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Devonian–Carboniferous boundary global correlations and their paleogeographic implications for the Assembly of Pangaea

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Abstract

The Devonian–Carboniferous transition was a time of diverse geologic change including both active orogeny and glacial–eustatic lowstands. Recent progress on biostratigraphy allows of unconformities and hiatuses which yields important information about major geologic events. We take only into account the data (mainly miospore stratigraphy) that can be directly or indirectly connected to the 'standard' conodont stratigraphy, focusing therefore on western Gondwana and Euramerica. Three time slices are considered: (1) the Late Famennian *trachytera* to *expansa*, (2) the mostly latest Famennian *expansa* to *sulcata*, and (3) the early to mid Tournaisian *sulcata* to *crenulata* Zones.

During the Late Famennian the miospore and megaflora distribution reflects contrasting climates for two regions that are at about the same latitude. We suggest that the emerging Variscan Mountains probably intercept the trade winds, making the northwestern Europe more arid than the eastern USA. A compilation of many localities around the Old Red Sandstone Continent demonstrates that the Late Famennian unconformities are mostly linked to the Antler and Hercynian orogenies but that Latest Famennian unconformities, being present almost everywhere, should be linked rather to glacial–eustatic lowstands. The Eohercynian orogeny in South America might well have induced a climate with glacial/interglacial cycles corresponding to expansion and melting of mountain glaciers. However, without any lithological evidence for glacial deposits, the three first Famennian transgressions (interglacials?) can only be explained by mountain glacier variations. On the contrary, significantly abundant biostratigraphic data have now become available in the Bolivian and Brazilian basins where all diamictites are dated in the latest Famennian LE–LN miospore Zones. We have therefore to accept that the only certainty of glaciers reaching the sea-level corresponds to the Early to Mid *praesulcata* Zone time-span. Geochemical work on brachiopods and matrix covering the *praesulcata* to *sulcata* shows large positive excursions in $\delta^{13}C$ and $\delta^{18}O$, suggesting changing oceanographic conditions of the global oceans during the Mid *praesulcata* Zone. The oxygen isotope trend may reflect a climatic response of a rapid, short and distinct glacial event during the upper Middle *praesulcata* Zone. The $\delta^{13}C$ maxima, more or less contemporaneous with the $\delta^{18}O$ maxima, might correspond to a major episode of C_{org} sequestration in the oceans analogous to those associated with increased marine fertility during Quaternary glacial phases. In East Greenland, which was low latitudes, in the southern hemisphere arid zone, latest Famennian climatic cycles can be recognised in terrestrial sediments. Here the arid conditions were temporarily replaced by some three successive humid episodes each defined by the flooding of the basin to give a lake. The last one corresponds to the Late *praesulcata* and *sulcata* Zones, i.e. to the eustatic global sea-level rise associated with the DCB. The coincidence of high sea and lacustrine conditions in the arid climatic zone is interpreted as resulting from a strengthening of the monsoon. The worldwide early-mid Tournaisian transgression is locally affected by the disintegration of the continental margin of northwestern Gondwana. From northern Brazil to the Middle East, erosional or nondepositional gaps occur almost everywhere.

Due to latitudinal climatic conditions, there were many more floral differences between northern and southern Euramerica than between southern Euramerica and western Gondwana. It is not until the (Late?) Visean and a new important glacial episode that the emergence of distinct floral provinces prevents intercontinental correlations.

Keywords: Famennian, Tournaisian, D–C boundary, stratigraphy, palaeogeography.

Introduction

This paper is a contribution to the IGCP Project 499 'Devonian land-sea interaction: Evolution of ecosystems and climate' (DEVEC) and it is based on the exceptional amount of biostratigraphical data available in the western part of the assembling Pangaea during the Devonian–Carboniferous transitional time. For other less well-documented regions we refer to Kalvoda (2002) for Siberia and Central Asia, and to Lethiers (1983), Metcalfe (2001), Young (2003) and Coen et al. (1996) for Eastern Gondwana, East and SE Asia.

When do we start and when do we end?

The Devonian–Carboniferous Boundary (DCB) is fixed at the first occurrence of the conodont *Siphonodella sulcata* at the locality of La Serre in the Montagne Noire in the south of France (Paproth et al., 1991). It marks a short event within the transition that started during the Late Devonian and ended during the Mississippian with a duration of less than 20 My. The variation of palaeogeography is not very sharp during this time interval as Pangaea had only moved some 10° to the north. In contrast, the climate, the faunas and the floras changed dramatically. To evaluate their merit for stratigraphic correlation, it is necessary to focus on these changes.

The Late Devonian map of Scotese (2000) that displays the lithologic climatic indicators shows a narrow 'tropical' (intertropical) zone bounded by two wider arid belts (Fig. 1). The Malvinokaffric faunas, isolated in Western Gondwana, had disappeared (Boucot, 1999). On the miospore evidence, two floristic realms are parallel to the latitude. Initially in equatorial Northern Euramerica there was the Frasnian *Archaeoperisaccus* Realm that was succeeded by the Mid Famennian *Cornispora varicornata* Realm (Streel et al., 2000). In arid tropical Southern Euramerica to cool temperate Western Gondwana an area with similar microfloras defines a further realm, the *Cymbosporites* Realm (here proposed for the first time) and permits easy correlation across the Rheic Ocean.

The equatorial province is not exactly centred on the equator of the Scotese map suggesting that, maybe, Pangaea has not shifted so far north at that time (Fig. 1). Additionally the similarity of the floras of Southern Euramerica and Western Gondwana implies the close proximity of these continents. Obviously the Rheic Ocean has not reopened during the Famennian as widely as suggested by Van der Voo (1988).

The Lower Carboniferous (Tournaisian and Viséan) map of Scotese (2000) shows lithologic climatic indicators which indicate that the 'tropical' belt has enlarged and the arid belts narrowed. The floral realms (Clayton, 1985) are much more diverse than during the Late Devonian. They are parallel to the palaeolatitude but also somewhat parallel to the emerging Variscan Mountains (Fig. 2).

Do we have a detailed biostratigraphy covering the Late Famennian to Early-Mid Tournaisian timespan, a period marked by the greatest changes?

Faunal and floral changes are the most obvious from the late Famennian to early-mid Tournaisian, a period of time (~10 My, Kaufmann et al., 2004) with a very detailed available biostratigraphy based on microfossils. Correlation between biozonations during this timespan was recently given by Matyja et al. (2000, fig. 17) for conodonts, miospores and ostracods, by Kalvoda (2002, fig. 26) for conodonts and foraminifera, by Legrand-Blain (2002) for conodonts, miospores, foraminifera, by Becker & House (2000) for ammonoids and by Melo & Loboziak (2003, fig. 17) for conodonts, miospores and chitinozoans (Fig. 3). The subdivision of the Famennian Stage is as suggested in Streel et al. (2000, fig. 6).

When does the Acadian Orogeny interrupt the east–west exchange between the Paleotethys and the remnant of the Rheic Ocean?

Copper (1986) proposed that the Appalachian Mountains significantly interrupted the east–west circulation as early as the Famennian using its cessation as an argument to explain the climate cooling and even the glaciation on Western Gondwana. Sandberg et al. (2002) assume that continental suturing would have already occurred in the Mid Devonian (at the Taghanic Onlap or the start of transgressive–regressive cycle IIa of Johnson et al., 1985). However, Sablock's map (1993, fig. 3) suggests that the east–west exchange was still operating during Famennian times. This was also shown by Becker & Kullmann (1996) where biotic exchange could only occur at times of high sea level between Euramerica, northern Gondwana and Prototethys.

The faunal data are rather in favour of the Copper hypothesis. Indeed foraminifera data (Kalvoda, 2001, 2002) indicate that the Frasnian *Eonodosaria* Complex which extended along the southern margin of Euramerica was succeeded by the Famennian *Quasi-*

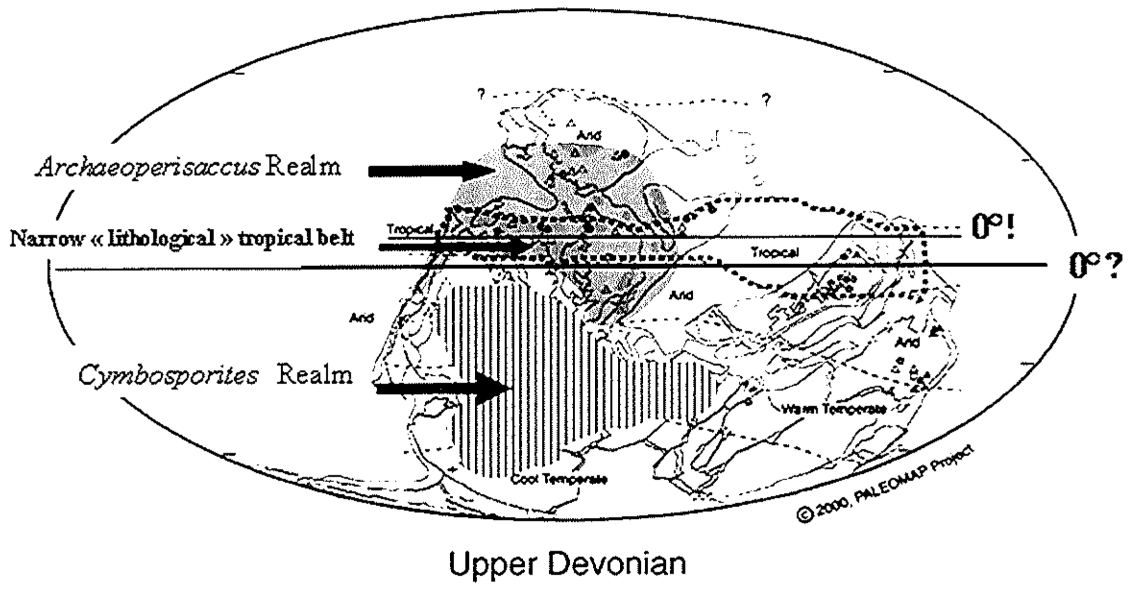


Fig. 1. Frasnian paleophytogeographical reconstruction after Streel et al. (1990, fig. 3b) based on Scotese Paleomap Project 2000: *Archaeoperisaccus* Realm is equatorial, *Cymbosporites Realm* is (sub-) tropical to sub-polar. The lithological data show a narrow 'tropical' belt and wide arid belts. Elements of the flora show an intertropical pattern, centred on the 'tropical' belt as delineated by lithological data, not exactly on the equator as proposed by the map. The *Cymbosporites* Realm occurs on both Southern Euramerica and Western Gondwana, which implies close proximity of continents.

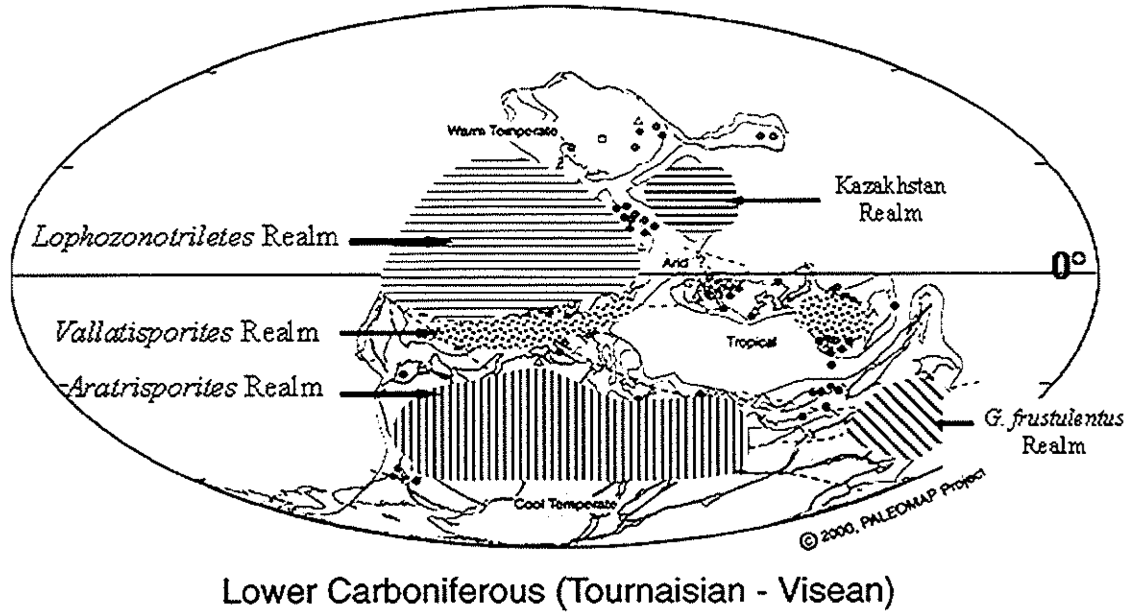


Fig. 2. Tournaisian–Visean paleophytogeographical reconstruction after Clayton (1985), slightly modified, based on Scotese Paleomap Project 2000. Pangaea has moved northward. The 'Tropical' belt has enlarged and the arid belts narrowed. The five floral realms are rather parallel to latitudes but might be somewhat controlled by the emerged Variscan Mountain Range.

endothyra Complex restricted to the east and south-eastern part of that continent (Fig. 4). The same is true for both ostracods and stromatopores which show south-eastern Euramerica distribution patterns during the Famennian (Mistiaen et al., 1998). The view is supported by Averbuch et al. (2005) where a synthesis of indicators of continental deformation and compression shows the widespread onset of tectonism in the Frasnian and continuing through the Famennian. The emergence of mountain ranges during the Famennian might also explain a climatic dependence of these distribution patterns. (See also below.)

How to explain the Late Famennian arid climate in south-eastern Euramerica?

The Late Devonian map of Scotese (2000) displays lithologic climatic indicators that show many occurrences of calcrete on the south-eastern margin of Euramerica. This is especially true in and around the

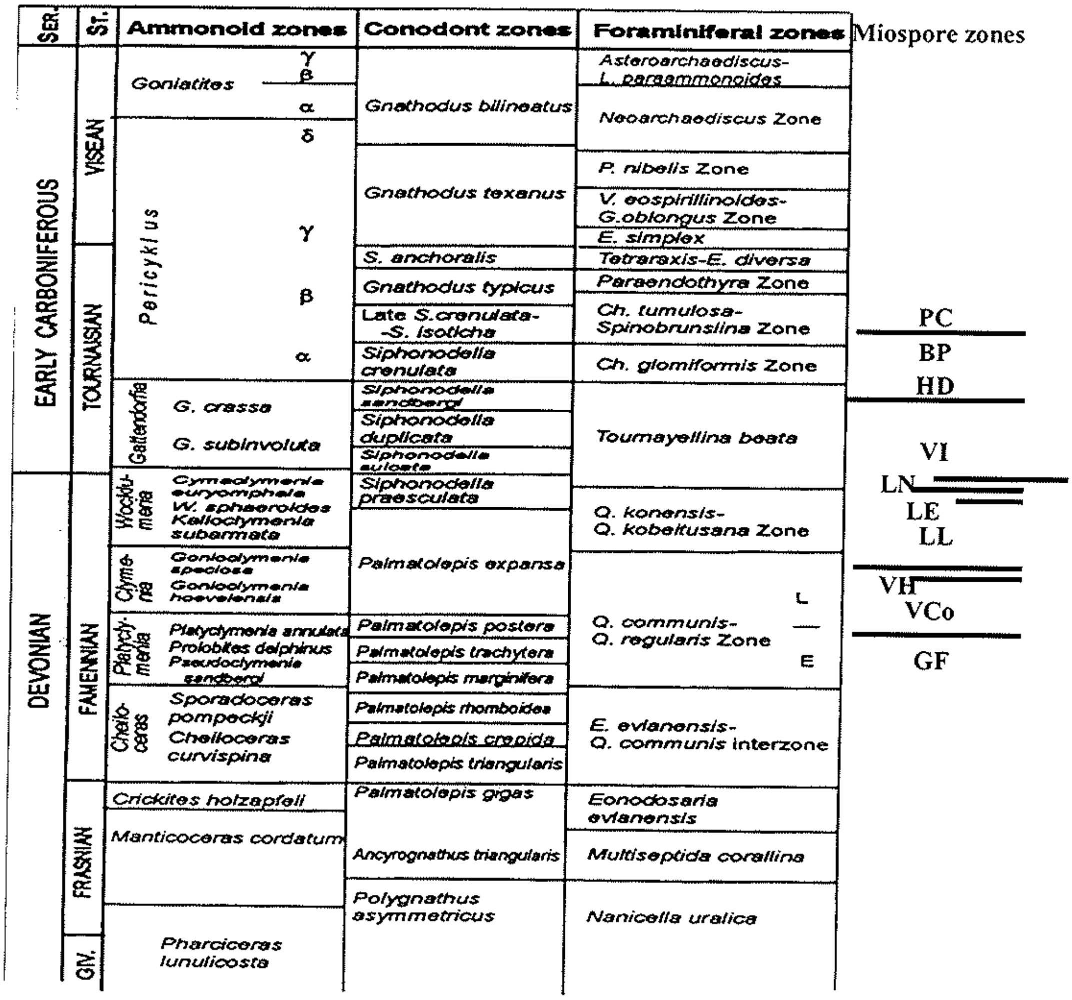


Fig. 3. A very detailed biostratigraphy is available across the Devonian–Carboniferous Boundary. Data on ammonoid, conodont and foraminiferal zones from Kalvoda (2002, fig. 26). Ammonoid zonation should be updated (see, for instance, Becker & House, 2000). Miospore zones after Streel et al. (1987) and Higgs et al. (1988). Late Famennian from *trachytera* to *expansa* Zones, Latest Famennian from *expansa* to *sulcata* Zones, early to mid-Tournaisian from *sulcata* to *crenulata* Zones.

present British Isles area. On the southern margin of Euramerica (Pennsylvania and New York State), there is very little evidence for this aridity. The miospore and megaflora distribution reflects these contrasting climates for two regions that are at about the same latitude.

The first occurrences of the VCo (*Diducites versabilis–Grandispora cornuta*) and the VH (*Apiculiretusispora verrucosa–Vallatisporites hystricosus*) Zones are delayed in the arid area. In Pennsylvania and New York State the VCo and VH Zones occurred first in the *marginifera* conodont Zone (Richardson & Ahmed, 1988), but appeared much later in Belgium (The VCo Zone in the *postera* conodont Zone and the VH Zone in the Mid *expansa* conodont Zone) (Fig. 5).

During the Late Famennian, the famous tree *Archaeopteris* is known from four species in Virginia and West Virginia (Hampshire Formation) but only by one in Belgium (Evieux Formation, Fairon-Demaret et al., 2001).

The explanation of this odd situation might be found in the Late Devonian orographic map of Blakey (2003) which suggests that the emerging Variscan Mountains probably intercept the trade winds, making the northwestern Europe more arid than the eastern USA. They probably play the same role as on Madagascar today whose mountains are partially responsible of the aridity of coastal Mozambique.

Why and how starts the change of climate?

During the latest Devonian new species of *Archaeopteris* occur in Ireland (Connery, 1999) but the most spectacular change in the vegetation cover is obviously the first occurrence of the cosmopolitan miospore *Retispora lepidophyta* which spread over most of the world during the late Late and Latest Famennian (LL to LN spore zones) suggesting a more equable climate than before (Fig. 6). As *R. lepidophyta* is considered to be produced by a plant living in near swamp conditions (Maziane et al., 2002), a wetter climate was probably responsible for a rather sudden large extension of these swamps. These changes are also seen in

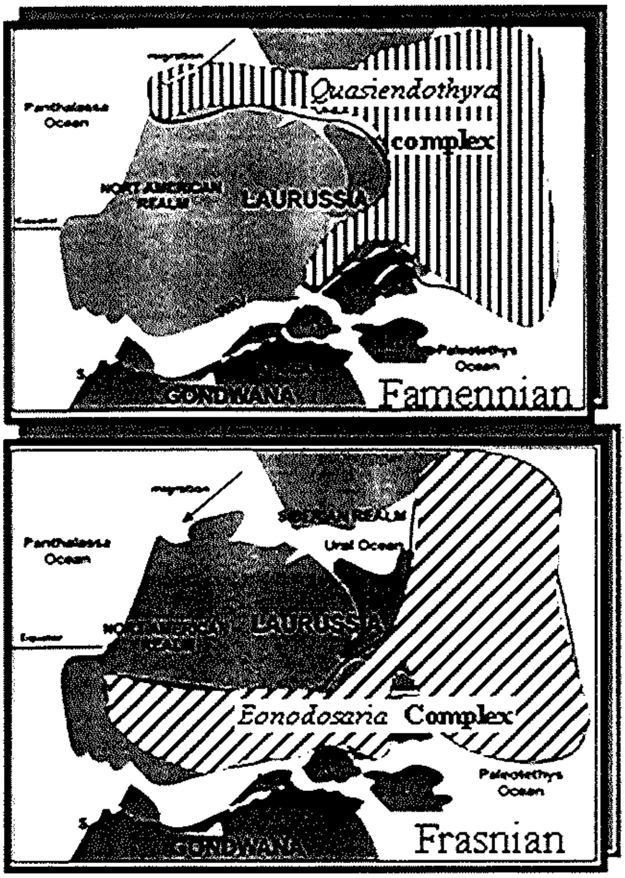


Fig. 4. Closure of east–west oceanic (Rheic-Paleotethys) circulation. Foraminiferal realms from Frasnian to Latest Famennian after Kalvoda (2001, 2002). The Frasnian *Eonodosaria* Complex has a wider east–west distribution than the Latest Famennian *Quasiendothyra* Complex.

East Greenland that lay at the heart of the Old Red Continent where *R. lepidophyta* occurs in very significant abundances (Marshall et al. 1999), during humid episodes but effectively disappears during arid intervals.

The change of climate has probably induced also a change in the rhythm and characteristics of sedimentation. One can differentiate Late and Latest Famennian climates in this respect using the suggestion of Ettensohn & Pashin (1997) that attention to the timing and plan of the unconformities may provide ways of discerning tectonic and climatic controls on their respective origins. Unconformities generated by pure eustasy are ideally of interregional extent, whereas unconformities generated by tectonism reflect more local factors associated with the evolution of sedimentary basins. Checking more than 30 localities around the Old Red Sandstone Continent (Fig. 7) one can demonstrate that the Late Famennian unconformities are mostly linked to the Antler and Hercynian orogenies but that the Latest Famennian unconformities, being present almost everywhere, should be linked rather to glacial–eustatic lowstand(s) (Fig. 8).

Dalmayrac et al. (1980) have demonstrated the importance of the Eohercynian orogeny in the South American Basins. Dating unconformities around the DCB in the same area, Lopez-Gamundi & Rossello (1993) have shown that all they have in common is their Famennian age.

Abrupt unconformities are recorded in the Late Devonian–Early Carboniferous from southern Peru (Isaacson & Sablock, 1990), Bolivia (Diaz-Martinez et al., 1999) and northern Argentina (Lopez-Gamundi & Rossello, 1993).

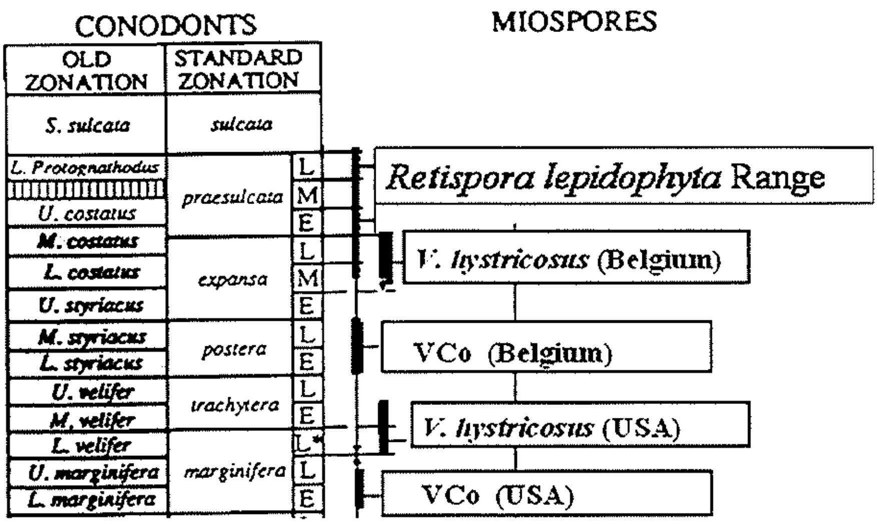


Fig. 5. Miospore data during the Late and Latest Famennian, after Richardson & Ahmed (1988) and Streel & Loboziak (1994, 1996). Conodont zonations after Ziegler & Sandberg (1990). Below the *Retispora lepidophyta* Range Zone, the diachronism of the first occurrence of the VCo (*Diducites variabilis–Grandispora cornuta*) Zone and the *Vallatisporites hystricosus* species, respectively in Belgium and USA, is demonstrated.

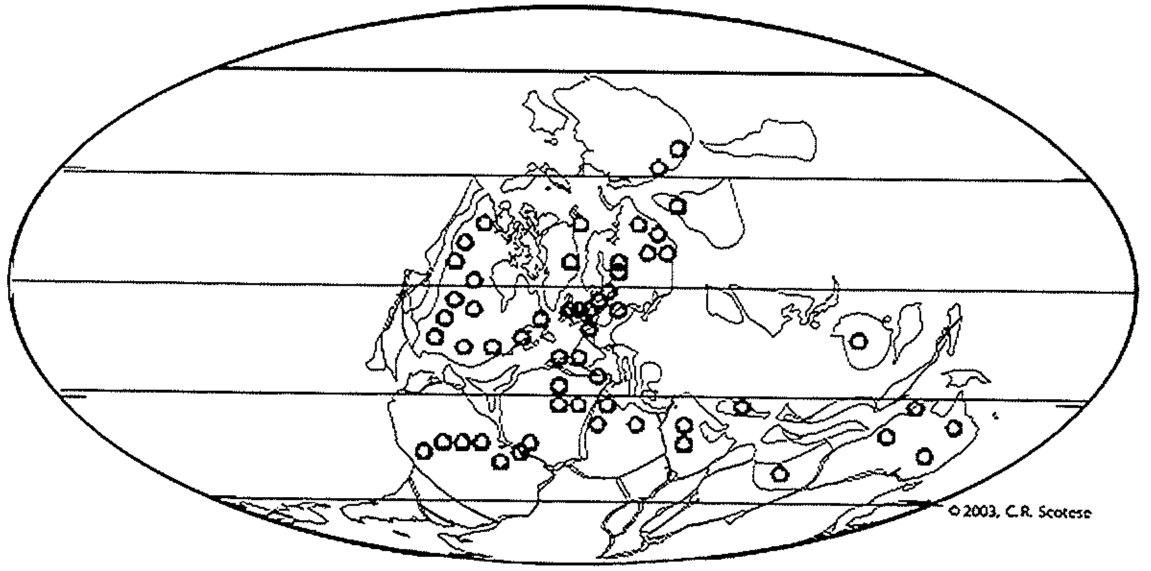


Fig. 6. Late Late and Latest Famennian *Retispora lepidophyta* assemblages distribution on Early Carboniferous (mid Mississippian) 2003 Scotese map. Cosmopolitan miospore assemblages spread over most of the world corresponding to a more équable climate.

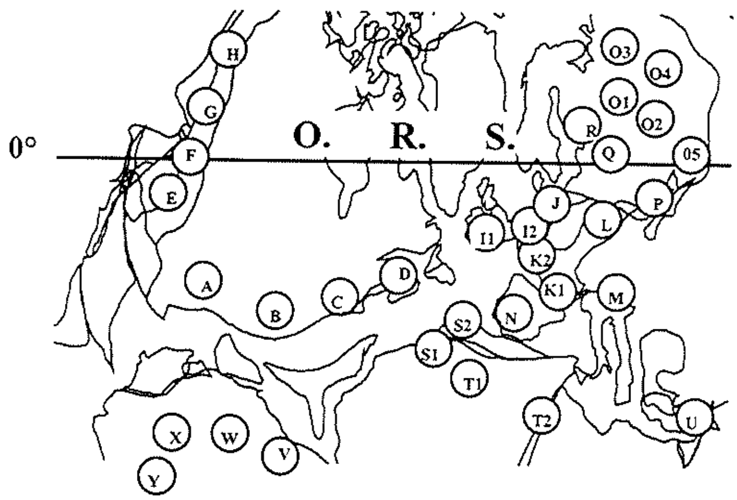
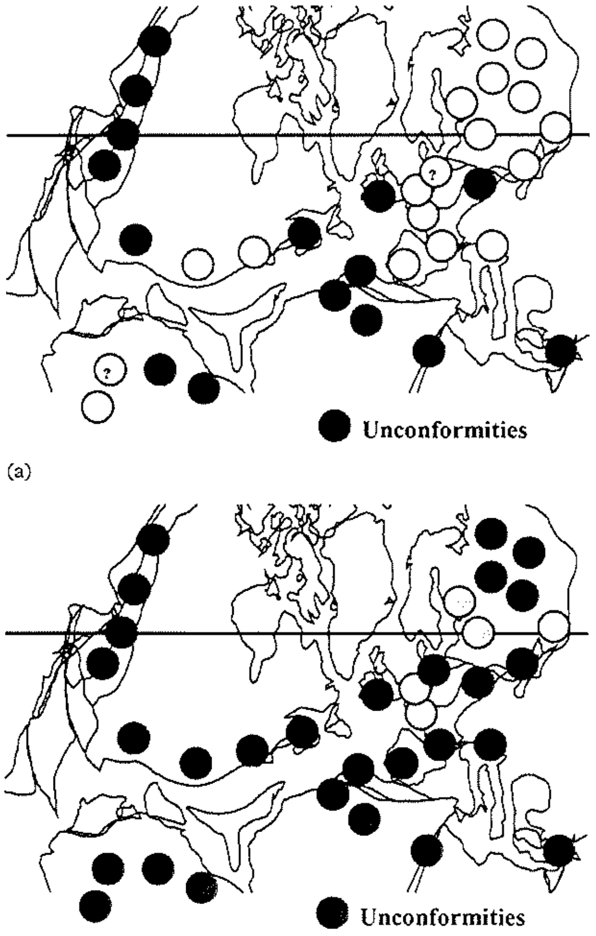


Fig. 7. Areas with available detailed sedimentology/biostratigraphy around the Old Red: Sandstone Continent during *trachytera* to *crenulata* timespan on Early Carboniferous mid Mississippian 2003 Scotese map. **A**: Ettensohn & Barron (1981), Ettensohn & Elam (1985), Ettensohn et al. (1988); **B**: Coleman & Clayton (1987), Eames (1974), McGregor (1979), Molyneux et al. (1984); **C**: Cecil et al. (2002), Ettensohn & Pashin (1997), Faill (1997), McGregor (1970), Richardson & Ahmed (1988), Streel & Traverse (1978); **D**: Martel et al. (1993); **E**: Gutschick & Rodriguez (1979), Gutschick et al. (1980), Sandberg et al. (1972), Sandberg & Gutschick (1979), Sandberg et al. (1980, 1989); **F**: Savoy (1992); **G**: Playford & McGregor (1993), Richards et al. (2002); **H**: Braman & Hills (1992); **I1, I2**: Friend et al. (2000), Higgs et al. (1988), Marshall & House (2000), Naylor et al. (1989), Streel (1986), Streel et al. (1987); **J**: Burmann (1975, 1976, 2001), Carson & Clayton (1997), Matyja (1993), Matyja et al. (2000), Olempska (1997), Racki & Turnau (2000), Stempien-Salek (2002), Turnau (1978); **K1, K2**: Morzadec et al. (1975), Morzadec & Streel (1980), Morzadec et al. (2000); **L**: Chlupac et al. (2000), Isaacson et al. (1999), Kalvoda & Kukal (1987); **M**: Kreutzer et al. (2000); **N**: Garcia-Alcalde (1998), Garcia-Alcalde et al. (2000), Gonzales et al. (2002), Oliveira et al. (1986); **O1** to **O5**: Byvsheva et al. (1984), Rzhonsnitskaya (2000), Sapel'nikov et al. (2000), Simakov (1992); **P**: Rzhonsnitskaya & Mamedov (2000); **Q**: Avkhimovitch et al. (1988, 1993), Golubtsov et al. (2000); **R**: Luksevics et al. (1999), Mark-Kurik (2000); **S1-S2**: Bultynck & Walliser (2000), El Hassani & Benfrika (2000), Rahmani-Antari & Lachkar (2001), Vachard et al. (1991), Wendt (1985), Wendt et al. (1984); **T1, T2**: Boumendjel et al. (1988), Conrad et al. (1986), Coquel & Latreche (1989), Grignani et al. (1987), Legrand-Blain (1993, 2002), Moreau-Benoit (1988), Moreau-Benoit et al. (1993), Paris et al. (1985); **U**: Higgs et al. (2002), Ravn et al. (1994); **V–Y**: Loboziak et al. (2000), Melo & Loboziak (2003).

The Eohercynian orogeny in South America might well have induced a climate with glacial/interglacial cycles corresponding to expansion and melting of mountain glaciers.

Whether enough water was stored as ice in mountain ranges to have induced significant regression marine cycles is as yet unknown. Whatever the answer, Sandberg et al. (2002) did not hesitate in subdividing the TR cycles IIe and IIf of Johnson et al. (1985) into interglacial/glacial events. However, there is as yet, no direct evidence for such mountain glaciers from the sedimentary record. In addition, the low lati-



(a)

(b)

Fig. 8. Timing and plan of unconformities provide ways of discerning tectonic versus climatic control. (a) Late Famennian (*trachytera* to *expansa*) unconformities are mostly linked to Antler and Hercynian orogenies. (b) Latest Famennian (*expansa* to *sulcata*) unconformities, being present almost everywhere, should be linked to glacial–eustatic lowstand.

tude isotope record provides no evidence for an earlier (pre-*expansa*) glaciation (Joachimski et al., 2004) nor do the results of a Late Devonian General Circulation Model (GCM) of climate (Ormiston & Oglesby, 1995).

What are the lithologic evidences of Gondwanan glaciation at the end of the Devonian?

Available evidence of glaciation in Brazil includes diamictites with striated, faceted and polished pebbles, rhythmites with dropstones, erratic boulders, striated pavements, and glacially deformed sedimentary rocks (Caputo in Streel et al., 2000). Diamictites were first accurately dated, (1) in western Bolivia in the LE (*Retispora lepidophyta–Indotriradites explanatus*) – LN (*Retispora lepidophyta–Verrucosisporites nitidus*) Zones from the lower part of the Cumana Fm at Lake Titicaca by Vavrdova et al., 1991, 1993 and (possibly LL?) (*Retispora lepidophyta–Knoxisporites literatus*) – LN Zones in the Torrega Fm in the Madre de Dios Basin by Vavrdova et al. (1996), (2) in south-east Bolivia in the LE Zone, or more accurately the LN zone (from the presence of the *Vallatisporites vallatus* group), in the Saipuru Fm, by Perez-Leyton (1991).

But no biostratigraphic data are available for the so-called Early Famennian (Colpacuchu Formation) or Early Tournaisian (Cumana Formation) diamictites as suggested by Isaacson et al. (1999, fig. 1).

The same is true for the Niger tillites mentioned by Scotese et al. (1999, p. 110) which are only known to be 'older than Viséan' (Lang et al., 1991) and for the so-called Devonian–Carboniferous Mambéré Formation from central Africa (cited by Isaacson et al., 1999, p. 241) that has only been indirectly dated by palaeomagnetic observations as 'close to Middle Devonian–Lower Carboniferous poles' (Censier et al., 1995, p. 24).

Significant and abundant biostratigraphic data have now become available from the Brazilian Basins, mostly from Amazonas (37 wells, hundreds of samples examined) where all diamictites (Upper Cururi and Oriximina Formations) are dated as LE–LN Zones (Loboziak et al., 2000, Melo & Loboziak, 2003). The diamictites of the Solimões and Parnaíba Basins can be dated as in the same miospore zones (Melo & Loboziak, 2003, fig. 16).

However, it is important to recognise that the difference between the successive LL, LE, LN miospore zones are based on the successive first occurrence of *Indotriradites explanatus* and *Verrucosisporites nitidus*. All taxa of the LL Zone range upwards into the succeeding LE and LN Zones. Since *V. nitidus* was less widely distributed than *I. explanatus*, the distinction between the LE and LN Zones is uncertain in the absence of the nominal species of the LN Zone (Melo et al., 1999). Another consequence is that the noted absence of the LL Zone in the Brazilian basins (Melo & Loboziak, 2003) does not imply that the glacial episode did not exist at that time. Indeed the abundant reworked palynomorphs found in Brazilian LE–LN Zones, which contain many Early and Middle Devonian miospores, might also include material from the LL Zone (that lacks its own characteristics), formerly deposited but eroded by the last glacial episode of the LE–LN major glacial deposits. But we know that the abundant reworked miospores present in diamictites do not contain characteristic species of the Famennian pre-*lepidophyta* miospore assemblages older than the VCo Zone (Streel et al., 2001) i.e. older than the *marginifera* or *postera* conodont Zones. Streel et al. (2000) attribute the overall scarcity of Early Famennian miospore species in Brazilian and North

African basins to a climatically-driven crisis of the coeval land plant cover.

Consequently, in South America, we have to accept that the only certainty of glaciers reaching the sea-level corresponds to the Early to Middle *praesulcata* time-span. In South Africa and the Falkland Islands there are more subtle lines of evidence for high latitude glaciation. This includes a very persistent regressive sandstone (the Perdepoort or Stanley Quartzite) that occurs in the Witpoort Formation and immediately beneath the DCB as defined by Theron (1994) and Streel & Theron (1999). This sandstone indicates a drawdown of sea level and is believed to mark the pre-DCB glaciation. The subsequent transgression that marks the end of this regression is associated with rare diamictites and black shales. There is also evidence for an earlier glacial–interglacial cycle within the Witpoort Formation.

When did the glacial cycles start and how many are there?

Veevers & Powell (1987) suggested that the transgressive–regressive sequences in Euramerica reflect glacial episodes in Gondwanaland. Sandberg et al. (2002) have proposed several events relating to glacial cycles (Table 1). How far all these rises and falls of sea level are really of eustatic rather of tectonic origin is still difficult to assess.

In middle (tropical) latitude, events 14 and 15 (2nd and 3rd interglacial episodes) correspond to a rather arid period in south-eastern Euramerica. Corresponding transgression phases are not very sharp in Belgium for instance, but the *annulata* black shale equivalent is tentatively matched (by conodont correlation) to the Bon-Mariage Event (Thorez & Dreesen, 2002).

Event 16 in Western USA is obviously mainly controlled by the Antler orogeny. Indeed the Early *expansa* major transgression is followed by two minor transgressive pulses: the fifth and sixth transgressions of Sandberg et al. (1989) corresponding to the Mid and Late *expansa* Zone respectively. In Belgium, the Early *expansa* transgression is almost non-existent (see the Beverire Event in Thorez & Dreesen, 2002) but the Mid *expansa* Fontin Event in the Evieux Formation and, chiefly, the Mid to Late *expansa* Epinette Event, immediately below the Etroeungt Formation, are well developed. This Mid *expansa* event is well defined by the ammonoids and is referred to as the Dasberg Event (Becker, 1993). Another transgressive pulse (Van Steenwinkel, 1993a,b), the seventh transgression (?) i.e. the Hangenberg Black Shales (HBS) in Germany (the true Hangenberg Event of Walliser, 1984), known all over the world (Caplan & Bustin, 1999) in the Mid *praesulcata* Zone, should belong also to the Event 16 but seems to be poorly understood in the Western USA. Evidently, Event 17, the Mid *praesulcata* eustatic fall, refers to the Hangenberg post-event (Streel et al., 2000), the Hangenberg Shales and Sandstones in Germany.

Without any lithological evidence of glacial deposits, the three first 'interglacials' of Sandberg et al. (2002) could probably better explained by mountain glacier variations. The tectonic influence on the TR curves is however important. In contrast, there is lithological evidence for the major glacial deposits of the LE–LN zones and the fourth transgression can be subdivided in several phases (transgressions 5 to ?7 of Sandberg et al., 1989).

More evidence for temperature change within this important interval is becoming available from isotope studies of conodont apatite. That of Joachimski et al. (2004) record fairly stable temperatures through much of the pre-*expansa* Famennian interval that they studied. Kaiser et al. (2004) but as yet only available in abstract, only find evidence for cooling from the upper *presulcata* followed by warming into the *sulcata* that may represent Gondwana deglaciation. But these data are, to date, not yet fully available through the entire DCB sequence.

Table 1. Several events relating to glacial cycles, after Sandberg et al. (2002)

Event 18 Late Famennian mass extension within eustatic fall (loss of many Devonian species including Lazarus fauna).
Event 17 start of the major eustatic fall at the climax of Southern Hemisphere glaciation (biotic decline begins) in the Mid *praesulcata* Zone.
Event 16 eustatic rise at start of the 4th interglacial episode (Etroeungt Lazarus fauna suddenly appears in Northern Hemisphere) in the Early *expansa* Zone. This is also the base of the TR cycle IIf of Johnson et al. (1985).
Event 15 eustatic rise at start of 3rd interglacial episode (*annulata* black basinal shales in Germany) in Early *postera* Zone.
Event 14 eustatic rise at start of 2nd interglacial episode (Baelen mudmound formed in Belgium) in the Early *marginifera* Zone.
Event 13 eustatic rise at start of the 1st interglacial episode (*Cheiloceras* dark shales deposited) in the Mid *crepida* Zone.
Event 12 eustatic rise (producing initial post-extinction biotic radiation after melting of initial ice cap) in the Mid *triangularis* Zone. This is also the base of the TR cycle IIe of Johnson et al. (1985).

Were the Events 16 (interglacial) and 17 (glacial) of Sandberg et al. (1989) homogeneous periods?

In the tropical climate that, at that time, characterized what is now NW Europe, the Auxiliary Stratotype Section of the Devonian–Carboniferous Boundary (DCB) at Hasselbachtal and the Stockum section, both in the Sauerland in Germany, display well-documented sections from the HBS to the DCB. T-R cycles (Bless et al., 1993, fig. 4) or holostratigraphic intervals (Becker, 1996, tab. 2) and quantitative miospore analysis (Streel, 1999) allow a detailed subdivision of this probably short timespan. Evidence for a wetter climate may have started in the Drewer Sandstone in the upper part of the Wocklum Limestone (LE Zone) immediately below the overlying HBS. Following Van Steenwinkel (1993a, p. 678), the HBS (LN typical Zone) is a condensed unit created by sediment starvation and corresponds to a worldwide event that, in the Sauerland area, was characterized by basinal condensation during an interval when the rate of eustatic sea-level rise was at its maximum. In the overlying lower part of the Hangenberg Shales and Sandstones, miospores are more abundant and better preserved than in the HBS. Possibly the deposition was very rapid preventing oxidation of the palynodebris. In the upper part of the Hangenberg Shales and Sandstones, sandstones are progressively more abundant and the palynofacies corresponds to oxic conditions. Coarser sediment and oxic conditions result probably from river run-off after the development of a wetter climate. These wet climatic conditions matched the lowest sea-level condition. These probably adverse climatic and edaphic environmental changes strongly reduced the coastal lowland swamp and consequently reduced the plant community producing *R. lepidophyta* (the LN transitional Zone of Higgs et al., 1993). This plant community completely vanished immediately before the Stockum Limestone deposition, being replaced by a very different, poorly defined miospore zone, the VI (*Vallatisporites vallatus–Retusotriletes incohatus*) Zone.

Geochemical work (Brand et al., 2004) on brachiopods and matrix covering the *praesulcata* to *sulcata* Zones in the DCB GSSP exposed at La Serre, Montagne Noire (France) shows large positive excursions in $\delta^{13}C$ and $\delta^{18}O$ of the unaltered brachiopods, reflecting changing oceanographic conditions of the subtropical La Serre Sea and possibly of the global oceans during the Mid *praesulcata* Zone. The oxygen isotope trend may reflect a climatic response of a rapid, short and distinct glacial event during the upper Middle *praesulcata* Zone i.e. during the glacial event 17. (Further data from conodont apatite will become available from Kaiser et al., 2004.) The $\delta^{13}C$ maxima, more or less contemporaneous of the $\delta^{18}O$ maxima, might correspond to a major episode of C_{org} sequestration in the oceans analogous to those associated with increased marine fertility during Quaternary glacial phases. Increased dust supply to the oceans during these glacials results in increased plankton productivity and the preferential sequestration of $\delta^{12}C$ into marine sediments. The relation between such events and the establishment of icehouse conditions has been discussed by Caputo (1995) and Streel et al. (2000).

The eustatic global sea-level rise corresponding to the Tournaisian *sulcata* Zone is associated with rapid negative shift of the $\delta^{13}C$, a result that is paralleled by the terrestrial $\delta^{13}C$ record (Strauss & Peters-Kottig, 2003) where it is interpreted as reflecting a sudden spike in O_2/CO_2 ratio.

In Gauss Halvø, East Greenland which was low latitudes in the southern hemisphere arid zone (Marshall et al., 2002), latest Devonian climatic cycles can be directly recognised in terrestrial sediments. Here the arid conditions were temporarily replaced by some three humid episodes accompanied by flooding of the basin to give a lake. The first humid event corresponds to the transition from the LL (*Retispora lepidophyta–Knoxisporites literatus*) to the LE (*Retispora lepidophyta–Indotriradites explanatus*) miospore Zones, within the Early *praesulcata* conodont Zone, a time-span which corresponds in Eastern Belgium to a Highstand or slow regressive phase following the Epinette transgression (Maziane et al., 2002). The last and most intense humid episode is that of the 3 m thick Obrutschew Formation where the basin became occupied by a wide and deep lake. This event is coincident with the boundary between the LN and VI spore zones and marks the turn over in the terrestrial vegetation immediately below the DCB (Fig. 9). It corresponds to the Late *praesulcata* and *sulcata* Zones, i.e. to the eustatic global sea-level rise across the DCB. The coincidence of high sea and lacustrine conditions in the arid climatic zone is interpreted as resulting from a strengthening of the monsoon. The change to the higher insolation that caused the glacial collapse also being responsible for forcing the monsoon further into the arid zone. The analogy used is the temporary extension of wetter conditions into present day Saharan Africa and Arabia with each post glacial climatic maximum. This East Greenland continental climatic record suggests three humid-arid cycles occurrences in the late Latest Famennian (LE to LN spore Zones). Since the last and most intense of these coincides with the DCB and is hence implicated with a major deglacial event the implication is that the two earlier cycles also rep-

Fig. 9. Glacial cycles in an equatorial area (East Greenland, the Gauss Halvøarea) during the Early *praesulcata* to *sulcata* Zones after Marshall et al. (2002). A series of aridity events succeeded by intense lacustrine flooding episodes as the post-glacial monsoon reached its maximum. The last lacustrine flooding might correspond to the Late *praesulcata-sulcata* Zones. One of the aridity events corresponds to the transition from the LL (*Retispora lepidophyta–Knoxisporites literatus*) to the LE (*Retispora lepidophyta–Indotriradites explanatus*) miospore Zones, within the Early *praesulcata* conodont Zone.

resent episodes of cooling and warming. The intensity of the final event being rather a result of the strength of the warming rather than that of the cooling. What this evidence also shows is that these cycles occurred within the LL, LE and LN spores zones and show a greater resolution of climatic events than can be determined from the rather condensed marine sections in NW Europe. There is no evidence for any earlier Famennian humidity events of similar magnitude within the East Greenland section with the climate being generally arid. By analogy with other Devonian lacustrine intervals, the duration of humid events were of the order of several hundred thousand years.

When summarizing (Table 2) the results given above, it appears that the number of suggested glacial/interglacial cycles first of all depends on the time resolution level of the investigations. Events 12 to 16 occur within a much longer period of time (Sandberg & Ziegler, 1996) than the succeeding local events and these local events, a much longer period of time than the duration of climate phases recorded from E Greenland to S France. Obviously the Events 16 (interglacial) and 17 (glacial) were not homogeneous periods but were rather comparable to the step by step Hirnantian glaciation (Late Ordovician) which is well demonstrated by Ghienne (2003).

When did the deglaciation process start and what was its duration?

The deglaciation process started in the Late *praesulcata* Zone and continued in the *sulcata* Zone across the DCB. There is no evidence of some interruption in the process as suggested by Savoy (1992, fig. 3) who starts the marine transgressions of Ross & Ross (1985) at the base of the *duplicata* conodont Zone.

Until recently the isotope stratigraphy was much better known in the Mississippian than for the end of the Devonian. Data given by Bruckschen et al. (1999), Mii et al. (1999) and Saltzman (2002) show that there is no evidence of glacial cycles in the earliest Mississippian. A positive excursion of the $\delta^{13}C$ during the Late Mid Tournaisian conodont *isosticha* – late *crenulata* Zone in Western USA (Saltzman et al., 2000, Sandberg, 2002) is the earliest suggestion of a subsequent (local ?) Tournaisian glacial episode, triggered by the Antler Orogeny. Unconformities in Western Canada (part of the Banff Formation in Richards et al., 2002) during the *sulcata* to *sandbergi* Zones are better explained by local tectonism (see also Whalen et al., 2000).

The worldwide early-mid Tournaisian transgression (Lower Alum Shale Event of Becker, 1993) was locally affected by the disintegration of the continental margin of northwestern Gondwana. From Northern Brazil to the Middle East, erosional or nondepositional gaps occur almost everywhere (Legrand-Blain, 2002; Melo & Loboziak, 2003).

Consequences

During the Devonian–Carboniferous transition Euramerica and western Gondwana were always connected enough to allow direct exchange of floras. In the Early to Middle Tournaisian there were only a few species that were not common to both areas and these enable the recognition of an *Aratrisporites saharensis* Microflora in the Western Gondwana Realm. In addition, and due to latitudinal climatic conditions, there were many more floral differences between northern and southern Euramerica than between southern Euramerica and western Gondwana. It was not until the

Table 2. Suggested Famennian glacial/interglacial cycles (no time scale!). The events 12 to 16 occur within a much longer period of time (Sandberg & Ziegler, 1996) than the succeeding local events and these local events, a much longer period of time than the duration of climate phases recorded from E Greenland to S France. Obviously the Events 16 (interglacial) and 17 (glacial) were not homogeneous periods. B11 and B12 are based on D11 and D12 miospore facies (Streel, 1999); H9 and H10 are based on sedimentological arguments (Streel, 1999)

A	B	C	D	E	F	G	H	I	
Events in W USA	(? Global) TR	Conodont zones	Miospore zones	Local events in NW Europe	Formations in NW Europe	Equat. climate in E Greenland	Trop. climate in NW Europe	Subtrop. cl. in S France	
	eustatic rise (reduced cont. C_{org} supply)	*sulcata*	VI	(DCB)	Hangenberg Lmst in Germany	monsoon	wetter	less cold	16
	eustatic rise (floral crisis)	Late *praesulcata*	VI	in Germany	Stockum Lmst	monsoon	wetter	less cold	15
	start of eustatic rise	Late *praesulcata*	LN transitional	in Germany	Stockum Lmst	monsoon	wetter	less cold	14
	peak of major eustatic fall	Middle *praesulcata*	LN transitional	in Germany	Hangenberg Sh. & Sst.	rather arid	wet	coldest	13
	(transgressive)	Middle *praesulcata*	LN (hystricosus >)		Hangenberg Sh. & Sst. in Germany	monsoon ?	wet	less cold	12
Event 17	start of major eustatic fall	Middle *praesulcata*	LN (lepidophyta >)		Hangenberg Sh. & Sst. in Germany	rather arid ?	wet	?	11
	/?7th transgression in Germany	Middle *praesulcata*	LN	Hangenberg Ev.	HBS in Germany	monsoon	(lightly rainy)	warmer	10
	minor eustatic fall in Germany	Lower/Middle *praesulcata*	LE		Drewer Sst. in Germany	rather arid ?	(heavy rainy)	warmer	9
	start of minor eustatic fall	Lower *praesulcata*	LL/LE		Etroeungt in Belgium & N France	monsoon			8
	/6th transgression in W USA	Middle-Late *expansa*	VH-LL	Epinette Ev.	Evieux - Etroeungt in Belgium				7
	/5th transgression in W USA	Middle *expansa*	VCo	Fontin Ev.	Evieux in Belgium				6
Event 16	start of eustatic rise /4th interglacial	Early *expansa*	VCo	Beverire Ev.?					5
Event 15	start of eustatic rise /3rd interglacial	Early *postera*	VCo	Bon-Mariage Ev.?			rather arid		4
Event 14	start of eustatic rise /2nd interglacial	Early *marginifera*					rather arid		3
Event 13	start of eustatic rise /1th interglacial	Middle *crepida*							2
Event 12	eustatic rise	Middle *triangularis*							1

(Late?) Visean and a new important glacial episode that the emergence of distinct floral provinces prevents intercontinental correlations (Clayton, 1985; Van der Zwan, 1981; Van der Zwan et al., 1985). However, for Eastern Gondwana, its movement from a subtropical to subpolar position from the Late Devonian upwards accentuated the isolation of the flora characterized by the *Granulatisporites frustulentus* Microflora (Playford).

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