Pattern and Timing of the Late Devonian Biotic Crisis in Western Canada: Insights from Carbon Isotopes and Astronomical Calibration of Magnetic Susceptibility Data

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PATTERN AND TIMING OF THE LATE DEVONIAN BIOTIC CRISIS IN WESTERN CANADA: INSIGHTS FROM CARBON ISOTOPES AND ASTRONOMICAL CALIBRATION OF MAGNETIC SUSCEPTIBILITY DATA

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ABSTRACT: Carbon stable isotope data from western Canada, in combination with biostratigraphic control and astrochronologic constraints from magnetic susceptibility data, provide insight into the pace and timing of the Frasnian–Famennian (F–F; Late Devonian) biotic crisis. In much of the world, this event is characterized by two organic-rich shales, which display geochemical anomalies that indicate low-oxygen conditions and carbon burial. These events, commonly referred to as the Lower and Upper Kellwasser events (LKE and UKE), have been linked to the expansion of deeply rooted terrestrial forests and associated changes in soil development, chemical weathering, and Late Devonian climate. The δ13C data generated from organic matter record a 3 to 4% positive excursion during each event. These data and other geochemical proxy data reported elsewhere corroborate hypotheses about enhanced biological productivity, driven by terrigenous input under exceptionally warm climatic conditions. In this hypothesis, a boom in primary production leads to successive development of anoxic bottom water conditions, decreased biotic diversity, and net transfer of carbon from the atmosphere to the ocean floor.

Despite the importance of the F–F events, a precise chronology for the events is lacking due to limited biostratigraphic resolution. Each of the F–F events falls within one conodont zone, with durations estimated on the order of 0.5 to 1.0 Myr. The LKE occurs very high in Frasnian Zone (FZ) 12, while the UKE begins within FZ 13B, just below the F–F boundary. A previous analysis of high-resolution magnetic susceptibility data from the studied sections in western Canada identified 16.5 eccentricity cycles, each lasting 405 kyr, within the Frasnian strata and one in the earliest Famennian. The present study reports δ13C anomalies associated with the LKE and UKE in the same sections. The LKE and UKE intervals comprise 7 to 8 and 13 to 13.5 m of stratigraphic section, respectively. Based on our analysis, this implies that they represent only one 405-kyr eccentricity cycle or less. We estimate that the duration of the LKE was a bit more than half of a long eccentricity cycle (~200–250 kyr), while the UKE was more protracted, lasting a full long eccentricity cycle (~405 kyr). The onset of both events is separated by one-and-a-half 405-kyr eccentricity cycles, indicating that they occurred about 500 to 600 kyr apart. This work demonstrates the utility of magnetic susceptibility, or other long time-series proxy data, used in conjunction with astronomical calibration to provide insight into the pacing of significant events in geologic time.

KEY WORDS: Frasnian–Famennian, biotic crisis, magnetic susceptibility, carbon isotopes, astronomical calibration

INTRODUCTION

The precise timing and pacing of events in Paleozoic time are two of the fundamental issues in stratigraphic analyses, and great strides toward understanding these factors have been made in recent years through the application of time-series analysis to high-resolution proxy records. Several studies have successfully identified the imprint of astronomical climate forcing in Paleozoic records and calibrated portions of the geologic timescale (for a review on this topic, see Hinnov 2013). Magnetic susceptibility data are commonly used for this type of analysis.
in Cenozoic and Mesozoic rocks (Shackleton et al. 1995; Weedon et al. 1999; Bouilla et al. 2008a, 2008b; Hinnov 2013) and, more recently, have been applied to sequences of Paleozoic age (De Vleeschouwer et al. 2012a, 2012b, 2014a; Da Silva et al. 2013; Wu et al. 2013).


In this paper, we use the astrochronologic framework for the Frasnian of western Canada (as published in De Vleeschouwer et al. 2012b) to interpret new high-resolution sequence and biostratigraphic analyses and stable carbon isotopic data. This approach provides insight into the pattern and timing of the F–F biotic crises, and it permits inferences to be made about the global carbon cycle during the studied catastrophic events.

The stepwise F–F events are commonly referred to as the Upper and Lower Kellwasser events (UKE and LKE, respectively) based on characteristic organic-rich shales deposited during these intervals at many locations in Europe (Buggisch 1991, Joachimski and Buggisch 1993). The LKE and UKE are characterized by positive carbon isotope excursions (Fig. 1) commonly interpreted to result from carbon burial in low-oxygen marine environments (Joachimski et al. 2001). Strata representing the events are usually relatively thin and locally condensed, and the LKE interval is commonly thinner than the UKE (Table 1). Wang et al. (1996) documented a broad positive $^{13}$C excursion near the F–F boundary in western Canada; however, their sampling was rather coarse, and the thickness of the interval that records the excursion (~40 m) strongly suggests that it encompasses both the LKE and UKE. Many records of these Kellwasser events were documented in shallow epeiric sea depositional settings, and it remains unclear whether the global ocean experienced widespread low-oxygen conditions. Recent proxy data from redox-sensitive trace metals in Australia and south China (George et al. 2014, Whalen et al. 2015) imply that the latter may not have been the case. A model to drive high productivity, development of low-oxygen conditions, and eventually two successive and closely spaced biotic crises involves the expansion of deeply rooted terrestrial forests and associated changes in soil development, chemical weathering, nutrification of epeiric seas, and ultimately major Late Devonian climate perturbations (Algeo and Scheckler 2010, Algeo et al. 2010). The global carbon cycle appears to have played a central role throughout these events. Therefore, this paper aims to gain insight into the pattern and timing of the carbon cycle perturbations, to better understand the ultimate processes driving the Late Devonian events.

Geologic Setting and Stratigraphy

Devonian rocks exposed in western Alberta range from middle Givetian to Famennian in age and consist of isolated and attached carbonate platform complexes and associated slope and basin facies (Figs. 2, 3; Switzer et al. 1994). The strata were deposited at a low paleolatitude along the western continental margin of Laurasia (Witzke and Heckel 1988, Scotese and McKerrow 1990). A regionally extensive tropical carbonate ramp system first developed in the Western Canada Sedimentary Basin during the Middle (Late Givetian) Devonian transgression across a widespread subaerial unconformity (Figs. 2, 3; Switzer et al. 1994). Fine-grained siliciclastic sediment mixed with platform-derived carbonates in slope and basin settings (Switzer et al. 1994, Oliver and Cowper 1963, Stoakes 1980, Switzer et al. 1994, Whalen et al. 2000b, Wenzte and Uyeno 2005). During most of the Frasnian, carbonate-platform growth kept pace with prolonged second-order (~10 Myr) and third- or fourth-order (0.5–5.0 and 0.1–0.5 Myr, respectively) sea-level rises and outpaced basinal sedimentation (Johnson et al. 1985, 1996; Day et al. 1996; Day 1998; Whalen et al. 2000b). The platforms record a series of depositional sequences that developed in response to eustatic sea-level events during the Late Devonian (Johnson et al. 1985, 1996; Whalen and Day 2010). Two of these sea-level events (IId2 and IId3, Fig. 3) coincide with the F–F LKE and UKE (Joachimski et al. 2009; Stigall
Table 1.—Thicknesses of the LKE and UKE intervals at localities around the world. The thicknesses of the LKE and UKE were determined by the extent of the documented isotropic excursion in each cited publication (see text for further explanation). Note that those sections with very thick event intervals (>10 m for LKE and >20 m for UKE) are based on relatively coarse sampling.

<table>
<thead>
<tr>
<th>Location</th>
<th>LKE (m)</th>
<th>UKE (m)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Section C, W. Canada</td>
<td>7.5</td>
<td>13.5</td>
<td>This study</td>
</tr>
<tr>
<td>Section W4, W. Canada</td>
<td>8.0</td>
<td>13.0</td>
<td>This study</td>
</tr>
<tr>
<td>Fuhe, S. China</td>
<td>2.8</td>
<td>8.2</td>
<td>Chen et al. (2005)</td>
</tr>
<tr>
<td>Baisha, S. China</td>
<td>4.0</td>
<td>6.7</td>
<td>Chen et al. (2005)</td>
</tr>
<tr>
<td>Carnic Alps, Austria</td>
<td>0.2</td>
<td>1.2</td>
<td>Joachimski and Buggisch (1993)</td>
</tr>
<tr>
<td>Courniace, France</td>
<td>1.2</td>
<td>1.2</td>
<td>Joachimski and Buggisch (1993)</td>
</tr>
<tr>
<td>Benner, Germany</td>
<td>1.0</td>
<td>1.2</td>
<td>Joachimski and Buggisch (1993)</td>
</tr>
<tr>
<td>Schmidt, Germany</td>
<td>0.5</td>
<td>1.2</td>
<td>Joachimski and Buggisch (1993)</td>
</tr>
<tr>
<td>Vogelsberg, Germany</td>
<td>1.8</td>
<td>1.3</td>
<td>Joachimski and Buggisch (1993)</td>
</tr>
<tr>
<td>Kowala, Poland</td>
<td>1.5</td>
<td>8.0</td>
<td>Joachimski et al. (2001)</td>
</tr>
<tr>
<td>B¨udesheimer, Germany</td>
<td>6.6</td>
<td>5.0</td>
<td>Joachimski et al. (2002)</td>
</tr>
<tr>
<td>Bou Ouneboud, Morocco</td>
<td>0.5</td>
<td>0.7</td>
<td>Joachimski et al. (2002)</td>
</tr>
<tr>
<td>Devils Gate, Nevada</td>
<td>5.0</td>
<td>30.0</td>
<td>Joachimski et al. (2002)</td>
</tr>
<tr>
<td>West Valley core, New York</td>
<td>20.0</td>
<td>30.0</td>
<td>Murphy et al. (2000)</td>
</tr>
<tr>
<td>HS core, W. Australia</td>
<td>10.0</td>
<td>42.0</td>
<td>George et al. (2014)</td>
</tr>
</tbody>
</table>

2012; Whalen and Day 2008, 2010) and are linked to the Late Devonian biotic crisis.

The Miette and Ancient Wall platforms, in the western part of the Western Canada Sedimentary Basin, are 400 to 500 m thick, respectively (Figs. 2, 3). Ancient Wall was larger than Miette, and both were much smaller than the Southesk-Cairn and other attached reef complexes further to the east (Fig. 2; Geldsetzer 1989, Mountjoy 1989). Miette and Ancient Wall remained isolated platforms through the Frasnian until their demise, along with Devonian metazoan reefs worldwide, during the F–F biotic crisis (McLaren 1982, Stearn 1987, 2008; Cramer and Saltzman 2007). Therefore, they constitute the most suitable proxy for evaluating the pattern and timing of the F–F events. We adopt a similar procedure as the one that has been employed in the evaluation of other major positive carbon isotope excursions such as Steptoean positive carbon isotope excursion (SPICE) (Saltzman et al. 1998), Hirnantian carbon isotope excursion (HICE) (Bergstrom et al. 2006), or the Irivekian (Cramer and Saltzman 2007), Mulde (Cramer et al. 2006), or Hangenberg excursions (Cramer et al. 2008). In all of these cases, the onset and termination of the carbon isotope excursion mark the beginning and the end of the event, respectively (Buggisch 1991; Saltzman et al. 1998; Joachimski and Buggisch 1993; Joachimski et al. 2002; Bergstrom et al. 2006; Cramer et al. 2006, 2008; Cramer and Saltzman 2007). Therefore, in our analysis, the onset of the Kellwasser positive excursions indicates the initiation of the event. The decline of δ¹³C values to a background level, which may differ from the pre-event background, marks the termination of the event.

METHODS

Sequence Stratigraphy and Biostratigraphy

Previous high-resolution sequence stratigraphic analysis of the Miette and Ancient Wall platforms identified a series of Frasnian depositional sequences and documented sea-level events that controlled their deposition (van Buchem et al. 1996; Whalen et al. 2000a, 2000b). In this project, we employed conodont biostratigraphy (Klapper 1989; Ziegler and Sandberg 1990; Klapper et al. 1995; Day et al. 1996; Klapper 1989, 1997, 2007) and magnetic susceptibility (MS) stratigraphy (Crick et al. 1997, 2001, 2002; Ellwood et al. 1999, 2000, 2001) in conjunction with sequence stratigraphic methods (Handford and Loucks 1993, Posamentier and Allen 1999) to better constrain temporal correlations. Whalen and
Day (2008, 2010) documented the MS stratigraphy and conodont biostratigraphy according to the Montange Noire (MN) or Frasnian zonation (FZ) of Klapper (1989), with refinements to that zonation by Girard et al. (2005). Details of the sequence stratigraphic framework for these rocks were documented by van Buchem et al. (1996) and Whalen et al. (2000a, 2000b), and a thorough discussion of the biostratigraphic and MS data within the sequence stratigraphic framework was outlined by Whalen and Day (2008, 2010). Both the outer-ramp to slope succession at section C and the platform-interior section, W4, record seven parasequences over the F–F interval (Figs. 4, 5). Five parasequences make up sequence 8, and two parasequences are within the lowstand systems tract (LST) and transgressive systems tract (TST) of sequence 9 (Figs. 4, 5). Section C appears to record continuous deposition, but it is characterized by some redeposited carbonates, especially in the lower part of the section, and it also includes a thin black shale interval at the F–F boundary that is probably condensed (Figs. 4, 5). Section W4 records subtidal parasequences with normal marine fauna in most of sequence 8 that become more peritidal during regression with minor truncation indicating subaerial exposure (Figs. 4, 5), while sequence 9 is dominated by peritidal facies (Whalen et al. 2000a, 2000b). The cyclostratigraphic framework (Fig. 6), based on the MS data of the studied sections (De Vleeschouwer et al. 2012b), provides insight into the chronology of the Kellwasser events during the Late Devonian.

Magnetic Susceptibility

MS represents the degree of magnetization of a material in response to an applied magnetic field. The MS of a sediment varies in proportion to the concentration of detrital-dominated paramagnetic and ferrimagnetic minerals and forms a proxy for the rate of supply of impurities delivered to the sedimentary environment (Ellwood et al. 2000). In the studied section, large-scale MS variations can be ascribed to relative sea-level changes and million-year-scale climate variations due to plate tectonics or biosphere innovations (Mabille and Boulvain 2007; Da Silva et al. 2009; Whalen and Day 2008, 2010). On a shorter timescale, climate, through variations in precipitation and wind, determines the flux of magnetic minerals to the ocean through riverine or eolian processes (Hladil et al. 2006, Riquier et al. 2010). Sampling for MS was conducted at half-meter intervals, and all samples are keyed to stratigraphic sections measured with a Jacob’s staff or tape measure and compass (Whalen and Day 2008). MS samples were weighed to within 0.001 g and measured on a KLY-3 Kappa bridge MS meter at Brooks Ellwood’s laboratory (Louisiana State University). MS values represent an average of three measurements of mass-normalized bulk MS reported in units of m$^3$/kg, for each sample. Due to the high carbonate content of most samples, the mass-normalized bulk MS signature is weakly positive. Whalen and Day (2010, Fig. 14 therein) demonstrated that the MS signal in the studied section is mainly driven by detrital inputs, and, consequently, the role of diagenesis is considered negligible.

Fig. 2.—A) Palinspastically restored Middle Frasnian paleogeographic map, illustrating the locations of Upper Devonian isolated and attached carbonate platforms in the Western Canada Sedimentary Basin. Buildups located west of the easternmost barbed line, indicating the eastern limit of Laramide thrusting, are exposed in the Canadian Rocky Mountains (after Geldsetzer 1989, Mountjoy 1989, Switzer et al. 1994). B) Location map of the overthrust belt in western Alberta and eastern British Columbia, showing the locations of the Miette and Ancient Wall platforms (after Mountjoy 1965). Lettered stars indicate the location of measured stratigraphic sections: C = Thornton Creek, W4 = Poachers Creek.
Stable C Isotopic Analysis

Stable carbon isotopic analyses on organic matter were conducted at the Alaska Stable Isotope Facility at the University of Alaska–Fairbanks. Samples were prepared by acidifying 10-g subsamples of powdered material with an excess of 1 M HCl at 50°C for 24 hours. The acid-insoluble residues were rinsed, freeze-dried, and analyzed for their C contents using a Costech Elemental Analyzer (ECS 4010). Isotopic ratios were then measured using a Conflo III interface with a Delta+XP mass spectrometer, and ratios are reported using conventional delta (δ) notation in per mil (‰) (Coplen 2011) relative to the international Vienna PeeDee belemnite (VPDB) standard (Gonfiantini 1978; Fig. 4). Typical instrumental precision is ±0.2‰.

Time-Series Analysis

In this paper, power spectra of the studied MS signals were computed using the multitaper method (MTM; Thomson 1982) with 3-2π prolate tapers, which is the same method used in De Vleeschouwer et al. (2012b). However, in this paper, we used a different red noise model (conventional AR[1]; Gilman et al. 1963) compared to the original paper (“robust” AR[1]; Mann and Lees 1996), because Meyers (2012) demonstrated that the “robust” AR(1) approach often imposes false positives within the low-frequency (i.e., eccentricity) portion of the spectrum. Also, we applied two complementary techniques: the MTM harmonic F-test (Thomson 1982) and the average spectral misfit method (ASM; Meyers and Sageman 2007), as implemented in the R package “astrochron.”

Fig. 3.—Upper Devonian stratigraphy of western Canada (after Switzer et al. 1994). Platform units are light gray, while basinal facies are white. Note that there is different stratigraphic terminology for platform as opposed to basinal units. The right side of the diagram illustrates relative transgressive–regressive sea-level events (after Johnson et al. 1985; Day et al. 1996, Whalen and Day 2008, 2010) The LKE and UKE strata are located within the marked study intervals.
The ASM was calculated for the Simla Mbr. of section C using frequencies that exceeded the 90% F-test confidence level (Fig. 6). The ASM method evaluates the misfit between the selected frequencies and the expected astronomical spectrum for the Late Devonian, with 405- and 100-kyr eccentricity, 32.25-kyr obliquity, and 19.95- and 16.85-kyr precession (Berger et al. 1992), for a wide range of plausible sedimentation rates. The null hypothesis of ‘no astronomical influence’ was quantitatively evaluated using Monte Carlo simulation. This approach permits a null hypothesis test in the absence of high-resolution biostratigraphic or radioisotopic age constraints. It should be noted, however, that the above-listed methods are designed for depositional settings characterized by quasi-constant accumulation rates. Hence, some assumptions and restrictions must be explicitly stated and incorporated. On carbonate platforms and ramps, sediment production rate is, above other factors, a function of carbonate productivity. The main trend is a decrease in the carbonate productivity with increasing depth, with maximum carbonate production in reefal environments (James 1997, Da Silva et al. 2009). Given that section C contains a transition between slope and outer-ramp facies, the accumulation rate cannot be considered constant between the two lithologies. In depositional environments with different accumulation rates, cycles of equal duration are expressed by a sediment package of different thickness. However, if one assumes that accumulation rate within the same sedimentary environment is quasi-constant, the above-listed techniques can be carried out separately for different sedimentary environments. This assumption seems acceptable given that Whalen and Day (2010) demonstrated the absence of significant unconformities and only slight variations in sedimentation rate within depositional environments.

RESULTS

Biostratigraphy and Sequence Stratigraphy

The study intervals for sections C and W4 record the LST, TST, and highstand systems tract (HST) of Alberta sequence 8 and the LST and TST of sequence 9 (Whalen and Day 2008, 2010). Deposits of sequence 8 record a very late Frasnian sea-level deepening event that resulted in progradation of ramp facies of the Ronde Mbr. and upper-slope facies of the Simla Mbr. (Figs. 4, 5; Whalen et al. 2000a, 2000b). This deepening occurred at or near the top of Frasnian Zone (FZ) 12, as documented by the first appearance of Polygnathus...
lodiensis (Klapper 2007) at 21.5 m in the Simla Mbr. at section C (Fig. 7; Table 2). Conodont samples from the platform-top environment of the Ronde Mbr. in section W4 contain few age-diagnostic forms, and biostratigraphic zonation remains questionable (Figs. 4, 5).

The base of FZ 13 is determined by the first occurrence of *Polygnathus brevicarina*; *Palmatolepis bogartensis*, the index taxon for FZ 13A, is first found higher (Fig. 7; Table 2). *Palmatolepis linguiformis*, marking the base of FZ 13B, was recovered at 34 m in section C (Table 2). At Miette, late Frasnian faunas in the upper part of section W4 are from the *Polygnathus* biofacies and thus far do not allow recognition of FZ 13. The deepening event recorded by sequence 8 coincides with a major global sea-level rise designated as North American cratonic Devonian T–R cycle IId2 (Day 1998), and the LKE was initiated during the TST of this sea-level cycle (Fig. 4). The TST contains five parasequences, and the LKE initiated near maximum transgression during deposition of the fourth parasequence. The remainder of the LKE is within the lower portion of the HST of sequence 8 (Figs. 4–6). Parasequences in the LST of sequence 8 on the platform top (section W4) are 6 to 8 m thick and sharp based, and they thicken and coarsen upward from silty bioclastic wackestones to float–rudstones containing brachiopods and corals (Fig. 5). In the last parasequence of the TST, a 14-m-thick cycle shoals upward from silty mudstones into laminated, fenestral peritidal wackestones with an undulating truncated upper surface (Fig. 5). The HST is an 8-m-thick bed of relatively massive bioclastic wackestone. In the slope environment (section C), the base of sequence 8 is a 15-m-thick package of redeposited lithoclastic rudstone interpreted as the LST. Overlying parasequences of the TST and HST are 3 to 9 m thick and are characterized by flooding surfaces overlain by calcareous shale or wackestone, followed by redeposited carbonate float–rudstones or packstones with corals and lithoclasts.

Deposits of sequence 9 include the Sassenach Fm. (Figs. 3–6). The tops of the extinct Frasnian platforms in western Alberta were emergent during a latest Frasnian–earliest Famennian sea-level lowstand, although subtidal conditions persisted in off-reef settings. The early Famennian conodont fauna from an oncoid debris flow bed (Whalen et al. 2002), 1.8 m above the base of the Sassenach Fm. stratotype at section C (Table 2, sample 48.8), indicates that initial latest Frasnian transgression of T–R cycle IId3, and resumption of carbonate deposition on the flanks and tops of the older extinct Frasnian platforms, occurred during the Lower to Middle *Palmatolepis triangularis* Zone (Figs. 4–6). The Sassenach Fm. spans the interval of the Lower–Middle *P. triangularis* Zone (Table 2) through the Lower *Palmatolepis crepida* Zone, following brachiopod-based correlations by Raaasch (1989) (Famennian zones DFM1 to DFM3), and conodont faunas documented by Johnston and Chatterton (1991), which place the base of the overlying Palliser Fm. within the Lower *P. crepida* Zone. Only two parasequences were documented in the lower part of sequence 9. At section C, the lower parasequence is 2.4 m thick, appears to be...
FIG. 6.—Astronomical calibration of MS data from western Canada. A) Cyclostratigraphic correlation of the Frasnian of western Canada (modified from De Vleeschouwer et al. 2012b). MS stratigraphy (black) and band-pass filtered signal of the MS stratigraphy at the frequencies of the 405-kyr eccentricity cycles (E1, brown and orange lines). White and gray boxes indicate 16.5 405-kyr eccentricity cycles over the entire Frasnian stage, implying a duration of $6.7 \pm 0.6$ Myr (the counting error is estimated to be 1 E1-cycle). B, C) Astronomical calibration of Upper Frasnian sections W4 (B) and C (C) that span the LKE and UKE intervals. The orange band-pass filters represent 405-kyr eccentricity cycles (LEC), and the blue band-pass filters represent 100-kyr eccentricity cycles. D) MTM power spectrum and harmonic F-test of the MS data from the Ronde Mbr., section W4, with astronomical interpretation. E) MTM power spectrum and harmonic F-test of MS data from the Simla Mbr., section C. F) The astronomical interpretation of this spectrum, here refined by the result of the average spectral misfit method, suggests an optimal sedimentation rate of 3.43 cm/kyr. This value is close to the sedimentation rate (3.2 cm/kyr) previously obtained by De Vleeschouwer et al. (2012b).
condensed, and consists of black shale overlain by a redeposited oncoid rudstone (Figs. 4, 5). The second parasequence is about 10 m thick, has a thin black shale at the base, and consists of interbedded black to gray shale, and silty mud–wackestone. The first parasequence is interpreted as the LST, and its base records the initiation of the UKE, while the second records deepening and the beginning of the TST. On the platform top, the base of sequence 9 consists of two parasequences, 8 and 5 m thick, respectively, that contain bioturbated silty mud–wackestone that grade upward into laminated or ripple cross-bedded wacke–packstones, all deposited within a peritidal environment (Figs. 4, 5). The first parasequence is interpreted as the LST and contains the initiation of the UKE, which continues into the overlying TST (Figs. 4, 5).

**Magnetic Susceptibility**

MS data are reported for sections C and W4 that span the F–F boundary (Figs. 5, 6). The mean MS value for samples from the part of section C reported here is $2.63 \times 10^{-7}$ m$^3$/kg ($n = 362$ for 189 m of section; Fig. 6). The mean MS value for samples from the portion of section W4 reported is $1.07 \times 10^{-8}$ ($n = 136$ for 67 m of section; Fig. 6).

In section C, the MS values are relatively low at the bottom of the interval and increase upward for about 10 m (Figs. 5, 6), with two relatively high MS peaks ($3.41 \times 10^{-7}$ m$^3$/kg and $3.14 \times 10^{-8}$ m$^3$/kg) near the top of the LKE interval. MS values drop to their lowest point in the studied interval ($8.27 \times 10^{-10}$ m$^3$/kg) within the uppermost Simla Mbr. and then gradually increase again in the overlying Sassenach Fm., with several peaks over the next 15 m of section (maximum value of $4.09 \times 10^{-8}$ m$^3$/kg). This second relatively high MS interval corresponds to the UKE (Figs. 5, 6).

In section W4, MS is very low ($1.41 \times 10^{-9}$ m$^3$/kg) in the lower part of the studied interval, corresponding to the LKE strata in the uppermost Ronde Mbr. (Figs. 5, 6). MS levels increase precipitously by over an order of magnitude ($2.48 \times 10^{-8}$ m$^3$/kg) at the Ronde–Sassenach contact, marking the base of the UKE interval. MS values fluctuate but stay relatively high through most of the UKE but gradually drop in the upper part (Fig. 5, 6).

\[ \delta^{13}C_{org} \text{ Section C} \]

Stable isotopic analyses were only conducted on samples straddling the F–F interval. In section C, this zone spans 34 m in the upper part of the section (Fig. 4). The $\delta^{13}C_{org}$ values record two positive excursions of similar magnitude of about 4%o (Fig. 4). The first is a 4.2%o (−29.7 to −25.5%o) excursion between 351 and 359 m, and it marks the position of the LKE. There are two minor positive fluctuations at the base of the interval (351–354 m), and values progressively drop to more negative levels of ~−28%o at 359 m. There are minor fluctuations above, until a large positive excursion of 4.39%o (−27.7 to −23.3%o) between 368.5 and 373 m, marking the lower part of the UKE interval (Fig. 4). This second positive excursion is followed by a decrease and a return to more negative values of ~−27%o at 380.5 m, marking the top of the UKE (Fig. 4).

\[ \delta^{13}C_{org} \text{ Section W4} \]

In section W4, the interval spanning the F–F boundary covers 25 m of strata near the top of the measured section (Fig. 4). The $\delta^{13}C_{org}$ data are much more variable than in section C (Fig. 4). The lower portion of the isotopic curve displays two positive peaks. The first, from 41 to 42 m records a 2.2%o excursion (−28.1 to −25.9%o). Values then drop back to −27.4 before the second 1.7%o positive excursion (−27.4 to −25.7%o) from 42.5 m to 48.0 m. The strata from 41 to 48 m, which record these two positive excursions, are interpreted as the LKE interval (Fig. 4).
Above the LKE, there is a series of four relatively large and closely spaced fluctuations that vary between a maximum of \( C_0 \) -23.9\% and minimum of \( C_0 \) -28.1\% over 4.5 m (Fig. 4). The UKE interval spans 52 to 65 m. The base of the UKE is taken as the first significant positive excursion of 3\% \((C_0)27\) to \((C_0)24\%\) beginning at 52 m. Above the closely spaced fluctuations, the isotopic signature records a significant 3\% negative excursion \((C_0)25.7\) to \( C_0 \) 28.7\% between 57.5 and 61.5 m in the section. Above this, there is a pronounced positive excursion of 3.7\% \((C_0)28.7\) to \((C_0)25\%\) between 61.5 and 64.5 m. We interpret the relatively abrupt decline, to a slightly higher background level, above this positive excursion to mark the top of the UKE interval.

**Time-Series Analysis**

We applied the cyclostratigraphic framework for the studied sections C and W4 (Fig. 6), as presented in De Vleeschouwer et al. (2012b), to assess the timing and pacing of the F–F transition. The

**Table 2.**—Conodont biostratigraphic data from section C, Ancient Wall. Table illustrates first occurrences of conodont taxa, conodont zones (Frasnian—Klapper 1989, 2007; Girard et al. 2005; Famennian—Ziegler and Sandberg 1990), stratigraphic units, and depositional sequences (Whalen and Day 2008, 2010).

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The following abbreviations are used:

- **Ad.** = Ancyrodenella
- **An.** = Ancyrognathus
- **I.** = Icriodus
- **Po.** = Palmatolepis
- **Pelek.** = Pelekygnathus
- **Pa.** = Polygnathus

**Ad.** = Ancyrodenella; **An.** = Ancyrognathus; **I.** = Icriodus; **Po.** = Palmatolepis; **Pelek.** = Pelekygnathus; **Pa.** = Polygnathus.
interval examined in section C spans the upper Mount Hawk Fm., the Simla Mbr. of the Southesk Fm., and the lower Sassenach Fm. (279–469 m), and it includes the uppermost part of FZ 11, 12, and 13, and the early Famennian Lower and Middle *triangularis* zones (Figs. 4, 5) (Whalen and Day 2008). A strong low-frequency (9–25-m period) cycle that is persistent throughout the Mount Hawk Fm. and the Simla Mbr., but loses spectral power in the Famennian Sassenach Fm., is associated with the 405-kyr eccentricity cycle (orange line on Fig. 6). The higher-frequency blue line (2.5–5-m filter on Fig. 6) indicates a periodic component in the MS data that is associated with 100-kyr eccentricity. While we are aware that there are other longer-term eccentricity cycles (Laskar et al. 2011), as a short hand we will refer to the 100-kyr eccentricity as “short” and the 405-kyr eccentricity as “long” eccentricity cycles (LECs). Both low- and high-frequency cycles show a thickening of the cycles throughout these late Frasnian and earliest Famennian deposits. Moreover, neither low- nor high-frequency cycles can be identified above 420 m in section C, suggesting that the input of coarser-grained siliciclastic sediments during the early Famennian hindered the development of astronomical cycles (Fig. 6). The thickening of cycles is in line with geological expectations, as a relatively higher sedimentation rate is expected in the prograding slope and outer-ramp facies of the Simla Mbr. compared to the deeper-slope facies of the Mount Hawk Fm.

Section W4 includes the Ronde Mbr. of the Southesk Fm. and the lower Sassenach Fm. It spans FZ 13a to 13c and the early Famennian Lower and Middle *triangularis* Zone (Figs. 4, 5) (Klapper 1989, Girard et al. 2005, Whalen and Day 2008). Similar to section C, decameter-scale cyclicity with a period between 10 and 20 m is ascribed to long eccentricity, and meter-scale cyclicity between 3 and 4.5 m is attributed to short eccentricity (Fig. 6).

**F–F Eccentricity Cycles and Events**

De Vleeschouwer et al. (2012b) analyzed data for several continuous and composite sections in western Canada and documented 16 Frasnian LECs. After comparison with the contemporaneous Kowala section in Poland (De Vleeschouwer et al. 2013), it was apparent that one half-long eccentricity cycle needed to be added, bringing the total up to 16.5 Frasnian LECs (Fig. 6). This corresponds to a duration of 6.7 Myr for the Frasnian (De Vleeschouwer et al. 2012b). The 405-kyr LECs and the cyclostratigraphic correlation of the sections (Fig. 6) are consistent with biostatigraphic- and sequence stratigraphic-based correlations (Whalen and Day 2008, 2010). This cyclostratigraphic correlation also indicates that part of LEC 16 in section W4 is likely missing due to a minor exposure-related unconformity (Figs. 4, 5). Subsequently, the stratigraphic positions of the δ¹³C excursions are indicated to pinpoint the LKE and UKE within the astrochronologic framework (blue and orange boxes on Fig. 6). The onset of LKE is within Frasnian LEC 15 (close to the maximum of that cycle), and it terminates within the lower part of LEC 16. The onset of the UKE is in the lower part of the last Frasnian LEC 17 (Figs. 4, 5). In section W4, the minor unconformity at the top of the parasequence underlying the onset of the UKE hampers the cyclostratigraphic assessment of the onset of the UKE. It seems that the UKE onset appears a bit lower (within LEC 16) in section W4 than in section C, but we consider section C to reflect the cyclostratigraphic patterns better and therefore place the onset of the UKE in the lower part of Frasnian LEC 17 (Figs. 5, 6). LEC 17 also contains the F–F boundary, which implies that only the lower part of the cycle is Frasnian in age, while the upper part is Famennian. Consequently, Frasnian LEC 17 and Famennian LEC 1 refer to the same 405-kyr eccentricity cycle. According to this definition, the UKE clearly terminates in the uppermost part of Famennian LEC 1 (Figs. 5, 6).

**DISCUSSION**

Determination of the timing of geologic and paleontologic events in the rock record is one of the most vexing issues in sedimentary geology, especially in deep time. The LKE and UKE were originally defined by the occurrence of distinctive black shale facies in the Upper Frasnian of Europe (Buggisch 1991, Joachimski and Buggisch 1993). Since both the LKE and UKE record biotic crises, the bases of these intervals are defined biostatigraphically and occur at the top of FZ 12 (the base of the Upper *Palmatoilepis* [Pa.] *rhenana* Zone) and near the top of FZ 13b (upper part of the Lower *Pa. linguliformis* Zone), respectively (Girard et al. 2005). Gaps in the geographic distribution of *Pa. linguliformis* lead to difficulty in identifying the UKE; therefore, proxy taxa are often employed as a work-around (Girard et al. 2005). Girard and Renaud (2007) developed a quantitative conodont-based approach that appears to be capable of distinguishing the positions of the event intervals in a variety of lithofacies.

δ¹³C Excursions and Defining the F–F Events

In addition to the faunal changes, both the LKE and UKE are characterized by significant positive δ¹³C excursions usually on the order of 4 to 5% (Joachimski and Buggisch 1993, Joachimski et al. 2002, Racki 2005). These isotopic excursions appear to occur synchronously and fundamental to the ongoing ecosystem changes (Over et al. 1997; Riquier et al. 2006, 2007; Whalen and Day 2008; Boyer et al. 2011, Whalen et al. 2015). However, many of these proxies vary with depositional setting, hampering their use as global markers (Algeo and Rowe 2012). Carbon isotopic data are one of the most commonly reported records from Late Devonian sections (Fig. 1). While the LKE and UKE are first and foremost biotic crises (McGhee 1988, Bambach et al. 2004, Stigall 2012), assembling detailed species richness data from many sections would be labor intensive. Wide varieties of geochemical data are available across the F–F interval, including elemental data that serve as proxies for detrital input, productivity, and redox conditions (Over et al. 1997; Riquier et al. 2006, 2007; Whalen and Day 2008; Boyer et al. 2011). Perturbations of the carbon cycle appear to have been globally synchronous and fundamental to the ongoing ecosystem changes (Joachimski et al. 2002, Buggisch and Joachimski 2006). Therefore, they constitute the most suitable proxy for evaluating the pattern and timing of the F–F events. As discussed earlier, we define the onset of the carbon isotope excursion to mark the beginning of the events, while the decline of δ¹³C values to a background level, which may differ from the pre-event background, marks the termination of the event (Fig. 4), similar to other positive excursions documented in the geologic record (Buggisch 1991; Joachimski and Buggisch 1993; Saltzman et al. 1998; Joachimski et al. 2002; Bergstrom et al. 2006; Cramer et al. 2006, 2008; Cramer and Saltzman 2007).

Surveying the shape structure of the δ¹³C excursions, recorded in both carbonates and organic matter, the LKE is commonly characterized by an abrupt onset (i.e., most of the δ¹³C excursion occurs over less than 1 m of stratigraphy), one or two positive peaks, and a gradual decline to background values. This is the pattern documented at the Benner, Bädesheimer Bach, Steinbach Schmidt, and Vogelsberg localities in Germany; Coumouc, France (Joachimski and Buggisch 1993, Joachimski et al. 2002); Baisha and Fuhe sections, south China (Chen et al. 2005); Windjana Gorge, Australia.
The UKE carbon isotopic records are more variable and complicated compared to the LKE. Many sections appear to record a relatively abrupt onset of the UKE $\delta^{13}C$ excursion, such as those in south China (Chen et al. 2005); New York State, USA (Murphy et al. 2000); western Canada (Wang et al. 1996, this volume); B¨udesheimer Bach, Germany; Kowala, Poland; Nevada, USA (Joachimski et al. 2002); and western Australia (Stephens and Sumner 2003). However, the $\delta^{13}C_{carb}$ record at Kowala, Poland, appears to record a bit more protracted onset to the excursion, and several other localities share this more protracted onset, including Bou Ounebdou, Morocco (Joachimski et al. 2002), and the classic face at Windjana Gorge and the Oscar Range, Australia (Stephens and Sumner 2003). The UKE interval at several localities records multiple positive and negative fluctuations of 2 to 3%, such as in south China, Windjana Gorge, Kowala, Poland, and section W4, western Canada. These excursions may be related to relative sea-level change, that will be discussed later herein. In many sections, picking the termination of the UKE is difficult, because the isotopic background following the event is 1 to 2% higher than that before the onset of the event. Here, we define the top of the UKE interval as either the location where $\delta^{13}C$ values return to background, or at the top of the Middle triangularis conodont zone if no discernible return to background levels was recorded. The thickness of the LKE and UKE intervals is strongly influenced by sedimentation rates, condensation, and omission surfaces or unconformities (Table 1).

Implications for Timing of the F–F Events

The LKE interval in section C is 8 m thick, while the UKE interval is 13.5 m thick (Fig. 4; Table 1). In section W4, the LKE interval is 7 m thick, and the UKE interval 13 m thick (Fig. 4; Table 1). The onset of the LKE occurs within Frasnian LEC 15, about half (section W4) to two thirds (section C) of the way through the cycle (Figs. 4–6). This is in good agreement with the cyclostratigraphic framework that has been constructed for the Kowala section in Poland, where the LKE black shale, which corresponds to an abrupt positive isotope excursion, occurs in the middle of LEC 15 (De Vleeschouwer et al. 2013). The end of the LKE isotope excursion in the studied sections occurs about one quarter into Frasnian LEC 16. This interpretation is solely based on section C, as it is likely that part of LEC 16 in section W4 is missing due to a minor exposure-related unconformity (Fig. 5).

The onset of the UKE occurs close to the start of Frasnian LEC 17. The UKE isotope excursion roughly covers this entire cycle, which can also be referred to as Famennian LEC 1. In the Kowala section (Poland), the UKE black shale also occurs right at the onset of the last Frasnian LEC. Hence, the cyclostratigraphic position of the onset of the UKE isotope excursion in western Canada is in perfect agreement with the position of the UKE black shale in Kowala, Poland (De Vleeschouwer et al. 2013). However, De Vleeschouwer et al. (2013) suggested that the time difference between the onset of the LKE and UKE amounts to ~800 kyr. Based on the results presented here, this estimate should be revised to 500 to 600 kyr. This new estimate is in very good agreement with a cyclostratigraphic study from south China by Chen et al. (2005). These authors indicated that the LKE represents ~250 kyr, the UKE represents ~500 kyr, and the time difference between the onset of both events was ~650 kyr. In the Kowala section (Poland), $\delta^{13}C_{carb}$ takes ~600 kyr to return to background values after the UKE (De Vleeschouwer et al. 2013).

The UKE isotope excursion, as delineated in the studied sections, occurs within one LEC and appears to have been ~250 kyr in duration (Figs. 5, 6). The UKE, as delineated in the studied sections, appears to have had a duration of about one full LEC, i.e., about twice the duration of the LKE (Figs. 5, 6). Several other lines of evidence also suggest a longer duration for the UKE. Joachimski et al. (2004, 2009) provided $\delta^{18}O$ data from conodont apatite, and van Geldern et al. (2006) reported $\delta^{13}C$ and $\delta^{18}O$ data from brachiopods from many localities that record a positive isotope excursion over a very narrow interval for the LKE, while the excursion associated with the UKE is spread out over a thicker stratigraphic interval. Data from these studies are rather coarse and result in duration estimates on the order of 0.5 and >1 Myr for the LKE and UKE, respectively, which are similar to duration estimates based on the extent of carbon isotope excursions (Goddéris and Joachimski 2004) and conodont biostratigraphy from the Benner and Schmidt quarry sections in Germany (Joachimski and Buggisch 1993), the latter of which provides the only late Frasnian radiometric tie point in the most recent geologic timescale (Becker et al. 2012).

The cyclostratigraphic estimates from western Canada (this study), southern China (Chen et al. 2005), and Poland (De Vleeschouwer et al. 2013) agree quite well, given the different methods used to define and delineate the events and the varying accumulation rates and depositional settings between localities. These data, along with $\delta^{13}C$ trends from other localities, indicate a relatively rapid initiation and termination of the LKE, while the UKE locally records rapid initiation but requires a longer time for termination of the $\delta^{13}C$ excursion. Post-UKE $\delta^{13}C$ background levels remain higher than pre-event values, indicating continued carbon burial into the Famennian (Figs. 1, 4). The association of these $\delta^{13}C$ excursions with interpreted changes in sea level and tropical sea-surface temperature (from $\delta^{18}O$ of conodont apatite; Joachimski et al. 2004, 2009) implies a relationship with climate change (Fig. 4).

Relationship Between Sea Level, Climate, and Astronomical Forcing

The LKE and UKE both appear to have been intimately related to climate and associated sea-level changes. Alberta sequence 8 and the LST and TST of sequence 9 comprise seven parasequences, each indicating higher-frequency sea-level events (Fig. 5). Sequence 8 was initiated during T–R cycle Id2, while sequence 9 was deposited within cycle Id3 (Figs. 4, 5) (Johnson et al. 1985, 1996; Day 1998; Whalen and Day 2010).

Sections C and W4 in western Canada record upper-slope–outer-ramp and platform-top depositional environments, respectively. The different depositional processes and the likelihood of subaerial exposure of the platform top likely influence the records of these two events. Deposition of the Simla Mbr. along the slope and outer ramp at section C appears to have been continuous, although the thin black shale interval that marks the base of the Sassenach Fm., F–F boundary, and the UKE (Fig. 6) is very likely condensed compared to the rest of the section. The Ronde Mbr. at section W4 was deposited under initially subtidal environments that shoaled to peritidal, and it displays evidence of subaerial exposure, including minor truncation of beds and an undulating erosional surface at the top of one parasequence in the middle of the LKE interval (Figs. 4, 5). The UKE at section W4 is within the Sassenach Fm. and is entirely peritidal facies. The $\delta^{13}C$ record at section W4 was likely influenced by episodes of subaerial exposure, as parasequence boundaries usually record a positive shift, likely influenced by exposure-related meteoric diagenesis (e.g., Immenhauser et al. 2003), followed by a negative swing that could be related to transgressive reworking and oxidation of previously deposited organic matter (Fig. 4).

The record of the LKE at section C records a relatively abrupt initial positive excursion followed by a gradual decline to background levels. Section W4 records a small initial abrupt positive excursion followed by a gradual decrease to background levels. The $\delta^{13}C$ record at section W4 is within the Sassenach Fm. and is entirely peritidal facies. The $\delta^{13}C$ record at section W4 was likely influenced by episodes of subaerial exposure, as parasequence boundaries usually record a positive shift, likely influenced by exposure-related meteoric diagenesis (e.g., Immenhauser et al. 2003), followed by a negative swing that could be related to transgressive reworking and oxidation of previously deposited organic matter (Fig. 4).
by a gradual positive excursion that ends abruptly. This variability is likely related to the increasingly peritidal conditions in section W4 (Fig. 4).

The record of the UKE at section C is relatively well defined, with an abrupt initial positive excursion and a gradual drop to less positive values (Fig. 4). The UKE in section W4 displays much more variability, with significant closely spaced positive and negative shifts in the lower portion of the UKE interval (Fig. 4). Again, we interpret this difference as reflecting the different depositional settings between the two sections. The outer-ramp to slope environment at section C did not experience exposure-related meteoric diagenesis, and thus it does not display successive positive-negative isotopic shifts, as recorded in peritidal facies in section W4.

The similarity in the astronomical interpretation for the two sections reveals comparable patterns indicating that the record at W4, while influenced by exposure events, is largely complete. We note one incidence where a minor unconformity is indicated by truncated beds and an undulating erosional surface within the LKE interval, but it appears that only a portion of LEC 15 was lost during exposure-related weathering.

The LKE and UKE both begin during LST or TST, and their onsets occur close to a minimum in their respective Frasnian LECs (Figs. 5, 6). This is in excellent agreement with a Late Devonian numerical climate modeling study, suggesting that an eccentricity minimum is a favorable astronomical configuration for a relatively cool and dry global climate (De Vleeschouwer et al. 2014b). Hence, we suggest that rapid climatic warming and relative sea-level rise, incited by an increasing eccentricity of Earth’s orbit, drove the Kellwasser events. Several parasequence boundaries in the latest Frasnian interval correlate with the descending limbs of LECs (13, 14, 15, and 16), indicating eccentricity minima during sea-level lows (Fig. 6). Maximum transgression is associated with maximal eccentricity during both Kellwasser events (De Vleeschouwer et al. 2013).

According to our cyclostratigraphic framework, the UKE is clearly related to the 405-kyr eccentricity maximum of Frasian LEC 17 (Figs. 5, 6). The LKE on the other hand, which is of shorter duration, was probably related to a 100-kyr eccentricity maximum within Frasnian LEC 15 (Figs. 5, 6).

Late Devonian global warming was also associated with a wetter global climate and thus with higher rates of terrestrial weathering and delivery of detritus to epeiric seas (Riquier et al. 2006, Whalen et al. 2015). This, in conjunction with the increase in vegetated land area and total root mass and concomitant effects on weathering rates from the Frasnian into the Famennian (Algeo and Scheckler 2010), resulted in detrital-driven nutrification, high rates of primary productivity, and development of low-oxygen conditions, all contributing to the biotic crisis (Joachimski et al. 2001, Riquier et al. 2006, Bond and Wignall 2008, Whalen et al. 2015). Subsequent cooling during the early Famennian was the result of a net carbon transfer from atmospheric CO2 towards the types of organic-rich shales that exemplify the Kellwasser events. The δ13C records from conodonts and brachiopods lend support to this climatic interpretation (Joachimski et al. 2004, 2009; van Geldern et al. 2006). The two events, however, were qualitatively different in that the KKE records demonstrate relatively rapid warming and cooling, whereas the UKE indicates rapid warming followed by protracted cooling into the early Famennian (Joachimski et al. 2004, 2009; van Geldern et al. 2006). This chronology is supported by our cyclostratigraphic analysis as well as the general thickness trends for the positive isotopic excursions documented globally (Table 1).

Although the main triggering mechanism for the Kellwasser events remains under debate, the growing consensus on the relative timing and duration of these events demonstrates that astronomical climate forcing played a major role in the timing and pacing of these environmental perturbations. Indeed, a comparison between isotopic data and numerical climate modeling results suggest an important role of astronomical forcing in bringing the climatic and environmental system, prone to anoxic bottom water conditions, past its tipping point (Fig. 16 in De Vleeschouwer et al. 2014b).

CONCLUSIONS

The Late Devonian records some of the most profound geologic, climatic, and biotic events of the Paleozoic, including a greenhouse-to-icehouse transition, eustatic sea-level changes, orogenesis, widespread marine anoxia, and a severe stepwise biotic crisis that ranks as one of the “big five” mass extinctions.

We combine records of sequence stratigraphy, stable carbon isotopic data, and time-series analysis of high-resolution MS data to provide insight into the pattern and timing of the F–F events. Placing the isotopic excursions in a cyclostratigraphic framework (De Vleeschouwer et al. 2012b), we demonstrate that the LKE began within Frasnian LEC 15 and terminated in the lower portion of Frasnian LEC 16, implying a duration of 200 to 250 kyr. The onset of the UKE occurs in the lower part of Frasnian LEC17 and terminates at the end of that same cycle in the earliest Famennian (Frasian LEC 17 and Famennian LEC 1 are synonymous). Therefore, the duration of the UKE isotope excursion in western Canada is estimated to be ~400 kyr. According to this new estimate, the onset of the LKE isotope excursion and the onset of the UKE isotope excursion are separated by 500 to 600 kyr. Our analysis is supported by cyclostratigraphy from a coeval section—the cyclostratigraphic position of the onset of the UKE isotope excursion in western Canada is in perfect agreement with its position within black shale in Kowala, Poland (De Vleeschouwer et al. 2013).

This analysis, along with δ13C trends from other localities, provides a view of a relatively rapid initiation and termination of the LKE. The UKE locally records rapid initiation but required a longer time for termination of the δ13C excursion, and background levels remained higher than pre-event values, indicating continued carbon burial during the early Famennian. The associated changes in tropical sea-surface temperature (from δ18O of conodont apatite) imply a relatively rapid rise and drop in temperature during the LKE, followed by a rapid rise and protracted drop in temperature associated with the UKE (Joachimski et al. 2004, 2009; van Geldern et al. 2006).

The cause of the F–F biotic crisis remains contentious, but an emerging consensus proposes that it was linked to the development of terrestrial forests and associated changes in root and soil development that influenced weathering and, through eutrophication, the redox state of many epeiric seas around the world (Algeo et al. 1998; Joachimski et al. 2001, 2002; Algeo and Scheckler 2010). Relatively abrupt warming intervals and transgression of low-oxygen waters into shallow platform environments during these two latest Frasnian sea-level events appear to have been important drivers of the biotic crisis (Hallow and Wignall 1999, Joachimski et al. 2002, Buggisch and Joachimski 2006, Bond and Wignall 2008, Stiggall 2012). Transgressive events also potentially influenced the biotic diversity trends during the F–F interval by facilitating the emigration of invasive species, i.e., ecological generalists that outcompeted specialists, reducing overall diversity (Stiggall 2012). This points to the complex interrelationships among climate, weathering, oceanography, and ecosystem dynamics that were severely perturbed during the F–F biotic crisis.

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