different specimens of baculitids

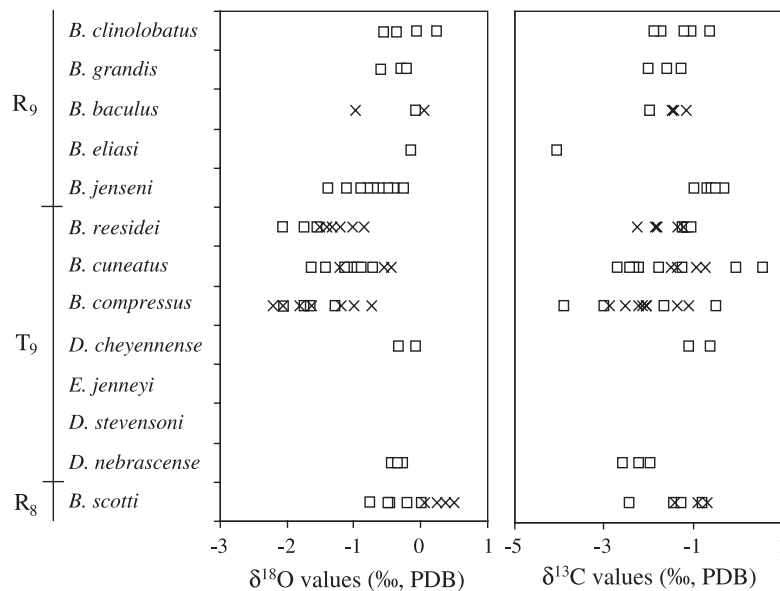


Fig. 5. Variation in  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values of baculitid shells with ammonite zones of the Bearpaw cyclothem. The samples were collected from places in the central portion of the Western Interior Basin. Open squares and "x" represent the samples from the United States and Canada, respectively. For ammonite genera, *B.*=*Baculites*, *D.*=*Didymoceras*, *E.*=*Exiteloceras*.  $T_9$  and  $R_9$  represent the transgressive and regressive phases of the Bearpaw cyclothem, respectively;  $R_8$  represents the regressive phase of the Claggett cyclothem.

from the same zone, and such differences can be ca. 1.0‰ for  $\delta^{18}\text{O}$  values and ca. 2.0‰ for  $\delta^{13}\text{C}$  values.

## 5. Discussion and implications

### 5.1. Vital effects

Stable isotopes of carbonate shells of mollusks have been widely used in paleoenvironmental reconstruction (Epstein et al., 1953; Milliman et al., 1974; Savin, 1982). However, carbonate shells of molluscs

may not always be precipitated in equilibrium with their ambient waters due to “kinetic isotope effects” or “metabolic isotope effects”, both referred to as vital effects (McConnaughey, 1989a; Dettman and Lohmann, 1993; Fastovsky et al., 1993; Mitchell et al., 1994; Carpenter and Lohmann, 1995). In such cases stable-isotope analyses may be unreliable for directly interpreting prevailing paleoenvironmental conditions.

Kinetic fractionation takes place during hydration and hydroxylation of  $\text{CO}_2$  in the process of shell calcification, normally discriminating against heavy

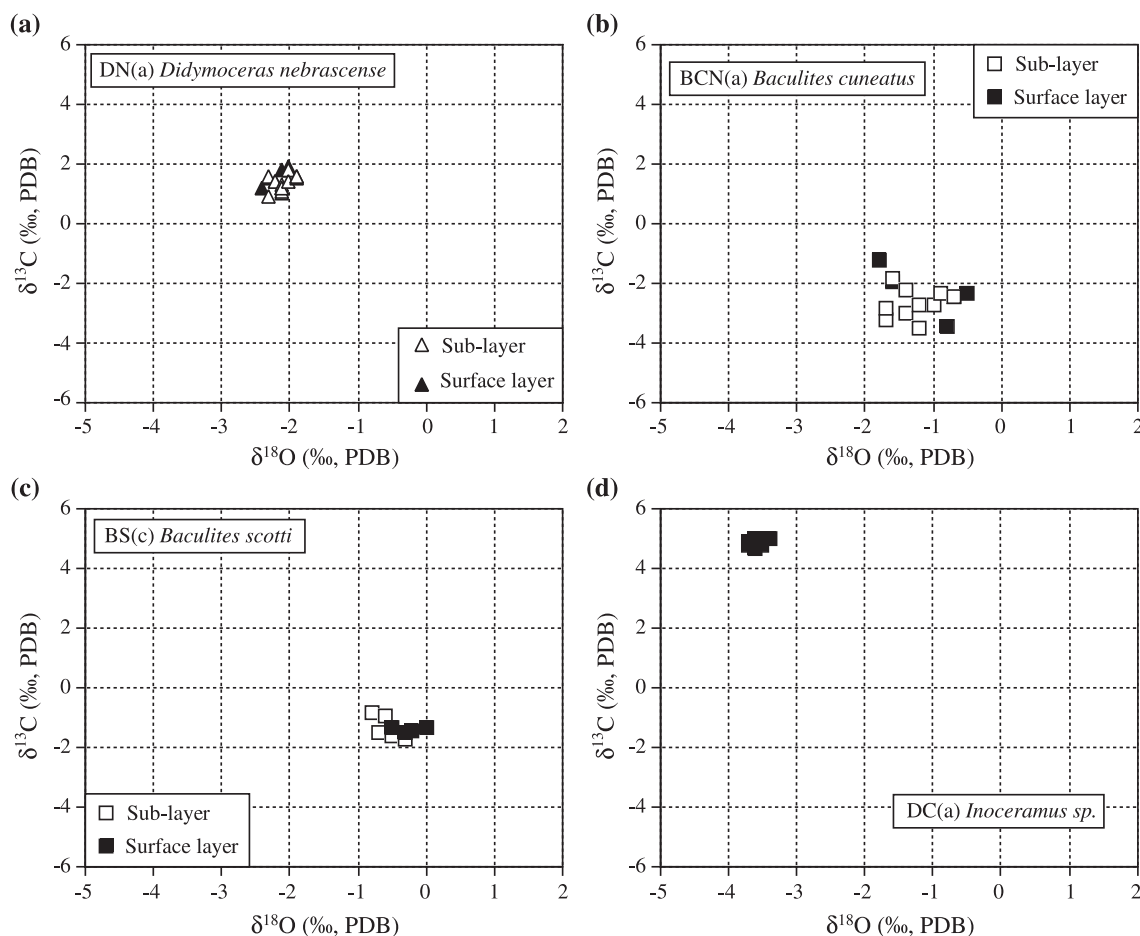


Fig. 6. Relations between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values of single molluscan shells from the Bearpaw cyclothem: (a) DN(a), *Didymoceras nebrascense* from the *D. nebrascense* zone; (b) BCN(a), *Baculites cuneatus* from the *B. cuneatus* zone; (c) BS(c), *Baculites scotti* from the *B. scotti* zone; (d) DC(a), *Inoceramus* sp. from the *Didymoceras cheyennense* zone; (e) BS(e), *Menuites* from the *B. scotti* zone; (f) BJ(a) and BJ(b), two *Baculites jenseni* from the *B. jenseni* zone; (g) *Inoceramus sublaevis* (DC(a)), *D. cheyennense* (DC(b)), and *Baculites* sp. (DC(c)) from the *D. cheyennense* zone; (h) *D. nebrascense* (DN(a)) and *Baculites* sp. (DN(c)) from the *D. nebrascense* zone. ‘Surface layer’ refers to samples from the uppermost part of the aragonite layer and ‘sub-layer’ refers to samples from the lower part of the layer.

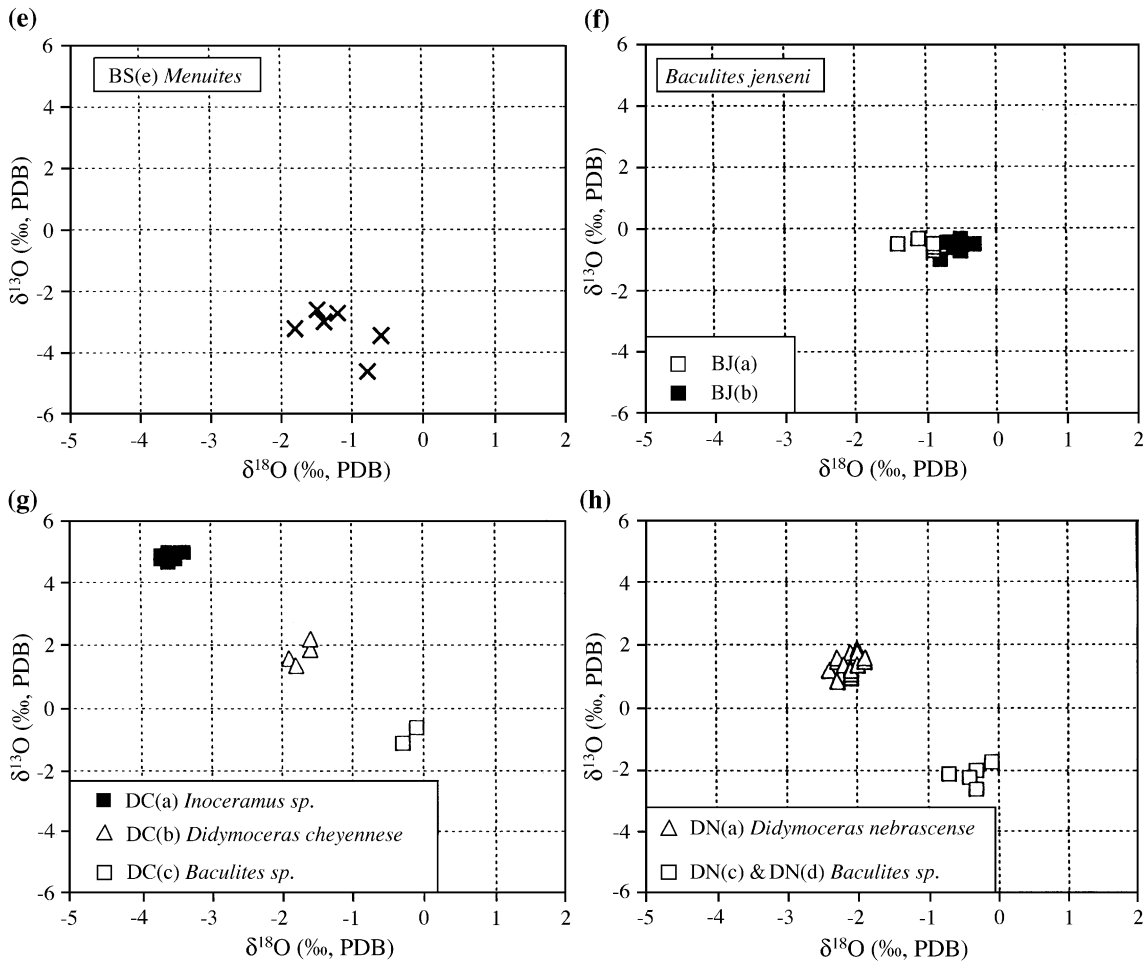


Fig. 6 (continued).

oxygen and carbon isotopes (McConnaughey, 1989a, b; Mitchell et al., 1994; McConnaughey et al., 1997). This process will lead to a strong linear correlation between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values (Keith and Weber, 1965; Land et al., 1975; Erez, 1977, 1978; Erez and Honjo, 1981; McConnaughey, 1989a,c; Dettman and Lohmann, 1993; Fastovsky et al., 1993; Mitchell et al., 1994; Carpenter and Lohmann, 1995). Metabolic effects are usually considered to influence carbon-isotope compositions through changes in the  $\delta^{13}\text{C}$  values of the dissolved inorganic carbon (DIC) in the internal reservoir from which the skeleton precipitates (Cummings and

McCarthy, 1982; Swart, 1983; Jones et al., 1986; McConnaughey, 1989a; McConnaughey et al., 1997; Zeebe et al., 1999; Heikoop et al., 2000; Smith et al., 2000). These changes primarily result from selective removal and addition of  $^{12}\text{C}$  by photosynthesis or chemosynthesis and by respiration. Some researchers (Erez, 1978; Carpenter and Lohmann, 1995) argue that higher metabolic activities of organisms may result in larger kinetic fractionation of both carbon and oxygen isotopes. However, a recent study by McConnaughey et al. (1997) showed that metabolic effects are minimal on the oxygen isotopes in shells.

vital effects  $\rightarrow$  linear relation  
between  $\delta^{13}\text{C}$  &  $\delta^{18}\text{O}$

Ammonites are generally considered to precipitate their shells in isotopic equilibrium because the shell of post-embryonic *Nautilus*, a possible living analogue of ammonites, is precipitated in isotopic equilibrium with seawater (Eichler and Ristedt, 1966; Taylor and Ward, 1983). Multiple analyses of different parts of single shells of ammonites in this study do not show any linear relationship between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values, the pattern normally produced by vital effect. In contrast, the values exhibit clustered or randomly scattered distributions (Fig. 6a to d). Fig. 6e to h shows the relationship between the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values in shells of different molluscan species from the same locality and, therefore, from the same ammonite zone. Intra-species variation in stable-isotope compositions of baculites from different ammonite zones may contribute to the variation in their  $\delta^{18}\text{O}$  values with biostratigraphic sequence. However, the investigation of modern brachiopods by Carpenter and Lohmann (1995) implies such intra-species differences in molluscs are probably imperceptible.

The higher  $\delta^{13}\text{C}$  values but lower  $\delta^{18}\text{O}$  values of inoceramids from WIS compared to the other co-existing molluscs cannot be attributed to a kinetic effect. On the other hand, the benthic life style of inoceramids eliminates the possibility of photosynthetic symbiotic effects to explain the values. MacLeod and Hoppe (1992, 1993) postulated that the inoceramids collected from DSDP cores, like chemosymbiont-harboring bivalves from a cold seep off coast of Oregon, probably hosted chemosynthetic symbionts that could produce the  $^{13}\text{C}$  enrichment in their shells. The basis for their hypothesis is the assumption that foraminifera provide reference values for surface and bottom water. As discussed by Grossman (1993), however, this assumption may not be justified so that MacLeod and Hoppe (1992) did not convincingly demonstrate that inoceramids hosted any chemosynthetic symbionts. Consequently, ammonites and inoceramids from the Bearpaw Sea could have precipitated their shells in stable-isotope equilibrium with the seawaters in which they lived so that their isotopic values should be reliable indicators of the paleoenvironment. This is consistent with many of the ammonites and inoceramids being autochthonous and endemic to the seaway. Variable paleoenvironmental conditions rather than vital

effects, therefore, likely account for the heterogeneity in stable-isotope compositions among the co-existing molluscs.

### 5.2. Paleoenvironmental implications

As stable-isotope values of ammonites and inoceramids are interpreted to record the original environments of the seaway, the heterogeneity in their stable-isotope compositions implies that the waters in which they lived had distinct stable-isotope compositions or temperatures. Baculitids, which were nektonic or planktonic animals, inhabited mostly the upper part of the water column in the seaway whereas inoceramids were benthonic. *Didymoceras*, with isotopic compositions similar to those of benthonic inoceramids, also probably lived very close to the bottom. Compared to the other Bearpaw ammonites, scaphitids have more variable stable-isotopic compositions, with some like those of *Didymoceras* but others in the range of the baculitids. Based on stable-isotopic compositions, therefore, scaphitids may have lived more or less throughout the main part of the water column.

A simple explanation for the isotopic heterogeneity is that the surface and bottom waters in the seaway had different temperatures. Tourtelot and Rye (1969) first reported differences in the stable-isotopic compositions of coeval molluscs from the Upper Cretaceous Pierre Shale of Montana, Wyoming, and the Dakotas. The most depleted value of their inoceramids is  $-5.56\text{‰}$ . Kyser et al. (1993) and Fisher and Arthur (2002) reported even lower  $\delta^{18}\text{O}$  values ( $-6\text{‰}$  to  $-10\text{‰}$ ) for inoceramids from the seaway. Assuming a  $\delta^{18}\text{O}$  value of  $0.0\text{‰}$  for the water in the seaway, the  $\delta^{18}\text{O}$  values of the benthonic inoceramids would imply temperatures in excess of  $40\text{ °C}$ , whereas the  $\delta^{18}\text{O}$  values of the ammonites would indicate more reasonable temperatures of ca.  $20\text{ °C}$ . These temperature differences between the surface and bottom waters are the reverse of the expected normal distribution of temperatures in the seaway. An inverse temperature stratification may have existed, but it would be unlikely to have persisted through much of the Late Cretaceous Epoch. In addition, a temperature of over  $40\text{ °C}$  is unreasonably high for molluscs to survive. Based on a convection model driven by turbulent buoyant plumes and observation of present

deep-water formation, Brass et al. (1982) suggest that bottom water with high salinity could be stable at temperatures higher than the surface waters, and the Cretaceous was much more favorable for the production of warm saline bottom waters. This may partially, but not fully explain, the lower  $\delta^{18}\text{O}$  values of the inoceramids from the WIS.

Large differences among the stable-isotope compositions of coeval benthonic and nektonic or planktonic animals from the Bearpaw cyclothem more likely points to the Bearpaw Sea being isotopically stratified, in much the same way as were the older Claggett and Greenhorn seas (Whittaker et al., 1987; Kyser et al., 1993) because the magnitude of the differences cannot be attributed to temperature alone. Assuming an average water temperature of 15 °C, the upper part of the water column would have had  $\delta^{18}\text{O}$  values of ca.  $-3.0\text{‰}$  to  $-1.0\text{‰}$  and  $\delta^{13}\text{C}$  values of ca.  $-5.0\text{‰}$  to  $-0.5\text{‰}$ , as indicated by nektonic or planktonic baculitids. In contrast, the stable-isotope compositions of the benthonic inoceramids suggest that the waters near the bottom would have had much lower  $\delta^{18}\text{O}$  values of ca.  $-6.0\text{‰}$  to  $-4.0\text{‰}$  and higher  $\delta^{13}\text{C}$  values of ca.  $0.0\text{‰}$  to  $4.0\text{‰}$  for waters of the same temperature. Although there may have been isotopic stratification of the seaway through much of the Late Cretaceous Epoch, the cause of this stratification and the source of lower  $\delta^{18}\text{O}$  values and higher  $\delta^{13}\text{C}$  values of bottom water are unclear.

Influx of relatively  $^{18}\text{O}$ -depleted fresh water to cause brackish-water conditions has been suggested to account for the lower  $\delta^{18}\text{O}$  values of molluscan fossils in the WIS (Scholle and Kauffman, 1975; Scholle, 1977; Pratt, 1983). If the low  $\delta^{18}\text{O}$  values of the seaway during the Late Cretaceous Epoch resulted from the presence of brackish water, the more  $^{18}\text{O}$ -rich and, presumably, denser and more saline waters in which the baculitids lived would have overlain the brackish and lighter bottom waters. Such a density stratification would have been unstable and therefore, unlikely to persist.

Wright (1987) postulated that fluvial run-off from the western highlands spread eastward, forming a low-salinity cap on the surface of the seaway. The surface water was decoupled from the underlying water and subsequently modified by evaporation on the eastern margin of the seaway until its density increased sufficiently to force sinking and a counter flow of

modified low-salinity or brackish water westward along the bottom of the seaway. Although this low-salinity cap may have existed from time to time for short periods, it is unlikely to have done so through all of Late Cretaceous time. In the short term, storms and waves presumably would have disrupted it. In the longer term, Wright's model places severe restrictions on Boreal waters entering the seaway, which contrasts with paleontological evidence strongly supporting the introduction of these waters at times up to and including the Late Campanian (Caldwell, 1968; Gill and Cobban, 1973; Kauffman, 1975; Eicher and Dinner, 1985).

Lawrence et al. (1979) have proposed that bottom water can become depleted in  $^{18}\text{O}$  by up to several per mil because of geochemical exchange across the sediment–water. During the breakdown of volcanic clasts to smectitic clays,  $^{18}\text{O}$  depleted water will be released into pore space. O'Neil and Kharaka (1976) and James and Baker (1976) also demonstrated that oxygen-isotope exchange might accompany cation exchange between detrital clays and seawater. Whittaker et al. (1987) and Kyser et al. (1993) have invoked this process to explain the difference in stable isotopes between inoceramids and ammonites in the WIS. On the other hand, the high  $^{13}\text{C}$  source of the pore water is likely microbial fermentation of organic matter in the sediments (Longstaffe, 1993; Colquhoun and Longstaffe, 2000). The modified pore water may have diffused and been expelled from the sediments, possibly enhanced by bioturbation, which would have affected the sessile benthonic organisms (e.g., inoceramids) more than the mobile benthonic animals (e.g., *Didymoceras*). Isotopic exchange reactions between the bottom waters and sediments might have some contribution to the heterogeneity in their stable-isotope compositions of molluscs observed.

Hay et al. (1993) hypothesized that Tethyan waters with high temperature and salinity and modified Boreal waters mixed in the WIS to form a new deepwater mass. This new water mass could have a  $\delta^{18}\text{O}$  value of ca.  $-4.0\text{‰}$  and a temperature of 22.5 °C. Fisher and Arthur (2002) propose a similar mechanism to explain the very depleted  $\delta^{18}\text{O}$  values of foraminifera and inoceramids observed. The newly formed  $^{18}\text{O}$ -depleted deepwater mass may account for, at least partially, the stratification.



Variations in  $\delta^{18}\text{O}$  values of baculitid shells in relation to biostratigraphic sequence (Fig. 5) could be due to changes in the temperature,  $\delta^{18}\text{O}$  values or both of the seawater in which the organisms lived. Volume of continental ice and influx of fresh water are the two main factors that could have affected the  $\delta^{18}\text{O}$  values of the water in the seaway. Other factors are likely to have been negligible or only locally important. Change in volume of continental ice is potentially the most important factor because ice in polar areas has much more negative  $\delta^{18}\text{O}$  values than modern seawater (Shackleton and Kennett, 1975). Paleoenvironmental data suggest, however, that there was negligible continental ice during the Late Cretaceous Epoch (Savin, 1982; Frakes, 1979; Upchurch and Wolfe, 1993). If a polar continental ice cap existed, it was probably restricted to the core of the Antarctic landmass. Hence, either influx of fresh water into the seaway or change in temperature with time is more likely to account for the variations in stable-isotope values of the baculitid shells.

Influx of fresh water, however, is irreconcilable with the patterns of change in the  $\delta^{18}\text{O}$  values of the baculitid shells with biostratigraphic sequence. Baculitids with the lowest average  $\delta^{18}\text{O}$  values are from the *Baculites reesidei* and *Baculites compressus* zones, with baculitids from the *Baculites jenseni* and *Baculites cuneatus* zones also yielding lower  $\delta^{18}\text{O}$  values than those of other zones. These four zones span the time when transgression of the Bearpaw Sea was at its peak or near peak (Caldwell, 1968; Gill and Cobban, 1973), and therefore, the time when marine environmental conditions would have been most typically open sea. If influx of fresh water into the seaway were significant during the Bearpaw cycle, one would have expected it to be strongest during ammonite zones coincident with significant regression. The baculitid shells from these zones then should have more negative  $\delta^{18}\text{O}$  values because the seaway was narrower during these times and tectonic deformation more active along the western margin. However, the oxygen-isotope compositions of the zonal baculitids from regressive and early transgressive phases are among the highest and are similar to those of fossils from the contemporary open oceanic environments (Shackleton and Kennett, 1975; Kyser et al., 1993).

Paleontological evidence, in general, suggests that, during the whole Bearpaw cycle, most of the seaway was negligibly brackish, if brackish at all. In contrast to the older cycles, the cheilostomate bryozoans—fairly stenohaline organisms (Ryland, 1970)—are fairly common throughout the upper Pierre Shale in the United States and are present in the Bearpaw Formation of the Southern Saskatchewan River valley (Gill and Cobban, 1966; Caldwell, 1968). Corals have been found in sedimentary rocks from several parts of the WIB (Gill and Cobban, 1966; Caldwell, 1968); and a rudist bivalve, *Ichthyosarcolithes coralloidea* (Hall and Meek), was recovered from the Bearpaw Formation in southwestern Saskatchewan (Caldwell and Evans, 1963). All this evidence supports the contention that, during the Bearpaw cycle, the fresh-water influx into the seaway was relatively low and its effect on the stable-isotopic compositions of the baculitids was minimal. Variations in the stable-isotopic compositions of the baculitids, therefore, are primarily a response to fluctuations in temperatures of the seaway. The lowest  $\delta^{18}\text{O}$  values of baculitids occur in the middle of the biostratigraphic sequence, which would correspond to the highest temperature of the seaway during peak or near-peak transgression. Kauffman (1984) argues that temperature may rise by  $\pm 5^\circ\text{C}$  at peak transgression. Differences between the  $\delta^{18}\text{O}$  values of different baculitids from the same ammonite zone reflect relatively small temperature fluctuations of the Bearpaw Sea during each ammonite zone because these baculitids no doubt lived at different times during the temporal span of the zone. The similarity in stable-isotope compositions of molluscan fossils from the Canadian and northern American portions of the basin supports the inference that temperature differences between them were minimal, such that the latitudinal temperature gradient in this portion of the seaway was much lower during the Bearpaw cycle than it is in the oceans today (Kauffman, 1977; Savin, 1982).

Prevalence of somewhat brackish-water conditions in both surface and bottom waters has been considered one of the probable reasons for the poorly diversified, mollusc-dominated fauna of the seaway during the Late Cretaceous Epoch (Kauffman and Caldwell, 1993). As discussed, however, the stable-isotope compositions of the molluscs imply



that any appreciable widespread, brackish water-conditions must have been very limited. Other environmental factors, such as oxygen-deficiency, should be sought, therefore, to explain the rarity of the numerous groups of stenohaline organisms (sponges, bryozoans, brachiopods, corals, and echinoderms) in the Bearpaw cyclothem relative to the contemporaneous faunas of the open seas covering non-epicontinental regions, such as the Gulf Coast for example.

### 5.3. Paleotemperature of the Bearpaw Sea

Paleotemperatures were calculated using the empirical temperature scale developed by Grossman and Ku (1986), based on naturally occurring aragonitic foraminifera and molluscs in normal seawater:

$$T = 20.6 - 4.34(\delta_A - \delta_W) \quad (2)$$

where  $T$  refers to temperature in  $^{\circ}\text{C}$ ,  $\delta_A$  is the  $\delta^{18}\text{O}$  value of aragonite relative to PDB, and  $\delta_W$  is the  $\delta^{18}\text{O}$  value of the ambient water relative to PDB.

As discussed above, influx of fresh water to the seaway probably was relatively low during the Bearpaw cycle, so that the mean  $\delta^{18}\text{O}$  value of open

oceans can be used as an approximation for that of the Bearpaw Sea. In addition, no significant permanent ice cover existed during the Late Cretaceous Epoch (see the previous discussion), and thus the  $\delta^{18}\text{O}$  value of the Cretaceous oceans must have been more negative than that of modern oceans. The overall average  $\delta^{18}\text{O}$  value of ocean water today is about  $-0.3\text{‰}$  (Craig, 1965). Using a value of  $-50\text{‰}$  for the  $\delta^{18}\text{O}$  value of polar ice today (Shackleton and Kennett, 1975), the  $\delta^{18}\text{O}$  value of the ocean prior to the formation of the presently existing ice sheets would be  $-1.3\text{‰}$ . This value is used as the mean  $\delta^{18}\text{O}$  value of the Bearpaw Sea to calculate the temperatures of the seaway.

Because it is difficult to quantify the isotopic composition of the bottom waters, which might have been modified by multiple processes as discussed above, the temperatures of bottom-waters are not calculated. Variations in near-surface temperatures of the Bearpaw Sea, calculated from the oxygen isotopic compositions of the baculitids, are indicated in Fig. 7. Assuming no variation in the  $\delta^{18}\text{O}$  value of the Bearpaw Sea, the surface temperature generally rose during transgression and fell after peak transgression. In the early transgressive phase (*Didymoceras nebras-*

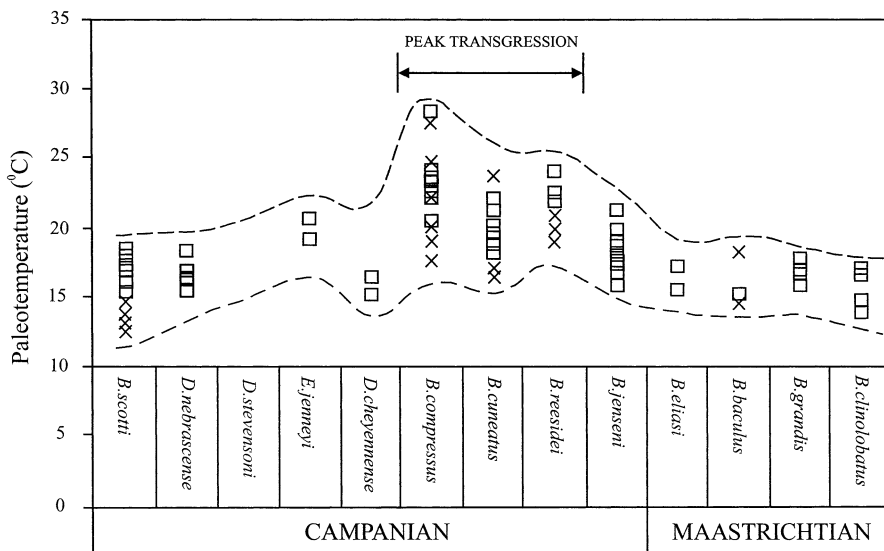


Fig. 7. Paleotemperature of the seaway during the Bearpaw cycle calculated from the  $\delta^{18}\text{O}$  values of baculitid shells from the central portion of the Western Interior Basin, assuming the  $\delta^{18}\text{O}$  value of the water in the seaway is  $-1.3\text{‰}$ . *B.*=*Baculites*, *D.*=*Didymoceras*, *E.*=*Exiteloceras*. Open square and 'x' refer to the samples from the United States and Canada, respectively. The Bearpaw Sea reached peak transgression during *Baculites compressus*, *Baculites cuneatus*, and *Baculites reesidei* times (Caldwell, 1968; Gill and Cobban, 1973; Cobban et al., 1993).

cense and *Didymoceras stevensoni* times), the temperature was ca. 13–17 °C and increased toward the highest values typical of peak transgression. It remained at nearly 20 °C through that period. In the following regressive phase, the temperature decreased to ca. 14 °C at the end of the cycle (*Baculites clinolobatus* time).

To some extent, the temperature trend of the seaway during the Bearpaw cycle may be related to CO<sub>2</sub> degassing from the increased tectonic activity and metamorphism west of the basin. These events likely increased the CO<sub>2</sub> content of the atmosphere, consequently elevating the temperatures of the area. Towards the middle of the transgression, the proto-Laramide orogen intensified (Kauffman and Caldwell, 1993) and subduction generated extensive metamorphism in the proto-Cordilleran region. Pelagic and slope carbonates, subducted beneath the continental plate, and carbonates in the orogen underwent regional metamorphism. A fraction of the carbonates would be decomposed to produce a large volume of CO<sub>2</sub>, released into the atmosphere through volcanism, fractures, and faults. It is postulated that this kind of degassing occurred in the early stages of the collision and orogen, when the temperature was high enough to decompose carbonate minerals (Kerrick and Caldeira, 1993). Kerrick and Caldeira (1993) related the warmest climate during the Early Eocene Epoch to extensive CO<sub>2</sub> degassing from the Himalayan orogen through metamorphism. They demonstrated that decarbonation of large volumes of impure carbonates by rapid prograde metamorphism, and transportation of the CO<sub>2</sub> produced to the surface by fluids through volcanism, fractures, and faults, could have released up to 10<sup>19</sup> mol of CO<sub>2</sub> into the atmosphere, causing a significant global warming. After the peak transgression of the Bearpaw cycle, the intensity of tectonic activity and the temperature of metamorphism probably decreased gradually, reducing the emission of CO<sub>2</sub> from the decarbonation of carbonates. Therefore, the temperature of the region also decreased, as seen in the temperature trend of the seaway during the Bearpaw cycle. The intensified tectonic activity expressed by increased volcanism is recorded by bentonites (originating as volcanic ashes produced during explosive eruptions) in the Bearpaw Formation and its equivalents. The numbers of

bentonite beds increase in the transgressive hemicyclothem and decrease again in the regressive hemicyclothem (Kauffman and Caldwell, 1993).

## 6. Conclusions

The lack of a linear relationship between the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values in ammonite shells indicates that these animals probably deposited their shells in isotopic equilibrium with the ambient waters in which they lived. Comparative study of the stable-isotope compositions of coexisting molluscan fossils suggests that ammonites could be planktonic or nektonic animals (e.g., *Baculites*), the shells of which have the highest average  $\delta^{18}\text{O}$  values and lowest average  $\delta^{13}\text{C}$  values, and probably also vagile benthonic animals (e.g., *Didymoceras*), the shells of which have the lowest average  $\delta^{18}\text{O}$  values and highest average  $\delta^{13}\text{C}$  values, similar to those of the inoceramids. Differences in stable-isotope compositions among coeval ammonites and inoceramids imply that the waters within the seaway were isotopically stratified during the Bearpaw cycle. This stratification is likely caused by multiple processes, including isotopic exchange between sediments and waters and formation of a new water mass by mixing of the Boreal and Tethyan waters.

Baculitids from zones spanning the peak transgression have the lowest average  $\delta^{18}\text{O}$  values, whereas those from other zones have higher average  $\delta^{18}\text{O}$  values. Changes in isotopic compositions through the zonal sequence suggest rising and falling temperatures during transgression and regression, respectively. Temperature fluctuations during the Bearpaw cycle may be partially related to intensified tectonic activity west of the basin. The pattern of the  $\delta^{18}\text{O}$  values cannot be readily explained by variations in freshwater influx. The Bearpaw Sea was not brackish, and thus, other environmental factors are responsible for the rarity of many invertebrate groups prevalent in coeval open oceans and seas.

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