

Cenomanian–Turonian and $\delta^{13}\text{C}$, and $\delta^{18}\text{O}$, sea level and salinity variations at Pueblo, Colorado

Gerta Keller^{a,*}, Zsolt Berner^b, Thierry Adatte^c, Doris Stueben^b

^aDepartment of Geosciences, Princeton University, Princeton NJ 08544, USA

^bInstitut für Mineralogie und Geochemie, Universität Karlsruhe, 76128 Karlsruhe, Germany

^cInstitut de Géologie, II rue Emile Argand, 2007, Neuchatel, Switzerland

Abstract

Stable isotopes of the surface dwelling planktic foraminifera *Hedbergella planispira*, its abundance variations, and mineralogical analysis of the Cenomanian–Turonian at Pueblo, CO, reveal cyclic variations in surface salinity due to changes in precipitation, freshwater influx, marine incursions and long-term sea-level fluctuations. *Hedbergella planispira* is a proxy for salinity variations, as indicated by 2–4‰ more negative $\delta^{18}\text{O}$ values in intervals of peak abundances as compared to intervals with reduced populations. Negative $\delta^{18}\text{O}$ values reflect periods of brackish surface waters caused by freshwater influx during wet humid periods, accompanied by increased clastic transport. More positive $\delta^{18}\text{O}$ values reflect more normal marine salinities as a result of arid periods and/or marine incursions and correlate with intervals of increased biogenic carbonate deposition. The magnitude of salinity variations during the low sea-level of the Hartland Shale is twice that during the sea-level transgression of the Bridge Creek Limestone. The rapid positive $\delta^{13}\text{C}$ shift that marks the onset of Oceanic Anoxic Event 2 (OAE 2) at Pueblo occurred over a period of about 100 ky (93.90–94.00 Ma), and coincided with the major sea level transgression that culminated in the deposition of the basal Bridge Creek Limestone. A positive $\delta^{13}\text{C}$ shift also occurred in the *Rotalipora cushmani* zone prior to OAE 2 and coincided with a sea level rise and enhanced preservation of terrestrial organic matter. The likely cause for OAE 2 is depletion of ^{12}C in the water column as a result of high primary productivity, whereas an earlier *R. cushmani* zone event was primarily caused by increased input of terrigenous organic matter. Both $\delta^{13}\text{C}$ events are associated with enhanced organic matter preservation and anoxic or dysoxic bottom waters.

Keywords: Cenomanian–Turonian; Stable isotopes; Salinity; Sea level; Pueblo, CO

1. Introduction

During the late Cenomanian to early Turonian, the Western Interior basin of North America experienced a major sea-level transgression due to a combination

of subsidence and a tectono–eustatic highstand (Caldell and Kauffman, 1993). As a result, a shallow seaway, the Greenhorn Sea, extended from the western Tethys north to the polar ocean, with the deep central axis passing from New Mexico north through the Rock Canyon area of Colorado (Fig. 1). Rivers drained the highlands to the west of the seaway, and to a lesser extent the eastern lowlands, transporting abundant siliciclastic sediments. Sediment deposition

* Corresponding author. Fax: +1-609-258-1671.

E-mail address: gkeller@princeton.edu (G. Keller).

varied between shales and marls during the low sea levels of the middle to late Cenomanian, and between marls and limestones during the high sea levels of the latest Cenomanian to early Turonian. The limestone/marl couplets are commonly interpreted as changes in clastic dilution and watermass stratification due to Milankovitch forcing of precipitation and runoff (e.g., Barron et al., 1985; Elder and Kirkland, 1985; Pratt et al., 1993; Sageman et al., 1998), or changes in clastic dilution and productivity (Eicher and Diner, 1989, 1991; Ricken, 1991, 1994).

Warm temperature to subtropical climates prevailed in the Cenomanian–Turonian Western Interior with humid to subhumid conditions (Pratt, 1984; Upchurch and Wolfe, 1993), and monsoonal circulation led to fluctuations in the hydrologic cycle (Glancey et al., 1993; Park and Oglesby, 1994). Surface salinity was variable with hyposaline waters at times of shallow seas and high freshwater influx, and more normal marine salinity at times of higher sea levels and the incursion of subtropical water masses from the

western Tethys. With these marine incursions, salinity sensitive planktic and benthic foraminifera and invertebrates invaded the hyposaline interior sea, particularly during the major transgression and global perturbation of the carbon system, Oceanic Anoxic Event 2 (OAE 2), that marks the lower part of the Bridge Creek Limestone Member (Eicher, 1969a,b; Eicher and Worstell, 1970; Elder, 1985; Leckie et al., 1998).

Palaeoenvironmental conditions and the OAE 2 of the Western Interior Seaway have been analyzed in a number of stable isotope studies based on organic matter, bulk rock carbonate, inoceramid and ammonite shells (e.g., Arthur et al., 1985; Pratt, 1984, 1985; Pagani and Arthur, 1998). All of these studies noted the anomalously light oxygen isotopic compositions and concluded that this depletion is likely due to freshwater influx to the seaway.

In this study, the palaeoenvironment, sea-level and salinity variations during the middle Cenomanian through early Turonian of the Greenhorn Formation

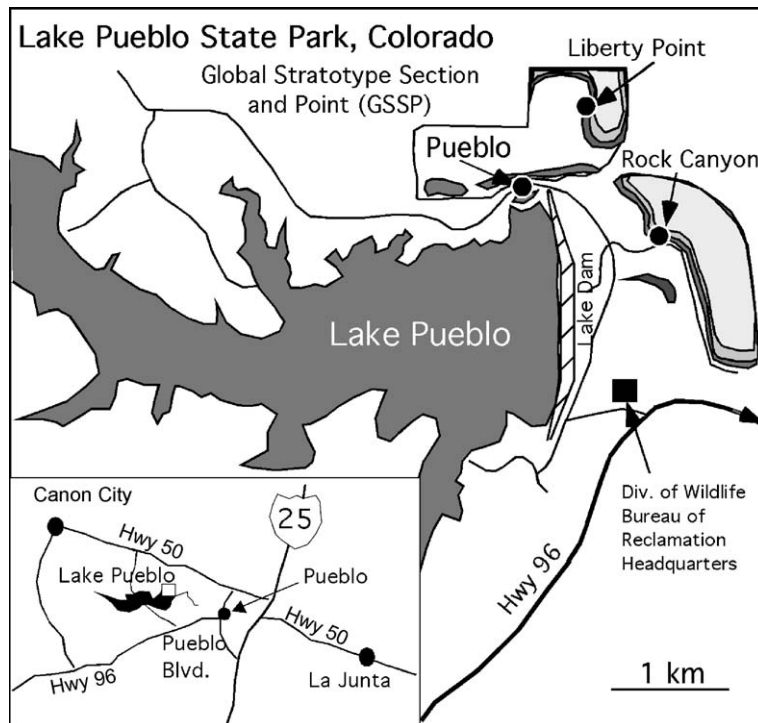


Fig. 1. Location of the Pueblo section in the Rock Canyon Anticline area of Lake Pueblo State Park, Colorado. The section is exposed along the road on the lakeside (see labels).



Fig. 2. Outcrop of the Pueblo section exposed on the road along Lake Pueblo showing exposure of the Hartland Shale and Bridge Creek Limestone.

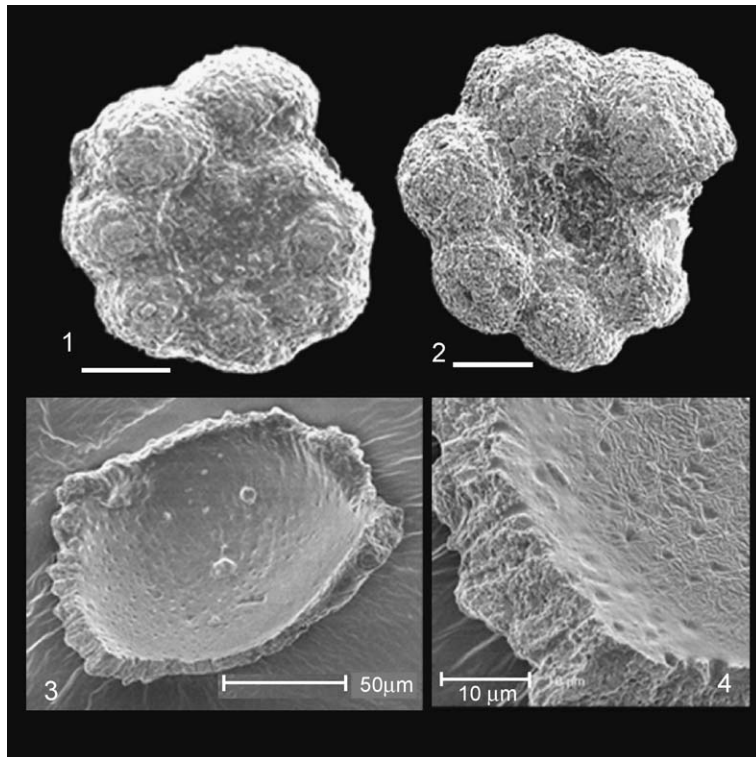


Fig. 3. Scanning electron micrographs (SEM) of *Hedbergella planispira*, which was analyzed for stable isotopes at the Pueblo section. 1 and 2: scale bar = 100 μm . 3 and 4: surface structure details. Note that although there is some recrystallization of the test calcite, the original pore and test structure are preserved and the foraminiferal shells are clear of calcite infilling.

Table 1

Mineralogical (calcite, quartz and phyllosilicates) and stable isotope measurements of *Hedbergella planispira* (63–100 μm , size fraction) and *Cibicoides* spp. at the Pueblo stratotype section

Pueblo	Sample	Height (m)	$\delta^{13}\text{C}$	$\delta^{18}\text{C}$	Calcite (%)	Phyllosilicates (%)	Quartz (%)
Marker beds			<i>Hedbergella</i>				
	PC1	0.00	0.25	− 9.98	28.16	18.00	25.26
	PC2				1.67	36.98	4.67
	PC3				72.19	14.86	10.95
	PC4	0.30	− 0.16	− 8.48	55.69	14.33	20.65
	PC5				76.52	4.78	18.14
	PC6				0.00	71.95	0.74
	PC7	0.85	0.04	− 8.59	26.85	20.54	13.60
	PC7	0.85	− 0.02	− 8.68			
	PC8	1.10	− 0.07	− 9.22	56.32	13.27	22.28
	PC9				31.84	24.07	18.70
	PC10	1.45	− 0.57	− 10.21	26.07	19.44	23.89
	PC11	1.65	− 0.23	− 9.53	34.00	15.40	15.76
	PC12				39.19	10.83	12.78
	PC13	1.95	− 0.22	− 11.14	21.94	20.92	27.23
	PC14				0.00	58.97	4.61
	PC16	2.15	0.07	− 9.00	56.07	12.71	30.38
	PC17				93.38	0.00	6.20
	PC18				40.94	16.49	14.46
	PC19	2.45	0.53	− 8.27	16.25	21.62	9.63
	PC20				36.09	12.16	12.11
	PC21				27.09	17.49	7.78
	PC22	2.60	0.69	− 8.04	39.41	12.14	14.59
	PC23				2.20	46.95	1.30
	PC24	3.00	0.47	− 8.39	69.45	12.15	14.85
	PC25				81.04	9.33	9.46
	PC26	3.55	0.12	− 9.58	32.39	17.10	19.55
	PC27	3.95	0.36	− 9.33	29.50	16.39	18.51
	PC28				41.26	15.38	20.44
	PC29	4.35	0.24	− 9.71	14.68	44.28	3.10
	PC30	4.50	− 0.10	− 10.65	43.55	15.30	20.67
	PC31	5.40	0.11	− 9.80	17.51	13.56	15.47
	PC32	5.70	− 0.24	− 12.06	49.30	7.71	8.24
	PC33	6.25	0.45	− 8.59	28.40	16.12	17.10
	PC34	6.85	− 0.06	− 11.55	32.86	26.07	15.74
	PC35	7.15	0.22	− 10.92	31.27	13.89	7.94
	PC36	7.40	0.73	− 7.26	47.88	13.82	16.70
	PC37	7.95	0.33	− 10.00	29.58	15.54	16.26
	PC38	8.25	0.22	− 10.64	45.05	12.18	15.58
	PC39	8.55	1.35	− 10.58	77.13	0.00	5.09
	PC40	8.75	0.70	− 7.94	25.53	5.84	8.92
	PC41	8.85	0.75	− 7.99	64.10	17.70	16.07
	PC42	8.95	0.21	− 11.00			
	PC43	9.30	0.81	− 9.34	17.88	18.85	55.57
	PC44	9.75	0.13	− 10.68	7.09	8.61	83.65
	PC45	9.85	1.02	− 8.29	8.71	8.60	19.22
	PC46	10.20	− 0.29	− 11.53	14.52	11.95	19.08
	PC47	10.50	0.80	− 9.21	25.54	7.69	8.37
	PC48	10.75	0.64	− 9.72	38.59	21.28	15.23
	PC49	11.00	0.92	− 9.82	72.08	9.52	17.05
	PC50	11.25	1.44	− 7.53	34.05	17.40	14.57
	PC51	11.75	2.07	− 6.50	62.71	21.36	14.75

Table 1 (continued)

Pueblo	Sample	Height (m)	$\delta^{13}\text{C}$	$\delta^{18}\text{C}$	Calcite (%)	Phyllosilicates (%)	Quartz (%)
<i>Marker beds</i>							
<i>Hedbergella</i>							
	PC52	11.80	2.19	-8.36	40.85	10.42	12.44
	PC53	11.85	1.64	-8.93	41.41	15.51	14.48
	PC54	11.90	1.62	-9.07	79.25	10.14	8.05
63	PC55	12.10	2.01	-8.55	87.63	4.47	6.48
63	PC56	12.25	2.52	-7.60	87.32	4.12	7.93
	PC57	12.30	1.60	-8.86	76.75	11.98	10.50
	PC58	12.60	2.20	-8.35	88.83	0.00	10.51
	PC59	12.80	2.07	-8.25	47.54	16.02	11.46
	PC60	12.95	1.60	-8.96	56.91	18.93	11.62
	PC61	13.20	1.91	-7.08			
67	PC62				70.23	4.96	8.44
	PC63	13.35	1.70	-7.45	23.36	6.83	3.75
	PC64	13.35	1.83	-7.31	9.44	26.98	5.22
	PC65	13.55	2.11	-8.57	34.12	21.35	4.76
69	PC66				0.49	35.11	1.44
	PC67	13.75	2.11	-8.75	49.44	19.62	11.18
	PC68				86.11	7.04	6.54
	PC69	14.00	1.63	-8.56	37.21	18.60	6.58
	PC70	14.25	2.13	-6.71	45.31	16.73	20.76
	PC - 71				79.83	4.57	2.84
	PC72	14.45	2.35	-8.52	14.36	42.60	2.14
	PC73				84.86	7.77	3.69
	PC74	14.45	2.41	-8.22	15.28	6.59	76.28
	PC74	14.75	1.86	-8.67	15.28	6.59	76.28
	PC75				73.23	10.67	15.57
	PC76	14.75	1.78	-8.30	56.05	7.92	15.82
	PC77	15.30	1.67	-9.81	20.66	2.93	7.05
	PC77A				60.43	3.65	5.37
	PC78	15.55	1.96	-7.47	33.54	18.99	16.24
	PC79	16.00	1.82	-8.65			
	PC80	16.45	1.81	-8.64	66.86	11.55	2.99
	PC80	16.65	1.66	-7.97	66.86	11.55	2.99
	PC81	16.80	1.61	-8.77	71.98	8.93	7.30
	PC81	17.35	0.94	-7.71	71.98	8.93	7.30
	PC81A				58.49	4.40	16.44
	PC82				75.84	8.39	4.77
	PC83	17.75	1.18	-9.87	64.38	11.69	5.76
	PC83A				60.30	7.29	5.69
	PC84	18.00	1.29	-9.47	84.47	7.15	1.18
	PC85				86.56	8.70	4.32
	PC86				82.36	10.61	5.86
	PC86A				62.78	8.74	10.16
	PC87				75.97	9.54	10.85
	PC87A				63.37	3.97	6.05
Pueblo	Sample	Height (m)	$\delta^{13}\text{C}$	$\delta^{18}\text{C}$	Calcite (%)	Phyllosilicates (%)	Quartz (%)
<i>Marker beds</i>							
<i>Cibicidoides</i>							
	PC88	11.85	1.91	-6.38	71.05	13.99	12.86
	PC89	12.10	2.34	-5.92	81.14	12.47	6.04
	PC90	12.25	2.43	-4.96			
	PC-91				56.45	7.70	5.92

(continued on next page)

Table 1 (continued)

Pueblo	Sample	Height (m)	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Calcite (%)	Phyllosilicates (%)	Quartz (%)
Marker beds			<i>Cibicidoides</i>				
	PC-92				55.57	4.93	6.02
	PC-93				33.50	12.37	4.43
	PC94	12.30	1.86	-6.71	62.51	10.74	7.25
	PC95				90.30	7.20	2.18
	PC96	12.80	1.79	-6.54	41.71	16.72	5.68
	PC97				81.00	10.35	7.08
	PC98				48.30	8.77	3.50
	PC99				76.18	4.19	5.26
	PC100				42.04	4.39	2.94

Note that benthic foraminifera are rare or absent, except in the oxic phase after the $\delta^{13}\text{C}$ excursion. Height (m) in the section is measured from the base of the outcrop and have been indexed to the key marker beds for orientation.

at Pueblo, CO, are analyzed based on mineralogy and stable isotope compositions of the surface dwelling planktic foraminifera *Hedbergella planispira* and the benthic taxon *Cibicidoides* spp. The Pueblo section is the Global Stratotype and Point of the Cenomanian/Turonian (C/T) boundary (Kennedy and Cobban, 1991; Kennedy et al., 2000).

2. Methods

A total of 100 samples were collected at 10-cm intervals in the changing lithologies of the Bridge Creek Limestone, and at 20- to 25-cm intervals in the more monotonous shale of the underlying Hartland Shale Member (Fig. 2). All samples were analyzed for bulk-rock mineralogy and microfossil content. Preservation of planktic foraminifera is relatively good, though tests are recrystallized to varying degrees depending on size and morphology of the species. Large species tend to be completely recrystallized with chambers frequently infilled with blocky calcite. In contrast, small species, such as *Hedbergella planispira*, tend to be only partially recrystallized, with the shell calcite and pore structure preserved and no infilling of test chambers (Fig. 3). This indicates that small species at Pueblo are more likely to retain primary environmental signals. Stable isotope analysis was thus performed on monospecific samples of the surface dwelling planktic foraminifer *H. planispira* using 40–75 specimens per sample in the size fraction 63–100 μm . Limestone samples were excluded from this analysis. Of the remaining lithologies (shales,

marls, bentonites, calcarenites), a total of 68 samples yielded sufficient specimens for isotope analysis (Table 1). Benthic foraminifera are nearly absent in the Hartland Shale and are rare in the Lower Bridge Creek Limestone, except for five samples that contained sufficient *Cibicidoides* for analysis.

The samples were analyzed at the stable isotope laboratory of the Department of Mineralogy and Geochemistry at the University of Karlsruhe, Germany, using an Optima (Micromass, UK) ratio mass spectrometer equipped with an online carbonate preparation line (Multi Carb) with separate vials for each sample. The results were calibrated to the PDB scale with standard errors of 0.1‰ for $\delta^{18}\text{O}$ and 0.05‰ for $\delta^{13}\text{C}$ (Table 1). Replicate sample analyses for $\delta^{13}\text{C}$ were within 0.06–0.13‰ and for $\delta^{18}\text{O}$, they ranged from 0.09‰ to 0.37‰. The higher values may reflect varying degrees of diagenetic alteration and/or seasonality effects.

Bulk rock mineral analyses were conducted at the Geological Institute of the University of Neuchâtel, Switzerland, based on XRD (SCINTAG XRD 2000 diffractometer), following the procedure outlined by Kübler (1987) and Adatte et al. (1996). Bulk rock contents were obtained using standard semiquantitative techniques based on external standardization.

3. Lithology

The studied Pueblo section forms a road cut at the northeast end Lake Pueblo (Figs. 1 and 2) where about 18.5 m of gray shales, bentonites and tan-

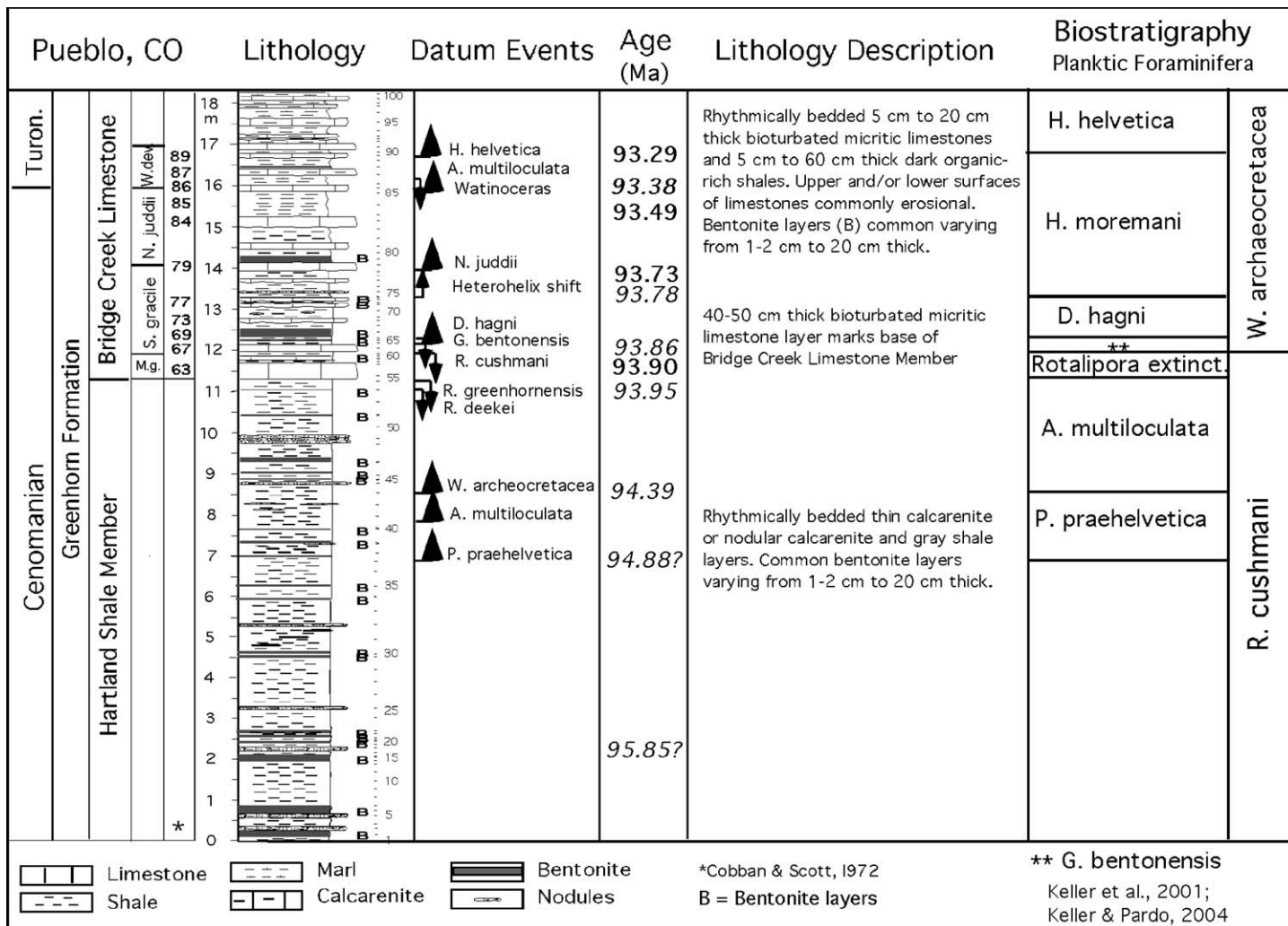


Fig. 4. Lithology, biostratigraphy and planktic foraminiferal and ammonite datum levels of the Pueblo section. Ages for datum levels are from Hardenbol et al. (1998), except for *Rotalipora deekei*, *Rotalipora greenhornensis* and *Globigerinelloides bentonensis*, which are extrapolated from age/depth and sediment accumulation rates. Planktic foraminiferal biozonation after Keller et al. (2001, Keller and Pardo (2004. Ammonite zonation from Cobban (1985).

colored limestones of the Greenhorn Formation are exposed. The Greenhorn Formation is divided into Hartland Shale and Bridge Creek Limestone Members. The Hartland Shale Member at this outcrop consists of 11.2 m of rhythmically bedded thin calcarenite or nodular calcarenite layers and 30- to 100-cm-thick gray shales. Bentonite layers are common and vary from 1–2 to 20 cm thick (Fig. 4).

The Bridge Creek Limestone Member exposed is about 6.5 m thick. A prominent 40- to 50-cm-thick bioturbated micritic limestone marks the base of the Bridge Creek Limestone and contains an Upper Cenomanian ammonite assemblage of the *Sciponoceras gracile* zone (Cobban, 1985) and a planktic foraminiferal assemblage indicative of the uppermost *Rotalipora cushmani* zone (Leckie, 1985; Keller and Pardo, 2004). Previous studies have identified this limestone facies as the main sea-level transgression near the end of the Cenomanian (Hancock and Kauffman, 1979; Kauffman, 1984; Sageman et al., 1998). Upsection, the lithology consists of rhythmically bedded 10- to 20-cm-thick bioturbated micritic limestones alternating with 10- to 60-cm-thick organic-rich dark shales (Figs. 2 and 4). Bentonite layers are common and are of variable thickness ranging from 1–2 to 20 cm, similar to the Bridge Creek Limestone have been labeled as marker beds by previous workers and in this study we follow the numbering system of Cobban and Scott (1972; Fig. 4).

4. Biostratigraphy and age control

The biostratigraphy of the Pueblo section is generally defined by ammonite and inoceramid zones (e.g., Cobban, 1985; Elder, 1985; Kennedy and Cobban, 1991; Leckie et al., 1998; Kennedy et al., 2000), though planktic foraminifera are abundant and provide the means for global correlations. Based on the Eastbourne and Pueblo sections, Keller et al. (2001) and Keller and Pardo (2004) identified new biostratigraphic markers and subdivided the *Whiteinella archeoretacea* and *Rotalipora cushmani* zones into three subzones for improved age control (Fig. 4). In both sections, planktic foraminiferal datum events and the structure of the $\delta^{13}\text{C}$ excursion (discussed below) are comparable. Age estimates for species first and last appearances have been calculated based on ages

of various ammonite and planktic foraminiferal datum events, which are extrapolated from the palaeomagnetic time scale and radiometric dates of Hardenbol et al. (1998) (Table 2). These ages are in good agreement with Eastbourne, except for *Helvetoglobaotruncana helvetica* which appears earlier at Eastbourne, possibly because this species is diachronous and/or because the evolutionary transition from *praehelvetica* to *helvetica* is difficult to determine.

Three $^{40}\text{Ar}/^{39}\text{Ar}$ ages determined from three bentonite layers (Obradovich, 1993; Kowallis et al.,

Table 2

Ages for planktic foraminiferal and ammonite first (FAD) and last (LAD) appearance datums, and $\delta^{13}\text{C}$ shifts (discussed below) at the Pueblo GSSP

Datum events	Age (Ma)	Reference
<i>Helvetoglobotruncana helvetica</i> FAD	93.29 ± 0.2	Hardenbol et al. (1998)
<i>Watinoceras devonense</i> FAD	93.49 ± 0.2	Hardenbol et al. (1998)
<i>Neocardioceras juddii</i> LAD	93.49 ± 0.2	Hardenbol et al. (1998)
<i>Neocardioceras juddii</i> LAD	93.59	this study at Pueblo
<i>Metoicoceras geslineanum</i> LAD	93.73 ± 0.2	Hardenbol et al. (1998)
<i>Neocardioceras juddii</i> FAD	93.73 ± 0.2	Hardenbol et al. (1998)
<i>Heterohelix</i> shift	93.78 ± 0.02 ^a	this study
OAE 2 $\delta^{13}\text{C}$ peak (2) excursion max.	93.86 ± 0.05 ^a	this study
<i>Globigerinelloides bentonensis</i> LAD	93.86 ± 0.06 ^a	this study
<i>Dicarinella hagni</i> FAD	93.86 ± 0.05 ^a	this study
<i>Rotalipora cushmani</i> LAD	93.90 ± 0.02 ^a	Hardenbol et al. (1998)
OAE 2 $\delta^{13}\text{C}$ peak 1	93.91 ± 0.02 ^a	this study
OAE 2 $\delta^{13}\text{C}$ excursion onset	94.00 ± 0.02 ^a	this study
<i>Rotalipora greenhornensis</i> LAD	93.95 ± 0.02 ^a	this study
<i>Rotalipora deekei</i> LAD	93.94	this study
<i>Whiteinella archeoretacea</i> FAD	94.50	this study
<i>Praeglobotruncana praehelvetica</i> FAD	94.88	this study
Mid-Cenomanian $\delta^{13}\text{C}$ shift (MCE)	95.71	this study
<i>Whiteinella paradubia</i> FAD	95.85	this study

^a Datum events at Pueblo that have been cross-correlated with Eastbourne, England. Error margins reflect uncertainty between these two sections.

1995) are compatible with these dates, but show considerable spread. Based on these $^{40}\text{Ar}/^{39}\text{Ar}$ data, the C/T boundary is estimated at 93.1 ± 0.3 to 93.3 ± 0.2 Ma, as compared with 93.49 ± 0.2 Ma based on the first appearance of *Watinoceras devonense*, which is the preferred marker for the C/T boundary, and 93.29 ± 0.2 Ma for the first appearance of *Helvetoglobotruncana helvetica*, the planktic foraminiferal marker species (Hardenbol et al., 1998; Kennedy et al., 2000).

5. Sedimentation rates

Sedimentation rates calculated from an age/depth plot for the Pueblo section (Fig. 5) indicate very low rates, suggesting a condensed record. The sedimenta-

tion curve can be interpreted as two linear segments intersecting at limestone marker bed 79. For the lower segment (Bed 63 to Bed 79) of the Bridge Creek Limestone, sediment accumulation rates average 1.23 cm/ky (1.11 cm/ky excluding bentonite layers), whereas for the upper segment they average 0.87 cm/ky (0.75 cm/ky excluding bentonite layers), assuming that limestone and shale layers were deposited at roughly equal rates. The break at limestone marker bed 79 may be related to the rising sea level and reduced terrigenous influx in the upper part of the Bridge Creek Limestone.

These sedimentation rates compare well with the average rates for the entire interval of 0.5–1.0 cm/ky estimated by Elder and Kirkland (1985), 0.9 cm/ky by Scott et al. (1998) and 0.57 cm/ky by Sageman et al. (1998). In comparison, the compa-

Pueblo, Colorado (GSSP)

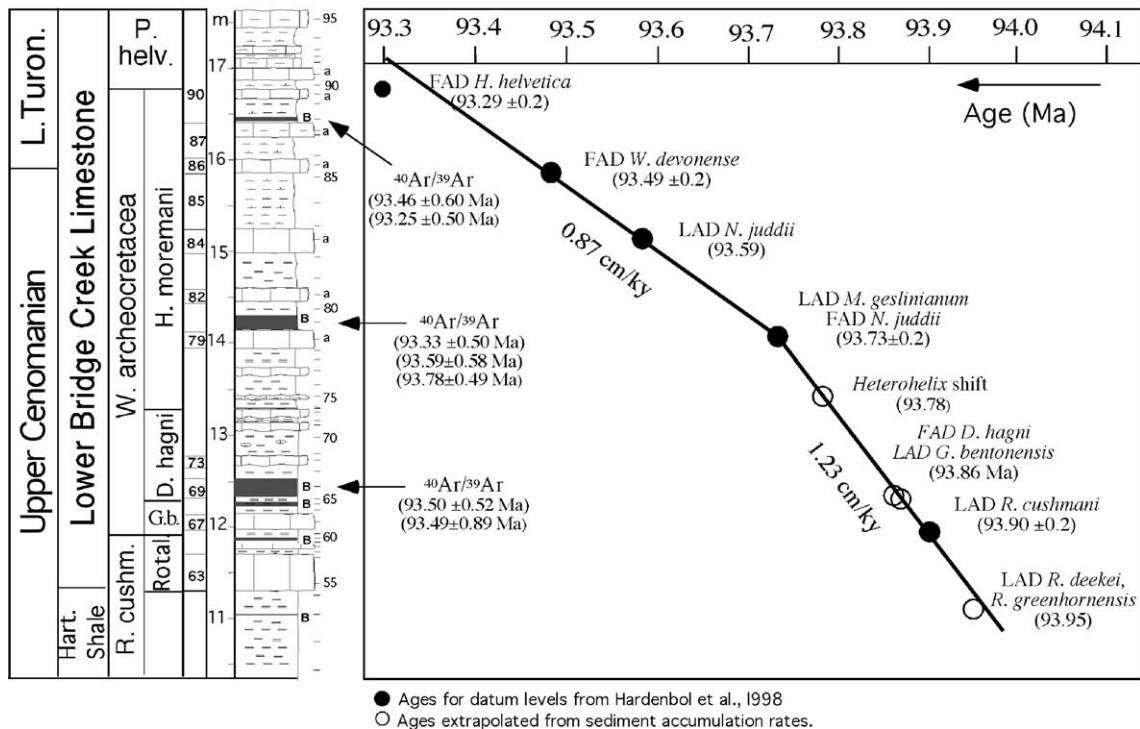


Fig. 5. Age/depth plot of the Pueblo section based on planktic foraminiferal and ammonite datum levels from Hardenbol et al. (1998) and extrapolated datum level ages for the *Heterohelix* shift, *Globigerinelloides bentonensis*, *Rotalipora deekei* and *Rotalipora greenhornensis*. Calculated datum level ages are coeval at Pueblo and Eastbourne. $^{40}\text{Ar}/^{39}\text{Ar}$ ages for bentonite layers are from Obradovich (1993) and Kowallis et al. (1995).

rable interval at Eastbourne averages 2.4 cm/ky. The low sediment accumulation rate is probably due to short periods of non-deposition between marl/limestone couplets and cyclic climate variations with humid wet periods accompanied by high sediment and freshwater influx depositing shales, and more arid periods accompanied by reduced sedimentary influx and increased production of biogenic carbonate (foraminiferal tests, coccoliths, invertebrates) depositing calcarenites and limestones. These dilution/productivity cycles have been interpreted as obliquity and precession cycles, respectively (Sage-man et al., 1998).

6. Stable isotopes

Oxygen and carbon isotope analyses of the Pueblo section are based on the surface dwelling planktic foraminifer *Hedbergella planispira* (Fig. 6). *Hedbergella planispira* populations range between -0.5% and 1.0% $\delta^{13}\text{C}$ for the Hartland Shale, and between 1.5% and 2.5% $\delta^{13}\text{C}$ for the Bridge Creek Limestone above the positive $\delta^{13}\text{C}$ excursion (Fig. 6). Similar values are observed in whole-rock $\delta^{13}\text{C}$ values of the same section (Pratt, 1984), and the overall trend is similar to the organic $\delta^{13}\text{C}$ curve (Pratt, 1985; Pratt and Threlkeld, 1994). In general, the observed $\delta^{13}\text{C}$ trends are very similar to published curves based on bulk-rock or fine-fraction carbonates across the late Cenomanian to early Turonian that mark the organic carbon-rich marine sediments known as Oceanic Anoxic Event 2 (OAE 2, Scholte and Arthur, 1980; Jarvis et al., 1988, 2001; Jenkyns et al., 1994; Accarie et al., 1996; Nederbragt and Fiorentino, 1999; Paul et al., 1999; Keller et al., 2001). However, at Pueblo, the absolute values of both bulk rock and planktic foraminifera are about 2.5–3‰ lighter. Pagani and Arthur (1998) observed a similar difference in unaltered aragonite and calcite shells of primarily early Turonian macroinvertebrates (ammonites, inoceramids and oysters) from other localities in the Western Interior, and attributed this to early diagenetic cementation and organic carbon degradation in the sediments. Because of the vital effects between macroinvertebrates and planktic foraminifera, differences in age and region, no direct comparison can be made.

6.1. Carbon isotopes

6.1.1. OAE 2 $\delta^{13}\text{C}$ excursion

The major features of the planktic foraminifera *H. planispira* $\delta^{13}\text{C}$ OAE 2 excursion at Pueblo are: (1) a rapid increase of 2‰ to reach the first peak, (2) a 0.6‰ decrease forming a trough, (3) a 1‰ increase to form a second peak, though followed by low values in the bentonite bed 69, and (4) prolonged high, but variable, $\delta^{13}\text{C}$ values into early Turonian (Fig. 6).

The same features are better seen in the more expanded coeval interval at Eastbourne, England (Paul et al., 1999; Keller et al., 2001) and suggest that this pattern reflects a series of widespread or global oceanographic events (Fig. 7). A decrease, or trough, occurs between peak 1 and peak 2 at Eastbourne and these intervals correspond to the *G. bentonensis* and *D. hagni* subzones, respectively. At Pueblo, the trough interval is very condensed though marked by a 0.6‰ decrease in $\delta^{13}\text{C}$ (Figs. 6 and 7). The $\delta^{13}\text{C}$ peak 2 is a sharp increase followed by a 1‰ drop in the bentonite marker bed 69 of Cobban and Scott (1972), and return to high values in the marl above. It is possible that the bentonite value is a diagenetic signal and peak 2 spans through the marl and limestone bed 73 above it, as suggested by biostratigraphy (see shaded interval in Fig. 7). Alternatively, the second peak may be condensed and/or partially missing at Pueblo. Above this interval, $\delta^{13}\text{C}$ values remain high with a temporary decrease near the first appearance of *H. helvetica* about 0.75 m above the C/T boundary at Pueblo (Fig. 7).

The onset of the rapid $\delta^{13}\text{C}$ excursion at Pueblo begins about 65 cm below the base of the Bridge Creek Limestone, and reaches a maximum in the marl layer above limestone marker bed 63 (Fig. 6). Based on sediment accumulation rates the $\delta^{13}\text{C}$ excursion began at 94.00 Ma and reached the first peak about 90 ky later at 93.91 Ma. The equivalent interval at Eastbourne is also estimated between 94.00 and 93.90 Ma (Keller et al., 2001).

6.1.2. Benthic oxic event

The second important feature of this dataset is the appearance of diverse benthic assemblages, including *Cibicidoides* species, indicating normal oxic bottom waters beginning with $\delta^{13}\text{C}$ peak (1) and continuing just above $\delta^{13}\text{C}$ peak (2), or about 100 ky (benthic

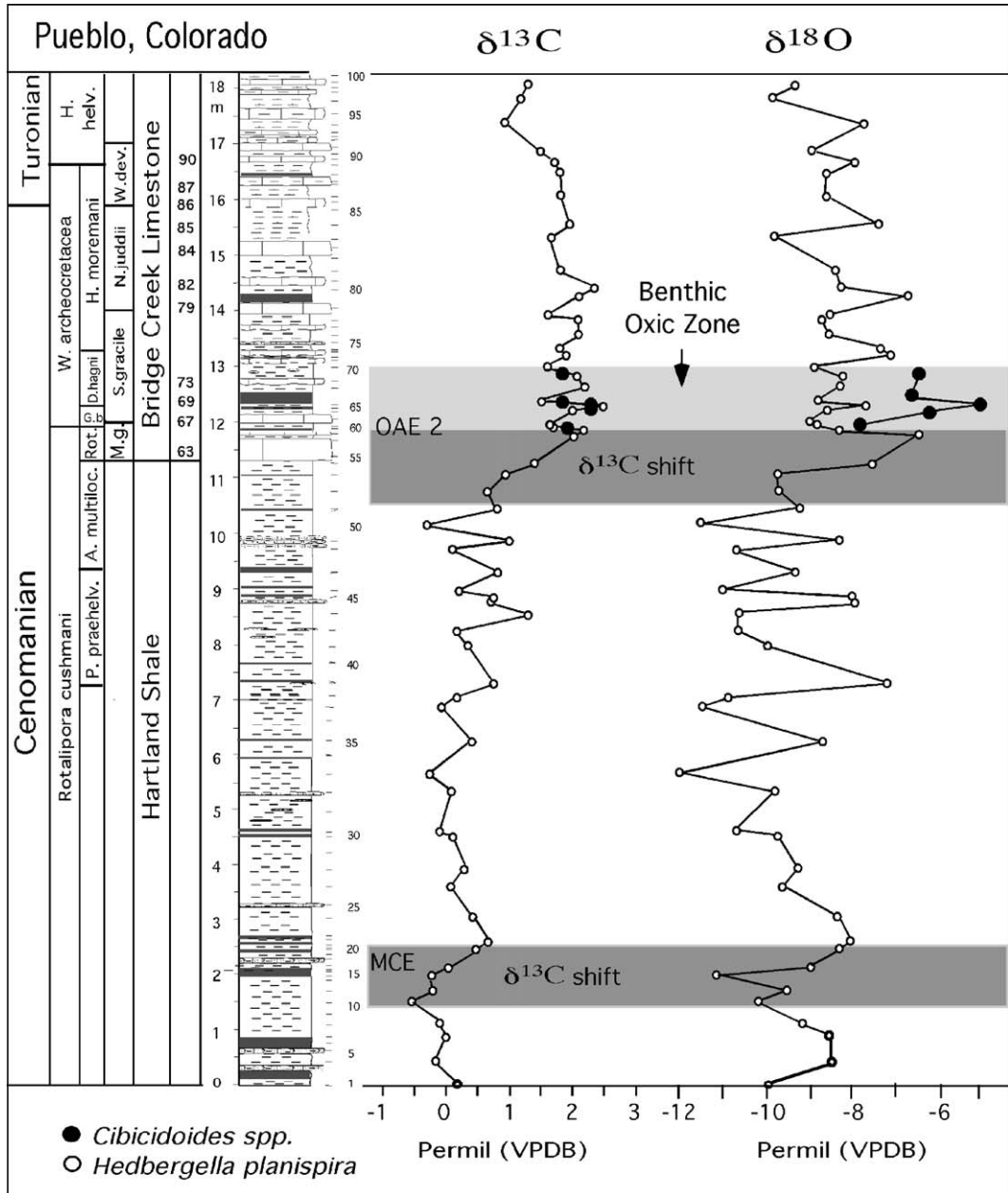


Fig. 6. Carbon and oxygen isotope curves of the planktic foraminifera *Hedbergella planispira* and the benthic foraminifera *Cibicidoides* spp. at Pueblo. Note that only a short interval in the benthic oxitic zone contained sufficient benthic foraminifera for analysis. Benthic foraminifera in the remainder of the section are rare or absent due to dysoxic or anoxic bottom conditions. There are two $\delta^{13}\text{C}$ shifts, the Mid-Cenomanian Event (MCE) and Ocean Anoxic Event 2 (OAE 2) of the late Cenomanian. OAE 2 has two distinct peaks separated by a trough, as also noted at Eastbourne (e.g., Paul et al., 1999; Jarvis et al., 2001; Keller et al., 2001). The unusually light $\delta^{18}\text{O}$ values are due to salinity variations. See text for explanation.

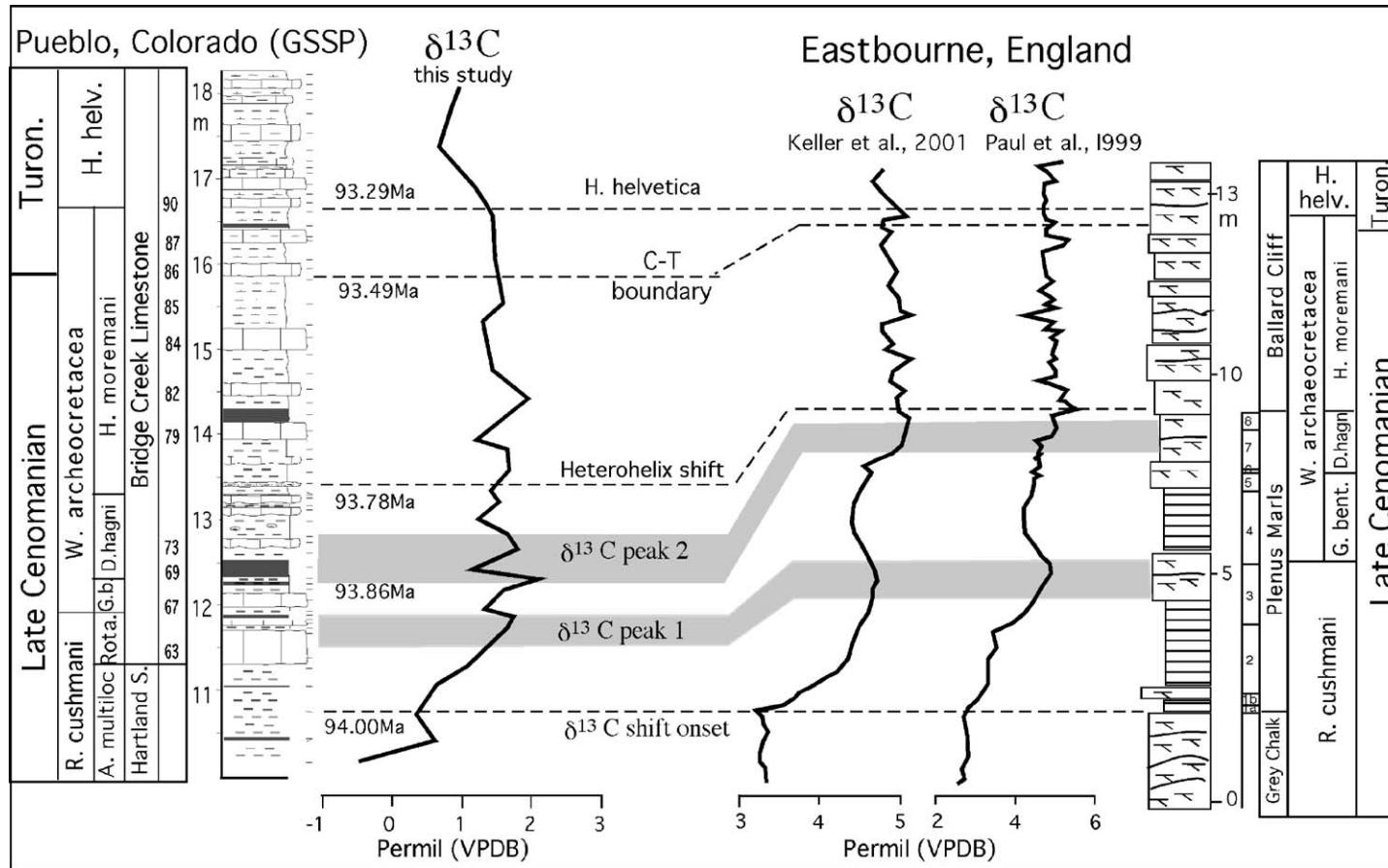


Fig. 7. Correlation of $\delta^{13}\text{C}$ isotope curves of Pueblo and Eastbourne based on planktic foraminiferal datum events shows the similarities of these curves including the sequence of rapid rise to peak (1), trough and peak (2). Note that the interval of the trough between the two peaks is condensed at Pueblo and suggests a period of non-deposition encompassing about 50 kyr.

oxic zone, Fig. 6). This is the only interval that yielded sufficient benthic species for stable isotope analysis and the only interval with benthic species that indicate oxic bottom waters. $\delta^{13}\text{C}$ values of *Cibicides* are nearly the same as for the surface dweller *Hedbergella planispira*, possibly due to the shallow water depth. This Interval of diverse benthic assemblages is probably equivalent to the “benthic zone” of Eicher (1969a).

6.1.3. *Rotalipora cushmani* zone $\delta^{13}\text{C}$ excursion

The third significant feature of the Pueblo $\delta^{13}\text{C}$ data is 1.5‰ positive $\delta^{13}\text{C}$ excursion, immediately following a negative excursion in the Hartland Shale about 8.5 m below the major $\delta^{13}\text{C}$ shift (Fig. 6). Above the positive excursion, $\delta^{13}\text{C}$ values gradually decrease and species diversity significantly increases. Two $\delta^{13}\text{C}$ excursions mark Mid-Cenomanian Events MCE I and MCE II in sections of northwestern Europe (Paul et al., 1994; Jenkyns et al., 1994; Mitchell et al., 1996; Mitchell and Carr, 1998). It is possible that the positive $\delta^{13}\text{C}$ excursion in the *Rotalipora cushmani* zone at Pueblo may be the upper MCE II, though correlation is uncertain in the absence of planktic foraminiferal control in the northwestern European sections. However, at least one of the MCEs is within the upper *R. cushmani* zone, about 17 m below OAE2, in pelagic limestones of the Umbria–Marche Basin of Italy as reported by Coccioni and Galeotti (2003). In the same area (Bottaccione Gorge, Gubbio) Jenkyns et al. (1994) also documented the MCE positive $\delta^{13}\text{C}$ excursion in the *R. cushmani* zone. The stratigraphic distance also seems comparable with some sections (e.g., Speeton) in the Anglo–Paris basin, where MCE I and MCE II are about 14 and 18 m below the OAE2, respectively.

Nevertheless, the positive $\delta^{13}\text{C}$ excursion at Pueblo may be a younger, late middle or late Cenomanian event. This is suggested by the ammonite stratigraphy. At Pueblo, the positive $\delta^{13}\text{C}$ excursion is in the *Metoicoceras mosbyense* ammonite zone of the lower Hartland Shale Member. In southern England, the MCEs occur within the lower middle Cenomanian *Acanthoceras rhotomagense* ammonite zone. By ammonite correlations this would place the MCEs about 30 m lower in the section in the Graneros Shale (Hancock et al., 1993).

6.2. Oxygen isotopes

Oxygen isotope ratios of the surface dweller *Hedbergella planispira* from the Pueblo section show a wide range of unusually low negative values that range from -7‰ to -12‰ $\delta^{18}\text{O}$ in the Hartland Shale and -6‰ and -9‰ $\delta^{18}\text{O}$ in the Bridge Creek limestone (Fig. 6). Whole-rock $\delta^{18}\text{O}$ values are also low though less depleted and range from -7‰ to -9‰ and -4‰ to -7‰ for the Hartland Shale and Bridge Creek, respectively (Pratt, 1985; Pratt et al., 1993). However, the higher sample resolution in our study prevents specific comparison of key features. The major features within the *H. planispira* dataset include: (1) cyclic high amplitude variation of 2–4‰ in the Hartland Shale, but only 1–2‰ in the Bridge Creek Limestone; (2) a 3‰ negative excursion preceding the positive MCE $\delta^{13}\text{C}$ shift; (3) positive $\delta^{18}\text{O}$ excursions associated with peak MCE and OAE 2 $\delta^{13}\text{C}$ shift; (4) positive benthic $\delta^{18}\text{O}$ shift during the “benthic oxic zone”; and (5) a negative excursion just below the ammonite C/T boundary (Fig. 6).

6.3. Diagenetic effects

Although $\delta^{18}\text{O}$ values of planktic foraminifera may be increased by early diagenesis (e.g., Schrag, 1999; Pearson et al., 2001), low $\delta^{18}\text{O}$ values and a covariance between the oxygen and carbon isotope ratios are generally considered to reflect late diagenetic overprinting caused either by interaction with meteoric ground waters, or by dissolution and recrystallization at higher temperatures during burial diagenesis (e.g., Schrag et al., 1995; Mitchell et al., 1997; Sakai and Kano, 2001). At Pueblo, the initial impression is that the overall low $\delta^{18}\text{O}$ values and a fairly good $\delta^{18}\text{O}/\delta^{13}\text{C}$ correlation ($R=0.45$, Fig. 8A) suggest that the primary signal has been diagenetically altered.

However, the depleted *Hedbergella planispira* values can, at best, only be partly explained by diagenetic alteration. Stable isotope analysis of well-preserved and recrystallized planktic foraminifera at Pueblo indicate that only about 1.5‰ can be attributed to diagenetic effects (Dal’Agnolo, written communication, 2003). However, even if 1.5–2‰ can be attributed to diagenetic effects, the overall values still remain unusually depleted (-5‰ to 10‰ and -4‰ to 7‰ for the Hartland Shale and Bridge Creek,

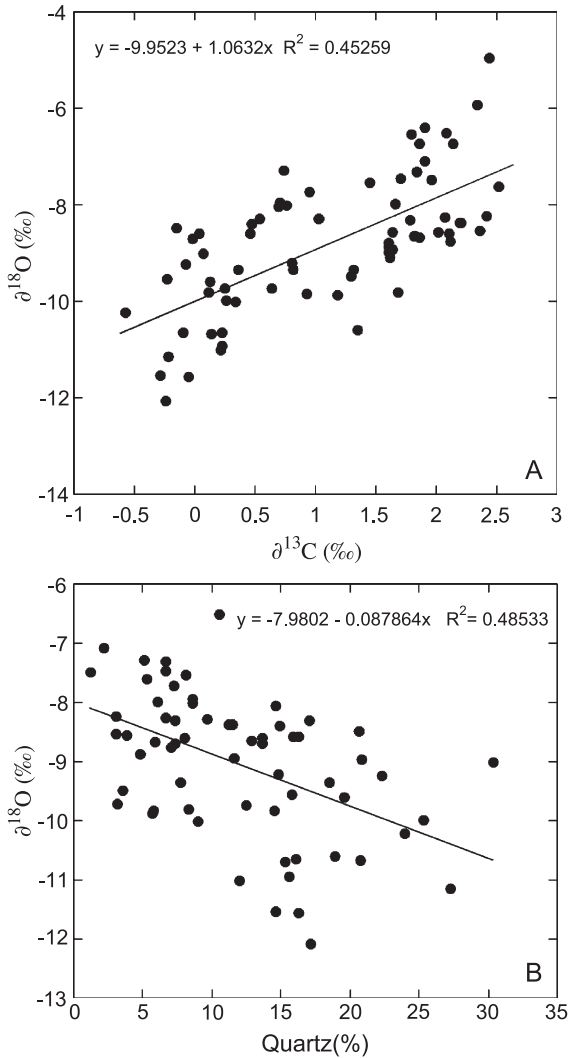


Fig. 8. Cross correlation of (A) $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, and (B) $\delta^{18}\text{O}$ and quartz in the Bridge Creek Limestone and Hartland Shale at Pueblo. (A) Note that there is some covariance between less negative $\delta^{18}\text{O}$ and more positive $\delta^{13}\text{C}$ values for the entire section ($R=0.45$), but no significant covariance for the Hartland Shale ($R^2=0.20$). This indicates that the negative $\delta^{18}\text{O}$ values can be explained only partly as late diagenetic overprint and that part of the primary signal is preserved. (B) There is a significant correlation ($R^2=0.48$) between high negative $\delta^{18}\text{O}$ values and maxima of quartz, which supports freshwater influx and increased erosion associated with high precipitation.

respectively), which cannot be attributed to diagenesis. Similarly, the cyclic fluctuations in $\delta^{18}\text{O}$ values are difficult to explain by differential diagenetic effects in these monotonous shales and marls (no

limestones were included in this data set) and these must have a different explanation. Moreover, there is a strong negative covariance between the $\delta^{18}\text{O}$ values and peak *H. planispira* abundance that cannot be explained by diagenesis (Fig. 9); (note that because of different background trends, a regression analysis is not useful to show this covariance). Although the $\delta^{18}\text{O}$ variations can be attributed to temperature, this is considered to be unlikely because of the unrealistically high temperature fluctuations required to generate such a large shift in $\delta^{18}\text{O}$.

Previous isotope studies on carbonate material concluded that the correspondence of whole-rock and inorganic $\delta^{18}\text{O}$ values precludes a major diagenetic effect (Pratt, 1985; Arthur et al., 1985; Pagani and Arthur, 1998). Based on a comparative study between the isotope composition of oxygen in calcareous remains of organisms and clay minerals in bentonites, Cadrin et al. (1995) inferred that the oxygen-isotope composition of the Western Interior Seaway was as low as -8‰ (VSMOW) during the Cenomanian, increasing with the transgression and the northward spread of Tethyan waters. Calcite in equilibrium with seawater of -8‰ $\delta^{18}\text{O}$ (VSMOW) at $26\text{ }^\circ\text{C}$ has a value of -10‰ (VPDB)—using the equation of Erez and Luz, 1983—which, taking into account an estimated diagenetic offset of 2‰ , is at our minimum value of -12‰ (PDB) for Pueblo. Consequently, instead of major diagenetic effects, the unusually low $\delta^{18}\text{O}$ values in the foraminifera *Hedbergella planispira* are more likely to be the expression of isotopic equilibration with the ^{18}O -depleted waters of the Greenhorn sea (e.g., Whittaker et al., 1987; Clarke and Jenkyns, 1999). In other words, the most depleted $\delta^{18}\text{O}$ values at Pueblo adjusted for diagenesis represent primary signals and reflect brackish water environments.

6.4. Salinity effects

In addition to diagenesis, there are several alternatives to explain the covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in carbonates (Fig. 8A). Among these, a change in salinity is most frequently invoked, which may occur due to either mixing between water masses with different salinities and isotopic compositions, or to changes in the precipitation–evaporation cycle (e.g., Wolff et al., 1999, Pierre, 1999; Rohling and

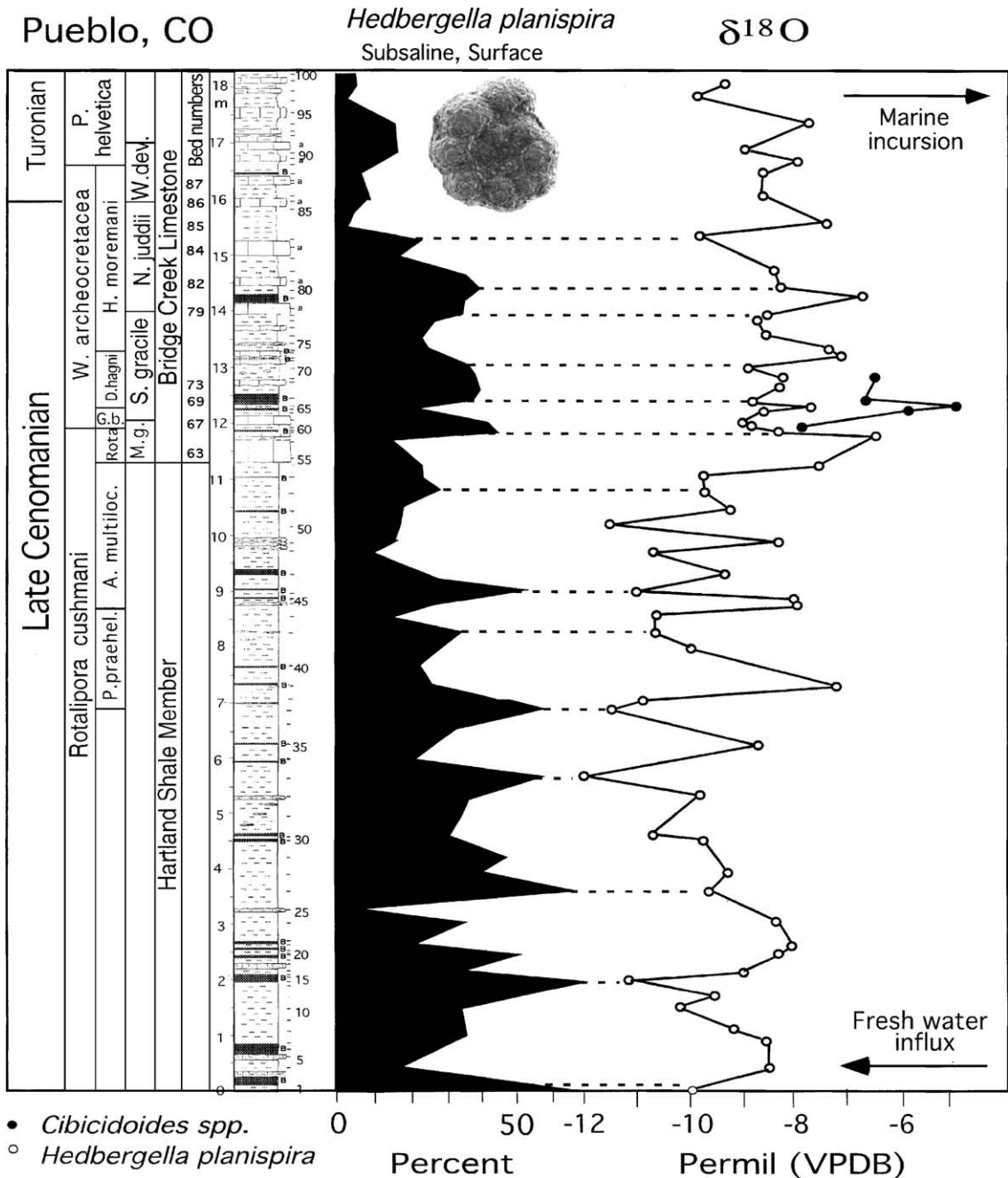


Fig. 9. Relative abundance of *Hedbergella planispira* compared with the $\delta^{18}\text{O}$ curve of this species at Pueblo. Note that relative peak abundances of this species generally correlate with the most negative $\delta^{18}\text{O}$ measurements that reflect subsaline surface waters. This indicates that *H. planispira* thrived in subsaline surface waters and can be used as proxy for salinity variations.

De Rijk, 1999). According to Cadrin et al. (1995), the isotopic heterogeneity of the Greenhorn Sea resulted not only from mixing of ^{18}O -depleted northern and normal-marine southern waters, but also from runoff of meteoric water from the developing Cordillera (Cadrin, 1992). Recently, Cochran et al. (2003) postulated the mixing of freshwater and seawater in a kind of “subterranean estuary” (i.e., a nearshore aquifer system) in order to explain salinity changes in the Maastrichtian Western Interior Seaway, as reflected by the O and Sr isotope records of well-preserved fossils.

The depleted $\delta^{18}\text{O}$ values of the planktic foraminifera *Hedbergella planispira* at Pueblo are in agreement with published estimates for the oxygen isotope composition of the seawater, ranging from -2‰ to -8‰ (VSMOW) for the Late Cretaceous Western Interior Seaway (Whittaker et al., 1987; Cadrin et al., 1995). As shown above, the value of -8‰ $\delta^{18}\text{O}$ (VSMOW) at the lower end of the range corresponds to the lowest oxygen isotope ratio registered at Pueblo (-12‰ VPDB). Considering that changes in the isotope composition of water are mainly due to mixing, it is not difficult to demonstrate that such seawater oxygen isotope values are compatible with realistic brackish salinities, based on reasonable assumptions of salinity and isotope compositions of the eligible mixing end-members.

Open ocean Late Cretaceous seawater, with a $\delta^{18}\text{O}$ value of about -2‰ (Cadrin, 1992) was slightly depleted in ^{18}O relative to the modern global ocean. Assuming a latitude-dependent isotopic gradient similar to present-day values (Schmidt et al., 1999), the southwards expanding arctic waters, may have been at least 3‰ lighter than mean ocean water at that time. The run-off and streams draining the high mountain ranges to the west must have been strongly depleted in ^{18}O due to the altitude effect, even at Pueblo’s moderately high palaeolatitude of 38°N (Pagani and Arthur, 1998). Glancey et al. (1993) estimated for this component a range between -21.5‰ and -11‰ (VSMOW). Similar values coupled with a strong latitude effect were also indicated by Phillips et al. (2000), based on the oxygen isotopic composition of meteoric–phreatic siderite and meteoric sphaerosiderites in the palaeo-aquifers and palaeosols of the Cretaceous Interior Seaway. In the shallow littoral areas of this Seaway, we assume a 40% freshwater

$\delta^{18}\text{O}$ component of -15‰ (VSMOW) and a marine component consisting equally of normal marine (-2‰) and arctic oceanic waters (-5‰). Based on these assumptions, a simple isotope mass balance calculation yields a $\delta^{18}\text{O}$ value of -8.1‰ (VSMOW). The salinity of such a watermass, assuming 35‰ for the marine component (Cochran et al., 2003) and 0.5‰ for freshwater, is with 21.2‰ practically identical with the salinity of 20‰ hypothesized by Cochran et al. (2003) for the brackish facies, based on the salinity tolerance of modern cephalopods. We are aware that such scenarios may be indefinitely varied. However, this example illustrates the plausibility of relating the low $\delta^{18}\text{O}$ values of the foraminifera to salinity changes.

Comparably low $\delta^{18}\text{O}$ values are not unusual in modern brackish facies. Surface waters of the Baltic Sea have average $\delta^{18}\text{O}$ values of about -7‰ (VSMOW) and minima as low as -8.2‰ at salinities of about 10‰ (e.g., Froehlich et al., 1988). Though the Baltic Sea is at a higher latitude than the palaeolatitude of the Western Interior Seaway (i.e., the decrease of $\delta^{18}\text{O}$ due to the latitude effect is stronger), at Pueblo this may be compensated by an altitude effect due to the proximity of the Cordillera to the west.

At Pueblo, there is a significant inverse correlation ($R=0.49$) between $\delta^{18}\text{O}$ values and the amount of quartz, which also supports freshwater influx and increased erosion associated with high precipitation (Fig. 8B) at times with low $\delta^{18}\text{O}$ values. In addition, the lowest $\delta^{18}\text{O}$ values correspond to high total organic content (TOC) and high oxygen index (OI) values, which indicate terrestrial organic matter influx due to increased precipitation and runoff.

Thus, the cyclic high amplitude $\delta^{18}\text{O}$ variations are interpreted to be the effects of mixing between marine and freshwater components. The Pueblo stable isotope data of the surface dweller *Hedbergella planispira* further indicate that this surface dweller thrived in subsaline waters and is thus a proxy for freshwater influx, as demonstrated below. The overall lower $\delta^{18}\text{O}$ trend of the Hartland Shale appears to reflect a subsaline shallow water environment, whereas the higher $\delta^{18}\text{O}$ values of the Bridge Creek Limestone reflect the sea-level rise and deeper marine environment, as supported by sedimentological and mineralogical data (Sageman et al., 1998; this study).

7. Salinity proxy—*Hedbergella planispira*

Stable isotope ranking indicates that *Hedbergella planispira* lived in surface waters (Price and Hart, 2002), as also indicated by its geographic distribution and abundance in shallow epeiric sea environments. In the Western Interior Seaway, *H. planispira* alternately dominated with low oxygen tolerant biserial benthic foraminiferal species (Eicher, 1969b; Eicher and Worstell, 1970; Leckie, 1987; Keller and Pardo, 2004). Assuming that *H. planispira* calcified its shell in equilibrium with seawater, similar to other planktic foraminiferal species (Corfield et al., 1990; Norris and Wilson, 1998; Price and Hart, 2002), and based on the discussion above assuming that a primary signal is still preserved, the cyclic variations in $\delta^{18}\text{O}$ values of this surface dweller record alternating periods of more normal marine salinity with freshwater influx in the Western Interior Seaway (Fig. 9). Hence, negative $\delta^{18}\text{O}$ excursions reflect freshwater influx at times of humid and wet periods with high runoff, which were associated with increased erosion (e.g., quartz maxima, Fig. 8B). Conversely, positively $\delta^{18}\text{O}$ excursions reflect dry periods with runoff, or marine incursions at times of rising sea levels.

Not only does *Hedbergella planispira* reflect salinity variations in $\delta^{18}\text{O}$ values of its shells, but there is also a 91% correspondence (20 out of 22 data points) between high population abundances and 2–4‰ negative $\delta^{18}\text{O}$ shifts that reflects a high tolerance for brackish surface waters (Fig. 9). A similarly high correlation is observed between relatively low *H. planispira* abundances and positive $\delta^{18}\text{O}$ shifts, which indicate a low tolerance for more normal marine surface water salinity. *Hedbergella planispira* is thus a generally excellent proxy for surface salinity based on $\delta^{18}\text{O}$ and high population abundances. The latter does not hold at times when relative abundances of *H. planispira* drop below 15%, signalling more normal surface salinity, as is the case in the *Anaticinella multiloculata* subzone and early Turonian (upper *Helvetoglobaotruncana moremani* to *Helvetoglobaotruncana helvetica* zones). There is also a notable exception in sample 79, where a high *H. planispira* population (35%) coincides with a positive $\delta^{18}\text{O}$ excursion (Fig. 9). The reason for this discrepancy is not clear, but it may be related to higher surface productivity and expansion of the oxygen minimum

zone. The validity of *H. planispira* as a surface salinity proxy can be further tested based on mineralogical data.

8. Terrigenous influx—mineralogy

Subsaline surface waters often result from freshwater influx due to wet periods with high rainfall, and are usually accompanied by increased terrigenous input, such as phyllosilicates and quartz. However, at Pueblo, phyllosilicates are not only of detrital origin, but are derived partly from in situ alteration of bentonite layers by seawater (Chamley, 1989; Weaver, 1989; Hillier, 1995), which has led to the formation of authigenic smectite and zeolite. This is evident in the different phyllosilicate versus quartz patterns of the three major Pueblo lithologies of bentonites, shales and marly limestones; phyllosilicates are enriched in bentonites (mainly smectite) compared with other lithologies) and quartz ($R^2 = 0.1$). However, there is a good correlation if phyllosilicates from bentonite layers are excluded ($R^2 = 0.59$). This indicates that phyllosilicates in shales and marly limestones reflect increased detrital input, whereas those in bentonite layers are largely due to in situ alteration of volcanic sediments. Due to the uncertainty factor introduced by bentonite alteration and subsequent smectite neoformation, we only use quartz and the ratio calcite/(quartz + plagioclase and K-feldspar) to infer detrital variations. Phyllosilicates are therefore excluded from this study.

High quartz and low calcite contents generally correlate with peak abundances of *Hedbergella planispira* and negative $\delta^{18}\text{O}$ shifts (Fig. 11), although this correlation is strongest during the deposition of the Hartland Shale when sea level was lowest and terrigenous influx highest. In this interval, freshwater influx is recorded by 3‰ negative shifts in $\delta^{18}\text{O}$ values, peak abundances of 60–70% *H. planispira*, and 20–30% quartz, but only about 25% calcite. Quartz gradually decreases in the uppermost Hartland Shale and Bridge Creek Limestone, reflecting the rising sea level as also indicated by increased calcite deposition and/or absence of erosion. However, there is also a period of increased freshwater and terrigenous influx in the Bridge Creek Limestone above the $\delta^{13}\text{C}$ shift and benthic oxic zone (bed numbers 74–

81), as indicated by 2% negative shifts in $\delta^{18}\text{O}$ values, peak abundances of 40% *H. planispira*, and 15–20% quartz, but only 25–40% calcite.

If high terrigenous influx of quartz, low calcite, negative shifts in $\delta^{18}\text{O}$, and peak abundances in *Hedbergella planispira* reflect subsaline marine conditions, then the opposite should reflect surface waters with more normal marine salinity. This is generally the case (shaded intervals, Fig. 11), but the correlation breaks down at the onset of the $\delta^{13}\text{C}$ shift (9.5–10.5 m) probably due to the rising sea level and an influx of Tethyan seawater from the south. Notable breakdowns in this correlation also occur in beds 64–65 at the peak of the $\delta^{13}\text{C}$ excursion and onset of the benthic oxic zone, and in bed 85 where low *H. planispira* abundance and a 2‰ negative excursion in $\delta^{18}\text{O}$ values suggest freshwater influx, and high calcite, but low quartz may indicate greater distance from the shoreline due to a sea level transgression.

9. Discussion

The age of the Pueblo section is well constrained based on planktic foraminiferal and ammonite biostratigraphies; the presence of all biozones, and details of the OAE 2 $\delta^{13}\text{C}$ excursion indicate that relatively continuous sediment deposition occurred across the Cenomanian–Turonian transition. These factors support the Pueblo section as an excellent Global Stratotype Section and Point (GSSP) for the C/T boundary (Kennedy et al., 2000).

The rapid $\delta^{13}\text{C}$ shift that marks the onset of the OAE 2 at Pueblo occurred over a period of about 90,000 years, similar to Eastbourne (Keller et al., 2001), and coincided with a major sea-level transgression that culminated in deposition of the basal Bridge Creek Limestone Bed 63 (Fig. 12). The likely cause of this major positive $\delta^{13}\text{C}$ excursion is the global depletion of ^{12}C in the water column as a result of high primary productivity and/or enhanced preservation of organic matter as a result of poorly oxygenated bottom waters globally (Arthur et al., 1985; Pratt, 1985). Both factors may have contributed at Pueblo and are associated with the sea-level transgression and an incursion of warm marine waters with normal salinity. High primary productivity and increased watermass stratification are indicated by the presence

of abundant low-oxygen tolerant planktic foraminifera (heterohelicids) that reveal expansion of the oxygen minimum zone (Leckie et al., 1998; Keller and Pardo, 2004). Enhanced preservation of organic matter in dysoxic or anoxic bottom conditions is indicated by the presence of rare low-oxygen tolerant benthic foraminifera (Eicher, 1969b; West et al., 1998). There is only one short interval of well-oxygenated bottom waters, which spans from above bed 63 to bed 73 (Fig. 10) and probably corresponds to the “benthic zone” of Eicher (1969a). The oxic bottom waters occurred at a time of maximum sea-level transgression, but are associated with subsaline surface waters, and may reflect the incursion of oxygenated deeper waters from the Tethys into the Western Interior Seaway.

The *Rotalipora cushmani* $\delta^{13}\text{C}$ shift is similar to OAE 2 in that it also appears to coincide with a sea-level rise at Pueblo, as suggested by benthic assemblages. This sea-level rise is preceded by freshwater influx, increased terrigenous input, including terrigenous organic matter, and dysoxic or anoxic bottom waters, as indicated by the presence of low-oxygen tolerant benthic species (common buliminellids). Higher planktic species diversity, including the appearance of *Rotalipora* and *Praeglobotruncana* species, indicate deeper water, increased watermass stratification, and enhanced productivity associated with and following the $\delta^{13}\text{C}$ shift. Thus, this smaller $\delta^{13}\text{C}$ shift may have had a similar cause as the OAE 2 $\delta^{13}\text{C}$ shift.

A similar magnitude $\delta^{13}\text{C}$ shift has been documented from the middle Cenomanian *Rotalipora cushmani* zone in sections from Italy, England and Europe by other groups and labeled MCE I and II for mid-Cenomanian events I and II (e.g., (Jenkyns et al., 1994; Paul et al., 1994; Mitchell et al., 1996; Mitchell and Carr, 1998; Jarvis et al., 2001; Coccioni and Galeotti, 2003). This suggests additional $\delta^{13}\text{C}$ shifts occurred in the low to middle latitude Tethys. However, there is some uncertainty in the timing of this event. In the Pueblo and Italian sections the $\delta^{13}\text{C}$ shift appears coeval within the *R. cushmani* zone, whereas ammonite correlations between Pueblo and English sections suggest that at Pueblo this may be a younger event. Further studies are in progress to determine the mid-Cenomanian stable isotope stratigraphy at Pueblo.

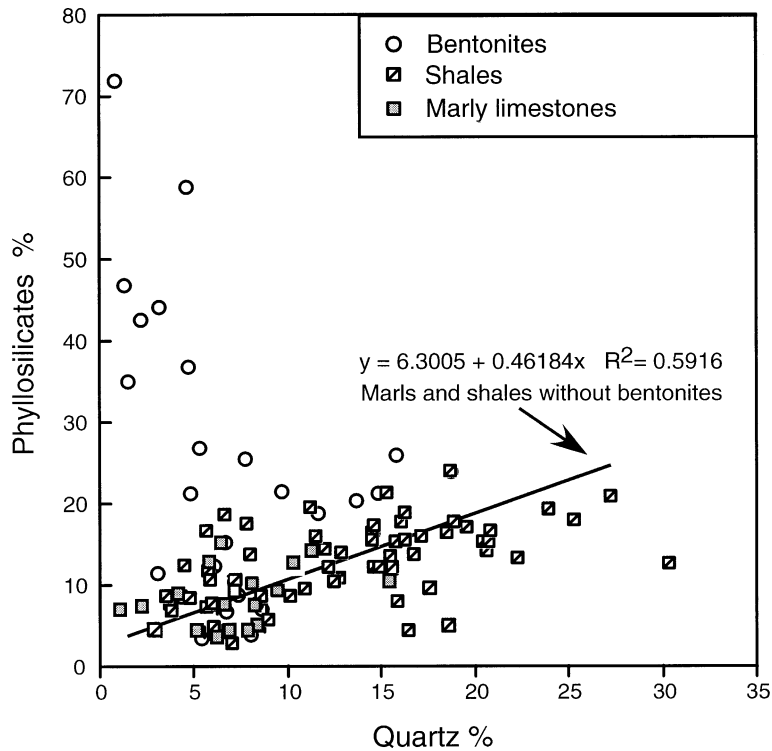


Fig. 10. Cross plot between phyllosilicates and quartz. Note that the three observed lithologies (bentonites, shales and marly limestones) show different patterns, especially for the bentonite and limestone layers. The shales appear to be more variable. There appears to be no correlation between phyllosilicates (in all lithologies) and quartz ($R < 0.1$). However, there is a good correlation if phyllosilicates from bentonite layers are excluded ($R^2 = 0.59$). This indicates an increase in phyllosilicates of shales and marls, which reflects an increase in the detrital input.

9.1. $\delta^{18}O$ salinity and sea level

The cyclic limestone/shale deposition of the lower Bridge Creek Limestone Member of the Pueblo section has long been interpreted as a product of orbital cycles with the obliquity index influencing shale deposition through high latitude precipitation and clastic influx, whereas evaporation and nutrient upwelling of the precessional index controls carbonate productivity (e.g., Pratt, 1985; Pratt et al., 1993; Sageman et al., 1998). This study supports this interpretation based on $\delta^{18}O$ measurements of the surface dweller *Hedbergella planispira*. This species is a proxy for salinity variations based on: (1) increased populations of *H. planispira* that generally measure 2–4‰ more negative $\delta^{18}O$ values than reduced populations; (2) peak abundances of this species and peak negative $\delta^{18}O$ values that correlate with high clastic influx; (3) low abundances of *H.*

planispira and high $\delta^{18}O$ values that correlate with decreased clastic influx, increased calcite deposition and high calcite/detritus ratios (Fig. 11).

Diagenesis can only partly explain the unusually depleted $\delta^{18}O$ values (–5‰ to –10‰ and –4‰ to –7‰ for the Hartland Shale and Bridge Creek, respectively) or the cyclic fluctuations, and cannot explain the strong negative covariance between $\delta^{18}O$ values and peak *Hedbergella planispira* abundances. We interpret the low $\delta^{18}O$ values and corresponding high *H. planispira* abundances as indicative of subsaline surface waters as a result of freshwater influx during wet humid periods, whereas the opposite reflect more normal marine salinity and more arid periods. The strong correlation between *H. planispira* abundances and $\delta^{18}O$ variations indicates that either can be used as proxy for salinity variations. Surface salinity variations can thus be inferred from the magnitude of the short-term $\delta^{18}O$ variations, or abun-

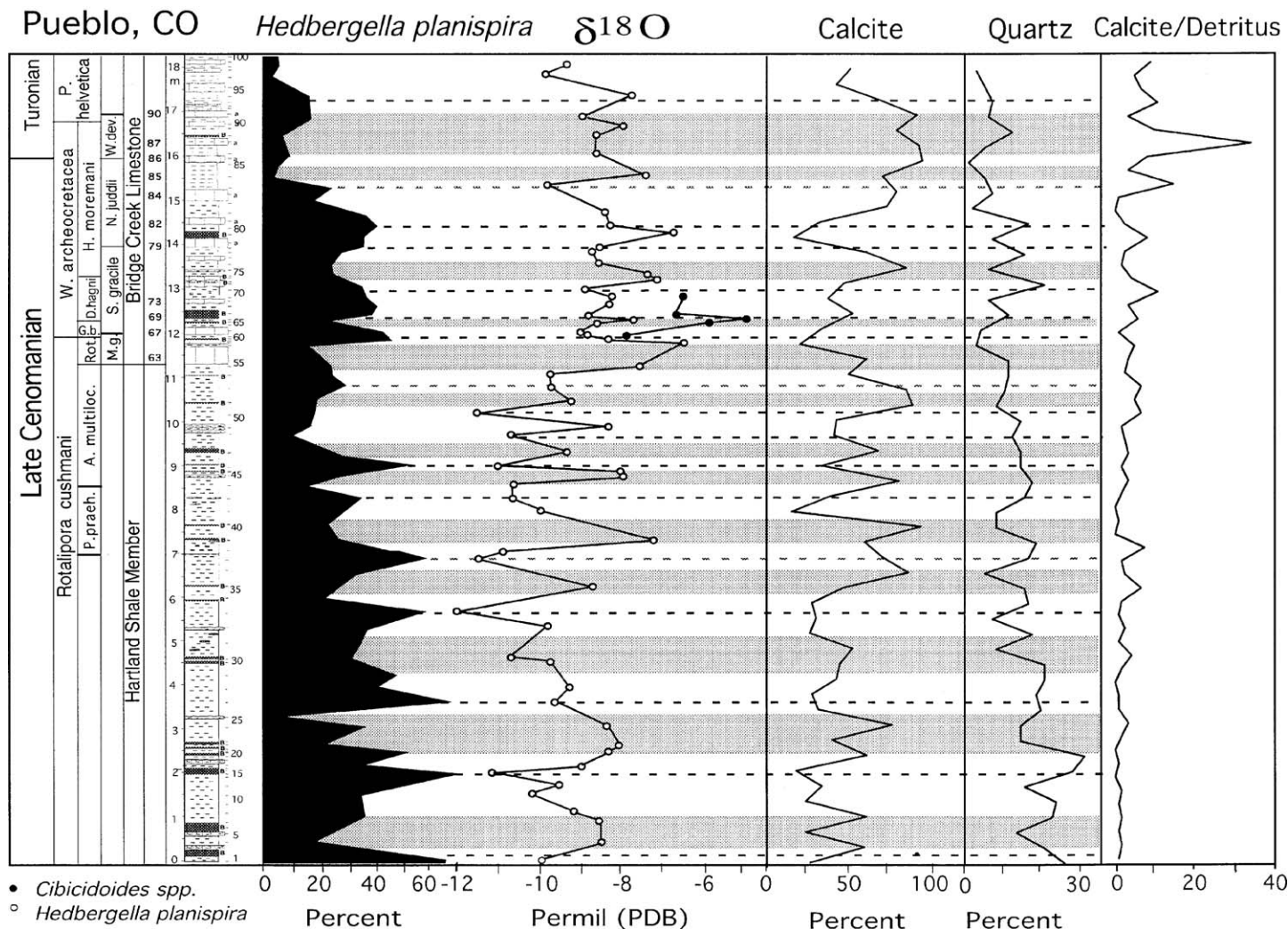


Fig. 11. Calcite, quartz and the calcite/detritus ratio compared with the $\delta^{18}\text{O}$ curve and *Hedbergella planispira* abundance at the Pueblo. Note that peak positive $\delta^{18}\text{O}$ values and low abundances of *H. planispira* broadly correlate with peak calcite deposition and high calcite/detritus ratios. This indicates that at times of more normal marine salinity productivity increased resulting in carbonate deposition. This increase is most notable during the transgression of the Bridge Creek Limestone. Quartz input is generally high during deposition of the Hartland Shale due to a lower sea level, and decreased with the sea-level transgression of the Bridge Creek Limestone.

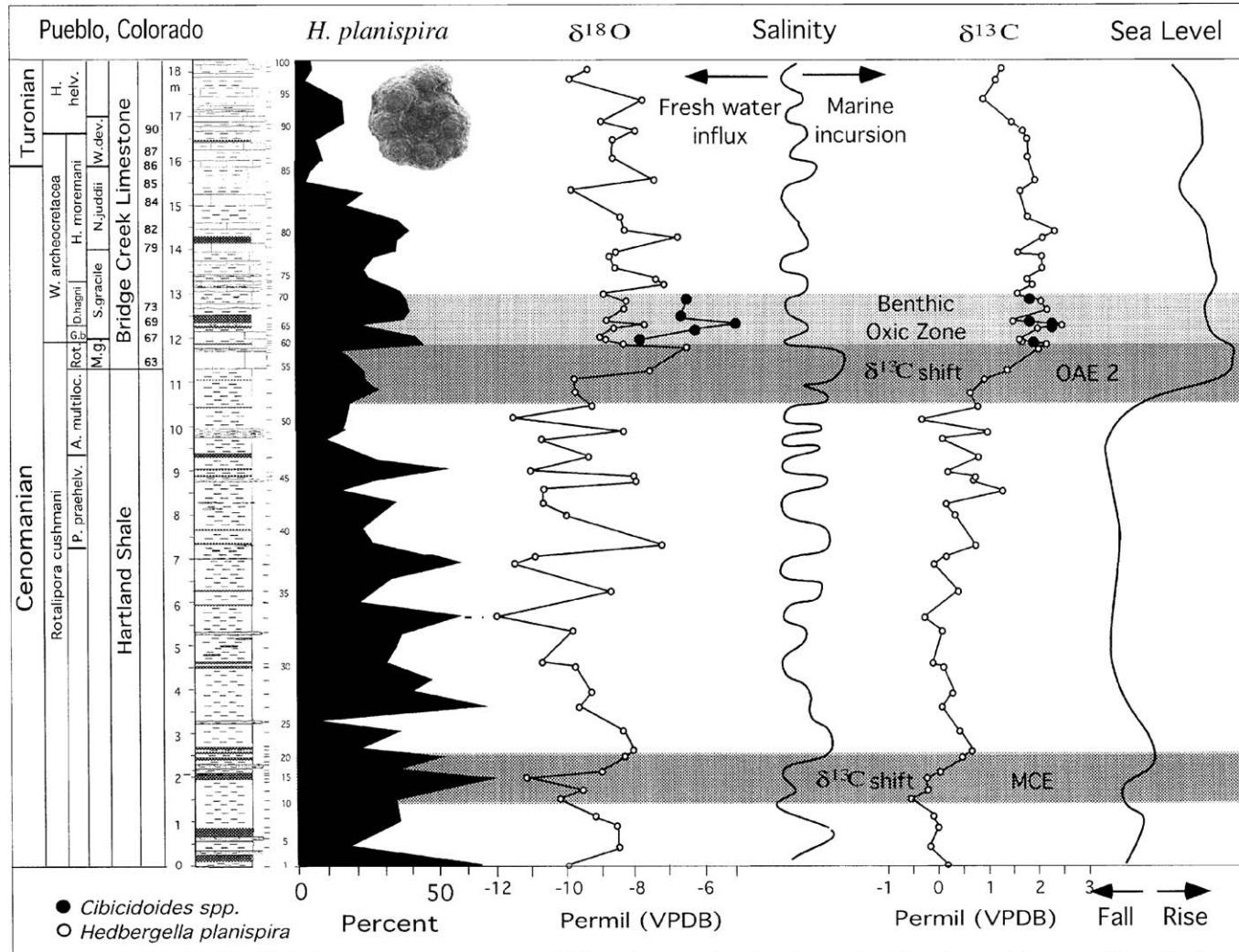


Fig. 12. Salinity and sea-level variations at Pueblo inferred from the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves and *Hedbergella planispira* abundances. Note that low $\delta^{13}\text{C}$ values in the Hartland Shale and high $\delta^{18}\text{O}$ variations indicate low sea-level and high salinity variations. A deeper water environment with more normal marine conditions and lower salinity variations is indicated for the Bridge Creek Limestone beginning with the positive $\delta^{13}\text{C}$ shift (sea level transgression) accompanied by lower $\delta^{18}\text{O}$ variations. Sea-level transgressions at the MCE and OAE2 $\delta^{13}\text{C}$ shifts are supported by the eustatic sea-level curve of Hardenbol et al. (1998).

dance fluctuations in *H. planispira*, as shown in Fig. 12. Salinity variations were significantly greater during the deposition of the Hartland Shale due to the shallower sea and proximity to shoreline, but may also reflect increased rainfall and higher freshwater influx. The magnitude of salinity variations in the Bridge Creek Limestone is about half that of the Hartland Shale and probably reflects mainly the sea-level transgression and marine incursion from the Tethys, as well as the greater distance to the shoreline, though overall more arid conditions cannot be ruled out.

If the unusually negative short-term variations in $\delta^{18}\text{O}$ values reflect low salinity due to freshwater influx, then the long-term trend in $\delta^{18}\text{O}$ reveals sea-level changes in the shallow epicontinental sea of the Western Interior Seaway. Overall low $\delta^{18}\text{O}$ values reflect the subsaline waters of a shallow sea, proximity to shoreline, and greater freshwater influx and clastic debris, whereas overall high values should reflect the opposite. The long-term $\delta^{18}\text{O}$ trend thus indicates a generally low sea-level during deposition of the Hartland Shale, a rapid rise at the top with the peak transgression in limestone bed marker 63 at the base of the Bridge Creek Limestone. The overall higher $\delta^{18}\text{O}$ values of the Bridge Creek Limestone indicate the generally high though fluctuating sea level (Fig. 12). Similar sea-level trends have been identified in previous studies based on palaeontological, mineralogical and sedimentological investigations (e.g., Hancock and Kauffman, 1979; Kauffman, 1984; Arthur et al., 1985; Pratt et al., 1993; Leckie et al., 1998; Jarvis et al., 2001).

10. Conclusions

- (1) The rapid $\delta^{13}\text{C}$ increase that marks the onset of the OAE 2 at Pueblo occurred over a period of about 90,000 years, and coincided with the major sea-level transgression that culminated in the deposition of the basal Bridge Creek Limestone. The likely cause for the OAE 2 is depletion of ^{12}C in the water column as a result of high primary productivity and enhanced preservation of organic matter as a result of poorly oxygenated bottom waters.
- (2) A positive $\delta^{13}\text{C}$ shift also occurred in the *Rotalipora cushmani* zone. This mid-Cenoma-

nian event also coincided with a sea-level rise and is associated with increased input of terrigenous organic matter and enhanced preservation in anoxic or dysoxic bottom waters. Further investigation is necessary to determine whether this $\delta^{13}\text{C}$ event is coeval with the Mid-Cenomanian Event identified in northern Europe.

- (3) $\delta^{18}\text{O}$ variations of the surface dwelling planktic foraminifera *Hedbergella planispira* reflect salinity variations despite a 1.5% to 2‰ diagenetic overprint. Low $\delta^{18}\text{O}$ values reflect periods of subsaline surface waters as a result of freshwater influx during wet humid periods accompanied by increased clastic transport. High $\delta^{18}\text{O}$ values reflect more normal marine salinity as a result of arid periods and/or marine incursions accompanied by increased biogenic carbonate deposition.
- (4) *Hedbergella planispira* is a proxy for salinity variations as indicated by 2–4‰ lower $\delta^{18}\text{O}$ values in intervals of peak abundances as compared with reduced populations. This species thrived in subsaline surface waters of epicontinental seas.

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