

# Utility of organic carbon isotope data from the Salina Group halite (Michigan Basin): A new tool for stratigraphic correlation and paleoclimate proxy resource

Andrew H. Caruthers<sup>1,†</sup>, Darren R. Gröcke<sup>2</sup>, Stephen E. Kaczmarek<sup>1</sup>, Matthew J. Rine<sup>1</sup>, Jeff Kuglitsch<sup>3</sup>, and William B. Harrison III<sup>1</sup>

<sup>1</sup>Department of Geological and Environmental Sciences, Western Michigan University, 1903 West Michigan Avenue, Kalamazoo, Michigan 49008-5241, USA

<sup>2</sup>Department of Earth Sciences, Durham University, South Road, Durham DH1 3LE, UK

<sup>3</sup>Department of Earth Sciences, North Carolina Wesleyan College, 3400 North Wesleyan Boulevard, Rocky Mount, North Carolina 27804-9906, USA

## ABSTRACT

Long-term global carbon isotope records ( $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$ ) for the Silurian have been largely derived from unrestricted open-marine carbonates and shales. Here, we demonstrate how organic carbon harvested from halite-dominated evaporite deposits in a restricted intracratonic basin can be used to produce a carbon isotope record. Inorganic and organic carbon isotope data were generated and compared from four subsurface cores from the Silurian Michigan Basin, representing unrestricted carbonate and restricted evaporite/carbonate deposition. The  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  records exhibit a number of long-term trends and major carbon isotope excursions (CIE) that are correlated with the globally identified Ireviken, Mulde, and Linde events. These data provide temporal and stratigraphic constraints in rocks where paleontological data are sparse or absent. They also potentially highlight the effect of enhanced local evaporation on isotope fractionation. This new technique for generating a long-term organic carbon isotope profile from Silurian halite sequences, which can be correlated to the global curve, is of broad interest to the geoscience and paleoclimate science communities. These data not only provide a valuable tool for understanding the chronostratigraphic framework within an evaporative interior basin, but they also provide a rare temporal link between periods of prolonged evaporite deposition and events of known paleoclimate change.

<sup>†</sup>andrew.caruthers@wmich.edu

## INTRODUCTION

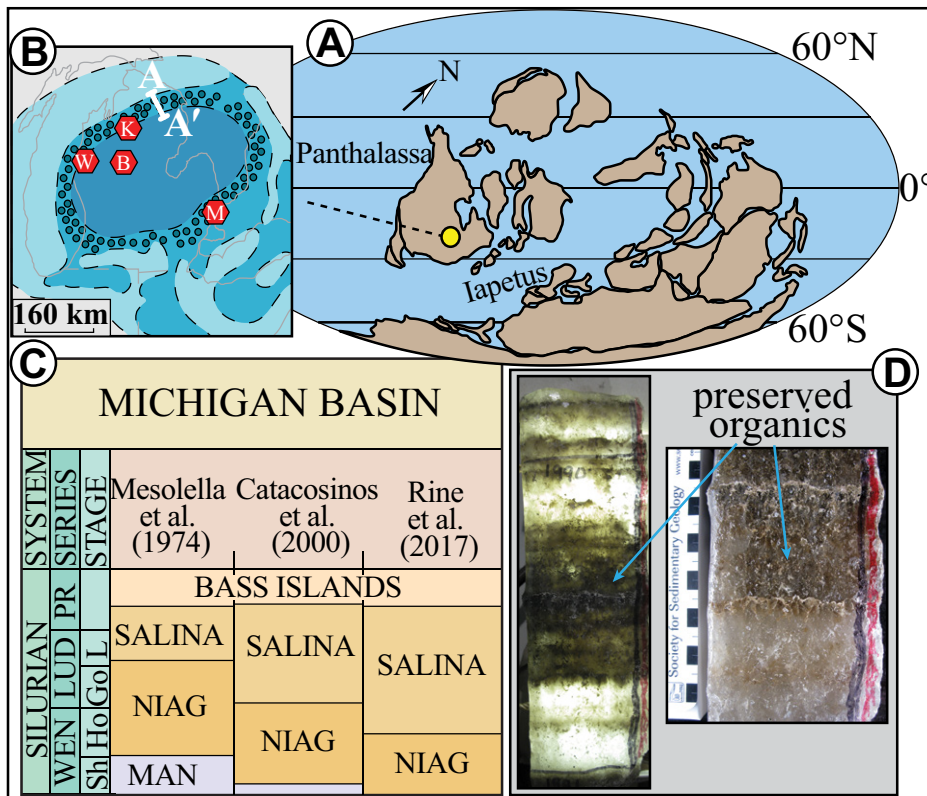
Evaporative basins are an integral part of science and society. They provide reservoir seals for over half of the world's petroleum reserves (Sarg, 2001, and references therein), and they are commonly mined for a variety of manufacturing and industrial applications (Warren, 2016). In an academic sense, large halite-dominated evaporative basins are especially important because they provide a potential rare link among the ancient atmosphere, hydrosphere, and lithosphere during long-term paleoclimate change (e.g., Fanlo and Ayora, 1998; Warren, 2016). However, in order to assess the potential effects of prolonged evaporation on the local paleoenvironment, it is necessary to first constrain the depositional timing of large-scale evaporation events.

One of the biggest obstacles in halite-prone sequences is obtaining reliable age constraint, because nearly all biostratigraphically important marine organisms are typically absent. In highly evaporative settings, it is also not possible to rely on inorganic sources, such as  $\delta^{13}\text{C}_{\text{carb}}$ , to compare with long-term global trends, because most halites contain very little associated carbonate (Warren, 2016). As such,  $\delta^{13}\text{C}_{\text{carb}}$  can, at best, provide an incomplete isotope record. Without temporal constraint in evaporite sequences, it is difficult (1) to establish stratigraphic correlations within and outside the basin, and (2) to discern the effects of changing global versus local climate conditions. These challenges become surmountable if temporally constrained and uniquely global chemostratigraphic signatures can be identified within the basin, thereby establishing a means of comparison with the global record.

In this study, we used a multiproxy data set approach. Local  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  records derived from normal marine carbonate and halite-prone deposits in the Silurian Michigan Basin (Fig. 1) were compared to an established global inorganic carbon isotope curve in order to establish depositional age constraints for the onset of halite deposition. During deposition, organic matter was trapped and preserved as inclusions within the halite (Fig. 1D). By collecting this organic material at regular intervals, it should be possible to produce a long-term geochemical record for the basin. The Salina Group halite of the Michigan Basin is an ideal candidate with which to test this approach, because it is thought to be structurally intact and it does not display large-scale doming, or postdepositional salt migration (Mesoletta et al., 1974; Cercone, 1984).

## SILURIAN PALEOCLIMATE

The Silurian was a period of intense climatic instability dominated by highly variable ocean temperatures, strong swings in eustasy, and frequent biotic turnover (Jeppsson, 1990; Samtleben et al., 1996; Azmy et al., 1998; Munnecke et al., 2003; Kaljo et al., 2003; Lehnert et al., 2010; Noble et al., 2012; McAdams et al., 2017; and others), as evidenced by eight biostratigraphically constrained carbon isotope excursions (CIEs; Cramer et al., 2011, 2015, and references therein). From oldest to youngest, these include the early Aeronian, late Aeronian, Valgu, Ireviken, Mulde, Linde, Lau, and Klunk events. Most CIEs range 2‰ to 4‰  $\delta^{13}\text{C}_{\text{carb}}$  relative to background values, with the Lau event ranging 7‰ to 12‰. Comparisons of these CIEs with small-scale extinction and recovery events exhibit variable degrees of correlation,



**Figure 1.** (A, B) Silurian paleogeography and schematic map of the Michigan Basin with location of cores at: K (State Kalkaska #2–15); B (Bruske #1–26); W (Weinert #2–6); and M (DC-Mead #1). Dark-blue circles—locations of pinnacle reefs (from Briggs et al., 1980). (C) Silurian chronostratigraphies for the Michigan Basin showing discrepancy in depositional timing for Niagara and Salina Groups. PR—Pridoli, LUD—Ludlow, WEN—Wenlock, L—Ludfordian, Go—Gorstian, Ho—Homerian, Sh—Sheinwoodian, NIAG—Niagara, MAN—Manistique. (D) Precipitated organic matter in halite from the A-2 Evaporite, DC-Mead #1 core at ~190.0 ft (58 m). Scale for D is in cm.

suggesting that no single repetitive set of paleoclimatic circumstances and/or controlling mechanism can explain these events (Munnecke et al., 2010; Noble et al., 2012; Cooper et al., 2014; Jarochowska and Munnecke, 2015; Trotter et al., 2016).

Until now, carbon isotope data ( $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$ ) for the Silurian have been derived primarily from marine carbonates (Cramer et al., 2011, and references therein) and, to a lesser extent, shales (Porębska et al., 2004; Noble et al., 2005, 2012; Lenz et al., 2006; Frýda and Štorch, 2014). We are unaware of any published study where such data were extracted from Silurian halite-dominated evaporite deposits to assess long-term trends in the organic carbon isotope record. Procurement of coeval carbon isotope data derived from both inorganic carbonate and organic matter offers an opportunity to compare the overall expression and timing of these events using a proxy that is influenced by highly variable causes of isotopic fractionation, and it also

offers the potential to assess variation in atmospheric  $p\text{CO}_2$  (i.e., changes in  $\Delta^{13}\text{C}$ ; e.g., Kump and Arthur, 1999). In a study focusing on the timing of the “two-prong” Mulde event in Arctic Canada, Noble et al. (2005, 2012) showed variability in CIE magnitude not only between adjacent stratigraphic sections, but also between the organic and inorganic carbon isotope records. This type of interrecord variability highlights the need to evaluate long-term trends and variations in  $p\text{CO}_2$  in these two systems in a variety of geographic locations and paleoenvironments.

## GEOLOGICAL SETTING

The Michigan Basin is a tectonically stable bowl-shaped intracratonic depression in North America (Fig. 1A) that contains ~5 km (16,000 ft) of Cambrian- to Pennsylvanian-age sedimentary strata (Howell and van der Pluijm, 1999; Catacosinos et al., 2000). Throughout the Silurian, the Michigan Basin was located at low-

lying (i.e., tropical) latitudes in the central part of the Laurentian Continental Seaway, which was connected to the Iapetus Ocean to the south and the Panthalassan Ocean to the west and covered much of present-day North America (Fig. 1A). During Wenlock (early Sheinwoodian–Homerian) time, elaborate barrier and pinnacle reef complexes developed along the margin and almost completely rimmed the basin (Fig. 1B). This barrier and pinnacle reef system comprises the upper part of the Niagara Group (Guelph Formation), which altogether ranges in thickness from 39 m (130 ft) to 213 m (700 ft) on the basin margin.

The Niagara barrier and pinnacle reef system is capped by up to 762 m (2500 ft) of interbedded anhydrite, halite (Fig. 1D), and organic-rich carbonate mudstone of the Salina Group, representing a major change in paleoenvironment and local sea level (Mesolella et al., 1974; Sullivan et al., 2016). The Salina Group is divided into repeating sequences of evaporites, carbonates, and shales identified as Salina units A-0 to G (Mesolella et al., 1974). Within this group, non-evaporite carbonate and shale units (e.g., A-0, A-1C, A-2C, C, E, and G) are known to have greater lateral continuity than evaporite-dominant units (Mesolella et al., 1974; Harrison and Voice, 2017).

Because the primary focus of this study centers on the subject of a new global correlation proxy, it is important to briefly discuss some confusing terminology in the local (i.e., sub-surface Michigan Basin) versus regional (i.e., eastern United States) stratigraphic assignment of Silurian units. Originally, the term Lockport was designated as a group-level term in the state of New York (Brett et al., 1995; fig. 1 in Cramer et al., 2011), whereas in the Michigan Basin, it is used as the formation level (e.g., formally as the Lockport Dolomite Formation in Catacosinos et al., 2000, 2001; or informally as the Lockport Formation). Similarly, the term Niagaran is a regional time designation in North America, temporally equivalent to the combined Llandovery and Wenlock Epochs (fig. 1 in Cramer et al., 2011). In the Michigan Basin, the term Niagara is used as a group-level designation that is only partially temporally equivalent to the Niagaran Series (see Catacosinos et al., 2000, 2001). The term Niagara has also been used as an informal designation for stratigraphic units within the Niagara Group (e.g., informally designated as “Gray” or “Brown” Niagara to correspond to the Lockport Formation and Guelph Formation, respectively, in Catacosinos et al., 2000, 2001). Until this confusing terminology is clarified, it is important to use the official stratigraphic terminology of Michigan, established by Catacosinos et al. (2000, 2001).

## Biostratigraphic Constraint

The base of the Salina Group was originally defined in New York State as the contact between the Vernon Shale and the underlying Guelph Dolomite, occurring at the Niagara-Cayugan boundary (fig. 4 in Brett et al., 1995). On the southern and western margins of the Michigan Basin, this lithological transition appears older, constrained to the upper Homerian (Wenlock) by the conodonts *Ozarkokina bohémica longa* and *Pseudooneotodus linguicornis* (NE Indiana and Ohio; Kleffner and Rexroad, 1999; Cramer, 2009; Swift, 2011; Kleffner et al., 2012) and by the onset of the Mulde event CIE (NE Indiana, Ohio, and Wisconsin; Cramer, 2009; Swift, 2011; Kleffner et al., 2012; Sullivan et al., 2016). However, this apparent difference in stratigraphic position for the base of the Salina Group between the eastern United States and the Michigan Basin area is not an implied diachroneity, rather, it is an indication that this lithological transition is complex and is less precisely known in the New York area.

In the Michigan Basin center, the base of the Salina Group is marked by the sudden change from thriving reef complexes of the Niagara Group to immense halite-dominant evaporite deposition of the Salina Group (Mesoellela et al., 1974; Catacosinos et al., 2000; Rine et al., 2017). Although the depositional timing of this lithologic transition between the Michigan Basin center and its margin is inferred through stratigraphic correlation, to date there have been no high-resolution geochemical studies and very few diagnostic fossils recovered from this uppermost Niagara and Salina Group transition in the Michigan Basin. Thus making direct age correlation difficult (Fig. 1C).

A conodont analysis of rocks lower in the Niagara Group by Kuglitsch (2013) identified *Kockelella walliseri* stratigraphically above co-occurring elements of *Kockelella ranuliformis* and *Ozarkodina sagitta rhenana* from the lower Lockport Formation. This suggests that platform carbonates (predating reef deposition) were deposited during the *O. s. rhenana* and lower *K. walliseri* zones of the middle Sheinwoodian (Wenlock). Mesoellela et al. (1974) also identified Wenlock brachiopods *Plicocyrta cf. arkansana* and *Pentamerus* sp. from the basal Salina Group. Although brachiopods are typically known to have longer stratigraphic ranges than conodonts, and they are therefore less valuable for age correlation, the presence of both fossil groups indicates a Wenlock age for the pre-Salina Group sediments. These observations suggest that the transition to evaporite deposition potentially occurred within the Wen-

lock, after the lower *K. walliseri* zone, which generally coincides with the known timing on the basin margin.

## SILURIAN $\delta^{13}\text{C}_{\text{ORG}}$ DATA FROM THE MICHIGAN BASIN

### Methodology

In total, 433 samples were collected and analyzed for  $\delta^{13}\text{C}_{\text{org}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  from the Miller Brothers-Weinert #2–6 (Permit #37731), Northern Michigan Exploration Corp–State Kalkaska #2–15 (Permit #29218), Roadrunner-Bruske #1–26A (Permit #59271), and Diamond Crystal Salt Co.–Saint Clair River Associates, a.k.a. DC-Mead #1 (Permit #28135), cores, representing the greater Niagara–middle Salina Group stratigraphy in the Michigan Basin (Figs. 2–5). In the Weinert #2–6 and State Kalkaska #2–15 cores, inorganic and organic carbon was analyzed from the Lockport and Guelph Formations of the Niagara Group, and in the State Kalkaska #2–15 core, only organic carbon was analyzed from the A-0C to A-1C units of the Salina Group. In the Bruske #1–26A core, inorganic and organic carbon was analyzed from the uppermost A-1E to A-1C units of the Salina Group (only organic carbon was analyzed from the uppermost A-1E unit). In the DC-Mead #1 core, only organic carbon was analyzed from the A-1C to C units of the Salina Group.

All organic carbon isotope samples were analyzed at the stable isotope biogeochemistry laboratory at Durham University, Durham, UK, and samples for inorganic carbon isotopes were analyzed at either the stable isotope laboratory at the University of Michigan (Weinert #2–6 and Bruske #1–26A well cores) or at the laboratory at Durham University (State Kalkaska #2–15 well core).

### Stable Isotope Laboratory, University of Michigan

Samples weighing a minimum of 10  $\mu\text{g}$  were placed in stainless-steel boats. Samples were then transferred to individual borosilicate reaction vessels and reacted at  $77\text{ }^\circ\text{C} \pm 1\text{ }^\circ\text{C}$  with 4 drops of anhydrous phosphoric acid in a Finnigan MAT Kiel IV preparation device coupled directly to the inlet of a Finnigan MAT 253 triple collector–isotope ratio mass spectrometer (IRMS). Limestones, dolomites, apatite, and siderite were reacted for 8, 12, 17, and 22 min, respectively.  $\text{O}^{17}$  data were corrected for acid fractionation and source mixing by calibration to a best-fit regression line defined by two standards (NBS 18 and NBS 19). All stable isotope data are reported in standard delta ( $\delta$ ) notation in per mil ( $\text{‰}$ ) relative to the Vienna Peedee

belemnite (VPDB) scale. Precision and accuracy were monitored through daily analysis of a variety of powdered carbonate standards, with at least four standards reacted and analyzed daily. Measured precision was maintained at better than 0.1 $\text{‰}$  for both carbon and oxygen.

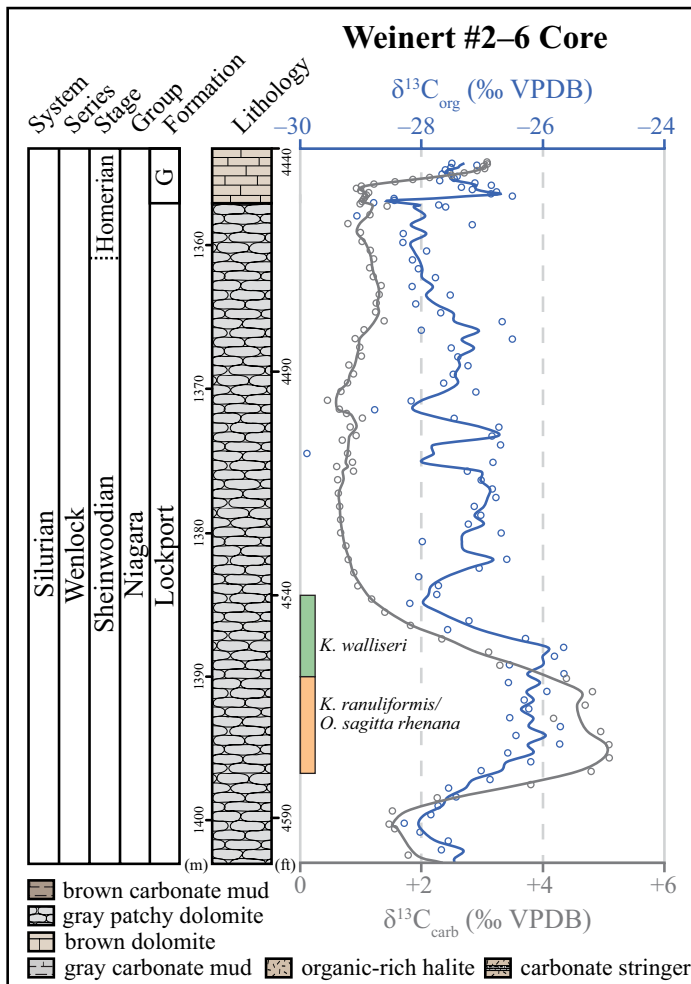
### Stable Isotope Biogeochemistry Laboratory, Durham University

Inorganic and organic  $\delta^{13}\text{C}$  data were obtained following the methods of Skrzypek and Debajyoti (2006) and Gröcke et al. (2009), respectively. Acquisition of  $\delta^{13}\text{C}_{\text{org}}$  data from halite included an additional desalination step. Approximately 200 mg aliquots of sediment were dissolved in deionized water for 1 h and centrifuged, and the supernatant was discarded. Wet samples were decalcified using 3 M HCl, left overnight at room temperature. Samples were then neutralized using deionized water and dried in an oven at 55  $^\circ\text{C}$ . Sample sets from the Bruske #1–26A and DC-Mead #1 cores were treated with hot HCl and then reanalyzed for  $\delta^{13}\text{C}_{\text{org}}$  in an effort to assess secondary effects of dolomitization (red text in supplemental data file<sup>1</sup>). The  $\delta^{13}\text{C}_{\text{org}}$  data were acquired using a Costech Elemental Analyzer (ECS 4010) coupled to a Thermo Finnigan Delta V Advantage IRMS. Carbon isotope ratios were corrected for  $^{17}\text{O}$  contribution and reported in standard delta ( $\delta$ ) notation in per mil ( $\text{‰}$ ) relative to the VPDB scale. Data accuracy was monitored through routine analyses of in-house standards, which were calibrated against international standards (USGS 40, USGS 24, IAEA 600, IAEA CH6): This provided a linear range in  $\delta^{13}\text{C}$  between +2 $\text{‰}$  and –47 $\text{‰}$ . Analytical uncertainty for  $\delta^{13}\text{C}_{\text{org}}$  was typically  $\pm 0.1\text{‰}$  for replicate analyses of the international standards and typically <0.2 $\text{‰}$  on replicate sample analysis.

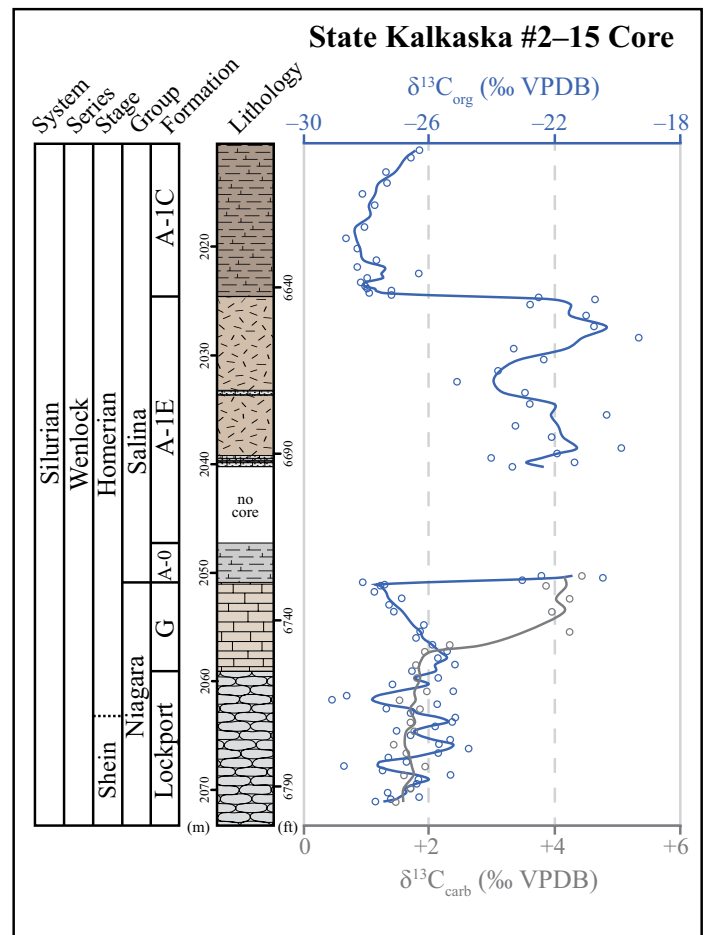
## RESULTS

Numerous time-correlative excursions are recorded within the inorganic and organic carbon isotope data. These excursions exhibit the same direction of offset, and they appear in multiple wells. In the Weinert #2–6 core, a positive CIE of 2 $\text{‰}$  ( $\delta^{13}\text{C}_{\text{org}}$ ) and 2.5 $\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) occurs from 4590 ft to 4540 ft (1399 m to 1384 m) (Fig. 2). Above the CIE, from 4540 ft to 4465 ft (1384 m to 1361 m) in the core,  $\delta^{13}\text{C}_{\text{carb}}$  values are consistent at approximately +1 $\text{‰}$ , and  $\delta^{13}\text{C}_{\text{org}}$  values show several smaller 1 $\text{‰}$ –1.5 $\text{‰}$  positive CIEs in the upper part of the Lockport Formation

<sup>1</sup>GSA Data Repository item 2018153, organic carbon isotope data from the Salina Group halite (MI Basin) USA, is available at <http://www.geosociety.org/datarepository/2018> or by request to [editing@geosociety.org](mailto:editing@geosociety.org).



**Figure 2.** Niagara Group (Lockport and lower Guelph Formations)  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  data from the Weinert #2-6 core, Michigan Basin, presented as a three-point moving average (solid line). Open circles represent actual data points. G—Guelph; VPDB—Vienna Peedee belemnite standard; *K.*—*Kockellella*; *O.*—*Ozarkodina*.



**Figure 3.** Niagara (Lockport and Guelph Formation) and Salina (A-0C, A-1E, and A-1C units) Group  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  data from the State Kalkaska #2-15 core, Michigan Basin, presented as a three-point moving average (solid line). Open circles represent actual data points. See Figure 2 for lithology key. G—Guelph; VPDB—Vienna Peedee belemnite standard.

that range between  $-28\text{‰}$  and  $-27\text{‰}$ . From 4465 ft to 4440 ft (1361 m to 1353 m), there is an abrupt and slightly offset  $+2\text{‰}$  CIE in  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  values that occurs near the base of the Guelph; the CIE in  $\delta^{13}\text{C}_{\text{carb}}$  occurs slightly above the CIE in  $\delta^{13}\text{C}_{\text{org}}$ .

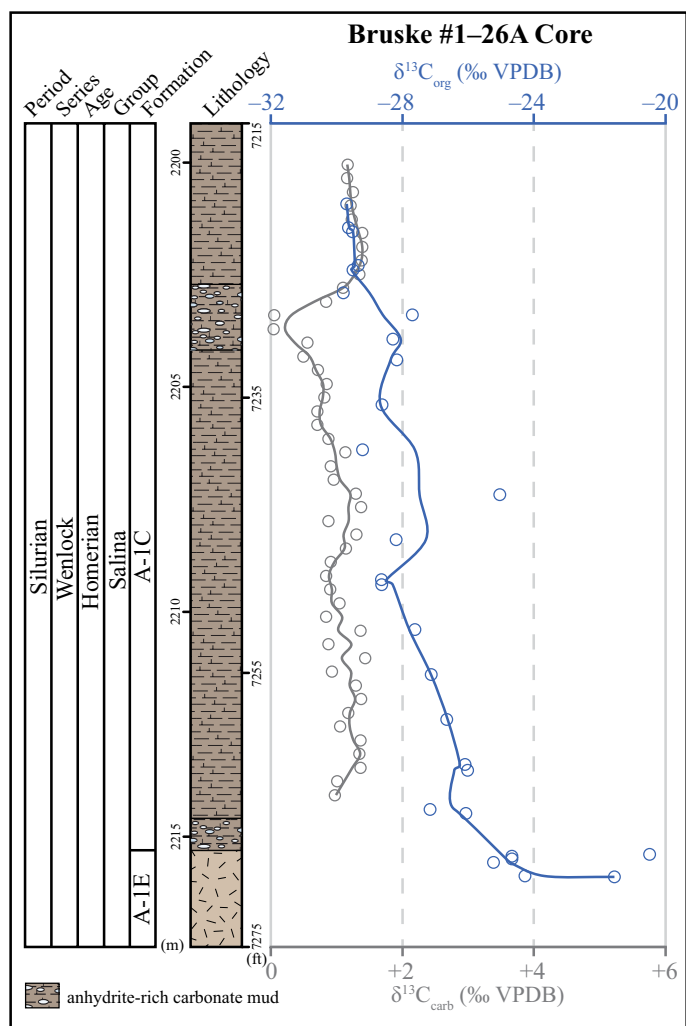
In the State Kalkaska #2-15 core, several  $3\text{‰}$ – $3.5\text{‰}$  CIEs in  $\delta^{13}\text{C}_{\text{org}}$  occur from 6792 ft to 6740 ft (2070 m to 2054 m), spanning the upper Lockport and lowermost Guelph Formations (Fig. 3). There is a  $+2\text{‰}$  CIE in  $\delta^{13}\text{C}_{\text{carb}}$  that occurs in the lowermost Guelph Formation at 6752 ft (2058 m) and extends to the boundary between the Niagara and Salina units. From 6732 ft to 6640 ft (2052 m to 2024 m), there is a large, abrupt,  $+8\text{‰}$  CIE in  $\delta^{13}\text{C}_{\text{org}}$  that begins in the lowermost Salina Group (A-0 unit), extends throughout the overlying A-1E unit, and

terminates in the lowermost A-1C unit. Within this large positive CIE,  $\delta^{13}\text{C}_{\text{org}}$  values show a high degree of variability, fluctuating  $2\text{‰}$ – $3\text{‰}$ , with maximum values reaching  $-20\text{‰}$ . Above this large positive CIE, from 6640 ft (2024 m) to the top of the core,  $\delta^{13}\text{C}_{\text{org}}$  values show a broad positive trend from  $-28\text{‰}$  to  $-26\text{‰}$  throughout the A-1C unit. Samples treated with hot HCl showed no change in  $\delta^{13}\text{C}_{\text{org}}$  values.

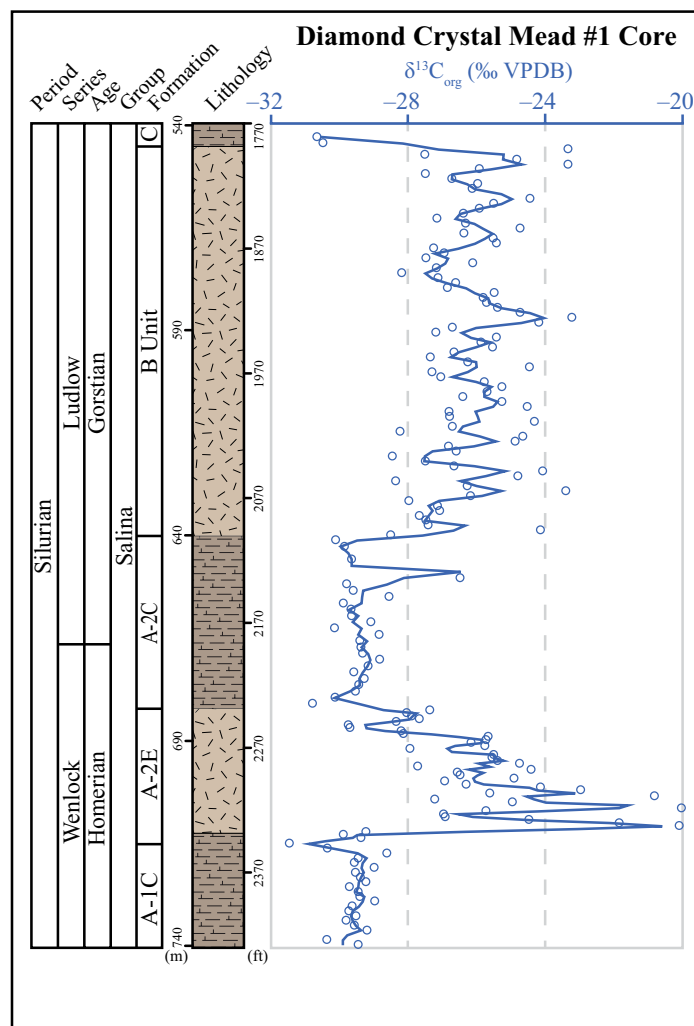
In the Bruske #1-26A core,  $\delta^{13}\text{C}_{\text{org}}$  values show a returning limb of a large positive CIE ( $-21\text{‰}$  to  $-27\text{‰}$ ) at 7272 ft (2217 m) that crosses the A-1E and A-1C unit boundary (Fig. 4). Above this returning limb,  $\delta^{13}\text{C}_{\text{org}}$  values record a steady, slightly negative, trend that spans the A-1C unit to 7218 ft (2200 m), with values that reach  $-29\text{‰}$ . From 7264 ft to 7218 ft (2214 m to 2200 m),  $\delta^{13}\text{C}_{\text{carb}}$  values are steady at

$+1\text{‰}$ , with a small  $0.5\text{‰}$  to  $+1\text{‰}$  negative CIE at 7232 ft (2204 m) of the A-1C unit, correlative with the appearance of anhydrite.

In the DC-Mead #1 core, there is a large, abrupt, positive  $10\text{‰}$  CIE in  $\delta^{13}\text{C}_{\text{org}}$  from 2349 ft to 2212 ft (716 m to 674 m), where values reach  $-20\text{‰}$  in the lowermost A-2E unit and then show a steady return to consistent values of approximately  $-30\text{‰}$  in the lower A-2C unit (Fig. 5). In the upper A-2C unit, there is a broad positive  $6\text{‰}$  CIE in  $\delta^{13}\text{C}_{\text{org}}$  that begins at 2135 ft (651 m) and spans the Salina B unit to 1784 ft (544 m), terminating very close to the contact with the overlying C unit. Organic carbon isotope values show a high degree of variation throughout the B unit, frequently fluctuating between approximately  $-27\text{‰}$  and  $-25\text{‰}$   $\delta^{13}\text{C}_{\text{org}}$ .



**Figure 4.** Salina Group (A-1E and A-1C unit)  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  data from the Bruske #1–26A core, Michigan Basin, presented as a three-point moving average (solid line). Open circles represent actual data points. See Figure 2 for lithology key. VPDB—Vienna Pee Dee belemnite standard.



**Figure 5.** Salina Group (A-1C, A-2E, A-2C, B, and C unit)  $\delta^{13}\text{C}_{\text{org}}$  data from the DC-Mead #1 core, Michigan Basin, presented as a three-point moving average (solid line). Open circles represent actual data points. See Figure 2 for lithology key. C—C unit; VPDB—Vienna Pee Dee belemnite standard.

## DISCUSSION

Carbon isotope data from the Michigan Basin show four large positive CIEs in the  $\delta^{13}\text{C}_{\text{org}}$  records from the Weinert #2–6, State Kalkaska #2–15, Bruske #1–26A, and DC-Mead #1 cores. Two of these CIEs are also identified in  $\delta^{13}\text{C}_{\text{carb}}$  records from the Weinert #2–6 and State Kalkaska #2–15 cores. Figure 6 presents data from the four wells in a single composite isotope curve. The four large positive CIEs include: co-eval  $\sim 3\text{‰}$ – $5\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) and  $\sim 3\text{‰}$ – $4\text{‰}$  ( $\delta^{13}\text{C}_{\text{org}}$ ) CIEs in the lower Lockport Formation; two successive  $\sim 8\text{‰}$ – $10\text{‰}$  ( $\delta^{13}\text{C}_{\text{org}}$ ) CIEs that span the upper Niagara Guelph Formation to lower Salina A-2 Evaporite (onset of the lower CIE in the Guelph Formation is also evident as a  $2\text{‰}$  perturbation in  $\delta^{13}\text{C}_{\text{carb}}$ ); and one broad  $4\text{‰}$ – $5\text{‰}$

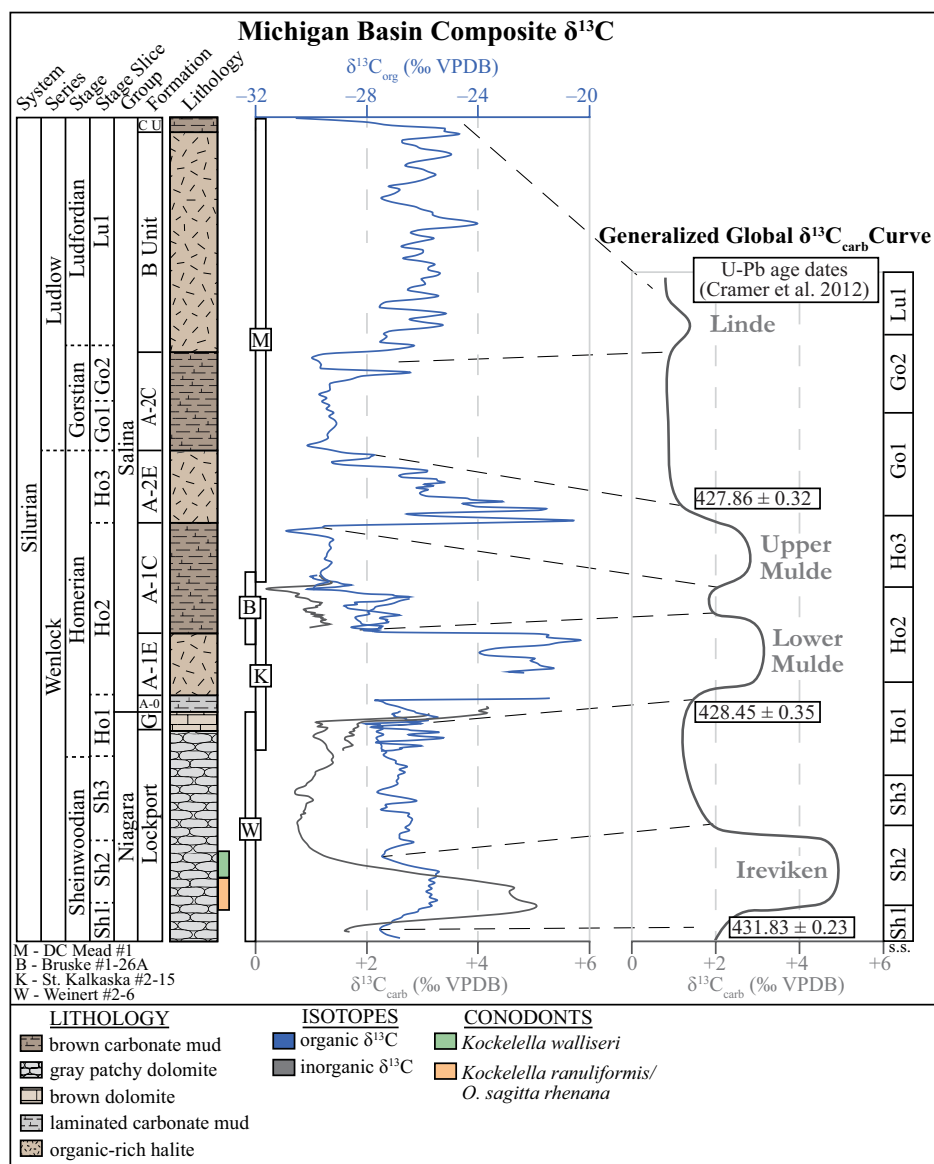
( $\delta^{13}\text{C}_{\text{org}}$ ) positive excursion spanning the uppermost Salina A-2 Carbonate Unit to uppermost Salina B unit (Fig. 6). In order to evaluate the geological significance and potential global correlation of these CIEs, it is important to first assess the potential impact of any local secondary processes that may have altered or affected the primary isotope signal.

### Secondary Effects on $\delta^{13}\text{C}$ Fractionation

Did diagenetic processes such as dolomitization and/or thermal maturation of buried organic carbon control the primary signal (or magnitude) of these CIEs? This is one of the most important questions to address because secondary effects could result in diagenetic overprinting that could potentially impact the carbon isotope rec-

ord and impede reliable correlation with global events. In order to answer this question, it is vital to first understand the physical characteristics of these CIEs (e.g., associated lithologies and magnitude of isotope perturbation).

Three observations suggest that the carbon isotope record in the Michigan Basin is a primary signal. First, carbon isotope data presented in Figures 3–5 show that the recorded CIEs spanned multiple lithologies and were not restricted to a particular interval (e.g., CIEs are not restricted to specific zones of halite or dolomite), as would be expected if secondary fractionation of carbon isotopes during diagenesis was the principal cause of the observed CIEs. Dolomitization, for instance, is quite common throughout the Silurian strata of the Michigan Basin and occurs throughout the Niagara and



**Figure 6.** Composite Niagara–Salina Group  $\delta^{13}\text{C}_{\text{org}}$  and  $\delta^{13}\text{C}_{\text{carb}}$  data from the Michigan Basin, presented as a three-point moving average, showing correlation with the global Ireviken and Mulde carbon isotope excursions (CIEs). Cores: K (State Kalkaska #2–15); B (Bruske #1–26); W (Weinert #2–6); and M (DC-Mead #1). Generalized global  $\delta^{13}\text{C}_{\text{carb}}$  curve and biostratigraphy are modified after Cramer et al. (2011, 2012) and Melchin et al. (2012). Sh—Sheinwoodian, Ho—Homerian, Go—Gorstian, Lu—Ludfordian, G—Guelph, E—Evaporite, C—Carbonate, C U—C Unit, O.—*Ozarkodina*, VPDB—Vienna Pee Dee belemnite standard.

Salina Group carbonates (Catacosinos et al., 2001). If dolomitization caused the observed CIEs, isotopic changes should be restricted to the dolomitized intervals of the A-0C and A-1C units. There would be no major fluxes in carbon isotope values (or CIEs) that extend into the A-1E or A-2E units. In order to evaluate this possibility, however, samples from the Bruske #1–26A and DC-Mead #1 cores were reprocessed using hot HCl (red text in supplemental

data [see footnote 1]). Hot HCl is commonly known to more fully dissolve dolomite crystals, thereby eliminating the mineralogical fractionation effect of dolomitization. The  $\delta^{13}\text{C}_{\text{org}}$  values showed no change from their original values, which reinforces the interpretation that the isotopes reflect a primary depositional signal.

Second, it has long been known that secondary salt crystal growth occurs throughout the Salina Group carbonate units (Catacosinos

et al., 2001). It could be the case that during thermal maturation, migrating fluids containing salt and hydrocarbons with isotopically light organics could have infiltrated sediment at various stratigraphic intervals, thus altering the original long-term carbon isotope record (Lewan, 1983). If this were the case, carbon isotope values would be expected to be lighter (more positive) wherever secondary reprecipitated salt crystal growth is present. However, our data show that organic carbon isotopes in the A-1C and A-2C units maintain more negative values ( $-29\text{‰}$ ) than either the pre-excursion levels of  $-27\text{‰}$  or the CIE levels of approximately  $-22\text{‰}$  to  $-20\text{‰}$  ( $\delta^{13}\text{C}_{\text{org}}$ ), despite the presence of minor post-depositional secondary halite crystal growths (Fig. 6).

Third, the uppermost (youngest) CIE in our composite curve has different physical characteristics in comparison to the preceding large positive CIEs. This could suggest a controlling mechanism that is independent. If a secondary process was affecting these data equally, then the CIEs would be expected to have a similar magnitude and character. As shown in Figure 6, the ascending limb of this uppermost CIE is broad, and isotope values reach a lower-magnitude value,  $-25\text{‰}$ , in comparison to the two older Salina Group CIEs, which reach values of  $-21\text{‰}$  and  $-20\text{‰}$ , respectively. The lack of apparent continuity between isotopic perturbations reinforces the claim that diagenesis or salt inclusions are unlikely to have caused these observed excursions.

Last, enhanced evaporation during Salina Group deposition could have had profound effects on the local inorganic and organic  $\delta^{13}\text{C}$  values in the Michigan Basin. The composite isotope curve (Fig. 6) shows that organic carbon values reached a maximum of  $-20\text{‰}$  during deposition of two thick halite Salina Group units (A-1E and A-2E units). Stiller et al. (1985), Schmid (2017), and others have contended that the process of evaporation leads to a loss of  $[\text{CO}_2]$  within evaporative brines, which in turn is reflected in the inorganic carbon isotope record as highly enriched  $\delta^{13}\text{C}_{\text{carb}}$  values in the associated carbonates. Documented Paleoproterozoic evaporite-related carbonates contain  $\delta^{13}\text{C}_{\text{carb}}$  values up to  $+17.2\text{‰}$  (Melezhik et al., 1999), whereas Permian and Triassic evaporite sequences have values up to  $+7\text{‰}$   $\delta^{13}\text{C}_{\text{carb}}$  (Schmid et al., 2006).

Interestingly, a similar positive shift is observed in  $\delta^{13}\text{C}_{\text{org}}$  values in organic biomarkers deposited during periods of enhanced evaporation. In a study of Miocene–Pliocene halite deposits in the Dead Sea, Grice et al. (1998; their fig. 2) showed large variation in  $\delta^{13}\text{C}_{\text{org}}$  values of various organic biomarkers, sug-

gesting  $\delta^{13}\text{C}_{\text{org}}$  values of n-alkanes range from  $-25\text{‰}$  to  $-33\text{‰}$ , hopanes range from  $-24\text{‰}$  to  $-25\text{‰}$ , diaromatic steroids range from  $-22\text{‰}$  to  $-24\text{‰}$ , and isoprenoids range from  $-16\text{‰}$  to  $-15\text{‰}$ . In a study of Ordovician carbonates from Iowa, Pancost et al. (2013) showed two correlative positive CIEs of  $2\text{‰}$ – $3\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) and  $4\text{‰}$ – $10\text{‰}$  ( $\delta^{13}\text{C}_{\text{org}}$ ), demonstrating that variability in the magnitude of this CIE in organic carbon can reflect different chemical compounds within the bulk organic matter. These studies show that our observed local changes in organic carbon values may have had a significant impact on the observed fractionation within the organic carbon isotope record, perhaps as a function of changing primary productivity brought on by changing paleoenvironment and  $p\text{CO}_2$ .

### Correlation with the Global Carbon Isotope Curve

A fundamental argument for the global correlation of carbon isotope excursions is underpinned by identifying similar chemostratigraphic patterns recorded in variable facies that are geographically far removed from one another (Shields et al., 2002; Oehlert et al., 2012; Saltzman and Thomas, 2012). This is observed in the new halite-derived carbon isotope data set from the Michigan Basin. The composite curve in Figure 6 is in general agreement with the global curves of Cramer et al. (2011, 2015) and Sullivan et al. (2016). As mentioned, Kuglitsch (2013) identified *Kockelella walliseri* slightly above co-occurring elements of *Kockelella ranuliformis* and *Ozarkodina sagitta rhenana* from the lower Lockport Formation, at 4540 ft–4580 ft in the Weinert #2–6 core, correlative with the lowest positive CIE (orange and green bars in Figs. 2 and 6). These conodonts are indicative of the lower *K. walliseri* and underlying *O. s. rhenana* zones of the Silurian (Jeppson et al. 2006) and suggest an early Wenlock (middle Sheinwoodian) age that is consistent with the Ireviken CIE of Cramer et al. (2011) and Noble et al. (2012).

In regards to the three overlying large positive CIEs (Fig. 6), temporal constraint is hampered by a lack of viable conodont data from the Niagara-Salina Group boundary in the Michigan Basin center (Klug, 1977; Kuglitsch, 2000; Sullivan et al., 2016), and a lack of age data from the Salina Group. However, as mentioned earlier, studies on the southern and eastern basin margin show that the Niagara-Salina transition occurred during upper Wenlock (Homerian) time (Cramer, 2009; Swift, 2011; Kleffner et al., 2012; McLaughlin et al., 2013; Sullivan et al., 2016). In a study from the western margin

(modern-day Wisconsin), Sullivan et al. (2016; their fig. 5) successfully constrained the Mulde event CIE from the correlative Niagara-Salina Group boundary by demonstrating Homerian  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios in conodont elements recovered from the acme of the positive CIE, thus constraining this event. Therefore, if the lithologic transition from carbonates to evaporites is generally coeval across the Michigan Basin, a similar large positive CIE at approximately the same stratigraphic position within the basin center would be expected. The composite isotope curve (Fig. 6) shows carbon isotope perturbation beginning within the uppermost Niagara Group (Guelph Formation) in the Weinert #2–6 and State Kalkaska #2–15 cores and continuing in the lowermost Salina Group. This correlation, along with the occurrence of Wenlock brachiopods *Plicocyrta* cf. *arkansana* and *Pentamerus* sp. in the basal Salina Group (Mesoella et al., 1974) at a level correlative with the ascending limb of the CIE, is strong evidence that the lowermost Salina group CIE is equivalent to the lower Mulde CIE.

Temporal constraint is not currently available for the two younger large CIEs because there have been no age-specific conodonts recovered from Salina Group sediments of the Michigan Basin center. Therefore, it remains possible that the overlying CIEs (spanning the Salina Group A-2E and B units) are an expression of the upper Mulde, Linde, or Lau events. Cramer et al. (2011, 2015) and Sullivan et al. (2016) showed the Mulde as a two-prong  $\sim 2.5\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) CIE that occurs in the upper Wenlock; the Linde as a small  $\sim 1.5\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) CIE that occurs in the upper Gorstian to lower Ludfordian (Calner et al., 2004); and the Lau as a large  $\sim 7\text{‰}$ – $12\text{‰}$  ( $\delta^{13}\text{C}_{\text{carb}}$ ) CIE that occurs in the middle Ludfordian.

There is no reason to suggest that the Mulde event would not have the same two-prong character as recorded in the global curve of Cramer et al. (2015). If the Niagara-Salina Group boundary CIE is the lower Mulde event, then it is possible that this overlying large CIE of similar magnitude is the basal expression of the upper Mulde event. The CIE in the A-2E unit is of similar magnitude as the lower Mulde event (within the A-1E unit), and its placement in the lower Salina Group appears too low to be consistent with the Linde or Lau event. However, no other study has documented carbon isotope values in the trough separating the two Mulde CIEs that return to a prolonged (and relatively stable) baseline that is more negative than pre-excursion levels. This might reflect local conditions within the Michigan Basin center. If this interpretation is correct, then it implies that the overlying broad positive CIE spanning the Sa-

lina B unit represents the smaller-magnitude Linde CIE (Fig. 6).

An alternative interpretation is that the large CIE spanning the Salina Group A-2E unit is an exceptionally large-magnitude example of the Linde event, or possibly even the Lau event. However, based on the location of the CIE within the general Salina Group stratigraphy in comparison to the timing and magnitude of perturbation expressed in the global curve, this interpretation seems less likely. Therefore, the interpretation that the lower Salina Group CIEs are examples of both Mulde excursions is most consistent with the available observations.

### CONCLUSIONS

Inorganic and organic carbon isotope data extracted from carbonate- and halite-dominated sequences in the Silurian Niagara and Salina Groups of the Michigan Basin record three large excursions in  $\delta^{13}\text{C}_{\text{carb}}$  and  $\delta^{13}\text{C}_{\text{org}}$  that can be correlated with the global records of the Ireviken, Mulde, and Linde events. Organic carbon isotope values reach a maximum of  $-20\text{‰}$  during deposition of two thick halite Salina Group units (A-1E and A-2E units), suggesting that local changes in organic carbon source material during periods of enhanced evaporation might have affected local fractionation, thereby influencing the observed values. The ability to establish a long-term carbon isotope record in halite sequences provides a new and important approach for investigating ancient evaporative depositional settings. This is of broad interest to the geoscience community because it delivers a new tool for chronostratigraphic correlation that is valuable for temporally linking periods of prolonged evaporite deposition to events of known paleoclimate change.

### ACKNOWLEDGMENTS

Support was given by the Midwest Region Carbon Sequestration Partnership program, Battelle Memorial, Columbus, Ohio, and by the U.S. Department of Energy. We thank the staff at the Michigan Geological Repository for Research and Education, part of the Michigan Geological Survey, for making core material available for study. We also wish to express our gratitude for the meticulous reviews and helpful suggestions of Mark Kleffner, Grzegorz Racki, and Brad Cramer, which greatly benefited this manuscript.

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SCIENCE EDITOR: BRADLEY S. SINGER  
ASSOCIATE EDITOR: BRADLEY CRAMER

MANUSCRIPT RECEIVED 15 NOVEMBER 2017  
REVISED MANUSCRIPT RECEIVED 5 MARCH 2018  
MANUSCRIPT ACCEPTED 4 APRIL 2018

Printed in the USA