Palaeogeography, Palaeoclimatology, Palaeoecology 392 (2013) 272–280

Contents lists available at ScienceDirect

Palaeogeography, Palaeoclimatology, Palaeoecology

journal homepage: www.elsevier.com/locate/palaeo

A forest fire and soil erosion event during the Late Devonian mass extinction

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ARTICLE INFO

Article history:
Received 19 March 2013
Received in revised form 11 September 2013
Accepted 15 September 2013
Available online 21 September 2013

Keywords:
Mass extinction
Atmospheric oxygen
Combustion
Soil erosion
Devonian

ABSTRACT

The Late Devonian mass extinction occurred in a stepwise manner and culminated close to the Frasnian–Famennian (F–F) boundary (372 million years ago). Organic-molecular indices from marine sedimentary rocks at the Sinsin section, Belgium, indicate that the sequence of combustion of land vegetation, soil erosion, and anoxia–euxinia occurred close to this boundary. The increased concentrations of biomarkers indicating forest fire and soil erosion measured in the Sinsin section suggest that fire became widespread at this time, leading to various damaging consequences (increased runoff and oceanic anoxia) that caused marine extinctions. Magnetic susceptibility data in the Sinsin section indicate a relatively dry climate spanning the F–F boundary, which would have encouraged forest fires. The study of organic biomarkers presents several lines of evidence to link forest fire and soil erosion to the Late Devonian mass extinction.

1. Introduction

The Late Devonian mass extinction, one of the five greatest during the Phanerozoic, was characterized by a stepwise extinction of marine organisms (McGhee, 1996; Rohde and Muller, 2005). The Late Devonian corresponds to a period of great reduction of reefal growth (Kiessling, 2002, 2009; Alroy, 2010). Following an initial early rise after the Late Givetian coral and stromatoporoid extinctions, reefs expanded for the last time during the mid-Frasnian but declined markedly in the Late Frasnian (*rhenana* and *linguiformis* conodont Zones), below the Frasnian–Famennian (F–F) boundary. Globally, metazoan reefs were wiped out by the end of the Frasnian (Copper, 2002). Other marine animals such as ammonoids, trilobites, ostracods, and brachiopods were also severely affected during the F–F mass extinction (Sandberg et al., 1988; McGhee, 1989; Casier et al., 1996; House, 2002; Casier, 2008; Feist et al., 2009). The F–F boundary (372 million years ago) marks the culmination of the stepwise Late Devonian mass extinction (Algeo and Scheckler, 1998; House, 2002; Chen and Tucker, 2003) and the extinction rates among marine invertebrates are highest in lower latitudes (e.g. McGhee, 1989, 1996).

The F–F mass extinction is related to two oceanographic events, the Lower Kellwasser and the Upper Kellwasser Events, thought to be caused by development of widespread marine anoxia in shallow epicontinental seas (Johnson et al., 1985; Bond et al., 2004; Bond and Wignall, 2008). The Upper Kellwasser Event is marked by a positive anomaly of $\delta^{13}C$ in marine carbonates (Joachimski, 1997; Chen et al., 2005; Buggisch and Joachimski, 2006) and by an increase in atmospheric oxygen (Goddéris and Joachimski, 2004) associated with a warm Earth period (late Frasnian to early Famennian; Joachimski et al., 2009). It includes a short-term cooling spanning the F–F boundary (Joachimski et al., 2009). Extinctions during the Upper Kellwasser Event are more abundant compared to those occurring during the Lower Kellwasser Event (Racki, 1999). For example, more than 75% of all ostracod species were wiped out close to the F–F boundary in lower latitudes (Lethiers and Casier, 1999; Casier and Lethiers, 2001).

Bolide impacts had been proposed to explain the Late Devonian mass extinction (Claeys et al., 1992; McGhee, 1996), but evidence is scarce, such as iridium anomalies very low in concentration (Playford et al., 1984; McGhee et al., 1986; Wang et al., 1991; Over et al., 1997) compared to the Ir anomaly measured at the Cretaceous–Paleogene boundary (Sawlowicz, 1993) and rare occurrence of microtektite like-glass (Claeys et al., 1992). Low Os and Re abundances at the F–F boundary are inconsistent with long-term volcanism and bolide impact as potential Late Devonian mass extinction mechanisms (Turgeon et al., 2007). Os isotopic compositions from the F–F boundary in two sections are inconsistent with a large meteoritic component (Gordon et al., 2009).

The Late Devonian mass extinction occurred after the entry of the first forests in the Frasnian but before the first seed plants in the late Fammenian during a period characterized by the spread of vegetation in swamps, not in dry uplands (Algeo et al., 1995). The marine anoxia

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http://dx.doi.org/10.1016/j.palaeo.2013.09.008

may have been induced by the development of vascular land plants, responsible for the production and delivery of dissolved nutrients extracted by plant roots to the oceans causing an increase in nutrient supply to the sea (Algeo et al., 1995; Algeo and Scheckler, 1998, 2010). The marine anoxic and extinction events are likely to have been linked causally through transient nutrient pulses that caused eutrophication of semirestricted epicontinental seaways, stimulating marine algal blooms (Algeo et al., 1995; Algeo and Scheckler, 1998, 2010). However, there is few evidence of an increase in nutrient supply to the sea at the F–F transition.

Here we present organic geochemical evidence of forest fire and soil erosion during the Upper Kellwasser Event close to the F–F boundary, suggesting that a combustion and soil erosion event contributed to the Late Devonian mass extinction.

2. Geologic setting

The Sinsin section is composed of limestone, marl, and mudstone. We recognized 17 beds, named Beds 1 to 17 (Fig. 2a). The F–F boundary may be fixed at the base of Bed 12 in the section studied based on the base of the Lower *triangularis* Zone (a conodont biozone) (Sandberg et al., 1988). Beds 7, 9, 10, and 11 are laminated marl and mudstone. Beds 9, 10, and 11 are barren of invertebrate fossils larger than 10 mm, and bear the highest total organic carbon content (TOC: 0.1 to 0.2%) in the section, suggesting the cessation of carbonate deposition or an increased supply of mud from land (Fig. 2). Ostracods present in Bed 11 are composed of *Myodocopida*, a proxy for hypoxic water conditions (Casier, 2004, 2008). Beds 3, 5, 6, 8, and 14 to 17 contain brachiopod shells. The size of invertebrate fossils significantly decreases in the limestone Bed 8 in the upper part of the *linguiformis* Zone and recovers in the limestone Bed 14 in the Lower *triangularis* Zone. The extinction horizon is estimated to be the top of Bed 8 based on these fossil occurrences (Streel et al., 2000a,b; Fig. 2b). The Sinsin section was located on a continental shelf below fair-weather wave base (Casier and Devleeschouwer, 1995) at low latitudes (Fig. 1).

In the Hony section, the upper Frasnian is composed of calcareous gray mudstones and coquina layers containing *Palmatolepis linguiformis* conodonts. These beds are overlain by a 1.5 m thick bed of dark gray mudstones that contain no conodonts. The first appearance of *P. triangularis* is just above this bed and indicates a Famennian Age. The exact location of the F–F boundary in the Hony section is therefore uncertain, but is often fixed in a very thin bed of white claystone containing microtektite-like glass, about 1 m above the base of the 1.5 m-thick bed of dark gray mudstones, and consequently is assumed to be the last occurrence of *P. linguiformis* (Claeys and Casier, 1994). Streel et al. (2000a,b) demonstrate the abrupt disappearance of palynomorphs at the Frasnian–Famennian transition at Hony.

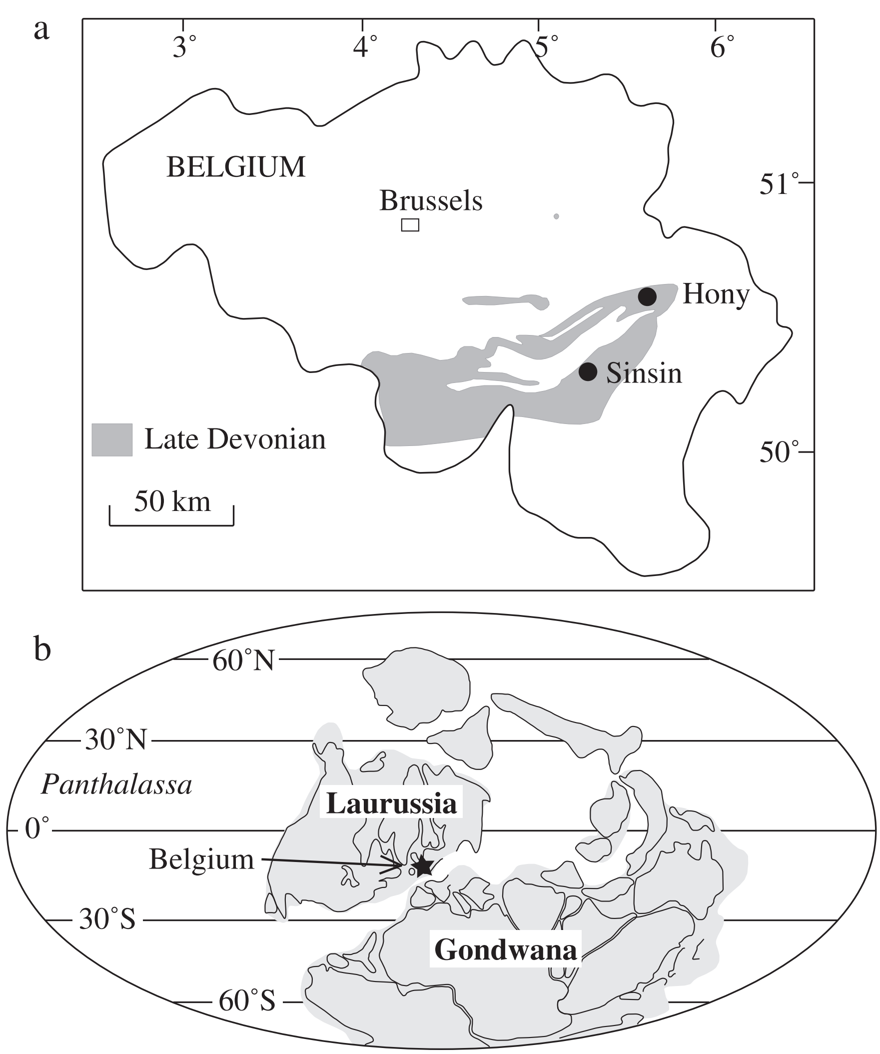

Fig. 1. Location of the studied sections. (a) Location of the Sinsin and Hony sections (modified from John et al., 2010). (b) Late Devonian paleogeography (Scotese and McKerrow, 1990) showing the position of Belgium.