


# The Devonian-Carboniferous Boundary around the globe: a complement

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## Abstract

This article compiles data on the Devonian-Carboniferous Boundary successions from countries and regions, which have not been dealt with separately in this special issue. Data derive from different palaeocontinents on a large palaeoclimatic gradient from the southern high latitudes of western Gondwana through the palaeotropics into the temperate/boreal northern latitudes of Siberia. The quality of the data is variable, but often surprisingly good, although major uncertainties can persist and traditional positions of the boundary do not withhold modern stratigraphic results and concepts. Not all these regions have been on the forefront of the DCB discussions in recent years, but they provide invaluable insights into the regional and global dimensions of uppermost Devonian and lowermost Carboniferous changes in the biosphere and geosphere. Almost globally documented are faunal turnovers and collapse of ecosystems in the Hangenberg extinctions. Those allow the placement of the DCB even if index fossils are absent or scarce. However, it also shows the regional differences, which advocate for a holistic approach to the boundary by ensuring global correlation without relying on the conceptual dominance a single taxon, which might be less powerful outside its usual facies.

**Keywords** Africa · Asia · Australia · Europe · South America · Famennian · Tournaisian

## Introduction

In order to fulfil the ambition of this special issue of presenting a global perspective on the Devonian-Carboniferous Boundary (DCB) succession, this contribution presents brief overviews and data for regions and countries (Fig. 1), which have not been treated in the other contributions. The reader is referred to those contributions for the Montagne Noire (Feist et al. 2021, this issue), Belgium (Denayer et al. 2021, this issue), the Carnic Alps (Spalletta et al. 2021, this issue), China (Qie et al. 2021,

this issue), Czech Republic (Kumpan et al. 2021, this issue), Germany (Becker et al. 2021, this issue), Greenland (Marshall 2021, this issue), Iran (Königshof et al. 2021, this issue), Ireland (Graham and Sevastopulo 2021, this issue), Poland (Matyja et al. 2021, this issue), some parts of Russia and Kazakhstan (Kulagina et al. 2021, this issue), Sardinia (Corradini et al. 2021, this issue) and the USA (Over 2021, this issue).

These certainly short and partly incomplete overviews presented in this paper are also thought to give the interested reader a starting point for a more in-depth research on the DCB interval in those regions.

This does not mean that these regions and sections can be neglected for our understanding of the DCB interval. In fact, they are important cornerstones and test cases that a global picture for the understanding of the geological and biological history of the DCB interval can be developed, including our search for dividing it and defining the base of the Carboniferous. This overview also shows that it seems to be likely that important discoveries are still to be made in those regions, which sometimes have only been superficially studied, and/or modern data are not available or only limited to single sections. Although presented according to modern geography, these regional overviews encompass various

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This article is a contribution to the special issue “Global review of the Devonian-Carboniferous Boundary”

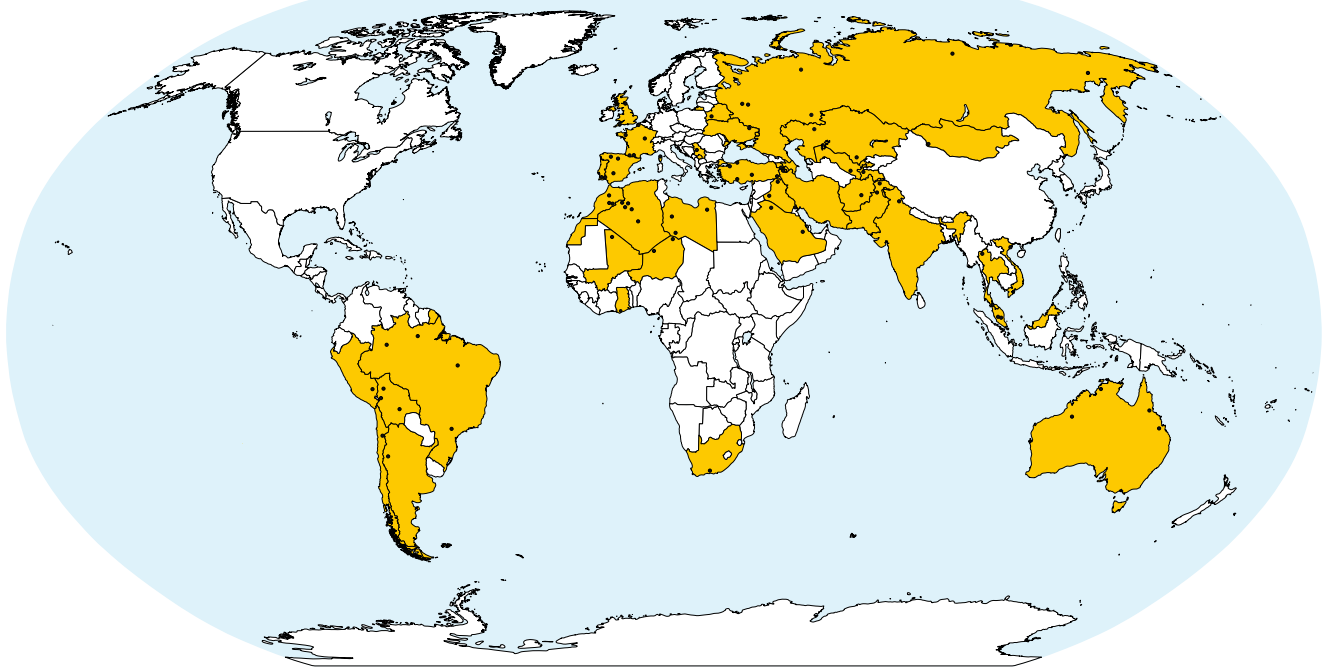
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**Fig. 1** Geographic overview on a modern world geography: the countries dealt with in this contribution are indicated in orange. The black dots indicate approximate positions of DCB sections in those countries. Note that a dot may represent several geographical rather close sections

palaeocontinents and palaeoenvironments, from the high latitudes of western Gondwana to the equatorial and temperate seas in the northern palaeohemisphere (Fig. 2).

The disparity in the abundance and precision of stratigraphic data has been a challenging problem for the authors of this contribution. Where possible, older data were re-interpreted and translated into modern taxonomic and stratigraphic nomenclature. The following abbreviations of genera are used in this paper. Conodonts: *Bi.*: *Bispathodus*; *Br.*: *Branmehla*; *Gn.*: *Gnathodus*; *M.*: *Mehlina*; *Pa.*: *Palmatolepis*; *Po.*: *Polygnathus*; *Pr.*: *Protognathodus*; *Ps.*: *Pseudopolygnathus*; *Rh.*: *Rhodalepis*; *Si.*: *Siphonodella*; Spores: *K.*: *Knoxisporites*; *R.*: *Retispora*; *Ru.*: *Rugospora*; *V.*: *Verrucosisporites*; *W.*: *Waltzispora*, Miospore zones: LL (*lepidophyta–literatus*), LE (*lepidophyta–explanatus*), LN (*lepidophyta–nitidus*), Rle (*lepidophyta*), LVa (*lepidophyta–vallatus*).

## South America

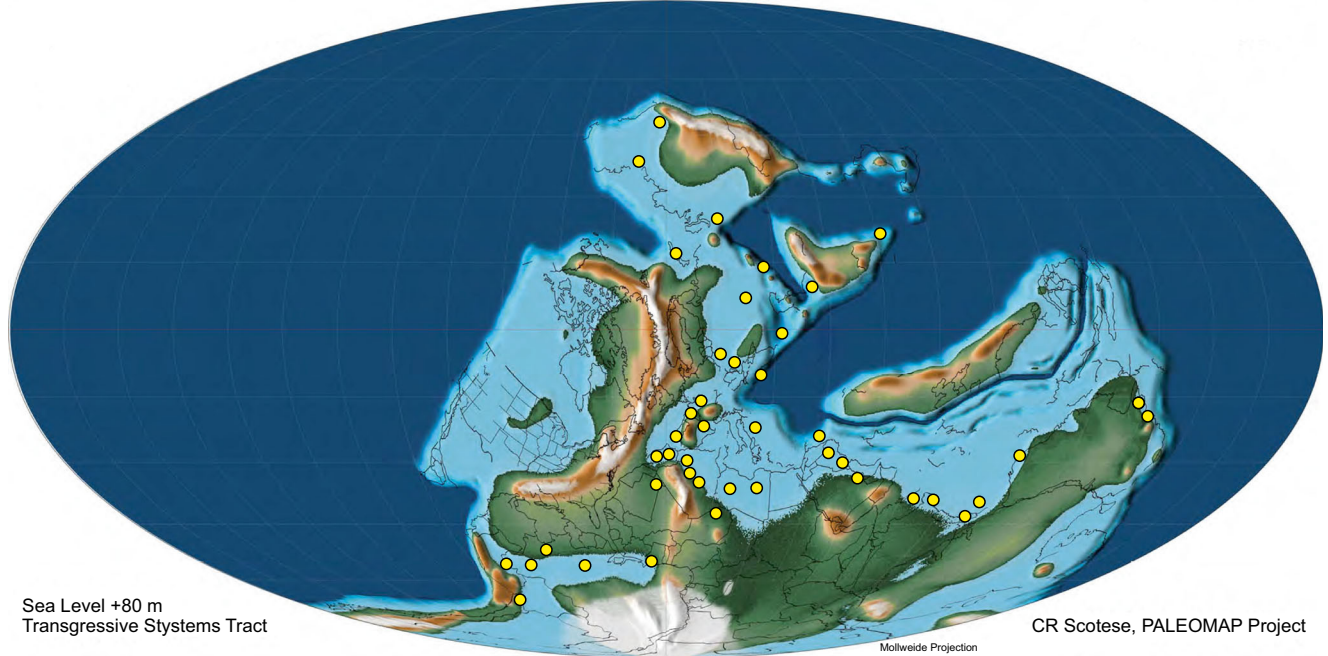
Due to the palaeogeographic position of South America in high southern latitudes during the DCB interval (Fig. 2), the correlation to the biostratigraphic schemes developed in the low palaeolatitudes is often difficult. Correlation between these two realms is mainly achieved thanks to the apparent cosmopolitan nature of some miospores (e.g. Melo and Loboziak 2003; Playford 1990). Upper Devonian–lower Carboniferous strata is exposed in the intercratonic basins covering parts of western Gondwana, and arc-related and retroarc basins related to the active western Gondwana margin, nowadays mainly

situated east of the Andes (Precordillera). Uppermost Devonian glacial records (e.g. diamictites) are found all over the region (e.g. Caputo et al. 2008; Isaacson et al. 2008; Lakin et al. 2016; Streele et al. 2013). Dating them has been a crucial step as they are the first markers of the climate deterioration leading to the Late Palaeozoic Ice Age. In the context of the DCB, these glacial deposits have been considered to represent time equivalents of the Rhenish Hangenberg Sandstone (e.g. Kaiser et al. 2015). However, this correlation is too simple and may only hold up for the top the glacial deposits, since their bases are often dated as LE miospore zone, which clearly predates the Hangenberg Sandstone.

## Brazil

The Upper Devonian–Carboniferous successions have been described from several intercratonic basins. The successions of the Amazon Basin have played a key role not only in the establishment of a miospore biostratigraphic scheme for South America (e.g. Loboziak et al. 1995, 2000; Loboziak and Melo 2002; Melo and Loboziak 2003; Melo and Playford 2012), but also for the correlation into other parts of Gondwana (e.g. Loboziak et al. 1999) and most important with the palaeotropical miospore biostratigraphy of Western Europe (e.g. Clayton et al. 1977; Streele et al. 1987).

In the Amazon Basin, the Upper Cuirri Formation (diamictites, siltstones and sandstone lenses) has a Rle-LVa age (Melo and Loboziak 2003), which is the equivalent of the LE-LN interval in the zonal scheme for Western Europe. The



**Fig. 2** Palaeogeographic map of C. R. Scotese (Palaeomap Project map 65) for the time interval of the Devonian-Carboniferous Boundary with approximate position of the main sections mentioned herein. Note that the depositional settings of several sections do not correspond to the land-sea distribution of the map

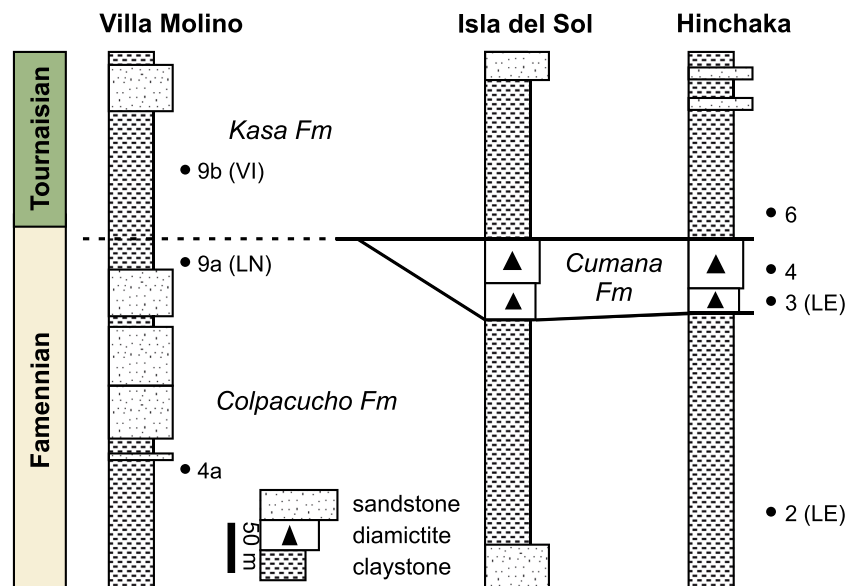
glacial pulse ends below the base of the Carboniferous. The DCB is situated in the sandy non-glacial deposits of the basal Oriximiná Formation. In other boreholes, the boundary is placed between the two formations (Melo and Loboziak 2003). It is important to note that in the Amazon Basin, the boundary between the youngest Famennian and oldest Tournaisian miospore biozones (LVa–AL) corresponds to an erosional surface (Melo and Playford 2012). Hence, the record of the DCB interval is incomplete, and differences in the ages of the basal Oriximiná Formation (Melo and Loboziak 2003) may be rooted in this problem. From the subsurface of the neighbouring Solimões Basin, Caputo et al. (2008) correlated glaciomarine diamictites (Jaraqui Member of the Jandiatura Formation) with the upper Curiri Formation of the Amazon Basin. However, as outlined by Lakin et al. (2016), the primary sedimentological and biostratigraphic data are difficult to access.

In the Parnaíba Basin, the uppermost Devonian–lowermost Carboniferous succession (Loboziak et al. 1992) has some interesting similarities to those of the Amazon Basin. Indeed, the upper part of the Cabeças Formation attributed to the LVa miospore Zone is composed of glacial and periglacial deposits (Streel et al. 2000). An unconformable contact is found when the following Longá Formation is entirely Tournaisian in age (Streel et al. 2013). However, a conformable contact is developed when the base of the Longa Formation corresponds to ‘end-Devonian transgressive marine shales’ (Streel et al. 2013). The boundary between both formations is also indicative for the retreat of the glacial facies, which took place

immediately below the DCB (Caputo et al. 2008). In southeastern Brazil, uppermost Devonian diacmitites dated as LN Zone (Loboziak et al. 1995) are found in the subsurface of the Paraná Basin (Lakin et al. 2016). These so-called Ortigueira Diamictite form the top of the Ponta Grossa Formation, which is unconformably overlain by the Moscovian or younger rocks of the Itararé Group (Milani et al. 2007).

#### Bolivia and Peru

In the Bolivian Altiplano, around the Lake Titikaka, the Upper Devonian–lower Carboniferous strata are regrouped in ascending order into Colpacucho, Cumaná and Kasa formations (Fig. 3). The Upper Devonian Calpacucho Formation is composed of marine mudstones and sandstones and, in its upper part, deeper-water offshore facies dominates. The following Cumana Formation is composed of up to 100-m-thick glacial marine diamictites and shales with dropstones of latest Famennian palynology-based age (Díaz-Martínez and Isaacson 1994; Lakin et al. 2016). The Kasa Formation represents a complex deltaic progradational sequence of Tournaisian–Visean age (Díaz-Martínez et al. 1999). As previously described from Brazil, the glacial pulse occurred just below the DCB. Palynological data (Díaz-Martínez et al. 1999; Vavrdová et al. 1993) show that the Cumaná Formation belongs to the LE Zone. The absence of *Verrucosiporites nitidus* does not allow to identify the LN Zone. However, Lakin et al. (2016) concluded that the extinction of *Retispora lepidophyta* and marine phytoplankton



**Fig. 3** DCB sections in the vicinity of Lake Titikaka, Bolivia. The sections are famous for the record of glacial deposits, which are all Famennian in age (Marshall et al. 2020). Figure slightly modified from

Lakin et al. (2016). Biostratigraphic data according to Vavrdová et al. (1993), Díaz-Martínez and Isaacson (1994), Díaz-Martínez et al. (1999), Lakin et al. (2016)

occurred within or immediately above the diamictites, and hence allows to place the DCB at this extinction level (= base of VI Zone). The position of the DCB above the glacial levels can be seen on the northeastern side of the Lake Titikaka (Marshall et al. 2020; Vavrdová et al. 1993). In the Chaguaya section, Marshall et al. (2020) used the last appearances of *R. lepidophyta* and *Umbellasphaeridium saharicum* to place the DCB 12 m above a 60–140-m-thick interval of glaciogenic sediments (coarse-grained sandstones, striated/grooved hardgrounds and diamictites). Within the glaciogenic sediments, they found a single specimen of *V. nitidus*, which may indicate the LN Zone. Older data from Vavrdová et al. (1993) and Díaz-Martínez et al. (1999) for the Villa Molina section show the presence of the LN Zone few metres (sample MM 9a) above the level interpreted as glaciogenic sediments by Marshall et al. (2020).

The most controversial Upper Devonian glacial records in Bolivia are from the Bermejo West section in the Sub-Andean Zone (see summary in Lakin et al. 2016). The section starts with black shales attributed to the late Famennian (?VCo Zone). This succession is topped by 18 m of diamictites and sandstone lenses of the Itacua Formation. The data of Wicander et al. (2011) showed that the base of the Itacua Formation in this section is within the LL Zone, and the top in the LN Zone. Hence, they concluded that the latest Devonian glaciation was a 1-3-My-long multiphase event. This interpretation was challenged by Streel et al. (2013) who proposed an alternative interpretation of the ages and spore assemblages, resulting in a shorter and simpler glacial event.

Isaacson et al. (1995) report a very unusual record of the DCB in 2 boreholes from the Bolivian part of the Madre de Dios Basin (Subandean Zone). In these boreholes, diamictites regrouped in the Toregua Formation are thought to be of Late Devonian and early Carboniferous age. Vavrdová et al. (1996) indicated the presence of the biozones LL, LN and VI in the Toregua Formation, shortly above the (?) lower Famennian (TG Zone). Lakin et al. (2016) indicate that the lower formation boundary is close to the base of the late Famennian VH/Hys Zone. Therefore, they attributed a VI age for the unconformable base of the overlying Yaurichambi Formation. Melo (2005) studied sections in the western part of the Subandean Zone. There, neither the Toregua Formation nor the undifferentiated Retama Group provided in situ latest Famennian and early Tournaisian palynofloras. Miospores of these ages are found as reworked elements in younger strata.

In SE Peru, the Ccatcca Formation shares many lithological similarities with the Bolivian Cumaná Formation and is considered to be age-equivalent (Díaz-Martínez 2004; Isaacson et al. 2008). If this correlation is correct, the DCB may be found in the lower part of the Ambo Group just above the Ccatcca Formation.

## Chile

In northern Chile, Upper Devonian to lower Carboniferous siliciclastic rocks are found in two major facies belts, which Bahlburg and Breikreuz (1993) termed western and eastern basins. The eastern basin (Sierra Almeida) contains a 2700-m-thick succession of shallow-marine platform deposits ranging

from Middle Devonian to Visean (Zorritas Formation) (e.g. Boucot et al. 2008; Isaacson et al. 1985; Niemeyer et al. 1997).

A first hint for the position of the DCB in the Middle Member of the Zorritas Formation was the record of *Wocklumeria* (Breitkreuz 1986). Rubinstein et al. (1996) recognised the VI Zone (oldest Carboniferous miospore biozone), but the study was based on a single sample. According to a detailed palynological study, palynological assemblages were assigned to the latest Famennian (Strunian)?, Tournaisian and Tournaisian-Visean (Rubinstein et al. 2017). The position of the DCB falls into a 15-m-thick mudstone interval between their samples 7051 and 7050. However, characteristic for the assemblages is the high number of reworked spores, which make precise attributions to palynozones impossible; e.g. the marker of the latest Devonian *R. lepidophyta* is only found in the middle of the youngest assemblages.

In the western basin (Costal Cordillera), several kilometres-thick turbiditic facies developed (e.g. Hervé et al. 1981). Biostratigraphic data are rare, and the different formations are attributed to the Devonian and Carboniferous based on plants, brachiopods and conodonts (see data and references in Hervé et al. 2007).

## Argentina

The DCB in Argentina is in general characterised by an unconformity, showing Devonian folded sediments overlain by less disturbed Carboniferous beds (López Gamundi and Rossello 1993). However, a biostratigraphic control of the succession is scarce in many areas, and data are available only in the central and northern Precordillera.

In the Rio Blanco Basin (San Juan province), a hiatus, possibly spanning from Frasnian to the early Tournaisian, is documented (Amenábar et al. 2009). The green-brown shales, with limestone nodules and lenses and subordinate sandstones of the Chigua Formation are followed by thick-bedded paraconglomerates with a few intercalated fine-grained sandstones and shales of the Malimán Formation. The units are separated by an angular unconformity. Palynomorphs indicate a Givetian age for the Chigua Formation, and a late Tournaisian-Visean for the lower part of the Malimán Formation (Amenábar et al. 2009). A similar situation is documented slightly to the south in the Uspallata-Callingasta Basin, where the green greyish sandstones and pelites of the Codo Formation, dated to the Frasnian, are followed by the sandstones, conglomerates and pelites of the El Ratón Formation of ?late Tournaisian-Visean age (Amenábar and di Pasquo 2008; Milana and Di Pasquo 2019).

A possibly different context occurs in the Sierra de Las Minitas (northern Precordillera, La Rioja province). There, a wide range of sedimentary siliciclastic lithotypes is assigned

to a poorly defined Jagüel Formation (González and Bossi 1986), which age encompasses most of the Devonian and Mississippian. These sediments are strongly folded and faulted and include a number of glacial diamictites (Prestianni et al. 2015, and referenced therein). In the lower part of the unit, Lower Devonian trilobites are documented (e.g. Holloway and Rustán 2012), but no stratigraphic data are available above until the diamictite levels, interpreted as Late Devonian Glacial Event (see discussion in Prestianni et al. 2015). In a section located in the southern part of the Sierra de Las Minitas, Prestianni et al. (2015) documented the miospore *Waltzispora lanzonii* from horizons located about 35 m above the diamictites in a palynological association mainly composed of Devonian taxa. However, *W. lanzonii* is documented only from the Tournaisian of north-northwestern Gondwana, and therefore the Devonian taxa should be considered reworked by glacial erosion (Prestianni et al. 2015).

## Africa

### South Africa

The correlation of strata of the large intracontinental Cape-Karoo Basin to global stratigraphic units is often complex. The Witteberg Group consisting of marginal marine deposits of mudstones, siltstones and sandstones encompasses the Givetian to Tournaisian interval. Nowadays (Gess 2016), the DCB can be precisely placed in this succession based on a rich and diversified fish fauna. The stratigraphic marker is well documented by the extinctions of placoderms and of most of acanthodians and sarcopterygians at the top of the sandstone-dominated Witpoort Formation. The succeeding mudstone- and siltstone-dominated Kweekvlei Formation is characterised by the radiation of the actinopterygians (Gess 2016). This faunal signal confirms the earlier idea of a regressive sandy Famennian and transgressive muddy to silty Tournaisian strata (Cooper 1986) and the position of the DCB close or at the contact between the Witpoort and Kweekvlei formations (Streel and Theron 1999; Theron 1993). An interesting element for the correlation with South America is the glacial signal in the Peerdepoort Member (upper part of the Witpoort Formation) (Almond et al. 2002).

### Ghana

In equatorial Africa, Ghana is the only country with a precise record of the DCB. The palynological records place the DCB within the Takoradi Shale Formation at Essipon Beach near Sekondi (Atta-Peters 1996; Atta-Peters and Anan-Yorke

2003). The formation is part of the Ordovician to Lower Cretaceous Sekondi Group.

The c. 200-m-thick Takoradi Shale Formation is composed of black or dark grey compact mudstone and shales with intercalations of sandstones. Atta-Peters and Anan-Yorke (2003) illustrate an about 20-m-thick section exposing the uppermost Devonian and lowermost Carboniferous. The palynological content of 23 samples have been correlated to Western European standard palynozones by Atta-Peters (1996). Samples 1–6 are correlated to the LE Zone, sample 7 to the LN Zone and samples 8 to 23 to the VI Zone. Sample 7 was taken from the top of a 40-cm-thick sandstone bed, and sample 6 1 m up-section in a 2.5-m-thick unit of shales and mudstones. Atta-Peters and Anan-Yorke (2003) added more samples to the critical interval around the DCB, and they could place the boundary about 45 cm above sample 7. Besides the very precise palynology-based position for the DCB in the Takoradi Shale Formation, it is interesting to note that as in South Africa the DCB is at or near the lithological changes from coarse to fine sediments.

#### Northern Africa

In northern Africa, the South Atlas Fault and its eastward prolongation separate the mobile Variscan belt in the north from a tectonically less deformed domain in the south, the so-called Northern African or Sahara Platform, which formed on the cratonic basement of Gondwana. Uppermost Devonian and basal Carboniferous strata crop out in both domains, but most information are from the remnants of the epicontinental sea, often called basins, which covered the Sahara Platform in southern Morocco, Algeria and Libya. North of the fault, almost all Variscan structural units contain strata attributed to the Upper Devonian and/or lower Carboniferous, but a lot of those units are affected by early Variscan deformation, called Eovariscan (370–360 Ma after Hoepffner et al. 2005), in the Late Devonian and early Tournaisian. Overall, unfavourable facies (e.g. unfossiliferous shales), reworked and re-sedimented material, important stratigraphic gaps and syndimentary tectonics hamper the precise determination of the DCB. Continuous stratigraphic sections are often missing and data remain often scattered over the entire succession. Hence, the DCB is often approximately placed by dates coming from much further down and/or up in these successions. The currently most-studied outcrops are situated in the Western Moroccan Meseta.

#### Morocco

Morocco is the African country, where the most modern biostratigraphic data for the DCB interval are available. Cephalopod faunas of the Upper Devonian successions of the eastern Anti-Atlas (e.g. Tafilalt, Maïder, Dra Valley) have been intensively

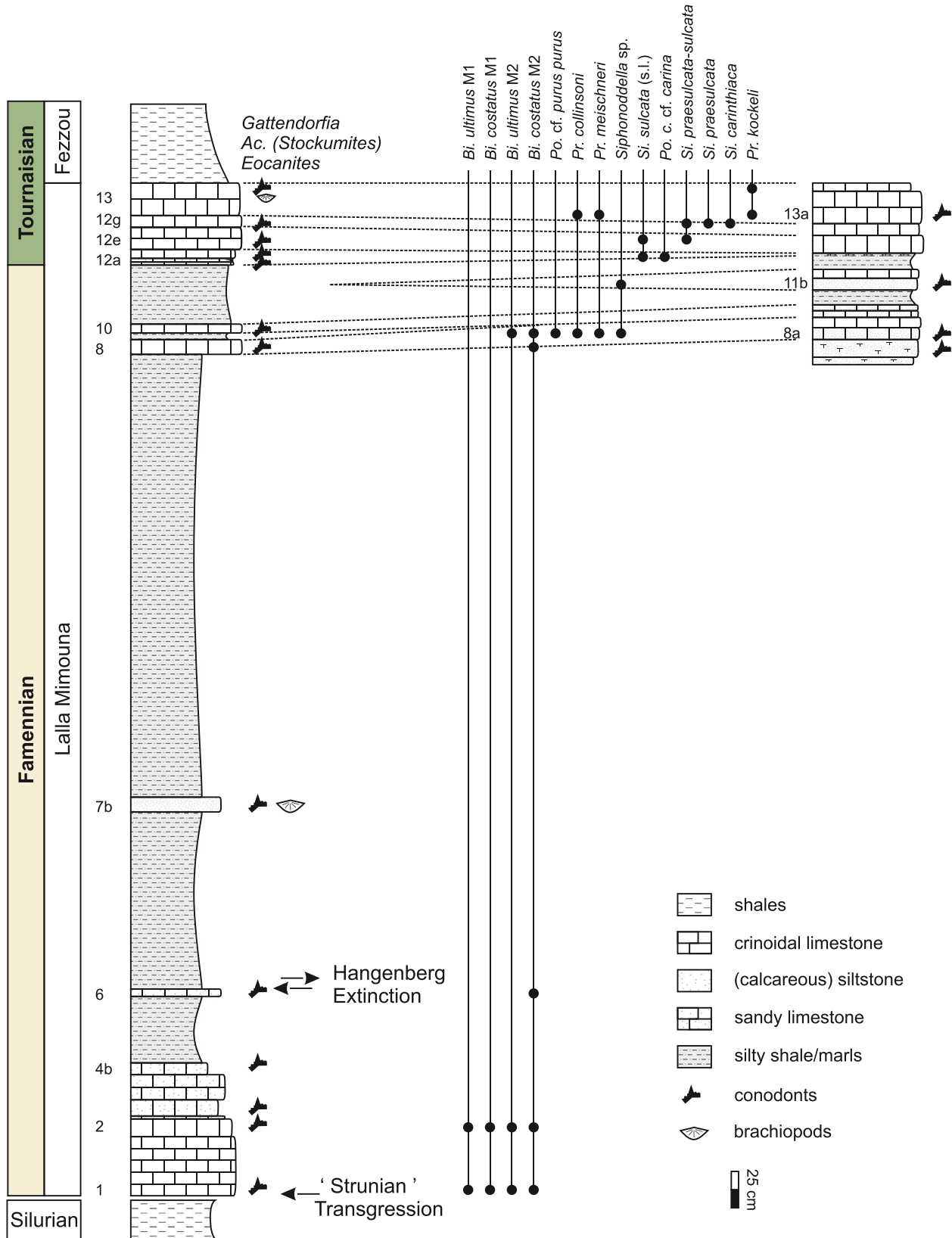
investigated since the pioneer studies of Termier and Termier (1950) and Petter (1959, 1960). With additional data for conodonts, a rather robust stratigraphic framework for the sections straddling the DCB has been established (e.g. Becker et al. 2013; Kaiser et al. 2007, 2011; Korn et al. 2004) in the different facies realms defined in the Tafilalt and Maïder (Wendt 1988; Wendt et al. 1984). In this region, the black shales or their weathered equivalents at the base of the Fezzou and Aoufital formations (e.g. Becker et al. 2013; Kaiser et al. 2011, 2013; Korn 1999) are considered to be equivalent to the Rhenish Hangenberg Black Shale.

In the northern Maïder, the fossiliferous upper Devonian-lower Carboniferous succession around the Lalla Mimouna hill has gained a lot of interest in recent years (Becker et al. 2012, 2013; Fröhlich 2004; Hartenfels and Becker 2016; Korn et al. 2004). Works concentrated on two sections called Lalla Mimouna A and B by Korn et al. (2004) and respectively Lalla Mimouna South and North by Becker et al. (2012, 2013). The sections are important since they are the only ones with abundant conodont assemblages across the DCB interval in this area.

The research interest in this area was triggered by the findings of Korn et al. (2004), on the northern slope of Lalla Mimouna, of an ammonoid-bearing horizon at the top of a 1.1-m-thick unit of sandy and marly limestones. These contain *Acutimitoceras hilarum* and *Postclymenia evoluta*, and conodonts indicative for the Upper *praesulcata* Conodont Zone (= *Pr. kockeli* Zone). Hence, in this section, the *Acutimitoceras-Postclymenia* Ammonoid Assemblage has a latest Devonian age.

Later, Becker et al. (2013) coined for the latest Devonian and earliest Carboniferous strata the name Lalla Mimouna Formation. This formation is 9-m-thick and consists of a silty shales/marls package sandwiched between two crinoidal limestone units (Fig. 4). The lower member (1.4-m-thick) is composed of coarse-grained crinoidal limestone (beds 1–4b), a poorly exposed silty/marly interval (bed 5), topped by a sandy limestone bed (bed 6). This last bed contains the conodonts *Palmatolepis gracilis gracilis*, *Bispathodus stabilis vulgaris*, *Bi. aculeatus aculeatus* and unnamed siphonodellids. According to Becker et al. (2013), the entire lower member belongs to the *Si. praesulcata* Zone although *Si. praesulcata* is absent from that member. This lower member is considered to represent the pre-Hangenberg sedimentation. The middle member (bed 7, 7.5-m-thick) consists of silty shales/marls with several intercalations of thin-bedded, calcareous siltstones, which have been individualized as sub-beds. One of those, Bed 7b, contains a rich brachiopod fauna typical for the latest Famennian. Becker et al. (2013) correlate the base of this unit with the onset of the Hangenberg crisis, but could not find the anoxic level as equivalents of the Hangenberg Black Shale.

The upper member (0.57 thick) is three-folded. The lower sub-unit (beds 8–10) comprises rather sandy, crinoidal pack-



**Fig. 4** DCB in Lailla Mimouna North section, Morocco. The DCB is placed at the base of bed 12 (12a) by Becker et al. (2013), although *Si. sulcata* appeared slightly higher. The slightly modified log from Becker et al. (2013) shows the main lithological and stratigraphic information

available for this section. The lateral section inserted shows the highly variable thicknesses of the beds around the DCB. This is most likely a depositional signal, and it certainly influences fossil content and hence the boundary position

rudstones with intercalated siltstone layers. It contains a rich conodont fauna including *Protognathodus meischmeri*, *Pr. collinsoni*, *Polygnathus* cf. *purus purus* and *Pseudopolygnathus* aff. *primus*. The middle sub-unit (bed 11) consists of deeply weathered calcareous siltstones, but laterally sandy or silty crinoidal pack-rudstones are intercalated in this subunit. Relatively pure crinoidal pack-rudstones separated by discrete siltstone layers form the upper sub-unit (beds 12–13). From thin limestones (beds 12b–c), Becker et al. (2013) mentioned a single specimen that falls into the variability of *Si. sulcata* (s.l.) sensu Flajs and Feist (1988). It resembles specimens regarded as transitional morphotypes between *Si. praesulcata* and *Si. sulcata* or as *Si. sulcata* (Ziegler and Sandberg 1996). Bed 12e contains a single specimen of *Si. sulcata* (s.l.) falling into group 5 of Kaiser and Corradini (2011). It suggests that bed 12e belongs to the *sulcata* Zone and hence it is undoubtedly of Carboniferous age. However, Becker et al. (2013) place the base of the *Si. sulcata* Zone, and in consequence the DCB at the base of bed 12. Bed 12g is attributed to the Upper *duplicata* Zone (= *Si. jii* Zone), which clearly hints to a, important stratigraphic gaps. A marked change in the conodont biofacies is observed in bed 13a, where siphonodelloids disappear and protognathodids (*Pr. meischmeri*, *Pr. collinsoni*, *Pr. kockeli*) re-appear. The Lalla Mimouna Formation is topped by deeply weathered, greenish silty shales of the Fezzou Formation.

The major problem of the Lailla Mimouna sections are the important lateral variable thickness of the beds, several beds wedging out laterally. In fact, some of the bed contacts may correspond to hummocky cross-stratifications, pointing to very dynamic depositional systems for which the reworked conodonts mentioned by Becker et al. (2013) are a further evidence. Overall, the sections at Lalla Mimouna share many characters with the succession exposed in the GSSP at La Serre (Feist et al. 2021, this issue), including its shortcomings.

In the southeastern Amessoui Syncline, two sections—El Atrous and Jebel Ouaooufilal—expose a several 100-m-thick DCB interval within the Aoufital Formation. The most recent detailed data are available for the El Atrous section (Kaiser et al. 2013). The missing *Wocklumeria* Zone (UD VI-D) indicates that the top of the Erfoud Formation corresponds to an unconformity. Above this unconformity, a 2.5-m-thick unit of reddish-white unfossiliferous shales (bed lower K3), which can contain sulphates from pyrite weathering, is considered to be the Hangenberg Black Shale equivalent. Up-section, the lower part of the Aoufital Formation (beds upper K3-K28) consists of a more than 260-m-thick succession of alternating unfossiliferous, greenish shales and thin-bedded siltstones and fine-grained sandstones. At the Jebel Ouaooufilal, this stratigraphic level contained a thin sideritic layer with the oldest *Acutimitoceras* (*Stockumites*), which indicates the lower UD VI-F Ammonoid Zone. The sandstones of bed K 28 are interpreted as laterally wedging mass-flow deposits, which are an equivalent of the Hangenberg Sandstone. The DCB is

placed at the base of bed K 31 into a coarsening-upward sequence (beds K29-36; 60-m-thick) from shales and intercalated thin sandstone beds to proximal sandstone deposits (K33-36). The position of the DCB is based by default on the presence of the rhynchonellid brachiopod *Hemiplethorhynchus* (?) sp. cf. *allani* and the absence of ?*Centrorhynchus*, which possibly indicates a Tournaisian age. The top of the Aoufital Formation contains about 30 m of olive-green shales, topped by black shales of the basal Oued Znaigui Formation correlated with the Lower Alum Shale Event (upper part of the lower Tournaisian or ‘middle’ Tournaisian in the older literature).

North of the South Atlas Fault, the allochthonous nappes at Mrirt at the eastern margin of the Western Moroccan Meseta are an example for several problems to locate the DCB (Becker et al. 2020; Walliser et al. 2000). Although several sections contain stratigraphic markers typical for the DCB interval (Becker et al. 2020), the precise position of the DCB remains unknown. More detailed data are available from the western part of the Western Moroccan Meseta. Kaiser et al. (2007) studied in the Oulmes region the Ain Jemaa sections. In the Ain Jemaa 1 section, basal clymeniid-rich nodular limestones (beds 1–2, middle member of the Bou Gzem Formation) contained conodonts indicating the *Bi. ac. aculeatus* Zone. Above these limestones follows the upper member of the Bou Gzem Formation composed in ascending order of a 3-m-thick unit of unfossiliferous, silty shales (bed 2b), a 7-m-thick unit of unfossiliferous pyritic black shales (bed 2c) and 15 m of shales with haematite-rich nodules and concretions (bed 2d). Up-section follow about 20 m of grey quartzitic sandstones (bed 2e, Ta'arraft Formation). Despite the absence of biostratigraphic evidence above the nodular limestones, bed 2c is interpreted as equivalent of the Hangenberg Black Shale, bed 2d as Hangenberg Shale equivalents and bed 2e as Hangenberg Sandstone equivalent. This interpretation was based on the application of a model for a supposed global sequence stratigraphic framework for the DCB interval (Becker 1996; Bless et al. 1993). The validity of this assumption cannot be verified without re-consideration of all data.

## Algeria

In Algeria, uppermost Devonian and lowermost Carboniferous strata crop out in several basins south of the South Atlas Fault (Conrad et al. 1986). A general problem is that many studies rely on the data calibrated to the then ground-breaking cephalopod stratigraphic schemes of Petter (1959, 1960) and Pareyn (1961), and not to modern stratigraphic zonations as, e.g. Korn et al. (2004, 2007).

The successions of the Béchar Basin *sensu lato* are often considered to be a reference. Those sections are fully marine and carbonate-dominated in its northern part (Djebel Antar and Horreit, Ben Zireg area), whereas in the southern part,



siliciclastic deposits define the boundary interval in the Saoura Valley and in the transition to the Ougarta (Marhouma).

In the NE of Djebel Antar (Soltane et Betoum Anticline), a condensed Upper Devonian-Tournaisian succession is only complete in the East. A 2–3-m-thick unit of shales is attributed to the latest Devonian (Strunian) based on the presence of *Bi. costatus* (Weyant 1976). This is followed by a 30-m-thick unit divided into a lower shaly sub-unit topped by a limestone sub-unit dated as late Tournaisian (*Si. anchoralis* Zone). The shaly unit contains at its base undeterminable goniatites and the conodonts *Pr. meischneri* and *Pr. kuehni*. Lemosquet et al. (1980) correlate this level with the Stockum level of the Rhenish succession. In the Ben Zirg area, the Viséan wildflysch contains blocks from which Alberti (1972) described latest Devonian trilobites (*Phacops* (*Ph.*) *accipitrinus*) and conodonts, which may occur from the *Bi. costatus* to the *Si. quadruplicata* zones. Weyant (1985) mentioned for the north of the Bechar Basin *sensu lato* ‘a *Protognathodus* fauna at the boundary between the upper Famennian and Tournaisian ...’, which most likely corresponds to the *Pr. kockeli* Zone placed at the base of the Carboniferous (Lemosquet et al. 1980), but also includes the basal Carboniferous conodonts mentioned by Alberti (1972).

At the southern end of the Bechar Basin *sensu lato*, a facies change from the sandstone-dominated Ouarourout Formation to the mudstone-dominated Olad-Bou Hadid Formation materialises the boundary interval. The DCB is placed at the base of the Olad-Bou-Hadid Formation after Conrad et al. (1970). In a section north of Ouarourout, they found a thin limestone bed near the formation base containing the conodonts *Po. inornatus*, *Po. communis*, *Ps. dentilineatus*, *Siphonodella* sp. and ‘*Ozarkodina*’ sp. The mudstones below this limestone bed yielded a cephalopod assemblage containing small specimens of *Gattendorfia*, *Imitoceras* and an undetermined clymenid. In consequence, Weyant (1985) placed the *sulcata* Zone at the base of the Olad-Bou Hadid Formation. Korn et al. (2007) correlated the Ouarourout ammonoid fauna with their *Gattendorfia-Eocanites* Assemblage. They indicate that this assemblage does not represent the earliest Carboniferous, but a higher stratigraphic level equivalent to the grey shales above the Hangenberg Limestone of the Rhenish succession. Hence, the actual DCB in the Algerian sections has to be positioned lower probably near the top of the Ouarourout Formation.

An important palynostratigraphic record for the DCB interval is known from the so-called argillaceous series of the basal Ioucha 1 Formation in borehole ut2 in the Western Grand Erg at the eastern side of the Bechar Basin *sensu lato*. In their update of the classical work of Lanzonih and Magloire (1969), Coquel and Abdesselam-Rouighi (2000) identified the oldest association (Biozone L7 of Lanzonih and Magloire, 1969) in –590 to –687.50-m depth. This rich assemblage contains *R. lepidophyta* and hence is of latest Famennian age. Coquel and Abdesselam-Rouighi (2000) pointed to the absence of *Lophozonotriteles cristifer* and the rareness of

*Rugospora flexuosa* compared to contemporaneous assemblages in the Sahara. The second assemblage (Biozone M2 of Lanzonih and Magloire 1969) in –540 to –590-m depth is less diversified. The new data show the presence of *Vallatisporites vallatus*, which combined to the absence/rareness of *Aratrisporites saharaeasis* indicates the lower to middle Tournaisian. Coquel and Abdesselam-Rouighi (2000) used the sudden disappearance of *R. lepidophyta* in ut2 as an argument to support a stratigraphic gap at the top of the Devonian (latest Strunian), and possibly the basal Carboniferous, but the later assumption lacks palynological arguments.

In the Timimoum region east of the Ougarta, the DCB interval is found in the argillaceous Kahla Sandstones, which are upper Devonian to lower Tournaisian (Conrad et al. 1986; Legrand-Blain 1979). The Kahla Sandstones are divided into the lower and upper Kahla Sandstones. The basal part of the lower Kahla Sandstones contains coquina-like limestone intercalations yielding a late Famennian brachiopod assemblage (Legrand-Blain 1979). This is topped by 180 m of fluvio-deltaic deposits of Strunian (age based on the report of *R. lepidophyta* by Lanzonih and Magloire (1969)). The upper Kahla Sandstones contains two major shaly intervals. The lower interval contains a *Gattendorfia-Kahlacanites* Assemblage (Ebbighausen et al. 2004; Korn et al. 2007), which is placed near the top of the *Gattendorfia-Eocanites* Ammonoid Genus Zone (Korn et al. 2004, 2007). In comparison to Central Europe (Ebbighausen et al. 2004), this assemblage points most likely to a stratigraphic level above the *Gattendorfia* Limestone. The upper shale interval contains a *Kazakhstania-Acrocanites* Assemblage and a *Pericyclus-Progoniatites* Assemblage, which indicate a late early Tournaisian to early late Tournaisian age for this shaly unit (Korn et al. 2010).

In other Algerian basins, the situation is comparable to the above-described successions. The uppermost Famennian is dominated by coarser-grained siliciclastic deposits, which on a N-S gradient show a trend from marine, deltaic to continental facies. Those deposits are often described as being regressive (see, e.g. Conrad et al. 1970, 1986; Legrand-Blain 1979). The following Carboniferous deposits are fine-grained and argillaceous intercalations become more common or the lithology is more variegated. They often contain a marine fauna of Tournaisian age, which leads to their description as a transgressive phase. Commonly, the base of the Carboniferous is equaled with the lithological change, e.g. lower and upper Khenig Sandstone, north of the Hoggar (Reggan-Ahnet-Mouydir area), but equivalent to the Timimoum region, the first Carboniferous fauna seems to occur stratigraphically a little bit higher than the earliest Carboniferous. Hence, it seems to be likely to assume that the DCB should be placed slightly lower as commonly done.

#### *Libya, Mali, Niger*

In Lybia, Devonian and Carboniferous strata have been of high interest for the hydrocarbon exploitation (e.g. Belhaj

2000). In eastern and western Libya, sedimentary rocks of latest Famennian age have been identified by palynomorphs. In the subsurface of eastern Libya (Cirenaica Platform), Streeel et al. (1988) attributed an assemblage of *R. lepidophyta*, *Ru. flexuosa*, *V. pusillites* and *K. literatus* to the LL Zone of the latest Famennian (*praesulcata* Conodont Zone). In western Libya (Rhadamès and Mourzouk-Djado basins), the palynozone 11 of Massa et al. (1979) seems to cover a similar stratigraphic level. In the traditional view (e.g. Belhaj 2000; Conrad et al. 1986), the latest Famennian rocks are unconformably overlain by the Marar Formation. Its base contains a diverse marine fauna including brachiopods (e.g. *Fustella*), ammonoids (e.g. *Muensteroceras*) and conodonts (e.g. *Gn. texanus*) of late Tournaisian age (Conrad et al. 1986; Mergl and Massa 2000).

However, a completely different scenario is reported by Mergl and Massa (2000). They suggest for a rich brachiopod-bivalve assemblage from the middle Tahara Formation in the Awaynat Wanin Area (Ghadames Basin) a Famennian age. In the upper part of the same formation, sandstones yield a brachiopod association attributed to the Hastarian (early Tournaisian). The lower part of the overlying Marar Formation also contains brachiopods attributed to the Hastarian. Mergl and Massa (2000) dated the so-called 'brachiopod sandstone marker' of Seidl and Röhlich (1984) in the upper Ashkidah Formation of the northern Murzuq Basin (Wadi ash Shati area) as definitely early Tournaisian.

In Niger, Devonian-Carboniferous succession are known from the Iullemeden Basin and Djado Sub-basin. The Djado Sub-Basin shares many similarities with the northward situated Murzuq Basin in Libya, but the upper Devonian strata is poorly studied and may even be largely absent. In the Carboniferous, the transgressive marine deposits of the Marar Formation are well-developed (Mergl and Massa 2000). The views on the upper Devonian and lower Carboniferous succession of the Iullemeden Basin (Tim Mersoi Sub-basin) are contradicting (see Conrad et al. 1986; Legrand-Blain 1985a). However, the DCB has so far not been precisely located and it may even be absent in this basin, which would agree with the strong regressive character of the uppermost Devonian, and the gradual flooding of the margin of the African craton during the Tournaisian. In Mali, the Devonian-Carboniferous transition is highly condensed and incomplete in the Taoudenni Basin (Legrand-Blain 1985b). The basal part of the Bir en Naharat Formation is not dated, but may be Devonian in age. Higher up in the formation, a level with the conodont *Ps. primus* is succeeded by a ferruginous conglomerate and fine-grained sandstones, both containing upper Tournaisian brachiopods. The rest of the Bir en Naharat Formation has been attributed to the Viséan.

## Australia

Strata straddling the DCB can be found in western and eastern Australia. One recurrent problem is the absence of precise biostratigraphic data and zonations, which often renders the placement of the boundary difficult or impossible. This problem is further enhanced, especially in the Carboniferous parts of the eastern Australian successions by a marked faunal provincialism (e.g. Jones 1995).

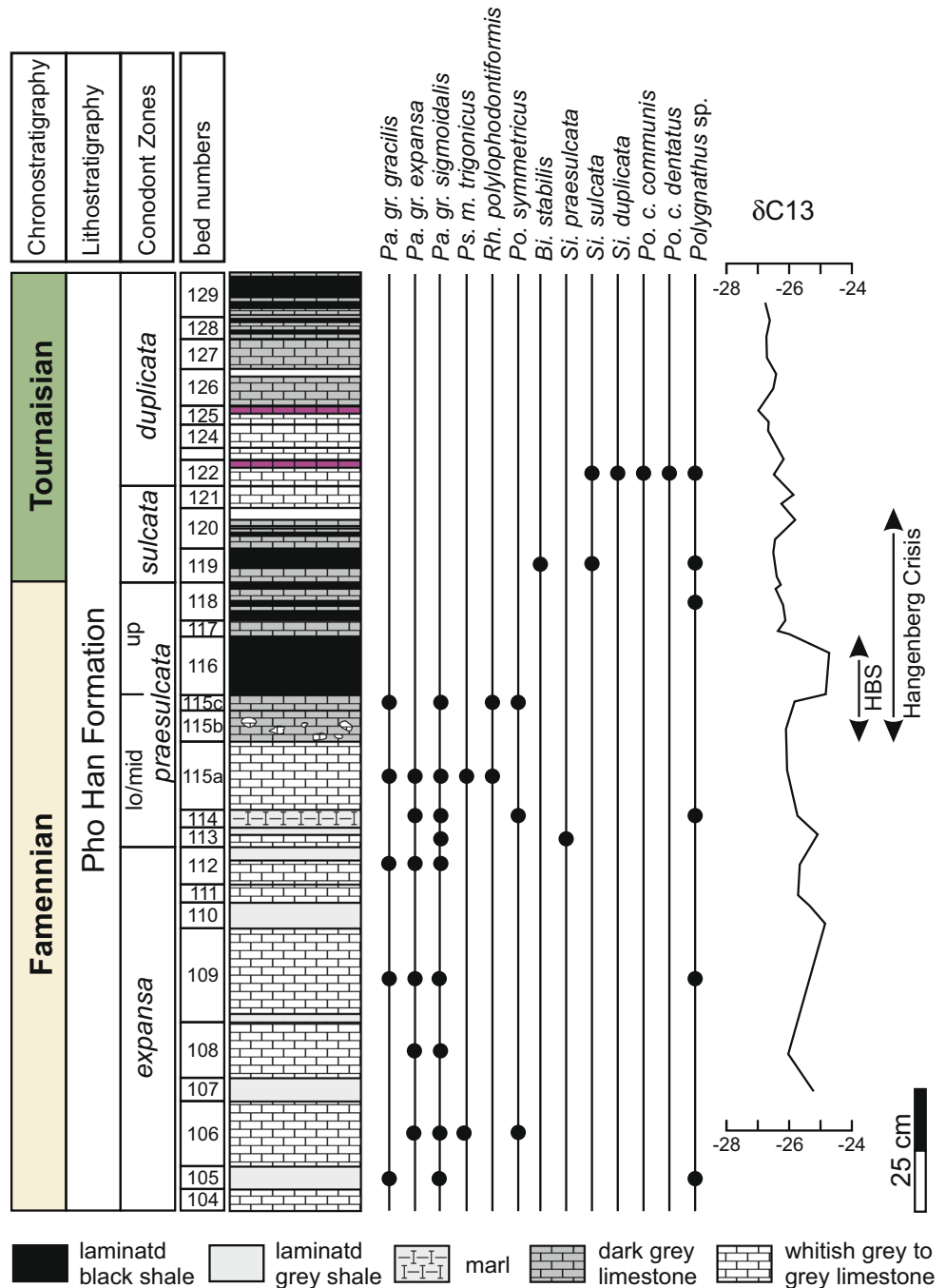
Around the DCB, this handicap can be overcome when palynomorphs are present. In many of the Australian basins, the recognition of the widespread, and almost cosmopolitan, *Retispora lepidophyta* Assemblage can be used to identify the latest Famennian (e.g. Jones 1995; Kemp et al. 1977).

In Queensland, the DCB can be situated in shallow-water carbonates (oolithic limestones), which are intercalated with thick volcanoclastic and siliciclastic complex (Bryan et al. 2003; Roberts et al. 1993). One example is the Gudman Formation (see Webb 2005 for details), west-southwest of Rockhampton in eastern-central Queensland. It is mostly early Tournaisian in age based on conodonts (*Si. sulcata* to *Si. crenulata* zones) (Mory and Crane 1982) and brachiopods (*Tulcumbella tenuistriata* Zone) (Roberts et al. 1993). However, the lower part of the formation contained conodonts of latest Devonian age. Mory and Crane (1982) reported an undifferentiated latest Famennian *Si. praesulcata* Zone. The data of Webb (2005), a posterior tip of a *Palmatolepis* and one element tentatively assigned to (?) *Bi. bispathodus* from the lower part of the formation, may extend the biostratigraphic range of the formation even deeper into the late Famennian. But up to now, the position of the DCB is not located precisely in the Gudman Formation. By default, it is located below the base of the first reefal unit, which plays an important role for the understanding of the recovery phase after the Hangenberg extinction (Denayer and Webb 2017; Webb 1998; Webb 2005).

Further north, in the hinterland of Townsville, the DCB is situated in several formations defined in different basins/regions (see details in Talent et al. 2002). Those formations are all dominated by mainly non-marine siliciclastic sedimentary rocks, but marine carbonated levels productive for conodonts, though rare and unevenly distributed, are intercalated into the succession. Mawson and Talent (1997) identified major marine pulses in the succession, one 'early' in the *Si. sulcata* Zone. This gives a relatively precise position for the boundary, especially if the underlying strata contains late Devonian conodonts (e.g. *Si. praesulcata* Zone in the Hardwick Formation of the Burdekin Basin).

In western Australia, in the Canning Basin, Talent et al. (1993) correlated the change from the mixed siliciclastic-carbonate succession of the Gumhole Formation to the overlying sandstones of the Yellow Drum Formation with the major global regressive phase of the Hangenberg Sandstone

**Fig. 5** DCB in the Cat Co section, Vietnam. Overview on stratigraphy, lithology,  $\delta^{13}\text{C}_{\text{org}}$  curve and occurrences of selected conodonts important for placing the DCB. The DCB is placed at the base of bed 122. *HBS* = Hangenberg Black Shale. (modified from Shizuya et al. 2020)



Event. Hence, these sandstones would (at least partly) be an equivalent to the Hangenberg sandstones of the Rhenish succession. Mory and Hocking (2011) summarised the available biostratigraphic data for the two formations. The best age for the upper part of the Gumhole Formation derives from the already abovementioned *Retispora lepidophyta* Assemblage. The local *Po. communis collinsoni* and *Icriodus platys* zones for shallow-water conodonts are recognised in the Gumpole Formation (Nicoll and Druce

1979), but their precise correlation to the international zonal schemes remains questionable.

Mory and Hocking (2011) place the DCB within the lower part of the Yellow Drum Formation. This position is based on the observations of a nowadays-filled trench next to Linesman Creek. There, the income of *Si. sulcata* marks the base of the Carboniferous, at a level also marked by a shift in  $\delta^{13}\text{C}$  (Andrew et al. 1994). Since the Gumhole Formation was not exposed in this trench, the precise

position of the DCB within the Yellow Drum Formation remains unknown.

## Asia

### Vietnam

The Cat Co section, located in the Cat Ba Island (northwestern Vietnam) and exposing rocks of the Pho Han Formation, has been studied in detail in the recent period with multidisciplinary studies including conodont biostratigraphy, facies analysis, trace elements, isotopes and organic geochemistry (Komatsu et al. 2012, 2014; Paschall et al. 2019; Shizuya et al. 2020). The section (Fig. 5) is very condensed around the D/C boundary and consists of micritic light grey to grey limestones, marls and interbedded thin dark grey limestones and organic carbon-rich, black, friable limestones (Komatsu et al. 2014). The Hangenberg Event is expressed by a c. 1 m (beds 116–118) of laminated black shales alternated with dark grey limestone (Komatsu et al. 2014; Shizuya et al. 2020).

Conodonts are well preserved, but the association lack of some typical elements of the DCB interval, bispathodids and protognathodids is not reported. The base of the Carboniferous is detected by the first occurrence of *Si. sulcata* in bed 119. Beside conodonts, calcareous algae, foraminifers, brachiopods and crinoids are reported. Geochemical studies (trace element proxies, total organic carbon, framboidal pyrite distributions, etc.) and a significant mercury enrichment suggest dysoxic/anoxic conditions from the latest Famennian through the early Tournaisian with severe anoxia (approaching euxinia) throughout the Hangenberg Event interval (Paschall et al. 2019; Shizuya et al. 2020). The authors interpret the mercury anomaly as the result of large-scale volcanic activity.

### Malaysia and Thailand

Devonian and Carboniferous sedimentary successions are known in SE Asia but the precise stratigraphy is still poorly known (Hamada 1967; Oliver et al. 1995). On both sides of the Malaysia-Thailand border near the western coast, the Sanai Limestone Member of the Jentick yielded diagnostic Famennian conodonts (Meor and Lee 2003) whereas the overlying Telegu Jatoh and Wang Kehan siliciclastic member yielded the trilobite *Macrobole kedahensis*, which indicates an early Tournaisian age (Cocks et al. 2005). From the Kathan Limestone Formation cropping out in Peninsular Malaysia, Metcalfe (2002) described conodonts from the upper Tournaisian *Sc. anchoralis* Zone, together with reworked late Devonian and early Tournaisian species, and evidenced a disconformity between Eifelian and upper Tournaisian strata. From Western Peninsular Malaysia, Tsegab et al. (2017) described a diverse conodont fauna ranging from the Famennian

to the Tournaisian, including abundant polygnatids and siphonodellids, from the Kintab Limestone Formation. However, the DCB markers are not described.

In the Chiang Dao area (northern Thailand), Randon et al. (2006) described conodonts from a thick-bedded chert sequence of the Fang Chert Formation. The authors documented late Frasnian, early Famennian (*Pa. gl. prima* Zone) and late Tournaisian (*Sc. anchoralis* Zone) conodonts from the ‘Chiang Dao Cherts’ section, without possibility to better constrain the stratigraphic context. This is a rather common situation in Thailand, where Late Devonian-early Carboniferous ages are attributed to black lydite/bedded chert sequences, based on very poor detailed stratigraphic control (see also discussions in Fontaine et al. 2005, 2009).

### NW India

The DCB is documented in the Pin Valley of Spiti (NW India) by Draganits et al. (2002). The authors studied several sections within the Middle Devonian-lower Carboniferous Lipak Formation, mainly represented by mixed siliciclastic and calcareous sediments in the lower part, followed by fossiliferous limestones and sandstones and, at place, rare dark mudstones and thinly bedded siltstones. The authors provided a reconstruction of the depositional environment in terms of sequence stratigraphy, supported by biostratigraphic data based on a few conodont samples.

C. 75 m above the base of the Lipak Formation, a 58-cm-thick unit of dark grey carbonaceous shales sandwiched between crinoidal grainstone is interpreted as a Hangenberg Shale equivalent. Conodonts from the samples just below and above this level yielded a fauna documented from both the latest Famennian and basal Tournaisian (e.g. *Bi. ac. aculeatus*, *Bi. spinulicostatus*, *Clydagnathus cavusformis* and *Ps. primus*).

### Pakistan and Afghanistan

Late Famennian and Tournaisian conodonts were illustrated by Molloy et al. (1997) from the Khyber region, north-west Pakistan. A few sections were measured in the area, and two of them (sections SM and SB) span the DCB. The sections are mainly constituted by limestone with interbedded shales and siltstones, and look continuous across the boundary, but a detailed log or a description is currently not available.

In the SM section, the boundary can be constrained in a few metres’ interval around a ‘shale and siltstone’ unit between the last occurrence of *Bi. costatus* and the first occurrence of *Po. rostratus*. In the SB section, the same interval looks thicker (more than 10 m) and includes several shaly beds. However, it should be noted that the conodont abundance is scarce in that part of the sections, protognathodids are not reported and only

one specimen of *Si. cooperi* is documented from the upper part of the SM section.

In its synopsis of the stratigraphy of the Carboniferous stratigraphy for the western Karakorum, Gaetani et al. (2004) defined for the Karambar Unit the Margach Formation. Although the fossil content of the formation is very poor and patchy, and stratigraphic diagnostic fossils apparently lack, the stratigraphic range has been indicated as mid Famennian to ?early middle Tournaisian.

Sections spanning the DCB in Afghanistan are mentioned by Mistiaen et al. (2015). However, detailed data are not available.

#### Iraq and Saudi Arabia

In the northern part of Iraq, the shallow-water uppermost Famennian-Tournaisian rocks belong to the Ora Formation. A rich spore association from the Kaista section was illustrated by Naquishbandi et al. (2010), but precise data on the DCB are not available, nor a detailed description of the section and on other fossil groups. The Ora Formation is also known in the subsurface of western and southern-eastern Iraq (Al-Juboury and Al-Hadidy 2008; Aqrabi 1998), but again, detailed data are not available.

In Saudi Arabia, the uppermost Devonian and lowermost Carboniferous strata are known from the subsurface in the NE and E of the country (Bassis et al. 2016). Conformable contacts between the two systems seem to be rare, and in many regions the strata are eroded and truncated by the pre-Khuff or pre-Unayzah unconformities. In this context, the presence of reworked latest Devonian miospores in the younger Tournaisian and Visean strata (e.g. Clayton 1995; Clayton et al. 2000) clearly indicates the erosion of the strata of the DCB interval.

The DCB lies within the upper part of the Jubah Formation. This formation consists of continental and fluvial depositional environments. Al-Hajri et al. (1999) and Clayton et al. (2000) provided slightly different lists of the miospore taxa from the uppermost Jubah Formation in the Harmaliyah field (eastern Saudi Arabia). However, both report the abundance of *R. lepidophyta*, which results in the identification of the regional Saudi Aramco palynozone D0 (Al-Hajri et al. 1999). Clayton et al. (2000) concluded that the upper part of this *Retispora lepidophyta* Assemblage corresponds to the LE and possible LN zones of Western Europe. The lower part could be older than those two zones. In the Abu Safah-29 borehole, Clayton et al. (2000) identified a *Indotriradites explanatus* Assemblage (also called Saudi Aramco palynozone C6, Hooker et al. 2011). They attributed a Tournaisian ('probably early Tournaisian', op. cit.), and indicate the similarities in the floral composition to Palynozone 12 of Grignani et al. (1992) in Libya.

Hence, the precise location of the DCB has so far not been identified, and Clayton et al. (2000) speculated about erosion or non-depositional as possible causes in their studied boreholes.

#### Turkey

Upper Devonian and lower Carboniferous rocks are widely exposed in various regions of Turkey, but the DCB is rarely precisely placed.

In the Istanbul-Zonguldak area (NW Turkey), the DCB is exposed in the shallow-water facies in the eastern part and deeper-water flysch facies in the western part (Okay et al. 2006). In the western part, conodont data from various localities (Çapkinoglu 2000, 2005) suggest that the boundary should be placed in the upper part of the Ayineburnu Formation, but precise studies are missing. However, a lenticular massive, poorly sorted sandstone-greywacke body with intraformational intraclasts (Küçükyalı Member) is locally equivalent to the upper part of the Ayineburnu Formation. This unit possibly represents an equivalent of the Hangenberg Sandstone event, but further evidence is needed.

In the eastern part, in the Topluca section (Bartın area), where the upper Strunian succession includes stromatoporoid biostromes, Denayer (2016) placed the DCB in a 1-m-thick poorly exposed interval between bed 33 that yielded the last *Quasiendothyra kobeitusana* (DFZ7 Foraminifera Zone) and bed 35 that yielded a depauperate fauna of unilocular foraminifers (MFZ1).

In the central and eastern parts of Turkey, Upper Devonian and lower Carboniferous successions are exposed in few localities of the Taurides. In the Geyik Dağı tectonostratigraphic unit, the DCB was investigated in detail in the Natlaş section. Here, the boundary lies in the Natlaş Member of the Ziyarettepe Formation. It starts with a 20-m-thick unit of yellowish sandy limestone with brachiopods, followed by an about 55-m-thick unit of dark grey, locally bituminous argillaceous limestone passing upwards to shales. Plodowski and Salanci (1990) placed the DCB about 30 m below the top of the Natlaş Member on the basis of the last occurrence of trilobites of the genus *Phacops*, whereas the Tournaisian genera *Cummungsellia* and *Piltonia* are reported less than 10 m above the boundary. Typical Strunian brachiopods occur in the same beds than the phacopid trilobites, while only *Aulacella* and *Ovatia* species occur in the upper part of the Natlaş Member, above the DCB (Plodowski and Salanci 1990). The overlying shales yielded the Tournaisian brachiopods *Rugosochonetes* sp. and *Schizophoria* sp., as well as the spores *Umbonatisporites distinctus* and *Raistrickia clavata* (among others) that indicate the upper lower Tournaisian PC Palynozone. Recently, Atakul-Özdemir (2017) described the conodont fauna of the member in its type section, precisely

documenting a continuous sequence from the *Bi. ultimus* to the *Si. bransoni* zones, and placed the DCB at the first occurrence of *Si. sulcata* in sample SN18, 25 m above the base of the Ziyarettepe Formation. The foraminifers were preliminary reported by Altiner et al. (2012) but no diagnostic taxon was identified near the DCB, with the notable exception of quasiendothyrids in the upper part of the Natlaş Mbr and tubular unilocular foraminifers 3 m above the boundary, based on conodonts.

In the Silifke area (Antalya tectonostratigraphic unit), the Korucuk Formation is the equivalent of the Ziyarettepe Formation. It starts with 15 m of greenish-greyish siltstone with plant fragments yielding the Strunian guide miospores *R. lepidophyta* and *V. pussillites* passing upwards to sandstone-shale-sandy limestone alternations rich in trilobites (Demirtaşlı et al. 1978). The upper part of the Korucuk Formation is a 22-m-thick unit made of dark grey platy limestones with dark shaly intercalations yielding Tournaisian brachiopods and foraminifers (*Earlandia* spp. and *Septabrunsiina* spp.) indicating an early Tournaisian age. The brachiopods are abundant in the uppermost Devonian part of the formation, where Plodowski and Salanci (1990) reported the same taxa as in the Natlaş Member, but without clear evidence of Tournaisian material.

In the Hakkari province (SE Turkey), close to the border with Iran and Iraq, the Devonian-Carboniferous succession is exposed in the Zap river valley. The Köprülü Formation starts with a c. 30-m-thick unit of dark sandy limestones interbedded with shales that include two horizons yielding late Devonian fish remains (Janvier et al. 1984). Above the last fish horizon, a 200-m-thick shale interval with small intercalated limestone beds yielded a diverse ostracod fauna indicating the change from restricted marine to intertidal or brackish conditions at the base of the formation to more open-marine conditions in the middle part. Janvier et al. (1984) placed the DCB about 80 m above the base of the Köprülü Formation based on a change in ostracod assemblages. A shaly horizon 50 m above this boundary yielded spores indicating the lower Tournaisian PC Palynozone (Higgs et al. 2002).

#### Azerbaijan and Armenia

The Geran-Kalasi section (Grechichnikova and Levitskii, 2011) is one of the most complete sections of the Upper Devonian–lower Carboniferous in Azerbaijan. It is located in the Nakhichevan Autonomous Region (west Azerbaijan) and mainly represent a marine continuous sequence with rich fossil associations: brachiopods, foraminifera, bryozoans, tabulate and rugose corals, brachiopods, ostracodes, crinoids, conodonts, algae and ammonoids are present. The section was described in detail by Grechichnikova and Levitskii (2011), who listed the fossil occurrences throughout the section and introduced brachiopod-based biozones. The same levels crop out in the Armenian flank of the Daralagez

mountain range, notably in the Arshaki-Arhbyur Gorge (Tolokonnikova 2016).

The DCB interval is represented by about 8 m of thinly bedded sandy limestones with bryozoans, brachiopods, ostracodes and conodonts (*Bi. costatus*, among others), followed by 5 m of ‘shaly silty, poorly exposed’ sediments and 7 m of sandy, thinly to medium-bedded sandy limestones. Grechichnikova and Levitskii (2011) attributed those beds to the upper part of the latest Famennian *Sphenospira julii*–*Spinocarinfera nigra* Brachiopod Zone. The lowermost Tournaisian is identified by the *Unispirifer praeulbanensis*–*Rhytiophora curtirostris* Brachiopod Zone, and is represented by 6 m of dark grey shales, with thin interbeds of siltstones and sandy limestones with brachiopods, tabulate and rugose corals, crinoids, conodonts and bryozoans (Abrahamyan et al. 1975; Alekseeva et al. 2018).

#### Uzbekistan

Preliminary data at the DCB in Uzbekistan are available from the Kule (Erina 2008; Narkiewicz et al. 2017) and Novchomok (Narkiewicz et al. 2017) sections. Both are located in the eastern part of the Kitab Geological Reserve area.

In the Kule section, two lithostratigraphic units are discriminated in the Famennian and Tournaisian: the Akbasay Formation composed of reddish grey cherty limestones and siliciclastics, followed by the Novchomok Formation, represented by ‘grey and light grey, fine-grained and micritic, sometimes brecciated crinoidal limestones’ (Yolkin et al. 2008). No shaly interval is reported near the boundary. A detailed sampling was carried by Erina (2008), who positioned the basal Carboniferous in the lower part of the Novchomok Formation based on the occurrence of questionable element of *Si. sulcata*. However, the presence in the same bed of an element of *Si. lobata* (figured as ‘*Si. angulata* n.sp.’ Erina) questions the position of the boundary. Narkiewicz et al. (2017) did not recover siphonodellids from the boundary beds, but placed the boundary in the uppermost part of the Akbasay Formation on the basis of the entry of *Po. p. subplanus*. Further studies confirmed the datum, thanks to the recovery of one element of *Si. sulcata* (Narkiewicz et al. 2021). Only one juvenile element of *Protognathodus* sp. is documented so far from the *Si. jii* Zone of the Kule section (Narkiewicz et al. 2021).

In the Novchomok section, the lower part is represented by marly shales and marls interbedded with pelitic and crinoidal limestones of the Yatavluk Formation, overlaid by the limestones of the Novchomok Formation. On the basis of a poorly preserved conodont fauna, including a few protognathodids, the boundary is tentatively placed in the uppermost part of the Yatavluk Formation (Narkiewicz et al. 2017). High mercury concentration in the latest Devonian were documented by Rakociński et al. (2021).

## Tajikistan

Successions spanning the DCB crop out in southern Tien-Shan (western Tajikistan), not far from the border with Uzbekistan (Bardashev et al. 2005, 2006; Bardasheva et al. 2004). Three parallel sections, named Shishkat 3, Shishkat 3a and Shishkat 3b, were measured and sampled in detail by Bardasheva et al. (2004). The DCB occurs within the Surkhobi Formation, a grey to light grey-cream limestone mostly made up by ‘clastic detrital and coarse-detrital rocks, as well as limestone breccia’ (Bardashev et al. 2006). Oolites were reported from the boundary beds, whereas mudstones or shales are not present. Foraminifers and conodonts are abundant; crinoids, brachiopods and ostracods are also present.

Despite the abundance of conodonts, the early siphonodellids are rare in the boundary beds, and only one element of *Si. sulcata* is reported both from sections Shishkat 3 and Shishkat 3a, and rare protognathodids are reported only from the upper part of the *Si. sulcata* Zone and above. Also, only an undifferentiated ‘*praesulcata* Zone’ is discriminated, and the scarce fauna makes further subdivisions difficult, even if the last occurrence of *Pa. gr. gonioclymeniae* about 11 m below the occurrence of *Si. sulcata* may indicate the beginning of the Hangenberg Event.

Detailed distribution data of foraminiferas are not provided, but faunal lists are provided for each lithological unit (Bardasheva et al. 2004). Several taxa of genera *Cribosphaeroides*, *Endoglossospiranella*, *Parathuramminites*, *Quasiendothyra*, *Septabrunsiina* and *Uralinella* are listed from a 10-m-thick interval approximately including the upper part of the *praesulcata* Zone, but their precise occurrence is not available. A less differentiated association, including *Parathuramina radiosphaerica*, *Parathuramminites mikhonoae*, *Bisphaera malevkensis*, *B. irregularis* and *Earlandia minima* is listed from a 5-m interval more or less coincident with the *Si. Sulcata* Zone.

## Kazakhstan

Sections exposing the DCB in Western Kazakhstan are treated in a dedicated chapter of the present special issue by Kulagina et al. (2021, this issue). However, a recent re-excavation and description of the Berchogur sections has been published by Nikolaeva et al. (2021). The first occurrence of *Si. sulcata* in Member 3 allows the localisation of the DCB. However, in their revision, Nikolaeva et al. (2021) identified a regressive event that most probably records the ‘Hangenberg Sandstone event’ concomitantly with a clear separation of Famennian and Tournaisian faunal associations (foraminifers, conodonts, ammonoids, etc.) within the uppermost part of Member 2b of the Dzhangana Formation. The position of the DCB should therefore be somewhat lowered in the succession compared to previously published data.

In southern Kazakhstan, the DCB is exposed in the Bolshoi Karatau Mountains, where upper Frasnian to lower Bashkirian

carbonates are exposed along platform-basin transects. Detailed biostratigraphic data for calcareous foraminifers (Zhaimina et al. 2017; Zhaimina and Kulagina 2019) allow to place the DCB, e.g. bed 15A in the Karamurum Section. As elsewhere in the former USSR, the base of the Carboniferous corresponds to the base of the *Tournayellina pseudobeata* Zone. Another interesting stratigraphic signal are breccias in the Balaturlan Member (Shukurganat Formation), which have been correlated with a global regressive trend (Zhaimina et al. 2017), supposedly correlatable to the Hangenberg Sandstone event.

## Mongolia

Rocks of Late Devonian and lower Carboniferous age crops out in several areas of Mongolia (Ariunchimeg et al. 2014; Ariunchimeg and Nyamsuren 2001), but only very few data are available. The DCB is not documented so far, as it corresponds to a formational boundary (e.g. Wang et al. 2005; Wang and Minjin 2004) or studies are limited to parts of the successions (e.g. Upper Devonian, Munkhjargal et al. 2021a, b, Suttner et al. 2020).

## Russia

This short overview on the DCB in Russia is necessarily incomplete. Information on the boundary are available from various regions in Russia (East-European Platform, Urals, Siberia, North-east Russia, Russian Arctic) and from a large suite of facies ranging from continental and coastal to basinal environments. The contribution of Kulagina et al. (2021, this issue) in this special issue provides a lot of details about DCB boundary sections with a special focus on the foraminifera biostratigraphy.

In the Moscow Basin, in the central part of the East-European Platform, the DCB corresponds to an unconformity between the Khovanshchinian and Malevkian Regional substages, and the correlation to global chronostratigraphy is complicated by abundant shallow-water conodont assemblages (e.g. Alekseev et al. 1996; Gatovsky et al. 2011). Although time is clearly missing, the region is important for latest Famennian vertebrate assemblages (e.g. Lebedev et al. 2018).

The Gumerovian Regional Substage, in which the DCB is situated, fills the stratigraphic gap in the Moscow Basin. It originates from the Zigan section on the western slope of the Urals (e.g. Pazukhin et al. 2009, Kulagina et al. 2021, this issue), and is today widely recognised in the stratigraphic schemes of the East European Platform and Urals. In its type section, the Gumerovian Regional Substage consists of c. 2 m of strata (bed 20 of the local nomenclature); its base corresponds to a sequence boundary. The base of the Carboniferous is placed at the base of bed 20c. However, Gatovsky (2015) reports an almost synchronous first occurrence of the conodont *Si. sulcata* and the miospore *V. pusillites* in the upper part of bed 20c. In the sequence

stratigraphic interpretation of Gatovsky (2015), bed 20a corresponds to a lowstand systems tract, bed 20b represents a transgressive systems tract and beds 20c to 20f to a highstand systems tract.

In the northern Urals, a lot of new data have been published from the Timan-Pechora region in recent years, especially from the Kamenka River and Vangyr River sections (e.g. Zhuravlev 2020; Zhuravlev and Sobolev 2019). The DCB is placed in the lower part of the Edzhid Formation. The latest Famennian is characterised by lagoonal, clayey carbonates, which contained conodonts as *Si. praesulcata*, *Si. bella*, *Si. quasinuda*, *Po. parapetus*, *Po. communis communis*, *Hindeodus penescitulus*, *Ps. graulichii* and *Ligonodina discreta*. The basal Carboniferous is represented by shoal and open marine carbonates containing *Si. sulcata*, *Si. bella*, *Si. quasinuda*, *Si. semichatovae* and *Patrognathus crassus*. The foraminifer *Tournayellina (Eotournayellina) pseudobeata* is found 0.5 m above the boundary (Vevl et al. 2012). The topmost Famennian carbonates are thought to have been deposited under regressive conditions, whereas the basal Carboniferous is transgressive. Tentative correlations to the Rhenish sequence stratigraphic model are discussed, especially to the relative sea-level curve. Vevl et al. (2012) indicated that the maximum regression occurs 0.3 m below the biostratigraphic DCB in the Kamenka River section.

In the northern Urals, important for the ongoing discussions for the revision of the DCB are records of *Pr. kockeli* in the Izyael River and Kozhim River sections (Nemirovskaya et al. 1993; Plotitsyn 2016).

The Omolon Massif in Far East Russia is often overlooked when discussing the DCB. The Upper Devonian-lower Carboniferous succession of the area had gathered a lot of attention in the 1980s (e.g. Bogush et al. 1987; Conil et al. 1982; Gagiev 1979; Shilo et al. 1984; Simakov et al. 1983; Swennen et al. 1986). Interestingly, the position of the DCB remains disputed. Depending on the interpretation of the fossil data and their correlation to other regions, the DCB of Gagiev (1979) is correctly placed after Kulagina et al. (2021, this issue) but has to be considerably moved up-section after Poty (1999) based on the distribution of both macro- and microfossils. This problem is important for our understanding of the DCB, the associated biological crises and recovery, and the Omolon Massif could play a central role in it. If the DCB remains in the 1980s position, the Omolon Massif can be considered a Carboniferous refuge for Devonian survivors of the biocrises. In this case, the Omolon Massif would be a starting point of the Carboniferous repopulation. However, this also means that the stratigraphic ranges for some fossil groups have to be extended at least locally, and that traditional biostratigraphic signals will fail in those refuges. If the proposed boundary of Poty (1999) is correct, then the faunal extinctions in the Omolon Massif follow the global scheme, but some organisms would have a (significant) earlier first occurrence than elsewhere if the original taxonomical attribution is correct. Doubts can be raised concerning the original

identification of the boundary solely based on a fragment of siphonodellid. Hence, these ‘Carboniferous’ newcomers would survive the end-Devonian extinction and provide the stock for the Carboniferous repopulation and diversification.

In Arctic Russia, Upper Devonian and lower Carboniferous rocks are exposed north of the Lena river. They belong to the Stolb, the Ebelakh and Bastakh formations (Mezhvilk 1958). The DCB should occur within the Ebelakh Formation, constituted by fine- to medium-grained feldspar-quartz sandstone with carbonate cement and small amount of dolomite and felsic volcanic matrix (Saraev et al. 2013). However, it is not possible precisely placing the boundary, as fossils are very scarce (Mezhvilk 1958). From the Lena River estuary area, Ivanovski (1967) noted that the oldest corals found in the Carboniferous strata are *Caninia cornucopiae* and *Uralinia multiplex*, which indicate probably an upper Tournaisian age, whereas older macrofauna seems to be lacking. Conodonts collected in the Stolb Formation (Izokh 2017) indicate an age not younger than the *Pa. rhomboidea* Zone, while the older assemblage from the Bastakh Formations is referred to the *isosticha*-Upper *crenulata* Zone (Izokh 2017; Izokh and Yazikov 2017).

The situation in Novaya Zemlya is also poorly known at a detailed scale. In the southern part of Novaya Zemlya, the uppermost Devonian-Tournaisian Lazarevskaya Formation includes neritic carbonate facies. The Famennian part of the Lazarevskaya Formation is composed of bioclastic grainstone and packstone and stromatoporoid boundstone (Schecoldin et al. 1994) that yield a rich and diverse assemblage of benthic organisms, including rugose corals (Gorsky 1938), brachiopods and foraminifers (Rzhonsnitskaya 1967). Shallow-marine carbonates extend at least up to the Viséan, with some intercalations of shale and bedded chert in the Tournaisian (Schecoldin et al. 1994). On the western coast of the southern area, Schecoldin et al. (1994) report in the Lazarevskaya Formation the occurrence of black shale and radiolarite that yielded lower Tournaisian conodonts allowing to place the DCB. However, only the Tournaisian species *Si. lobata*, *Si. crenulata* and *Si. obsoleta* were reported. In the Kara area (central-eastern part of the archipelago), the succession across the DCB consists of black bedded cherts, radiolarites, shales and rare carbonate intercalations. The conodont fauna consists of *Siphonodella* sp., *Patrognathus crassus* and *Clydogmathus cf. cavusiformis*.

## Europe

### Ukraine

In the Donets Basin, the DCB is located at the interface of the Devonian Porfiritovaya and Bazalievskaya formations. The uppermost Famennian is characterised by shallow-water carbonate facies, often dolomitized, with an abundant foraminiferal fauna



including *Quasiendothyra communis*, *Q. kobeitusana*, *Q. konensis*, *Cribrosphaeroides* sp., *Paracaligelloides florenensis*, *Tournayella* sp., *Septatournayella* sp., *Septaglomospiranella* sp. and *Septabrunsiina* sp., as well as the index conodont *Si. praesulcata* (Poletaev et al. 1990, 2011). Brachiopods and stromatoporoids are abundant (Berchenko and Poletaev 2007). An unconformity is developed at the base of the overlying Bazalievskaya Formation, but after Poletaev (1981) it might reflect local condition of sedimentation. However, this unconformity should probably be reconsidered under the perspective of the Hangenberg Sandstone event. This formation passes southwardly into the Porfiritiv Formation, which is composed of fine-grained, often dolomitic limestone with the previously reported foraminiferal assemblage, overlain by black shale with brachiopods and a more diverse conodont fauna composed of *Antognathus volnovachaensis*, *Bi. costatus*, *Bi. aculeatus*, *Si. praesulcata*, *Po. parapetus* and *Pseudopolygnathus* spp. (Poletayev et al. 2013), which likely indicate some intermixing of Devonian and Carboniferous species.

The lower Tournaisian is recognised regionally as the Buzinivian Horizon and corresponds to the Bazalievskaya Formation. The basal beds of this unit contain the Tournaisian conodonts *Siphonodella* aff. *sulcata*, *S. semichatovae* and *Patrognathus andersoni* (Poletayev et al. 2013). The foraminifers, though less diverse than in the underlying strata, are well characterised by *Bisphaera malevkensis*, *B. irregularis* and *Earlandia minima* and form the local ‘bispheric foraminifers zone’ (Poletaev et al. 2011) that possibly corresponds to the ‘unilocular zone’ of Poty et al. (2006).

## Belarus

The DCB is documented in two boreholes from the Pripyat Depression, in the southern part of Belarus (Avchimovitch et al. 1988, 1993). The authors studied a rich miospore association and noted a strong floral turnover across the boundary. Unfortunately, detailed lithological and stratigraphic data are not available. However, malformed spores of *R. lepidophyta* described by Avkhimovich (1993) allow the recognition of the *tener* event (sensu Prestianni et al. 2016) in this succession, i.e. the recognition of the HBS event equivalent.

## Serbia

The DCB is exposed in NW Serbia at the Milivojevic Kamenjar section, a large overturned olistolith spanning from latest Famennian to Serpukhovian (Korn and Sudar 2016; Sudar et al. 2018). Ammonoid and conodont fauna and biostratigraphy were recently documented by Korn and Sudar (2016) and Sudar et al. (2018), respectively.

The DCB interval ‘occurs in a continuous limestone succession, not interrupted by strata indicating the Hangenberg Event’

(Sudar et al. 2018, p. 172). However, a minor lithological change between dark-grey nodular and flaser limestone with high shale content (Famennian) to grey bedded and partly nodular limestone with minor shale content (Tournaisian) is documented. Ammonoids do not allow a precise biostratigraphy of this part of the section, as they are concentrated in a few levels dated to an early ‘*Wocklumeria* Stufe’ and to a generic ‘middle Tournaisian’ (Korn and Sudar 2016), i.e. the upper lower Tournaisian. Conodont data are more precise and a clear difference in the association is documented in connection with the lithological change: typical latest Famennian species occur in the topmost shaly limestone, whereas Carboniferous taxa as *Si. sulcata* and *Po. purus subplanus* are documented just above. A recent attempt to recover palynomorphs from the DCB interval shales failed as the material has a high thermal maturity (Prestianni, pers. com., 2020).

## Great Britain

In many parts of Great Britain, the DCB beds are found in the stratigraphic interval often called or associated with the ‘upper’ Old Red Sandstone (ORS). During the Upper Devonian to lower Carboniferous the palaeo-coastline moved from the south of the British Isles towards the north, and marine conditions gradually installed during this time slice. Hence, depending on the geographical position of the studied section, the DCB can be found in neritic marine, marginal marine or continental facies.

Marine conditions dominate the Devonian successions of Devon and Cornwall (SW Britain). There, a classical example for the DCB documented in neritic facies is the Pilton Mudstone Formation of the Culm Basin. In this formation, trilobites and cephalopods are documented ranging from the late Famennian *Wocklumeria* to the early Tournaisian *Gattendorfia* ammonoid zones (e.g. Goldring 1955). Further north, where the continental character of the deposits increases, especially palynomorphs have been demonstrated to be of critical stratigraphic importance (e.g. Clayton et al. 1986; McNestry 1988). In this special issue, the contribution of Graham and Sevastopulo (2021, this issue) illustrates this facies realm widely developed in the Upper Devonian and lower Carboniferous in southern Ireland, southern Wales and southern England (e.g. Mendips).

Further north, in the entirely continental facies realm, a recent biostratigraphic study in Scotland (Marshall et al. 2019) demonstrates the importance for the correct placement of the DCB. These new data for the stratigraphic distribution of palynomorphs show that the Upper ORS Kinnesswood Formation is Late Devonian in age, and not early Carboniferous as previously postulated. The DCB approximates with the base of the following Ballagan Formation and its coastal wetland environments (Millward et al. 2018), in which the oldest Tournaisian miospore zone was found at its base at Burmthouth.

These new dates change the perception of the vertebrate evolution in the late Devonian and early Tournaisian. The previously postulated range extension of typical Devonian fish into the Carboniferous can be refuted with the new age of the Kinnesswood Formation. This increases the amplitude of the magnitude of the Hangenberg event for the terrestrial realm, and the DCB becomes a clear barrier in the evolution of terrestrial vertebrates. This study also shows that the Ballagan Formation comprises the entire Tournaisian, and hence the tetrapods and arthropods of this formation fill important parts of the Roemer's Gap (Marshall et al. 2019).

## France

In France, Devonian-Carboniferous successions have been documented in different tecto-sedimentary contexts in various parts of the country, north and south of the future Variscan Chain. The best-known successions in the Montagne Noire and Avesnois are treated in this special issue respectively by Feist et al. (2021, this issue) and Denayer et al. (2021, this issue).

The Boulonnais area in northern France is the equivalent of the Belgian Brabant Parautochton as its western continuation emerging below the post-Palaeozoic cover near the Boulogne-sur-Mer area. The Devonian and Carboniferous successions of the Massif de Ferques are comparable to the one, e.g. in the Toumai drillings and Feluy area in Belgium (Denayer et al., 2021, this issue) but the transition is far less well dated. The Sainte-Godeleine Formation is a greenish to reddish sandy and shaly unit c. 50 m thick displaying shallow-water bioturbated facies almost devoid of fossil (Ager and Wallace 1967). Its age is extrapolated from the lower Famennian acritarch assemblage known at the top of the underlying Hydrequent Formation (Loboziak and Strel 1981). Rare limestone intercalation near the top of the Sainte-Godeleine Formation yielded ostracods that Lethiers (1982) attributed to the Tournaisian. A hiatus of the Strunian and possibly lower Tournaisian is possible after Hance et al. (2001) and Mansy et al. (2007).

South of the Boulonnais, the Palaeozoic strata equivalent to the Dinant Synclinorium are known through drillings below the Mesozoic-Cenozoic strata of the Paris Bassin. The Brevilliers drillings reached the Devonian and the DCB was documented by Combaz and Strel (1971). The Strunian part is typically an alternation of crinoidal grainstone and dark bioclastic shale and yielded the diagnostic miospores *Retispora lepidophyta*, *Raistrickia* spp. and *Vallatisporites* spp., including abnormal forms pointing to the *tener* event (Prestianni et al. 2016). The top of the Famennian (~195-m depth) is marked by a sandy limestone bed (probably the Hangenberg Sandstone equivalent basal bed of the Hastière Formation) followed by more massive crinoidal limestone

devoid of miospores, i.e. the lower part of the lower Tournaisian Hastière Formation.

In Brittany, the Vosges and the Morvan, the Devonian successions are usually well known, but in all these areas, the DCB has not been placed with certainty as the transition towards the Carboniferous is overwhelmed by coarse-grained siliciclastic deposits traditionally regarded as the record of the Bretonic phase of the Variscan orogeny.

In the Armorican Massif of Brittany, the Devonian and Carboniferous strata occur in the Chateaulin and Morlaix Synclines. A minimum latest Famennian age (late Strunian) for the youngest Devonian deposits has been documented (Zorn Formation, Morzadec and Strel 1980; Kermerrien Formation, Rolet et al. 1986; Dourduff Breccia, Coquel and Deunff 1977). The overlying mixed silicoclastic and volcanoclastic Kertangy Formation has yielded only badly preserved late Tournaisian miospores (Morzadec et al. 1975); the DCB and lower Tournaisian are either recorded in unfossiliferous volcanic tuff or in a hiatus.

In the northern Vosges Massif, the situation is similar to the Armorican one, with a thick sequence of Upper Devonian and lower Carboniferous flyschoid mixed siliciclastic and volcano-clastic sequence (Bruche Series, Blanalt and Lillie 1973) where the DCB cannot be placed with certainty (Fluck et al. 1987). In the southern Vosges, the Chagey and Stalbert formations represent the fossiliferous Strunian (Asselberghs 1926) and passed upwards to the flyschoid Bois de la Gigoutte Formation and its lateral equivalent Markstein Series, where Maass and Stoppel (1982) documented a conodont assemblage indicating the DCB. In the Morvan Massif (north-eastern tip of the French Central Massif), the facies are similar with the southern Vosges. The thick silicoclastic sequence exposed in Savigny-Poil-Fol starts with greenish siltstone that yield a (late) Strunian assemblage of brachiopods and trilobites (Weyer 1965), overlain by sandstone and conglomerate in which intercalated siltstone beds yielded the brachiopod *Ziganella* attributed to the basal Hastarian (basal Carboniferous) by Legrand-Blain (1997).

In the Pyrenees, the DCB is found in the upper part of the Barousse Formation (sensu Perret 1993), which crops out on both sides of the French-Spanish border. The currently best and most detailed studied sections (Milles, Saubette, Garcet) are found on the French side (e.g. Cygan et al. 2002; Perret 1988, 1993). These sections represent pelagic carbonate facies (grey micritic limestones), which near the boundary is interrupted by the intercalation of an argillaceous level up to several decimetres-thick, which can be correlated with the Hangenberg (Black) shale and maybe also the Hangenberg Sandstone. The general facies evolution, but also the faunal content, show striking similarities with many sections in the Montagne Noire (see Feist et al. 2021, this issue). In the current context of the position of a revised DCB, it is interesting to note that Perret (1988) highlighted the stratigraphic

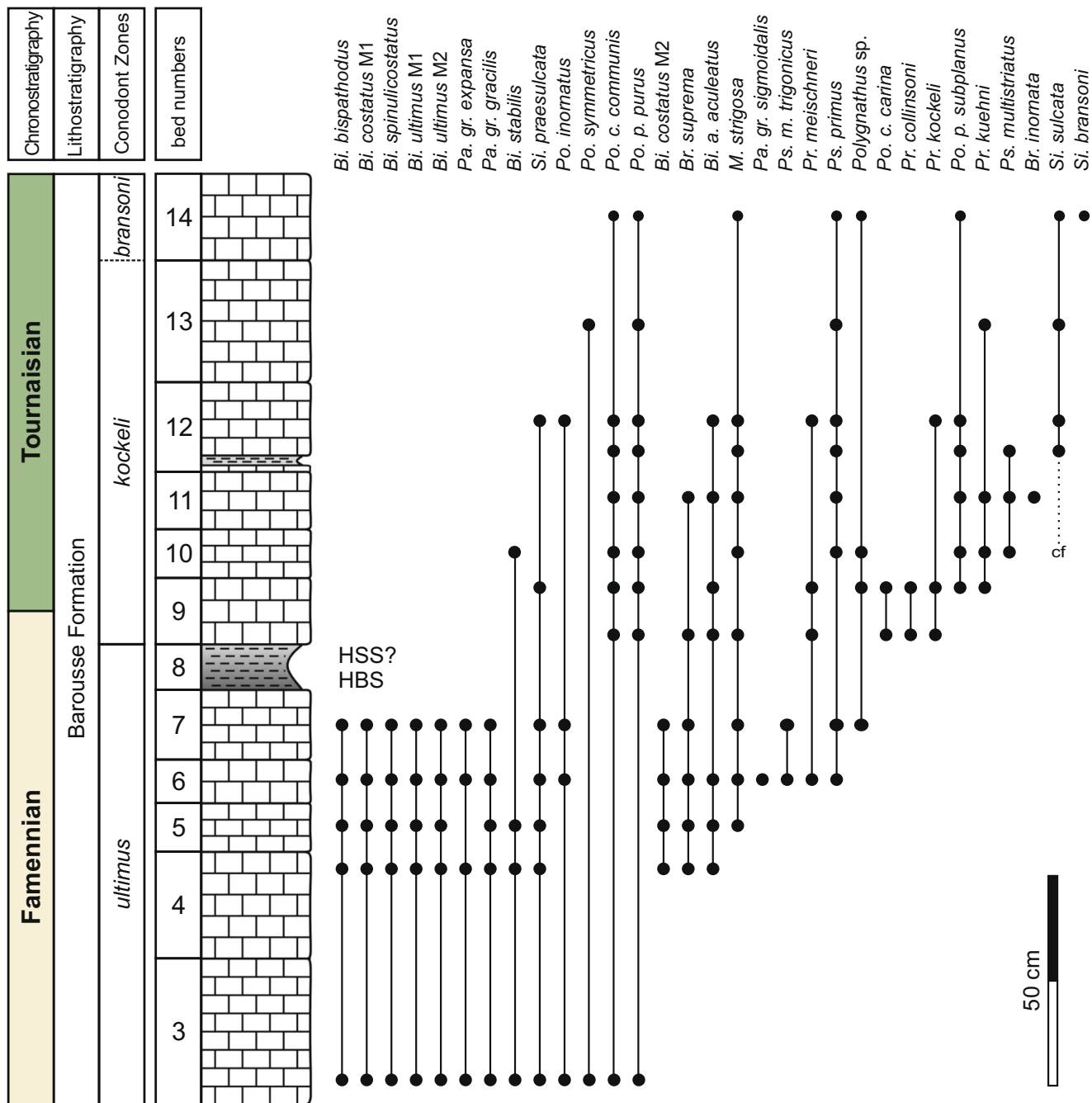


Fig. 6 DCB in the Milles section, French Pyrenees. Overview on stratigraphy, lithology and conodont distribution based on the data in Perret (1993), Cygan et al. (2002) and Kaiser et al. (2009). HBS =

Hangenberg Black Shale, HSS = Hangenberg Sandstone (modified from Perret 1993; Kaiser et al. 2009)

importance of protognathodids (*Pr. meischneri*, *Pr. collinsoni*, *Pr. kockeli*) in the interval between the top of the siliciclastic level and the first occurrence of *Si. sulcata* several decimetres higher in the nodular limestone facies.

In the Milles section (Fig. 6), based on rich conodont assemblages, the DCB was placed between samples 11 and 12 by Cygan et al. (2002) using the first occurrence of *Si. sulcata*. Later, Kaiser et al. (2009) lowered the DCB within bed 9 due

to the occurrence of *Pr. kuehni* in the upper part of the bed. *S. sulcata* was surely reported from bed 12, and as *Si. cf. sulcata* from bed 10. Hence, the DCB is now located only one decimetre above the level of the equivalent of the Hangenberg Sandstone. In the Saubette section, the situation is almost identical. The base of the Carboniferous is found 50 cm above the HBS equivalent (sample 14 in Perret 1988, 1993). The Garcet section is important for the absence of a lithological change in the DCB Interval. It

conodonts (western Palaeothethys)	South America									Africa				Australia							
	Miospores	revised	old	Amazon Basin	Solimoies Basin	Parnaiba Basin	Parana Basin	Brazil	Bolivia	Peru	Chile	Argentina	South Africa	Ghana	Morocco	Algeria	Lybia	W Aus.	E. Aus.		
lower <i>Si. crenulata</i> (part)																					
<i>Si. sandbergi</i>																					
<i>Si. jiji</i>																					
<i>Si. duplicata</i>																					
<i>Si. bransonii</i>																					
<i>Px. kockelli</i> <i>sulcata</i>																					
<i>Bi. ultimus</i>																					
<i>Bi. costatus</i>																					
<i>Bi. ac. aculeatus</i>																					
<i>P. gr. expansa</i>																					

conodonts (western Palaeothethys)	Asia														Europe						
	Miospores	revised	old	Vienna Basin	India	Iraq	Saudi Arabia	Turkey	Mongolia	Russia	United Kingdom	France	Spain	Portugal	Sw Portugal	W. Europe	old	new			
lower <i>Si. crenulata</i> (part)																					
<i>Si. sandbergi</i>																					
<i>Si. jiji</i>																					
<i>Si. duplicata</i>																					
<i>Si. bransonii</i>																					
<i>Px. kockelli</i> <i>sulcata</i>																					
<i>Bi. ultimus</i>																					
<i>Bi. costatus</i>																					
<i>Bi. ac. aculeatus</i>																					
<i>P. gr. expansa</i>																					

◀ **Fig. 7** Simplified correlation chart for the strata exposing the DCB discussed herein. The DCB is indicated with the red dotted line. The column Omolon Massif (Russia) proposes 2 boundary positions depending on the interpretation of the available data (see Kulagina et al. 2021; Poty 1999). Plain numbers make references to the beds, where the DCB has been identified precisely; numbers in italics are references to sample numbers or beds, which are used for placing the DCB. Chile: Quebrada Zamolana, section D, Rubinstein et al. 2017; Ghana: Essipon Beach section, Atta-Peters and Anan-Yorke 2003; Morocco: Lalla Maimouna section North, Becker et al. 2013, Amessoui Syncline: El Atrous section, Kaiser et al. 2013; Vietnam: Cat Co section, Shizuya et al. 2020; Turkey: Bartın Area, Topluca section, Denayer 2016; Taurides, Natlaş section, Atakul-Özdemir 2017; Azerbaijan: Geram-Kalas section, Grechichnikova and Levitskii 2011; Russia: Urals: Zigan River section, Gatovsky 2015; Petchora, Kamenka River section, Zhuravlev 2020; Iberia: Cantabria: Baleas Quarry section, García-Alcalde and Menéndez-Alvarez 1988; Pyrénées: Milles section, Kaiser et al. 2009; SW Portugal: Pedra Ruiva section, Pereira et al. 2007. The small blue triangles indicate glacial deposits after Lakin et al. (2016)

remains completely carbonate. The DCB is placed between samples F and G (Cygan et al. 2002), based on the FO of *Si. sulcata*. In contrast to the other Pyrenean sections, protognathodids have not been reported so far from this section.

In northern Corsica, an upper Famennian neritic succession contains a rich benthic fauna (corals, foraminifers, stromatoporoids) (Krylatov and Mamet 1966). The Capitello Limestone yielded a diverse assemblage of *Quasiendothyra* spp. indicating the Strunian. Baudelot et al. (1981) indicate the occurrence of conodonts pointing also to a Strunian age for the unit. The occurrence of *Po. ex. gr. marburgensis*, known to disappear during the Hangenberg Black Shale event, might tie up the age to the upper part of the Strunian. Up-section, the Capitello Limestone is silicified and passes to calcareous shales then to a Culm-like succession with sandstone and conglomerate intercalations. The DCB, though not placed with precision, should fall into the silicified limestone unit (Baudelot et al. 1976).

## Spain and Portugal

In the Cantabrian Zone in Northwestern Spain, the uppermost Devonian and lowermost Carboniferous strata are represented by the Ermita Formation and lateral equivalents. On the Cantabrian block, the Ermita Formation is missing or highly condensed. Towards the Asturian Arc, the sections become more complete. The uppermost Ermita Formation consists of channelled quartz arenites and conglomerates that cut into the underlying quartz sandstones (Keller et al. 2008). They called this the Llombera beds. Keller et al. (2008) pointed that crinoidal shoal deposits called Baleas Limestone are at least locally sandwiched into the Llombera beds. However, in a more distal setting, the Llombera beds are replaced by muddy limestone and dark shales, which can be topped by crinoidal limestones of the Bales Limestone (van Loevezijn and Raven 2017). The DCB is often placed in the top of the Ermita

Formation and the carbonated lateral equivalents. García-López and Sanz-López (2002) summarised available biostratigraphic data. In the classical Olleros de Alba section, the top of the Ermita Formation contains a conodont assemblage, which has been attributed to the *Si. Sulcata* Zone, but together with *Si. praesulcata* and *Si. sulcata* appears a single element, which may already indicate the *Si. duplicata* Zone (sample OLL-3 of García-López and Sanz-López 2002). A similar earliest Carboniferous age was reported for the top of the Ermita Formation at Santiago de las Villas. Here a conodont assemblage contains *Ps. primus*, *Bi. stabilis*, *Pr. kockeli*, *Pr. kuehni*, *Pr. meischneri* and *Siphonodella* sp. In the Baleas Quarry, García-Alcalde and Menéndez-Alvarez (1988) placed the DCB in the middle of the Baleas Limestone, using the first occurrence of *Si. sulcata* in their sample J-263. The bed J-259, 75 cm below, yielded a rich conodont assemblage including *Si. praesulcata* and *Pr. meischneri*. In the Palentine Domain, Sanz-López et al. (1999) tentatively placed the DCB in the upper part of the Vidrieros Formation using conodont assemblages containing protognathodids and siphonodellids.

In the Spanish Pyrenees, the Barousse Formation (Perret 1993) ranges from the middle Famennian *Pa. marginifera* Zone to the Tournaisian *Si. sandbergi* Zone (Sanz-López 2002a, b). The unit consists of nodular limestones with an increase in the abundance of crinoids and cephalopods in the upper part. A shale bed, 0.2 to 2-m-thick ('level B' of Boyer et al. 1974), occurs within the unit and is considered an equivalent of the Hangenberg Black Shales (Sanz-López 2002a). Conodonts collected from the limestones just above yielded conodonts from the *Si. sulcata* to possibly the *Si. crenulata* Zones (Boersma 1973; Perret 1993).

In the southern Central Iberian Zone, García-Alcalde et al. (2000) correlated in the lower part of the Casa de La Vega Formation unfossiliferous dark claystones with the Hangenberg (Black) Shale. These claystones are intercalated into limestones, which contained late Famennian to late Tournaisian conodonts.

In SW Spain and Portugal, the succession of the Iberian Pyrite Belt (IPB) and neighbouring structural units in the South Portuguese Zone exposes uppermost Devonian to lowermost Carboniferous strata (e.g. Moreno et al. 1996; Pereira et al. 2008). Within the stratigraphic section of the IPB, the Phyllite Quartzite Group (PQG) forms the sedimentary basement of the Volcanic-Siliceous Complex (VSC), which is composed of a complex patchwork of intrusive and extrusive volcanic rocks, siliciclastic sedimentary rocks and siliceous rocks. Strata of the PQG often contains late Famennian palynozones (VCo, VH), but in some cases, the latest Famennian palynozones (LL and LN) are found in shales attributed to the PQG (e.g. Nasceidos Member, Pomarão anticline, Oliveira et al. 2006; Mina de São Domingos section, Pereira et al. 2007). However, the earliest Carboniferous palynozones have so far not been found in the VSC, but the boundary is often placed within that lithostratigraphic unit.

In Southwest Portugal, the DCB is situated within the Tercenas Formation. Oliveira (1990) reported clymenids and brachiopods indicating a latest Famennian to early Carboniferous age. Pereira et al. (2007) showed the presence of the LN and VI palynozones in the Pedra Ruiva section. Hence, the DCB has to be placed in the c. 2.5-m-thick interval between the samples 279 and 280 of Pereira et al. (2007).

## Conclusions

This compilation shows that a lot of data on the knowledge of the DCB exist even in regions and countries (Fig. 7), which have not been on the forefront of discussions on the boundary in recent years. Detailed modern data are surprisingly available more often than expected, but there are also many regions, which requires new detailed looks to avoid any miscorrelations. Around the globe, when facies changes to dark shales or sandstones occur, those units have often been correlated with the well-known name Hangenberg of the Rhenish succession. Although this approach might be useful, without biostratigraphic control this seems to be an overoptimistic approach for boundary identification. However, many sections recorded faunal turnover and collapse of ecosystems in the Hangenberg extinctions. This allows the placement of the boundary even if index fossils are absent or scarce.

This compilation clearly shows the need to correlate the boundary through different facies and on the different palaeocontinents. This will not be achieved with a single criterion, but requires multiple lines of evidences. If the boundary definition should work globally, we have to step back from the focus on conodont-bearing carbonate pelagic facies, and better integrate the calendar of the often, dramatic changes in the latest Devonian–basal Carboniferous biosphere and geosphere.

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## Declarations

**Conflict of interest** The authors declare that they have no conflict of interest.

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