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The geological foundation of the Netherlands: the early Carboniferous, Devonian and older

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ABSTRACT

This chapter addresses the geological development of the Netherlands and its immediate surroundings during the older portion of the Paleozoic (Cambrian to early Carboniferous, 538-323 Ma). During this period northwestern Europe experienced two major phases of mountain building: the Caledonian and the Variscan phases. These laid the foundations for the geology of the Netherlands. The Caledonian phase is associated with the progressive closure of the Iapetus Ocean between Laurentia to the northwest and Baltica and Avalonia to the southeast and south, respectively. Although classically considered as a Caledonian phase, the Brabantian Deformation event is related to the progressive closure of the Rheic Ocean which existed south of Avalonia. This is represented by an angular unconformity surrounding the ancient Brabant Massif (Belgium). This unconformity separates deformed Silurian and older, (very) low-grade metamorphic units from overlying relatively undeformed non-metamorphic Middle Devonian deposits. The continued northward migration of microplates originating from Gondwana ultimately closed the Rheic Ocean, and formed the Variscan Mountains and foreland basin during the Carboniferous. It is against this backdrop that the mainly Middle Devonian to lower Carboniferous sedimentary succession of the Netherlands accumulated along the northern edge of the Rhenohercynian sea. Borehole and seismic data of this succession are still very scarce. However, over the last decade several research initiatives provided valuable new insights in this fundamental episode in the geological history of the region.

<< A tilted Famennian sandstone-shale alternation in the Trooz quarry (southeast of Liège, along the Vesdre River in Belgium). Photo: Geert-Jan Vis.

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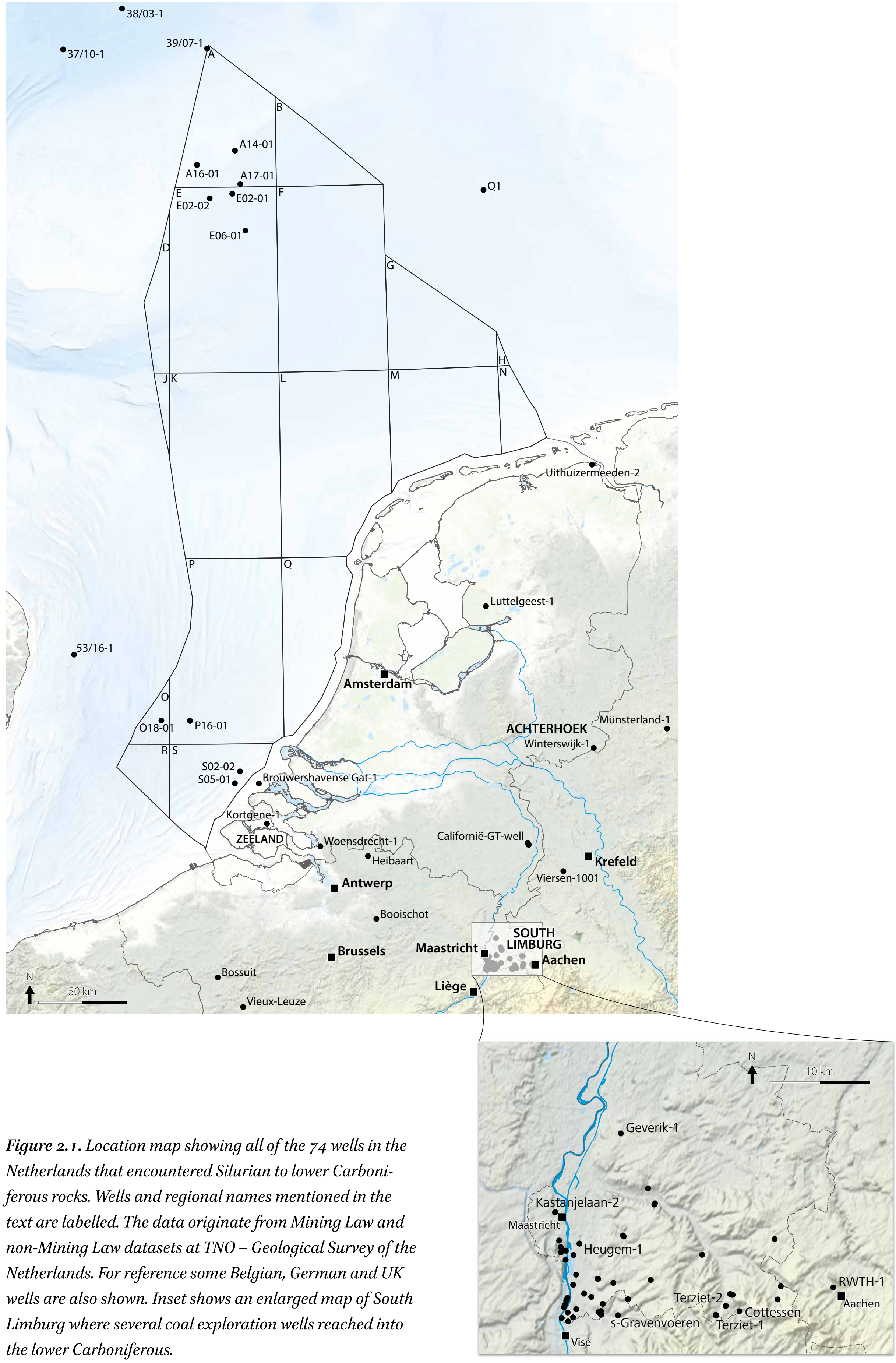

Figure 2.1. *Location map showing all of the 74 wells in the Netherlands that encountered Silurian to lower Carboniferous rocks. Wells and regional names mentioned in the text are labelled. The data originate from Mining Law and non-Mining Law datasets at TNO – Geological Survey of the Netherlands. For reference some Belgian, German and UK wells are also shown. Inset shows an enlarged map of South Limburg where several coal exploration wells reached into the lower Carboniferous.*

Introduction

The oldest strata encountered in the Netherlands date back to the Silurian period. In nearby Belgium, older lower Paleozoic (Cambrian to Ordovician) strata are widely encountered in outcrop and boreholes. In this chapter, the geological development during the Cambrian to early Carboniferous (539-323 Ma) of northwestern Europe is outlined and linked to the stratigraphic succession encountered in the Netherlands.

During the older part of the Paleozoic, many fundamental changes occurred, both biologically and geologically. After the first terrestrial vegetation emerged during the Ordovician (~480 Ma, Steemans et al., 2009), vascular plants radiated during the Silurian (~430 Ma) setting the stage for the spread of vegetation onto land during the Devonian. Two of the big five mass extinctions occurred during this period: the Late Ordovician (~444 Ma) and Late Devonian extinctions (375-360 Ma). Added to this there was a major tectonic reorganization along the northern margin of the Rheic Ocean.

In spite of the fundamental changes occurring during the Paleozoic, the expression of this period in the subsurface of the Netherlands is difficult to disentangle. None of the Paleozoic successions is exposed in the Netherlands, although ample outcrops are known from Belgium and Germany. In the Netherlands, pre-Serpukhovian (pre-Namurian) sedimentary rocks are only known from 74 boreholes (Fig. 2.1) or are inferred from seismic data. In the ultimate southwest and southeast of the country, this Paleozoic succession is found at relatively shallow depth (<500 m). Going northwards, the depth rapidly increases, reducing the number of boreholes and seismic surveys that reach it. Based on the interpretation of seismic data, the top and base of the lower Carboniferous limestone succession (Dinantian) have been interpreted in a patchy pattern across the onshore Netherlands (Ten Veen et al., 2019).

Interest in the occurrence of gas-prone source rocks in the northern offshore has grown over the last decades (Gerling et al., 1999; Doornenbal et al., 2019). Especially the coals in the upper Visean Elleboog, Scremerston and Yoredale formations have been considered as potential source-rocks, but others have been identified as well (Kearsey et al., 2018). About a decade ago, geothermal exploration activities increased, focussing on Upper Devonian siliciclastics and lower Carboniferous (Dinantian) limestones. Between 2017 and 2020 the Green Deal Ultra Deep Geothermal Energy programme explored the possibilities for ultra-deep, i.e. deeper than 4000 m, geothermal projects in the Netherlands, focussing on lower Carboniferous limestones. Over the last five years, activities related to the possible construction of the Einstein Telescope have generated, and continue to generate, new data for southern South Limburg and neighbouring Belgium (see Textbox 1). All these developments have generated new data and insights that are incorporated in this chapter.

Historic studies

Numerous studies have addressed the Paleozoic succession during the last 150 years. Because the Paleozoic occurs at shallow depths in and around the southeast of the Netherlands (South Limburg), most studies have focussed on that region. An overview of the most important contributions treating the pre-Serpukhovian (pre-Namurian), is given below (also see Vis & Houben, 2023).

The detailed geologic mapping by Staring (1869, map sheet 27) was the first to indicate Paleozoic rocks in or near the Netherlands. In the southeast of the Netherlands, between Maastricht and Aachen, his map shows various units which we now know as dating to the pre-Serpukhovian Paleozoic, such as 'Culm' (lower Carboniferous basinal shale), 'Bergkalk' (lower Carboniferous limestone) and 'Kolenlooze zandsteen' (Serpukhovian, Namurian sandstone). Nearly half a century later, a drilling campaign in the search for coal provided a two-dimensional component to the understanding of the distribution of the Paleozoic represented by a cross section (Van Waterschoot van der Gracht, 1918).

The first detailed Paleozoic subcrop map of South Limburg was published by Sax (1946). The mapping campaign used the enormous amount of data collected through the exploitation of upper Carboniferous coal in South Limburg. Whereas most previously published maps and profiles approach the geology from the surface, Sax (1946) compiled data from shafts and upward-directed borings from within the coal mines. The map does not include the pre-Serpukhovian Paleozoic succession, but the accompanying cross sections do indicate the presence of lower Carboniferous (Dinantian) limestones ('Kolenkalk') near Voerendaal (well SM-59). Bless et al. (1976) presented a more regionally extensive Paleozoic subcrop map for South Limburg and its surroundings. Some of the challenges posed by Bless et al. (1976) related to the distribution of Devonian, Dinantian, Namurian as well as late Westphalian strata of South Limburg, were subsequently addressed by Kimpe et al. (1978), who provide updated maps of the Paleozoic abrasion surface and its immediate cover.

The most recent Paleozoic subcrop map of South Limburg and neighbouring areas was published by the Geological Survey of the Netherlands (Rijks Geologische Dienst, 1995). This map, and the Geological Atlas of the Subsurface of the Netherlands (TNO-NITG, 1999), provide geological cross sections through South Limburg. Remarkably, these profiles show Paleozoic strata down to 6 km below sea level, while the deepest borehole (Geverik-1) only reaches 1580 m below sea level and seismic data have

Einstein Telescope: watching the universe from underground

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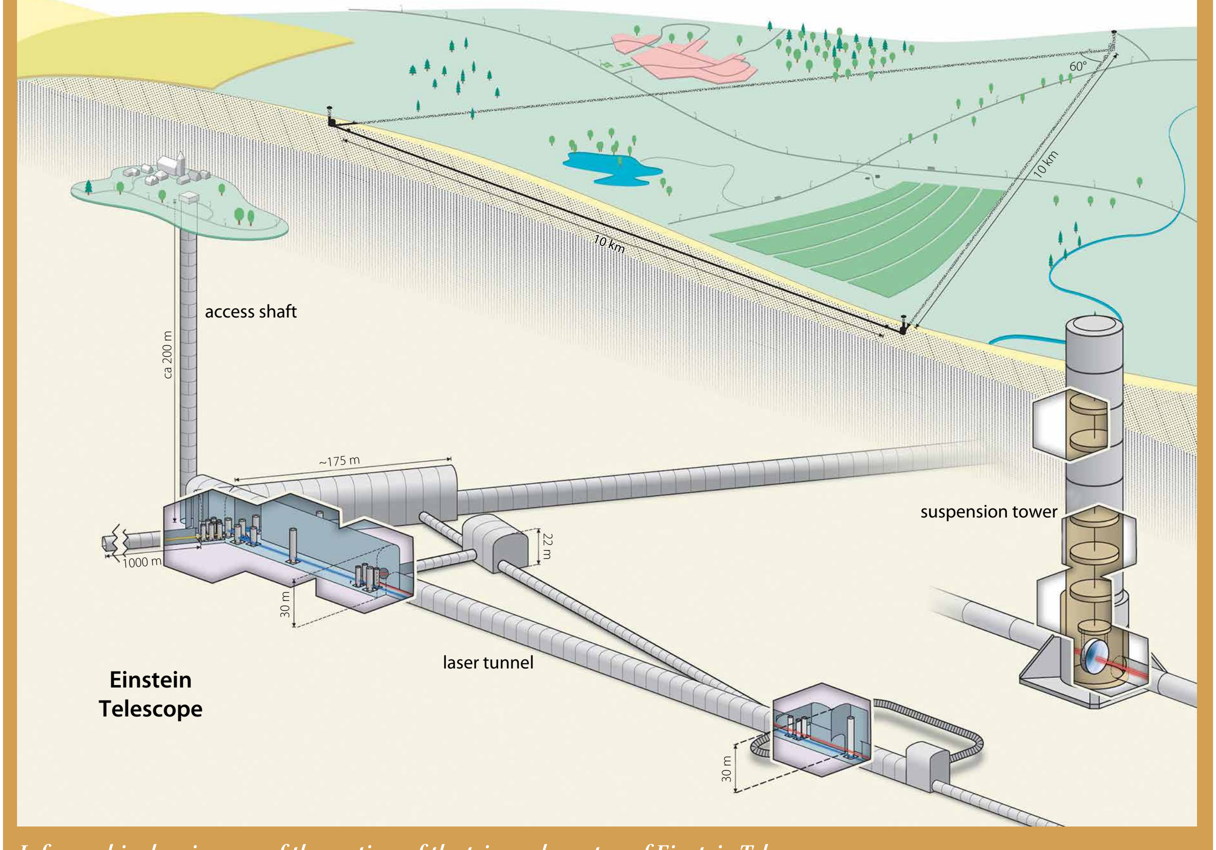
From 2035 onwards the third generation gravitational wave observatory may be operational 200-300 m below the rolling land surface of South Limburg and neighbouring Belgium. Gravitational waves are ripples in space-time (analogue: ripples at the water surface after a stone was thrown into a pond) caused by some of the most violent and energetic processes in the universe. Their existence was predicted by Albert Einstein in his general theory of relativity. The strongest gravitational waves originate from colliding black holes and neutron stars and from supernovae (Einstein, 1916, 1918).

On 14 September 2015, the gravitational-wave observatory LIGO in the United States made the first physical observation of a passing gravitational wave, using laser inferometry. These undulations in space-time were caused by two colliding black holes 1.3 billion light-years away. For better triangulation (determining the origin of a gravitational wave) and higher accuracy, the third generation gravitational-wave observatory will be built in Europe and underground. Detecting gravitational-waves underground has the advantage of co-registering less noise from activities at the surface and thereby increasing the accuracy.

A situation where more or less unconsolidated rock overlies competent rock, provides ideal conditions for this purpose. In the South Limburg-Belgium region, these conditions are met. Soft Cretaceous and Cenozoic rocks overly generally competent, consolidated Upper Devonian and lower Carboniferous sedimentary rocks. Unfortunately, the region in South Limburg straddling the border with Belgium is poor in both borehole and seismic data, due to limited exploration activities for resources such as coal and hydrocarbons. These data are necessary to position the enormous underground facility, which consists of a triangular tunnel system with tunnels measuring up to 10 km in length. At the vertices, large underground caverns will hold sensitive supercooled mirror systems. To support the case for placing the Einstein Telescope in this region, various studies and a data acquisition campaign are underway. Besides the South Limburg-Belgium region, also Sardinia and possibly southeastern Germany hope to attract this high-tech project.



Infographic showing one of the vertices of the triangular setup of Einstein Telescope.
Credits: Nikhef/Thijs Balder.

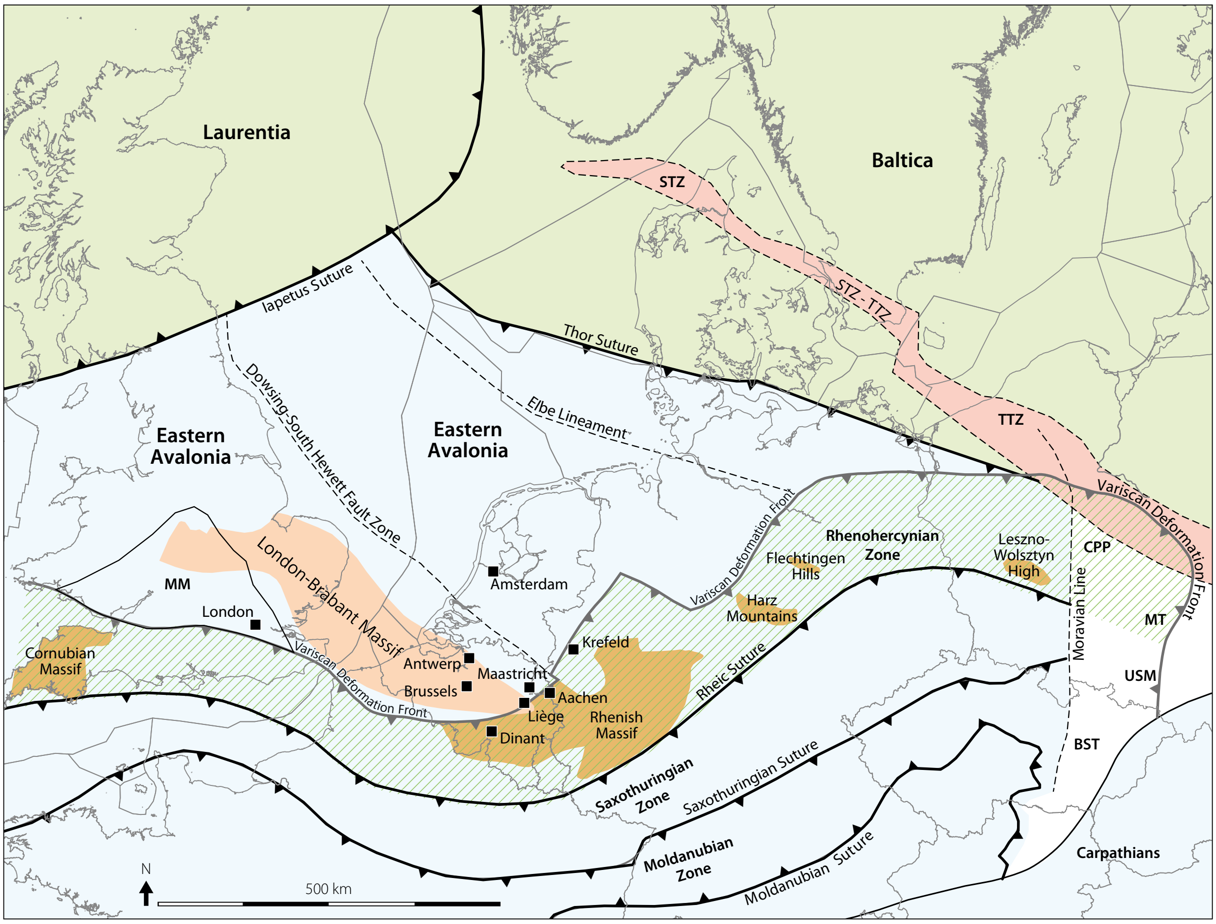


Figure 2.2. *The tectonic setting of the area around Eastern Avalonia, showing the amalgamated terranes and key terminology. STZ = Sorgenfrei-Tornquist Zone; TTZ = Teisseyre-Tornquist Zone; CPP = Central Polish Paleorift; BST = Bruno-Silesian Terrane; USM = Upper Silesian Massif; MT = Malopolska Terrane; MM = Midlands Microcraton. The map is compiled using various sources (Debacker, 2001; Verniers et al., 2002; Pharaoh et al., 2006; Woodcock et al., 2007; Sintubin et al., 2009; Bełka et al., 2010).*

poor resolution below the top-Paleozoic interface. Hence, some caution is advised when using these cross sections.

Updated lithostratigraphy

The limited number of wells penetrating the older Paleozoic succession in the Netherlands has hampered the definition of lithostratigraphic units. The current definitions are partially incomplete, are incorrect or are informal. During the last decades new wells became available and new studies have provided new insights. Therefore, an update of lithostratigraphic definitions is timely and is proposed in this chapter (see section Stratigraphy).

Tectonic setting

During the time-interval covered in this chapter (Cambrian to early Carboniferous), the plate-tectonic amalgamation of northwest Europe was completed (Fig. 2.2). The Netherlands was situated on continental crust of the eastern part of the microcontinent Avalonia. In the late Ordovician this microcontinent collided with Baltica, resulting in closure of the Tornquist Ocean and formation of the **Thor Suture**, and in the mid-Silurian with Laurentia, resulting in closure of the Iapetus Ocean and formation of the **Iapetus Suture** (Verniers et al., 2002; Pharaoh et al., 2006; Woodcock et al., 2007; Smit et al., 2016). Based on the timing of collision(s), the widespread occurrence of an angular unconformity overlain by red-coloured Devonian continental deposits and the paleogeographical position, this places these events within the **Caledonian cycle** (e.g. McKerrow et al., 2000). However, as explained in Textbox 2, strictly speaking true Caledonian deformation (i.e. the Scandian phase) is reserved for Silurian deformation in Scotland, Greenland and Norway, directly related to closure of the Iapetus Ocean. However, as in Belgium, where the effects of (a) late early Paleozoic deformation event(s) and the overlying unconformity can be observed in outcrop, the Netherlands occupies a south-central position within the eastern parts of Avalonia, at a considerable distance from the Iapetus suture to the NNW and the Thor suture to the ENE.

Based on faunal records, Cocks & Fortey (1982) showed that the Avalonian microplate was attached to Gondwana during the Cambrian in a southern high-latitude position, and migrated during the Ordovician to a lower latitude near Baltica, to which its fauna in the Late Ordovician is similar. Paleomagnetic data indicate that Avalonia occupied positions at 60°S in Early Ordovician Tremadocian, 40°S in Late Ordovician (Caradoc), 20°S in Silurian and 10°S in Late Devonian Famennian times (Torsvik et al., 1996; Cocks, 2000; Torsvik & Cocks, 2016).

In the late Early Ordovician, Avalonian crust rifted from Gondwana and rapidly drifted northwards. Based on faunal records, it is thought that it collided with Baltica by Late Ordovician (mid-Caradoc to mid-Ashgill) times (Vecoli & Samuelsson, 2001; Verniers et al., 2002). There is a general consensus that Tornquist Ocean subduction was towards the south but, because of the absence of a protuberant deformation belt and the absence of Tornquist Ocean subduction-related magmatism, the collision of Avalonia and Baltica has often been referred to as a soft-docking event, accompanied by a large amount of dextral displacement. Late Ordovician-early Silurian calc-alkaline magmatism documented in Belgium and the British Isles is slightly too young and far too distant to be attributed to the Avalonia-Baltica collision (e.g. Verniers et al., 2002; Linnemann et al., 2012). Nevertheless, there is clear evidence that significant deformation in the southern North Sea area (Mid North Sea High and southern

Caledonian, Scandian, Acadian, Brabantian, Ardennian – what's in a name?

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In northwest Europe red-coloured Devonian clastic deposits unconformably overlie deformed, metamorphic lower Paleozoic deposits. This reflects a deformation event of regional extent. In the British Isles this long-recognized deformation event was defined as the **Caledonian Orogeny**, named after the Roman name for Scotland, where the effects of this orogeny were first and best documented (e.g. Suess, 1885; Haug, 1900; Stille, 1924). With the advent of plate tectonics and the recognition of the existence of different 'Caledonian' events, other names were introduced, such as **Scandian**, **Acadian**, **Brabantian** and **Ardennian**. These are explained below.

The **Caledonian Orogeny** encompasses a series of early Paleozoic to Devonian events, including intra-oceanic events, associated with the progressive closure of the Iapetus Ocean between Laurentia to the northwest and Baltica and Avalonia to the southeast (present-day orientation). The **Scandian event**, that took place from ~440 to 410 Ma, was the regionally most important Caledonian event and reflects the closure of the Iapetus Ocean and associated continent-continent collision between Laurentia to the NW and Baltica and Avalonia to the SE in mid-Silurian (Wenlock) times. Its effects are documented in Scotland, Norway and East Greenland, and include thick-skinned thrusting with high-pressure metamorphism (e.g. Gee et al., 2012). Surprisingly, there is no significant deformation along the northwestern leading edge of Avalonia that can be linked to the Scandian event (Woodcock et al., 2007, 2019).

The **Acadian event**, originally defined in the Canadian Appalachians, was introduced in the British Isles by Soper et al. (1987) in order to explain Early to Middle Devonian deformation affecting the Lower Devonian lower Old Red Sandstone, often preserved in sinistral transtensional basins in England and Wales. This deformation event took place after closure of the Iapetus Ocean and resulted in sinistral strike-slip, transtension and transpression in Scotland, Northern Ireland, England and Wales. In Wales and Northern England, sinistral transpression along the NW side of the Midlands Microcraton (Avalonia) was associated with ~400-390 Ma cleavage development.

The **Brabantian event** and **Ardennian event** were defined in Belgium at the start of the 21st century as a response to potentially erroneous long-distance correlations of deformation events based merely on timing (e.g. Debacker, 2001; Verniers et al., 2002). In Belgium, two separate angular unconformities occur between lower Paleozoic slates and red-coloured Devonian continental deposits, one occurring in northern and central Belgium, and one occurring in southern Belgium and northern France (e.g. Sintubin et al., 2009). The angular unconformity surrounding the Brabant Massif (N-Belgium) separates deformed Silurian and older, low-grade metamorphic units from overlying relatively undeformed diagenetically-altered Middle Devonian deposits. As the timing of the deformation event in the Brabant Massif coincides with the Acadian event in the British Isles (see Debacker et al., 2005), the term Acadian event was introduced also in Belgium (Van Grootel et al., 1997). However, at the start of the 21st century the Acadian event in the British Isles was considered to be related to closure of the Iapetus parts of Ringkøbin Fyn High) and the North German-Polish Caledonides is associated with Avalonia-Baltica collision. This includes Late Ordovician-Silurian thrusting and amphibolite metamorphism around the Thor Suture and foreland basin development on the southern margin of Baltica (e.g. see De Vos et al., 2010 and references therein). Smit et al. (2016, 2018) attributed the absence of a protuberant deformation belt to Devonian-Carboniferous extension with collapse of the deformed Tornquist accretionary prism.

During the Ordovician and Silurian, the Iapetus Ocean progressively closed. From around the mid Ordovician onwards (i.e. after the Grampian Orogeny in Scotland) the Iapetus Ocean became subducted to the NW below Laurentia. The Southern Uplands accretionary prism in Scotland testifies of this subduction (e.g. Verniers et al., 2002; Woodcock et al., 2007 and references therein). The joint Avalonia-Baltica continent collided with Laurentia by Wenlock (mid-Silurian) times and gave rise to the continent Laurussia (e.g. Cocks, 2000). This collision, reflected by the Scandian deformation event, resulted in intense deformation with high pressure metamorphism in Scotland, Greenland and Norway (e.g. Gee et al., 2008, 2012), but only minor deformation along the suture between Avalonia and Laurentia (Woodcock et al., 2007, 2019). The main deformation along the northern (WSW-ENE trending) margin of Avalonia occurred later, during the Early Devonian, and was characterized by sinistral transpression, accompanied by low grade metamorphism and cleavage development at ~400-390 Ma (Soper et al., 1987;

Ocean, whereas observations in Belgium showed that the deformation within the Brabant Massif was difficult to link to either the closure of the Tornquist Ocean (between Baltica and Avalonia) or the closure of the Iapetus Ocean (between Laurentia and Avalonia-Baltica) (e.g. Debacker, 2001; Verniers et al., 2002; Sintubin & Everaerts, 2002). As a working solution, a different name, the **Brabantian event** or **Brabantian deformation phase**, was introduced (see Debacker, 2001 and Verniers et al., 2002). The angular unconformity surrounding the Ardennes Inliers (S-Belgium) separates deformed Ordovician and older, (very) low-grade metamorphic units from overlying deformed latest Silurian to Early Devonian (very) low-grade deposits, which were again deformed during the Variscan Orogeny (e.g. Sintubin et al., 2009). This event, only observed in the Ardennes Inliers now situated in the Variscan deformation belt, took place well before the Acadian and Brabantian events, and cannot be tied to a specific early Paleozoic shortening event. Hence, the term **Ardennian event** or **Ardennian deformation phase** was introduced (e.g. Verniers et al., 2002). However, this nomenclature is still under discussion (e.g. Herbosch et al., 2020; Herbosch, 2021) and there is no convincing evidence yet that the Ardennian event was compressional (Sintubin et al., 2009). The unconformity surrounding the Ardennes Inliers could also reflect a Silurian rifting event, later overprinted by the Variscan Orogeny (e.g. Sintubin et al., 2009).

Currently the Acadian deformation event is no longer considered to be associated with the closure of the Iapetus Ocean or the Tornquist Ocean, but is interpreted to be a proto-Variscan event along the northern margin of the Rheic Ocean. This interpretation reconciles the Acadian deformation in the British Isles and the Brabantian deformation in northern and central Belgium (Debacker, 2001; Woodcock et al., 2007; Sintubin et al., 2009).

Table summarizing the various terminologies on the 'Caledonian' events listed in the text.

Event name	Description
Scandian	Most important Caledonian event of regional extent reflecting the closure of the Iapetus Ocean and associated continent-continent collision in Wenlock times between Laurentia to the NW and Baltica and Avalonia to the SE. Deformation documented along Iapetus suture (Scotland, Greenland and Norway).
Acadian (in British Isles)	Deformation of lower Devonian Old Red Sandstone and underlying units in the British Isles. In Wales and Northern England sinistral transpression was associated with ~400-390 Ma cleavage development along NW side of Midlands Microcraton (Avalonia). Currently attributed to an event along the northern margin of the Rheic Ocean.
Brabantian	Deformation of Silurian and older deposits in Anglo-Brabant deformation belt, along NE side of Midlands Microcraton (Avalonia), documented in N Belgium (e.g. Brabant Massif). Deformation from Wenlock onwards, with ~400 Ma cleavage development. Currently attributed to an event along the northern margin of the Rheic Ocean.
Ardennian	Angular unconformity in N France and S Belgium between Ordovician and latest Silurian-Devonian. No convincing evidence for compressional event and potentially related to rifting event along southern side of Avalonia (Rheic margin).

Woodcock et al., 2007). This deformation event is called the Acadian Deformation event (Soper et al., 1987; see Textbox 2).

During the early Paleozoic a ~NW-SE trending basin developed along the NE side of the Midlands Microcraton (i.e. the core of the eastern part of Avalonia), called the Brabant Basin. This Brabant Basin, situated in northern Belgium and eastern England, consists of a thick, fault-bounded Cambrian succession, overlain by a thinner Ordovician succession, in turn overlain by an often very thick Silurian succession. During the Late Ordovician, around the time of Avalonia-Baltica docking, widespread slumping took place (Debacker et al., 2001; Debacker, 2012). Possibly, also the Asquempont Detachment System, an originally NE-dipping low-angle normal fault system that can be traced throughout the Brabant Massif, and that was formed between the middle Katian and the time of cleavage development (see Herbosch et al., 2008; Herbosch & Debacker, 2018, and original references therein), can be related to this time period. The Silurian succession shows evidence of foreland basin development, with a thick package of fine-grained turbidites, except where it overlies the northeastern parts of the Midlands Microcraton (Verniers et al., 2002 and references therein). The Brabant Basin was deformed from Wenlock times onwards, accompanied by Silurian foreland basin development, and resulted in the Anglo-Brabant Deformation Belt, of which the Brabant Massif in northern Belgium forms the principal outcrop area (e.g. Van Grootel et al., 1997; Sintubin, 1999; Debacker, 2001; Sintubin & Everaerts, 2002; Verniers et al., 2002; Debacker et al., 2005). An angular unconformity separates the anchizonal to epizonal, deformed units of the Anglo-Brabant Deformation Belt from overlying Givetian (Middle Devonian) and younger sedimentary deposits (De Vos et al., 1993). Deformation of the NW-trending Brabant Basin along the northeast margin of the Midlands Microcraton, with development of the Anglo-Brabant Deformation Belt, took place progressively over ~30 Myr, and culminated in cleavage development at ~400 Ma (Debacker et al., 2005 and references therein). This timing corresponds to the Acadian deformation in the British Isles, but because of the deformation belt's position and orientation at right angles to the Iapetus suture, the term Brabantian deformation event has been introduced (Debacker, 2001; Verniers et al., 2002; Sintubin et al., 2009).

The Brabantian deformation event is essentially an inversion of the pre-existing Brabant Basin, with the strongest inversion occurring in the central basin parts, surrounded by outward propagating (Silurian) foreland basins (e.g. Debacker, 2001; Sintubin & Everaerts, 2002; Debacker et al., 2005). Interestingly, along/above the northeast side of the Avalonia Microcraton, Silurian to lowermost Devonian deposits are diagenetically altered and hardly deformed (e.g. Verniers et al., 2002; cf. Welsh Borderland along NW-side of Midlands Microcraton). This implies that also in other parts of eastern Avalonia outside of areas of significant basin inversion, the presence of relatively undeformed Silurian deposits can be expected, as seen in wells Kortgene-1 and O18-01 in the southwestern Netherlands (Houben & Vis, 2021).

Deep seismic profiles in the southern North Sea (Blundell et al., 1991) show a southwest dipping deep reflector in the mantle below the Dowsing-South Hewett Fault Zone, interpreted as a possible remnant of an Ordovician intra-Avalonia subduction zone (Lee et al., 1993; Pharaoh et al., 1995; Pharaoh, 2018). The Dowsing-South Hewett Fault Zone (Fig. 2.2) may represent a suture between two different terranes within the amalgamated eastern Avalonia microplate. This terrane boundary continues to the southeast towards the Roer Valley Graben (e.g. Pharaoh et al., 2006; Pharaoh, 2018) and may have controlled later graben development.

A series of negative gravity anomalies within the Anglo-Brabant Deformation Belt are often interpreted as granitic bodies (Everaerts et al., 1996; Mansy et al., 1999), or are simply referred to as low-density bodies (Debacker, 2001; Sintubin & Everaerts, 2002). These bodies had a significant influence on Brabantian deformation and must have been present prior to this event (De Vos, 1997; Sintubin, 1999; Debacker, 2001; Verniers et al., 2002; Debacker et al., 2005). It is difficult to link these to Tornquist Ocean subduction but, if subduction-related, they may have been caused by subduction of a short-lived ocean within the eastern Avalonia terrane assembly along the Dowsing-South Hewett Fault Zone (e.g. Debacker, 2001; Verniers et al., 2002). As an alternative, Linneman et al. (2012) suggested an intracontinental origin.

During the Ordovician northward migration of Avalonia, the Rheic Ocean opened between Avalonia and Gondwana (Ziegler, 1990). After the collision of Avalonia with Baltica, the Rheic Ocean started to subduct, both southwards below the Armorican terranes and northwards below Avalonia (Franke, 2000). The ~Late Ordovician-Silurian 'Ardennian event' in Belgium is still disputed, and might even be related to Rheic extension, rather than reflecting a compressional event (see Sintubin et al., 2009; cf. Herbosch et al., 2020; Herbosch, 2021). The younger Brabantian event is regarded as the equivalent of the Acadian event, and is not related to the Iapetus Ocean closure to the north but instead represents a 'proto-Variscan' event related to progressive closure of the Rheic Ocean to the south (Debacker, 2001; Woodcock et al., 2007; Sintubin et al., 2009). Possible causes for the Brabantian and Acadian events are flat-slab subduction or impingement of another Gondwana-derived terrane (Woodcock et al., 2007).

During the Devonian and Carboniferous, the three-plate convergence gave way to a two-plate convergence (e.g. Ziegler, 1990). The latter involved the northward drift of Armorican terranes from Gondwana against Laurussia, with a major sinistral translation between Laurentia-Greenland and Fennoscandia-Baltica. During the Early Devonian a narrow (400-600 km wide) and short-lived oceanic basin opened before the Rheic Ocean had completely closed (Franke, 2000; Franke et al., 2017). This oceanic basin is referred to as Rhenohercynian Ocean (e.g. Von Raumer et al., 2009; Pharaoh et al., 2010; Franke et al., 2017). It was located south of Avalonia and north of the closing Rheic Ocean in what now is central Europe and southwest England. Multiple reconstructions of the complex plate configurations at that time have been postulated in literature (e.g. Shail & Leveridge, 2009; Von Raumer et al., 2009; Franke et al., 2017). To the north, this basin had a passive margin, while to the south its oceanic crust subducted beneath the Gondwana-derived terranes (Shail & Leveridge, 2009; von Raumer et al., 2009; Franke et al., 2017). The famous ophiolites of the Lizard Complex in southwest England are associated with early-formed oceanic crust in this basin (Kirby, 1979; Shail & Leveridge, 2009), as is the mid-ocean ridge basalt (MORB) in central Germany (Franke, 2000). In the Netherlands and Belgium no evidence for this small oceanic basin has been identified. Due to its small size and short-lived character, we here prefer the use of the term Rhenohercynian sea. The Rhenohercynian sea is not to be confused with the Rhenohercynian Zone, which is the Variscan fold-and-thrust belt that eventually formed after collision of the Variscan orogenic wedge with the passive northern margin of the Rhenohercynian sea (Fig. 2.2, Pharaoh et al., 2010).

The northward subduction of the Rheic Ocean below Avalonia was accompanied by periods of slab roll-back during the Late Devonian and Carboniferous. This resulted in a collapse of the former Tornquist Ocean accretionary prism around the Thor Suture (Smit et al., 2016, 2018). It also led to wide-spread formation of horsts and grabens around the British Isles and the southern North Sea, via reactivation of the pre-existing basement fabric (Smit et al., 2016, 2018). In addition to this back-arc extension, some expulsion of Baltica to the east may have taken place as well, causing further E-W extension in the southern North Sea area (Coward, 1993). The net result was a series of WNW-ESE trending, fault-bounded half-grabens in the southern North Sea, similar to the basins described in the English onshore (Fraser & Gawthorpe, 1990; Chadwick, 1993; Hollywood & Whorlow, 1993). The basins are separated by granite-cored horsts such as the one encountered in well A17-01 (Ziegler, 1990). Sedimentation resumed in the area of well A17-01 during the late Middle Devonian. Strong Middle Devonian to early Carboniferous block faulting, controlled by the basement fabric, has been documented from both the northern flank of the Anglo-Brabant Deformation Belt (Muchez & Langenaeker, 1993) and the Mid North Sea High (Quirk, 1993; Maynard & Dunay, 1999). Similar block-faulting is thought to exist below the Netherlands (Cameron & Ziegler, 1997) and evidence to support this can be found on good quality 3D seismic surveys in the Groningen area (Ten Veen et al., 2019) or the northern offshore (Ter Borgh et al., 2018b).

Continued subduction ultimately resulted in closure of the Rheic Ocean, with formation of the nappe complex of the Variscan Mountains during the late Carboniferous (Ziegler, 1990; Franke, 2000). The thrust front was situated south of the Brabant Massif and is represented by the Midi-Aachen Thrust (Fig. 2.2; Huis in 't Veld & Den Hartog Jager, 2025, this volume). The Rhenohercynian Zone, which includes the Ardennes and the Rhenish Massif (Fig. 2.2), has an Avalonian crust and Avalonian faunas, and was thrust upon more northerly areas of Avalonia during the late Carboniferous (Pharaoh, 1999; Franke, 2000; Oncken et al., 2000).

Climate and biotic evolution

Cambrian

The Cambrian Period (541-485 Ma) is widely known for its rapid diversification of life forms, known as the Cambrian explosion. This produced the first representatives of all modern animal phyla. Although diverse life forms prospered in the oceans, the terrestrial realm remained predominantly barren, not boasting more complex life than microbial soil crusts (Retallack, 2009). Initially stromatolites were the most abundant colonial marine organisms. The advent of grazing and bioturbation, and the proliferation of silicate-bearing sponge-like creatures like the archaeocyathids, led to the demise of stromatolites, microbial mats and associated organisms by the end of the early Cambrian (Hagadorn & Bottjer, 1997). Cambrian climatic conditions are not understood in detail. Recent oxygen-isotope analyses of phosphatic microfossils suggest temperatures that were comparable to late Mesozoic and early Cenozoic greenhouse climate conditions (20-25°C at southern high-latitudes, Hearing et al., 2018).

Ordovician

The Ordovician (485-444 Ma) marks another dramatic phase of evolutionary radiation: the Great Ordovician Biodiversification Event (GOBE, Munnecke et al., 2010). A true regime shift was the introduction of filter-feeding organisms. Groups that remained dominant during the rest of the Paleozoic arose; notably articulate brachiopods, molluscs (bivalves, gastropods and cephalopods), echino-

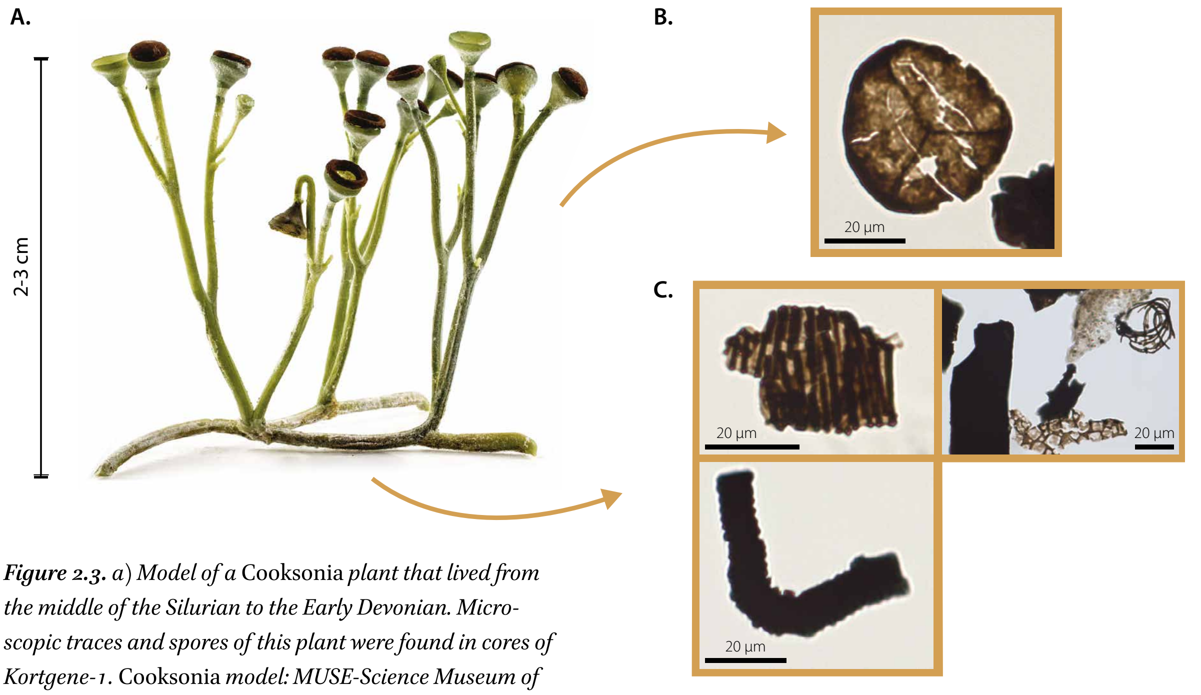


Figure 2.3. a) Model of a Cooksonia *plant that lived from the middle of the Silurian to the Early Devonian. Microscopic traces and spores of this plant were found in cores of Kortgene-1.* Cooksonia *model: MUSE-Science Museum of Trento (via Wikimedia Commons, CC BY-SA 3.0); b) Spore of* Cooksonia *(Kortgene-1, 1910.4 MD). These spores are also found in fossilized* Cooksonia *specimens; c) Microscopic tubular plant fragments of* Cooksonia*, that stabilized and anchored the plant (Kortgene-1, 1898.5 MD).*

derms (notably crinoids) and reef-building corals (Droser & Finnegan, 2003). The Ordovician also manifests the oldest unequivocal evidence for colonization of the terrestrial realm by arthropods (MacNaughton et al., 2002) and possibly by arachnids (spiders, Lozano-Fernandez et al., 2020). The first microscopic evidence (spores) of land-plants date back to the Early Ordovician (Steemans et al., 2009; Rubinstein et al., 2010; Strother & Foster, 2021). These plants were small, non-vascular, ephemeral and probably directly evolved from green algae (Strother & Foster, 2021).

The Early Ordovician was very warm and lacked continental-scale ice-sheets. Sea-surface temperatures as high as 45⁰C are suggested by some authors (Song et al., 2019). More temperate climatic conditions characterize the Middle Ordovician, with further cooling into the intense and relatively abrupt Hirnantian Glaciation (Late Ordovician). Sedimentological evidence for this glaciation is found across much of Gondwana, which was located in the high-latitude southern hemisphere (Delabroye & Vecoli, 2010; Cocks & Torsvik, 2021). Despite the polar position of Gondwana, the Hirnantian Glaciation remains remarkable given the very high atmospheric Ordovician CO_2-concentrations, which are estimated to be 14 to 16 times pre-industrial levels (Crowley & Baum, 1995). Associated with the Hirnantian Glaciation, the oldest of the 'big-five-mass-extinctions' occurred. This global extinction eliminated about 50% of marine genera and nearly 85% of marine species (Harper et al., 2014).

Silurian

The Silurian is known for the appearance of vascular land-plants, of which the first fossil records date to the middle Silurian (Wenlock, Edwards & Feehan, 1980). The earliest-known representative of this group is *Cooksonia* (Fig. 2.3). This relatively small (<10 cm high) and simple plant lacked leaves and an extensive root-system, essentially comprising a sporangium on a stem. Preferred habitats were along rivers and streams, forming micro-forests. Spores originating from *Cooksonia* have been recorded in Silurian to lowermost Devonian strata in the Netherlands (Houben & Vis, 2021, wells Kortgene-1 and O18-01; Fig. 2.1). Somewhat more complex plants that are considered the first club-mosses (lycopsids) known as *Baragwanathia* appear in the early Ludlow (Rickards, 2000). Perhaps in conjunction with the advent of land-plants, the earliest-known animals to have fully adapted to terrestrial conditions appeared during the middle Silurian, including the millipede *Pneumodesmus* (Selden & Read, 2008). In the marine realm, reef-building corals radiated during the Silurian, leading to the first extensive Paleozoic coral reefs (Copper, 2002). The early Paleozoic jawless fishes continued to thrive, and early jawed fishes, including the shark-like acanthodians and the armour-headed placoderms, evolved during the Silurian. Eurypterids were the apex predators of the Silurian oceans. These arthropods, also named sea scorpions, somewhat resemble modern horseshoe crabs.

The Silurian contains at least six large (up to 6‰) carbon isotope excursions (Melchin et al., 2020), which are interpreted to represent major carbon-cycle perturbations that are associated with minor mass extinction events and anoxia (Munnecke et al., 2010). Intense glaciations are not inferred for the Silurian, although there is sedimentological evidence for early Silurian glaciers (Díaz-Martínez & Grahn, 2007). Based on oxygen isotope data, it appears that the climate warmed immediately after the Hirnantian Glaciation, reaching a maximum towards the end of the Llandovery. The Wenlock was relatively cold and the late Silurian (Ludlow-Pridoli) is characterized by progressive warming.

Devonian

The Silurian-Devonian boundary coincides with a major carbon isotope perturbation, termed Klonk (Małkowski & Racki, 2009) which in terms of extinction-intensity was only a minor, relatively gradual event (Bambach, 2006). The drastic carbon-cycle perturbations that characterize the Silurian are no longer recorded in the Devonian (419-359 Ma).

The Devonian marks the 'mid-Paleozoic predator revolution' (Signor & Brett, 1984) and the related occupation of the free water column, the 'nekton revolution' (Klug et al., 2010). During the Devonian the first ammonoids and numerous fish groups (e.g. toothed sharks and giant placoderms) originated. More advanced vertebrates, including the first tetrapods developed (e.g. Blieck et al., 2010; Niedźwiedzki et al., 2010) and the most extensive reef complexes of the Phanerozoic (e.g. Copper & Scotese, 2003) flourished. The 'greening of land' by the diversification and spread of land plants, including the oldest true forests (e.g. Stein et al., 2012) had a profound effect on terrestrial ecosystems.

The Late Devonian is characterized by a series of biotic crises, which together comprise one of the 'big-five-mass-extinctions' (Becker et al., 2016). The most prominent of which are the Kellwasser Crisis near the Frasnian-Famennian boundary and the latest Famennian Hangenberg Crisis. These crises led to the extinction of the stromatoporoid reefs and placoderm and ostracoderm fishes (Kaiser et al., 2016). Also, the chitinozoans, stratigraphically-important marine palynomorphs, do not occur after the Devonian (Paris et al., 2000). The *Archeopteris* flora and associated *Retispora lepidophyta* spore assemblages (Streel, 1986), as observed in well Kastanjelaan-2 in Maastricht (Bless et al., 1981b), disappeared during the Hangenberg Crisis. The onset of the latter is marked by widespread black shale deposition (Hangenberg Black Shale in Europe; Kaiser et al., 2016). This black shale, which has not (yet) been identified in the Netherlands, coincided with the main extinction level for marine invertebrates (Kaiser et al., 2016).

During the Devonian, vascular land plants spread (Algeo et al., 2001), enhancing silicate weathering (Le Hir et al., 2011) and consequently leading to a long-term atmospheric CO_2-decline. Average temperatures dropped almost continuously from the Early Devonian all the way to the late Carboniferous (Buggisch et al., 2008; Joachimski et al., 2009). In spite of lacking sedimentological evidence of Devonian glaciation on Gondwana, a pronounced eustatic fall across the Frasnian-Famennian boundary is attributed to a glaciation (Isaacson et al., 2008).

Early Carboniferous

During the Mississippian (~early Carboniferous, 359-323 Ma), the Netherlands were situated near the equator (Torsvik & Cocks, 2016). The Tournaisian was a relatively warm, and largely ice-free period, starting with a major eustatic sea-level rise. Some authors link this rise to a deglaciation during the ultimate stage of the Hangenberg Crisis (Kaiser et al., 2016). The Netherlands gradually moved northward through the intertropical convergence zone, leading to the establishment of warm, monsoonal climatic conditions, with strongly seasonal run-off patterns (Falcon-Lang, 1999; Bennett et al., 2017). Following gradual cooling through the late Tournaisian and early-mid Visean (Buggisch et al., 2008), continental ice sheets started to grow in Gondwana (the UK-Brigantian Stage, Fielding et al., 2008; Limarino et al., 2014). During the following Serpukhovian (Late Mississippian, early Namurian), the ice-sheets progressively increased in volume. This increase in ice-volume was accompanied by conspicuous climatic cyclicity, ascribed to 100-kyr orbital eccentricity forcing (Waters & Condon, 2012). Climatic cycles have been described in detail for the Belgian succession by Poty (2016) who determined the onset of glacio-eustatic cyclicity already in the early Visean. Likewise, these third-order sequences have been determined in the Rhenish Kulm Basin in Germany (Herbig, 2016). Around the Early Pennsylvanian (late Namurian), maximum ice-volume stabilized, setting the stage for the glacio-eustasticy dominated depositional patterns of the Pennsylvanian and early Permian (see Bouroullec & Geel, 2025, this volume; Huis in 't Veld & Den Hartog Jager, 2025, this volume).

The Carboniferous Period is best known for its terrestrial ecosystems. Coastal peat swamps inhabited by the fern-like *Rhacophyton* had appeared by Late Devonian times, but a shift to dominance of arborescent lycophytes occurred during the early Carboniferous (Scheckler, 1986). Cyclic changes in coastal vegetation occurred in response to glacio-eustatic cycles. This is exemplified in the Visean Yoredale facies, which was deposited in the northern Dutch offshore (Ingrams et al., 2020).

Stratigraphy

Boreholes that were drilled over the last years have led to new insights regarding the pre-Serpukhovian Paleozoic stratigraphy of the Netherlands. When comparing with the previous edition of this chapter (Geluk et al., 2007), substantial additions and changes to the stratigraphy were needed. As work on Paleozoic stratigraphy is still ongoing, new insights have not completely matured and have not been ratified by the Stratigraphic Committee of the Geological Survey of the Netherlands. The revized stratigraphy below should be considered as a proposal for an updated lithostratigraphic subdivision of the Silurian, Devonian and lower Carboniferous successions in the Netherlands.

Chronostratigraphic terminology

The duration of the time interval discussed in this chapter spans approximately 200 million years. The oldest rocks within the Brabant Massif in Belgium are Cambrian (541-485 Ma) in age. The Ordovician spans the interval between 485-444 Ma, the Silurian between 444-419 Ma, the Devonian between 419-359 Ma and the older half of the Carboniferous between 359-330 Ma.

In northwestern Europe, numerous chronostratigraphic terminologies for the late Devonian and early Carboniferous are employed, particularly on the stage and substage levels (see the overviews in Hance et al., 2006a,b; Thorez et al., 2006; Poty et al., 2014; Schindler et al., 2018). Because this book follows the internationally recognized terminology as outlined by the International Commission on Stratigraphy (ICS, Gradstein & Ogg, 2020), we provide a brief overview of regionally used chronostratigraphic terms (Table 2.1). We discuss only the Devonian and early Carboniferous (Mississippian), because regional variations in older Paleozoic strata are of less importance.

According to the ICS, the Devonian System comprises three series. The Early Devonian comprises in ascending order the Lochkovian, Pragian and Emsian stages. The

Table 2.1. *Overview of regionally used chronostratigraphic terminology for the Devonian and lower Carboniferous of northwest Europe. German regional chronostratigraphic units after Herbig (2016). UK units are after Waters et al. (2011). References to Belgian regional stratigraphic nomenclature are provided in the text.*

	Global Series	Subseries	Regional Series	Global Stages	Global Substages	Regional Stages	UK Substages	Belgian Substages	German Substages
Carboniferous (pars.)	Pennsylvanian (pars.)	Early	Silesian (pars.)	Bashkirian (pars.)		*Namurian*	*Yeadonian*		
							Marsdenian		
							Kinderscoutian		
							Alportian		
	Mississippian	Late		Serpukhovian			*Chokierian*		
							Arnsbergian		
							Pendleian		
		Middle	Dinantian	Visean		*Visean*	*Brigantian*	*Warnantian*	
							Asbian		
							Holkerian	*Livian*	
							Arundian	*Moliniacian*	
							Chadian		
		Early		Tournaisian		*Tournaisian*	*Courceyan*	*Ivorian*	
								Hastarian	*Hangenbergian*
Devonian	Late			Famennian	Latest			*Strunian*	
					Late				*Dasbergian-Wocklumian*
					Middle				*Hembergian*
					Early				*Nehdenian*
				Frasnian					*Adorfian*
	Middle			Givetian					
				Eifelian			*Couvinian*		
	Early			Emsian					*Emsian*
				Pragian					*Siegenian*
				Lochkovian			*Gedinnian*		

Gedinnian is a Belgian equivalent of the Lochkovian, whereas the Pragian ICS-Stage overlaps with the German Siegenian stage. The Middle Devonian comprises the Eifelian and Givetian stages. The Couvinian is the equivalent of the Eifelian Stage in the Ardennes. The Late Devonian comprises two stages: the Frasnian and the Famennian. The German Upper Devonian Stufen I-VI, also referred to as the Adorfian, Nehdenian, Hembergian, Dasbergian, and Wocklumian stages are now also considered regional terms (Table 2.1; Schindler et al., 2018). The Famennian Stage is subdivided into four formal substages (Early, Middle, Late and Latest). The Latest Famennian corresponds to the Strunian regional substage (after Etroeungt in Belgium, Streel et al., 2006). The Hangenbergian in Germany is an equivalent to the Strunian and early Hastarian in Belgium (e.g. Paproth et al., 1983; Herbig, 2016).

The 'early' Carboniferous, according to the ICS is to be referred to as the Mississippian Series. The Mississippian consists of three subseries (Early, Middle and Late) which correspond to the Tournaisian, Visean and Serpukhovian Stages respectively. Historically, the Carboniferous succession of western Europe has been subdivided into two series. The lower part, dominated by marine depositional environments, was called the Dinantian and the upper part, dominated by terrestrial depositional environments, was called the Silesian. The Dinantian was divided into two stages; the Tournaisian and Visean. The Silesian was divided into the Namurian, Westphalian and Stephanian. This traditional scheme became obsolete with the definition of the mid-Carboniferous boundary, which corresponds to a boundary within the Silesian, corresponding roughly to the Namurian A-B boundary in traditional nomenclature (Dusar, 2006; Wagner & Winkler Prins, 2016). This also means that the Mississippian and Dinantian stages are not fully equivalent. In Belgium, Dinantian substages were defined by Conil et al. (1977); in ascending order Hastarian (early Tournaisian), Ivorian (late Tournaisian), Moliniacian (early Visean), Livian (middle Visean), and Warnantian (late Visean). Following revisions, these substages are now tied into a framework of modern biostratigraphic, lithostratigraphic, and sequence stratigraphic data (Poty et al., 2014). In Great Britain a similar subdivision on substage-level was formally proposed by George et al. (1976) and Ramsbottom et al. (1978) with a later update by Waters et al. (2011). For the Dinantian, these are in ascending order; the Courceyan (Tournaisian), the Chadian and Arundian (early Visean), the Holkerian (middle Visean) and the Asbian and Brigantian (late Visean). The lower two substages of the Namurian (Pendleian and Arnsbergian) correspond to the Serpukhovian ICS-Stage.

In this book, we use ICS-terminology and therefore mention the terms Dinantian, Namurian and Westphalian in parenthesis. For details of the Namurian development of the Netherlands, the reader is referred to Huis in 't Veld and Den Hartog Jager (2025, this volume). Because the lithostratigraphic nomenclature of the lower Carboniferous in the northern Dutch offshore is largely derived from the UK offshore, we here also refer to the regional UK Substages to describe detailed stratigraphic aspects of the Tournaisian and Visean succession.

Lithostratigraphy

In this section, some new lithostratigraphic units are introduced and some existing units are amended. As these units have not been ratified by the Stratigraphic Committee of Geological Survey of the Netherlands, and are thus still informal, they are used without capitals here (e.g. formation, not Formation).

Kortgene group – Boomvliet formation (proposed)

We propose the introduction of the Kortgene group comprising the Boomvliet formation for strata of Silurian or earliest Devonian age (Fig. 2.4). These are the oldest known strata encountered in the Netherlands (Houben & Vis, 2021), penetrated by wells Kortgene-1 and O18-01 in the southwest of the Netherlands (province of Zeeland) (Fig. 2.1) and have never been formally included in the lithostratigraphic nomenclature of the Netherlands. The names of the group and formation refer to the village and hamlet near the drilling site of well Kortgene-1.

The succession as encountered in the above mentioned wells consists of sub-horizontal dark grey shale to siltstone layers containing pyrite, alternating with intercalations of lighter grey, (very) fine grained, sandstone with poorly rounded grains (Gökdag et al., 1983; Krans, 1983). Mineralogically, the fine-grained sandstones consist of quartz (~60%) and feldspar (~9%), with a matrix consisting of clay, mica, rock fragments and clay fragments. Locally, dolomitization has occurred (Krans, 1983). The decimetre-scale sandstone layers typically have sharp basal contacts and the succession shows frequent evidence of flow (flute casts) and loading (Fig. 2.5a-d). The sediments in this succession are interpreted to have been deposited as turbidites sourced from a fine-grained source area, alternating with pelagic mud. The sediments show no signs of metamorphism but are affected by diagenesis, faulting and deformation. At least three fracture episodes have been identified based on observations in O18-01 (Swennen & Muchez, 1991). The fractures are filled up successively by ferroan dolomite, ankerite and quartz and dolomite and anhydrite satinspar (Fig. 2.5a,b; Gökdag et al., 1983). Miospores in the sedimentary rock are dark coloured and substantially affected by thermal alteration, oxic degradation and/or the imprint of diagenetic pyrite-growth structures (Houben & Vis, 2021).

The minimum age of this succession – late Silurian (Wenlock-Pridoli) to earliest Devonian (Lochkovian) – is determined based on the miospore association as recorded in core samples of Kortgene-1 (Houben & Vis, 2021). The deposits of the Kortgene group accumulated below wave-base in a marine foreland basin that started to develop over the northern part of present-day Belgium (Brabant Basin). In well Kortgene-1 the succession measures 210 m up to the bottom of the well. In Belgium, the turbiditic upper Silurian succession reaches a thickness of at least 1000 m (Verniers et al., 2001).

In UK offshore well 53/16-1, a similar succession as the one attributed here to the Boomvliet formation was encountered. The succession consists of a 223 m thick, steeply dipping grey turbiditic sandstone. The age for this succession was estimated to be pre-Carboniferous and probably latest Silurian based on the similarity with the (at that time) undated succession in O18-01 (Woodcock & Pharaoh, 1993; De Vos et al., 2010; Houben & Vis, 2021). The deposits of this newly proposed formation correspond to megasequence 3 of the Brabant Massif in Belgium (Verniers et al., 2002; Herbosch, 2021).

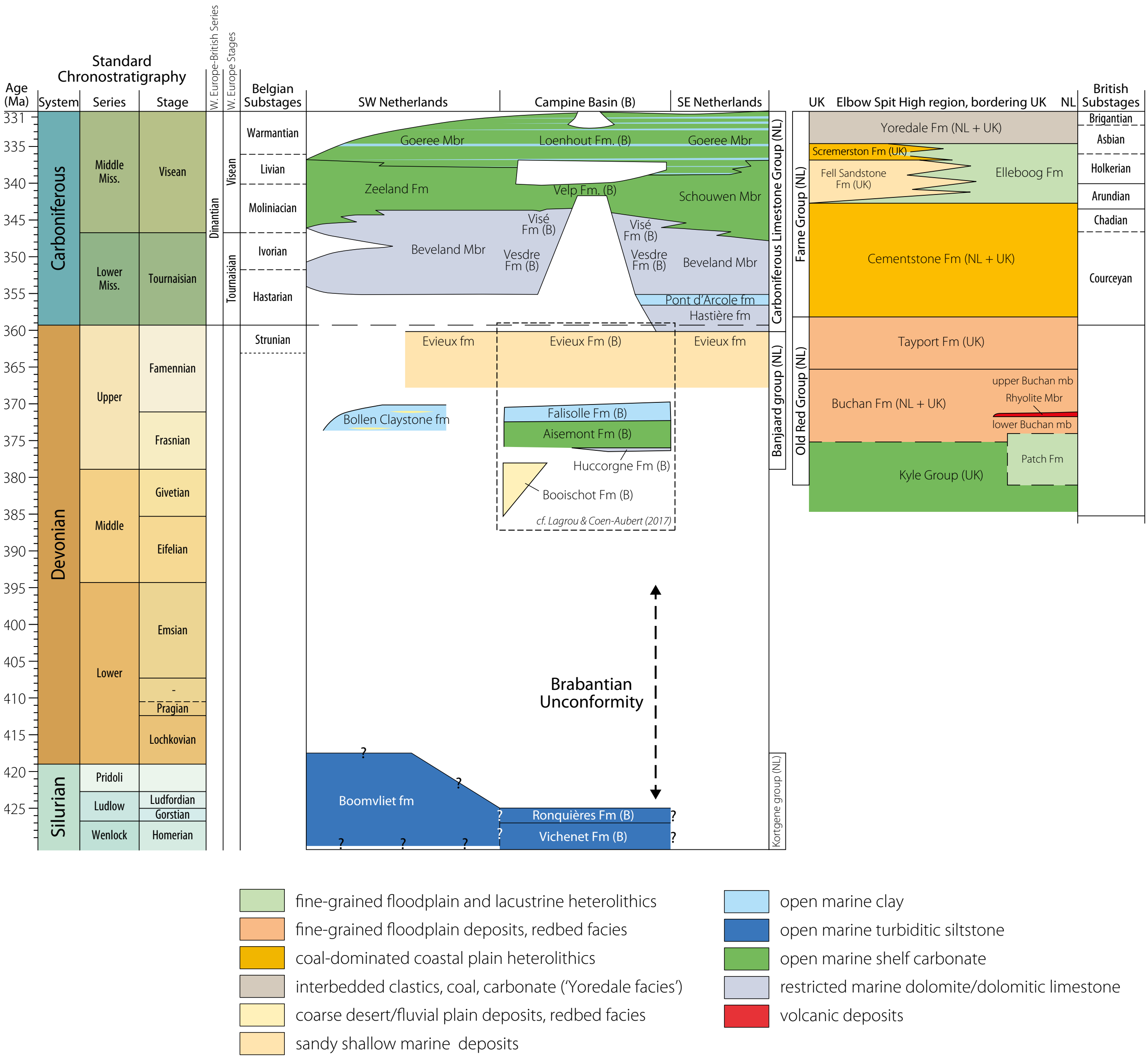


Figure 2.4. Chronostratigraphic (Wheeler) diagram for the upper Silurian to Middle Mississippian of the Netherlands. Belgian stratigraphy follows Verniers et al. (2001), Laenen (2003) and Lagrou & Coen-Aubert (2017); UK stratigraphy according to Cameron (1993b).

Banjaard group (emended)

The Banjaard group was introduced as an informal unit by Van Adrichem Boogaert & Kouwe (1994) for the upper Middle Devonian to lowermost Carboniferous, largely marine succession in the southern Netherlands. An overview of the Devonian lithological units in the Campine basin is provided by Dusar et al. (2015). Two informal formations were proposed: the Bollen Claystone formation and the overlying Bosscheveld formation. At the time of introduction only very little was known regarding the former formation, which was informally defined in Zeeland. The latter in essence represents the transition between Devonian siliciclastics and lower Carboniferous limestones. A recent re-evaluation of core material and new palynostratigraphic constraints now allow for a modification.

The Bosscheveld formation was completely cored in the Kastanjelaan-2 borehole in Maastricht and was extensively studied (Bless et al., 1981b). In addition, new boreholes drilled during the last decade near Venlo (CAL-GT-01 to CAL-GT-05) and in South Limburg close to the Belgian border (Terziet-1 (B62D1112) and -2 (B62D1113)), now provide a better appreciation of these Devonian-lowermost Carboniferous strata in the Netherlands. This leads us to abandon the Bosscheveld formation and to introduce (from old to young) the Evieux formation in the Banjaard group (Fig. 2.4) and the Hastière and the Pont d'Arcole formations in the Carboniferous Limestone Group. These modifications allow for a better synchronization with the stratigraphy of Belgium and Germany.

Bollen Claystone formation (emended)

The Bollen Claystone formation is encountered in the southwest of the Netherlands (Zeeland area, Fig. 2.4), where the formation was cored in two boreholes (Brouwershavense Gat-1 and S05-01). In these cored sections, it consists of dark shaly mudstones with some intercalated, thin, white to greenish-grey, fine-grained sandstones, which are interpreted as tempestites. The unit is in places slightly calcareous or dolomitic. Well log signatures of the unit appear fairly homogeneous and consequently, substantial limestone development is not foreseen. The Bollen Claystone was deposited on a low-energy shelf, shoreface to transitional-marine setting. Recently, miospores assignable to the genus *Diducites*, which roughly indicates a Frasnian or early Famennian age were identified in a core sample from well Brouwershavense Gat-1, at 2902.6 MD (S. Houben, pers. obs.). Based on this tentative age information, and the lithofacies character, the Bollen Claystone formation is likely to correlate to the Falisolle Formation of Lagrou & Coen-Aubert (2017), a unit that was previously referred to as the Lambermont Formation by Laenen (2003). This succession of green-grey shales becomes more sandy towards its top and spans the Frasnian-Famennian boundary (Lagrou & Coen-Aubert, 2017). In the Campine Basin, conspicuous reefal/biostromal limestones of the Aisemont Formation are encountered below the Falisolles Formation (Fig. 2.4). So far there is no indication for the presence of Frasnian limestone in the Netherlands (see also section Middle to Late Devonian), although Frasnian limestones are reported in the Visé area, just south of Maastricht (e.g. Poty & Delculée, 2011).

Evieux formation (proposed)

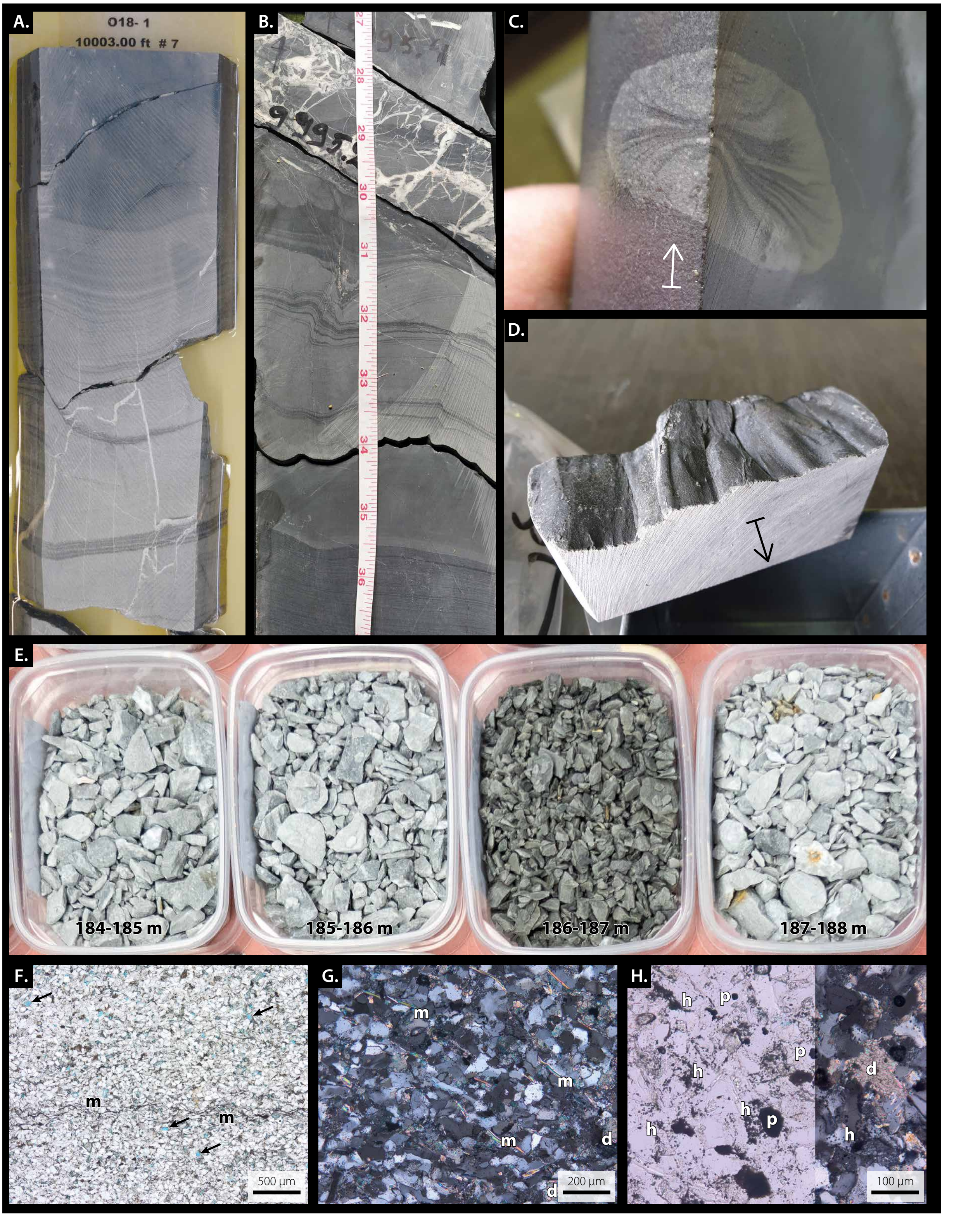
The Evieux formation is introduced for strata constituting alternations between red and grey-green shaly siltstones and strongly micaceous sandstones (Fig. 2.5e-h, 2.6a). The type-locality is well CAL-GT-01-S1 (2248-2708 MD = end of GR-log). The Evieux formation is derived from its namesake that is recognized in Belgium and Germany (Bultynck & Dejonghe, 2001; Laenen, 2003; Menning, 2018). Towards the top of the formation in well Kastanjelaan-2, carbonate-cemented sandstones occur, as well as mottled sediment, possibly indicating the development of soils (e.g. plant root horizons). The age of the unit is late to latest Famennian. However, the age of the base of the unit is not clearly defined (Fig. 2.4). In the Zeeland area (e.g. well Brouwershavense Gat-1) it is thought to pass downhole into the finer-grained marine siliciclastic sedimentary rocks of the Bollen Claystone formation. In South Limburg, in analogy to the Condroz Group of Belgium, a continuation to strata resembling the Montfort Formation (upper Famennian, dark-grey micaceous sandstones with abundant clayey intercalations, Bultynck & Dejonghe, 2001; Laenen, 2003) and the Esneux Formation (middle Famennian, green-coloured micaceous sand and siltstones, Bultynck & Dejonghe, 2001) is assumed to be present. The Evieux formation is also encountered in the Zeeland area (Fig. 2.4). Although a good understanding of the encountered lithofacies is lacking, similar strata appear present in wells that penetrate Devonian strata in the eastern onshore Netherlands, as far north as the Uithuizermeeden-2 well (Fig. 2.1)

The Evieux formation was deposited as part of a laterally extensive shallow marine system. This developed in the relatively narrow and shallow Rhenohercynian sea at the southern edge of the Old Red Continent. Sediments were continuously redistributed through strong along-shore currents, which are proposed to be related to a strong westward directed current within this confined basin (Paproth et al., 1986). The cyclic lithological nature was mitigated by small-scale eustatic base-level variations and, as a result, a whole spectrum of alluvial, estuarine, restricted marine (lagoonal) and shallow marine siliciclastic depositional settings may be encountered (Thorez et al., 2006).

By latest Famennian times, climatic aridification associated with the Hangenberg Crisis and a drop in sea-level led to subaerial exposure, as observed in the Kastanjelaan-2 borehole (483-468 MD). This sea-level drop may be associated with development of the Hangenberg Sandstone in Germany and Belgium (Denayer et al., 2021). This upper part of the Evieux formation in Kastanjelaan-2, correlates to the transitional strata referred to as the Dolhain, Comblain-au-Pont and Etroeungt formations in Belgium (Denayer et al., 2021).

Old Red Group

The Old Red Group comprises mainly fluvial and terrestrial sediments of Middle Devonian to early Carboniferous



(Tournaisian) age (TNO-GDN, 2022c). Sediments of this group were encountered in wells A17-01, E02-01 and E06-01 in the northern Dutch offshore (Fig. 2.1), but due to the low number of wells, the distribution of sediments of this group is not well known (Fig. 2.4). The sediments of this group are broadly equivalent to those of the Upper Old Red Group of the UK southern North Sea (Cameron, 1993b; Kearsey et al., 2015). The group has not been penetrated completely in one single well, but based upon a compilation of well data a thickness of up to 1500 m may be representative of the Mid North Sea High (Geluk et al., 2007). To the south, the group extends most likely under large parts of the Netherlands offshore and onshore, grading laterally into the marine Banjaard group.

Patch Formation

The Patch Formation was introduced by Van Adrichem Boogaert & Kouwe (1994) for a succession of grey and green-grey silty claystones, containing a few thin sandstone beds and some thin, brown, intercalated dolomite beds encountered in well A17-01 (Fig. 2.1). The depositional environment is interpreted to have been lacustrine or fluvial floodplain. Its thickness reaches nearly 300 m in well A17-01, but it is uncertain whether this formation can be recognized as a separate lithostratigraphic unit further away from well A17-01 (Fig. 2.4).

Biostratigraphic age determination is difficult for this formation, as is the case for the entire Old Red Group. A Givetian to Frasnian age is inferred (Van Adrichem Boogaert & Kouwe, 1994). The Patch Formation corresponds to the lower part of the Buchan Formation in the UK (e.g. Kearsey et al., 2015) and to the underlying limestones and mudstones of the Kyle Group in the UK (Fig. 2.4). Deposits of the Kyle Group have not been proven in wells in the Netherlands, but they have been interpreted to be present on seismic data near the Elbow Spit High (Ter Borgh et al., 2018a,b). The Kyle Group is characterized by white, buff, brownish grey, or pink limestones and thin grey shales, with an overlying, more variable unit of marine, lagoonal, and basin-marginal sabkha facies (Cameron, 1993a). The limestone contains tabulate and rugose corals, ostracods, brachiopods, bivalves, gastropods and crinoids, which demonstrate a shallow, warmwater marine environment (Marshall & Hewett, 2003). The Kyle Group is of Givetian to early Frasnian age (Cameron, 1993b).

Buchan Formation

The Buchan Formation (Fig. 2.4) comprises a succession of white to reddish sandstones, and red to red-brown and grey claystones and siltstones. The sediments are interpreted to have been deposited in a fluvial setting, where the sandstones represent stacked channel deposits. In well A17-01, nearly 140 m of felsic extrusive igneous rocks (rhyolite) occur within the formation (Rhyolite Member, Fig. 2.4). Based on the extrusive character, the radiometric age (341±30 Ma) is regarded to be unreliable (see Van Bergen et al., 2025, this volume). An older (Middle Devonian) rhyolite occurs in the Embla gas field (Norway, block 2/7; Marshall & Hewett, 2003).

In absence of biostratigraphic data from this formation in the Netherlands, a Frasnian to Famennian age is assumed. This matches with UK well 38/03-1 (Fig. 2.1), where palynological assemblages show discrete marine transgressive events of Frasnian age (Marshall & Hewett, 2003). The formation has been encountered with a thickness of over 600 m in the northwestern Dutch offshore and in the UK.

Tayport Formation

The Tayport Formation (Fig. 2.4) is composed of red to red-brown and grey claystones and siltstones, with well-developed sandstone intercalations (Fig. 2.6b.). The

← ***Figure 2.5.*** *Core photographs from selected boreholes. a) Boomvliet formation. Laminated fine-grained sandstone (grey) to siltstone (dark grey). Filled fractures are also visible, see Fig. 2.5b (O18-01, 10003 feet = ~3049 m). b) Boomvliet formation. Soft sediment deformation structures, flute casts and an intensely fractured interval resulting from three periods of fracturing according to Swennen and Muchez (1991) (O18-01, 9995.2 feet = ~3046.5 m, ruler in inches). c) Boomvliet formation. A peculiar ball-shaped sandstone in a siltstone matrix, interpreted to be the result of loading, i.e. soft sediment deformation (O18-01, 9999.4 feet = ~3048 m, finger for scale). d) Boomvliet formation. 3D view of flute casts or erosional scours on the sole of a bed, (O18-01, 9955 feet = ~3034 m). e) Cuttings from the Evieux formation illustrating the light grey, micaceous sandstone (Terziet-2 (B62D1113), 184-188 m). f) Evieux formation. Thin section of finely laminated fine-grained sandstone with laminae enriched in muscovite (m). Pores occur mostly as mouldic or oversized pores, which formed due to grain dissolution (arrows) (Terziet-2 (B62D1113), 187-188 m, 04x PPL). g) Evieux formation. Thin section of faintly laminated fine-grained litharenitic sandstone showing alignment of muscovite (m). The Authigenic Fe-dolomite crystals (d) in between the detrital quartz grains (white to black in XPL). Most of the blue colours are due to Fe-dolomite staining rather than (blue epoxy) pores (Terziet-2 (B62D1113), 151-152 m, 10x, XPL). h) Evieux formation. Thin section showing detailed image of the sandstone with pyrite (p), µm-scale microcrystals (h; possibly TiO_2). The sandstone is strongly cemented by quartz and authigenic dolomite (d), resulting in a very poor porosity (Terziet-2 (B62D1113), 149-150 m, 20x, PPL, XPL). Microphotographs (f-h) from Veeningen (2020).*

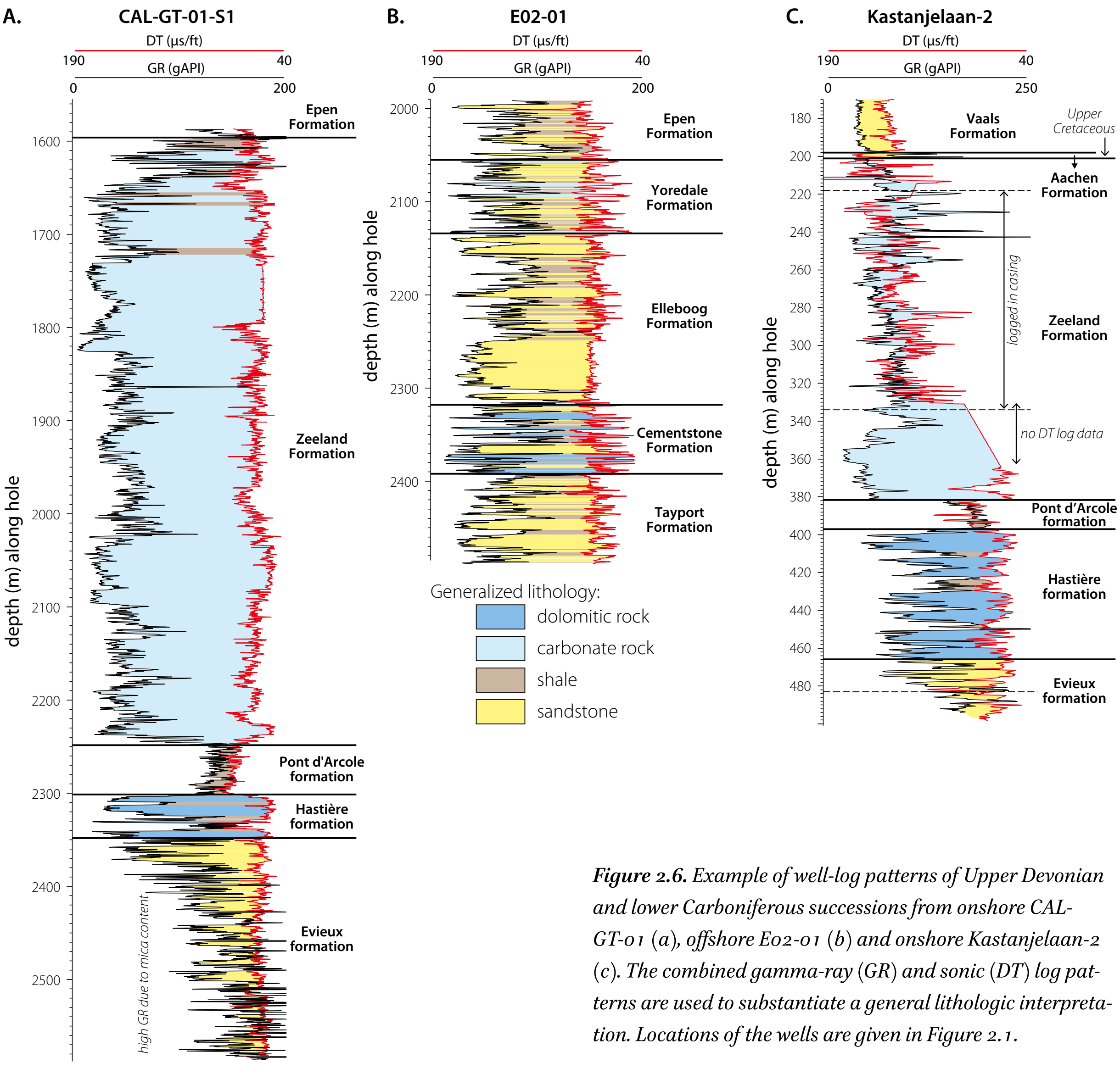


Figure 2.6. Example of well-log patterns of Upper Devonian and lower Carboniferous successions from onshore CAL-GT-01 (a), offshore E02-01 (b) and onshore Kastanjelaan-2 (c). The combined gamma-ray (GR) and sonic (DT) log patterns are used to substantiate a general lithologic interpretation. Locations of the wells are given in Figure 2.1.

presence of thin limestone beds has been inferred from wireline logs (Kearsey et al., 2015). The sandstones are thought to represent fluvial channel fills. The clay- and siltstones were deposited on a floodplain, at times well-drained, which caused a red colouration. This formation has been encountered in wells E02-01 (Fig. 2.6b) and E06-01. Based on the limited number of wells its total thickness is estimated to be at least 450 m in the Dutch sector, and 650 m in the UK sector (37/10-1). The top of the Tayport Formation is of uppermost Famennian age in well E06-01 (J.E.A. Marshall pers comm. in Kearsey et al., 2015). In UK-well 37/10-1 an earliest Tournaisian age has been assigned based on palynomorphs (McLean, 2012). The Tayport Formation correlates with the Kinnesswood Formation which is excellently exposed onshore Northumberland (UK). There, a remarkable hardpan calcrete horizon corresponds to the Devonian-Carboniferous boundary (Marshall et al., 2018), and possibly also the regressive phase of the Hangenberg Crisis (Kaiser et al., 2016). A similar phenomenon has not yet been observed in wireline log sections from offshore wells.

Farne Group

This group contains sedimentary rocks deposited in a paralic to shallow-marine environment. The variable lithology comprises claystone and sandstone, with minor coal seams and a variable amount of (dolomitic) limestone beds. The group contains three formations (Fig. 2.4), from bottom to top: the Cementstone, Elleboog and Yoredale (Cameron, 1993b; Van Adrichem Boogaert & Kouwe, 1994). The Cementstone and Yoredale formations are the equivalents of the formations defined in the UK, while the Elleboog Formation correlates with the Fell Sandstone and Scremerston formations in the UK. The group has not been penetrated entirely in one single well, but is assumed to reach a thickness between 1 and 2.5 km in the

UK sector of the North Sea (Cameron et al., 1992). Strong variations in thickness, of the order of hundreds of metres, occur over syndepositional fault zones south of the Mid North Sea High (Maynard & Dunay, 1999).

Cementstone Formation

The Cementstone Formation (Tournaisian to early Visean, Fig. 2.4) is composed of cyclically alternating carbonates, claystones and sandstones with minor coal seams (Fig. 2.6). The carbonates occur as limestones, dolomitic limestones and dolomites. The fine-grained sediments of this formation accumulated in desiccating, hypersaline lakes on an arid coastal plain (Leeder, 1992). The sandstone units represent stacked fluvial channels (Cameron, 1993b). The thickness of the formation in wells is up to 400 m (e.g. UK well 44/2-1; Cameron, 1993b).

Elleboog Formation

The Elleboog Formation (early to middle Visean, Fig. 2.4) comprises stacked fluvial sandstone bodies and claystones, with minor intercalations of coal and limestones, representing an overbank environment (Fig. 2.6b). To the west, the percentage of sandstones in the lower part increases, grading into the fluvial Fell Sandstone Formation of the UK parts of the southern North Sea (Cameron, 1993b). The sandstone is massive, pale-grey and white, fine to medium or coarse-grained and poorly sorted. It contains sporadic beds of grey, blocky to subfissile, non-calcareous mudstone up to 10 m thick and occasional beds of dark-grey, pyritic, carbonaceous mudstone.

Three sandstone units, each from 80 to over a 100 m thick, have been identified in the UK offshore blocks 43 and 44 (Cameron, 1993b). In a southward direction, the sandstones rapidly grade into mudstones with interbedded thin sandstones (e.g. E06-01). The sandstones are overlain by fine-grained deposits (UK Scremerston Formation, Kearsey et al., 2018) which comprise alternations of sandstone, siltstone, mudstone and coal, with occasional thin dolomite or limestone beds. These sediments accumulated in a range of delta-plain and overbank environments (Leeder et al., 1989, Cameron, 1993b). Their combined thickness is around 100 m.

Yoredale Formation

The Yoredale Formation (late Visean, locally earliest Serpukhovian, Fig. 2.4) comprises a cyclic alternation of marine limestones and mudstones, within a dominantly deltaic succession of claystones, sandstones and coal seams (Fig. 2.6b). The cyclicity within the limestone beds has been interpreted to reflect millennial-scale cycles, as a result of high-frequency arid-humid climatic fluctuations (Tucker et al., 2009). In the Netherlands, a thickness of 175 m was encountered in wells (e.g. A14-01, E02-01, E06-01), whereas in the UK sector, over 700 m were penetrated (41/24a-2; Cameron, 1993b).

Carboniferous Limestone Group

Sedimentary rocks interpreted as belonging to the Carboniferous Limestone Group occur in about 60 wells in the Netherlands, the majority of which are in South Limburg. At the time that the lithostratigraphic definitions for these lower Carboniferous sedimentary rocks were introduced (Van Adrichem Boogaert & Kouwe, 1994), several key wells had not been drilled or were still confidential. New data and new insights ask for an urgent revision of definitions of this group (see Mozafari et al., 2019). For the following description we rely heavily on new insights and interpretations presented by Mozafari et al. (2019). We also propose the introduction of the well-known Belgian Hastière and Pont d'Arcole formations as new units in this group in the Netherlands (Fig. 2.4).

The group generally comprises light-grey, brown and black carbonates, with variable intercalations of bedded chert and claystone, all deposited in a marine carbonate platform and slope environment. Karstic phenomena have been identified in several wells (Mozafari et al., 2019).

Hastière formation (proposed)

Well Kastanjelaan-2 was fundamental to the description of the Devonian-Carboniferous transition in South Limburg (Fig. 2.6c). The latter is thought to be located in a confined basin, the Visé-Puth Basin (Bless, 1989; Poty & Delculée, 2011) and served as the informal type-section for the Bosscheveld formation (Fig. 2.4). The fully cored 120 m thick succession below the limestones of the Zeeland Formation (500-395 MD) was completely assigned to the Bosscheveld formation by Van Adrichem Boogaert & Kouwe (1994). This cored succession comprises from bottom to top, (I) green-red coloured micaceous sandstones (500-483 MD, now included in the Evieux formation); (II) a fairly heterolithic succession of micaceous sandstones and siltstones with possible traces of soil development (mottling due to roots, concretions) and plant-remains (483-468 MD, now considered the uppermost part/top of the Evieux formation); (III) a dolomitized crinoidal bioclastic limestone (468-395 MD); and (IV) a dark grey fossiliferous shale with a distinct elevated gamma-ray signature (395-380 m). The extinction of the miospore *Retispora lepidophyta* is recorded at 447.5 MD (see Bless et al., 1981b and S. Houben pers. obs.), and serves as a good approximation for the position of the Devonian-Carboniferous boundary (Prestianni et al., 2016; Marshall et al., 2018).

This clear lithological fragmentation was also noted by Laenen (2003), who appointed the Kastanjelaan-2 well as the type-section for the Belgian Bosscheveld Formation in the Campine Basin. He placed the base of this formation

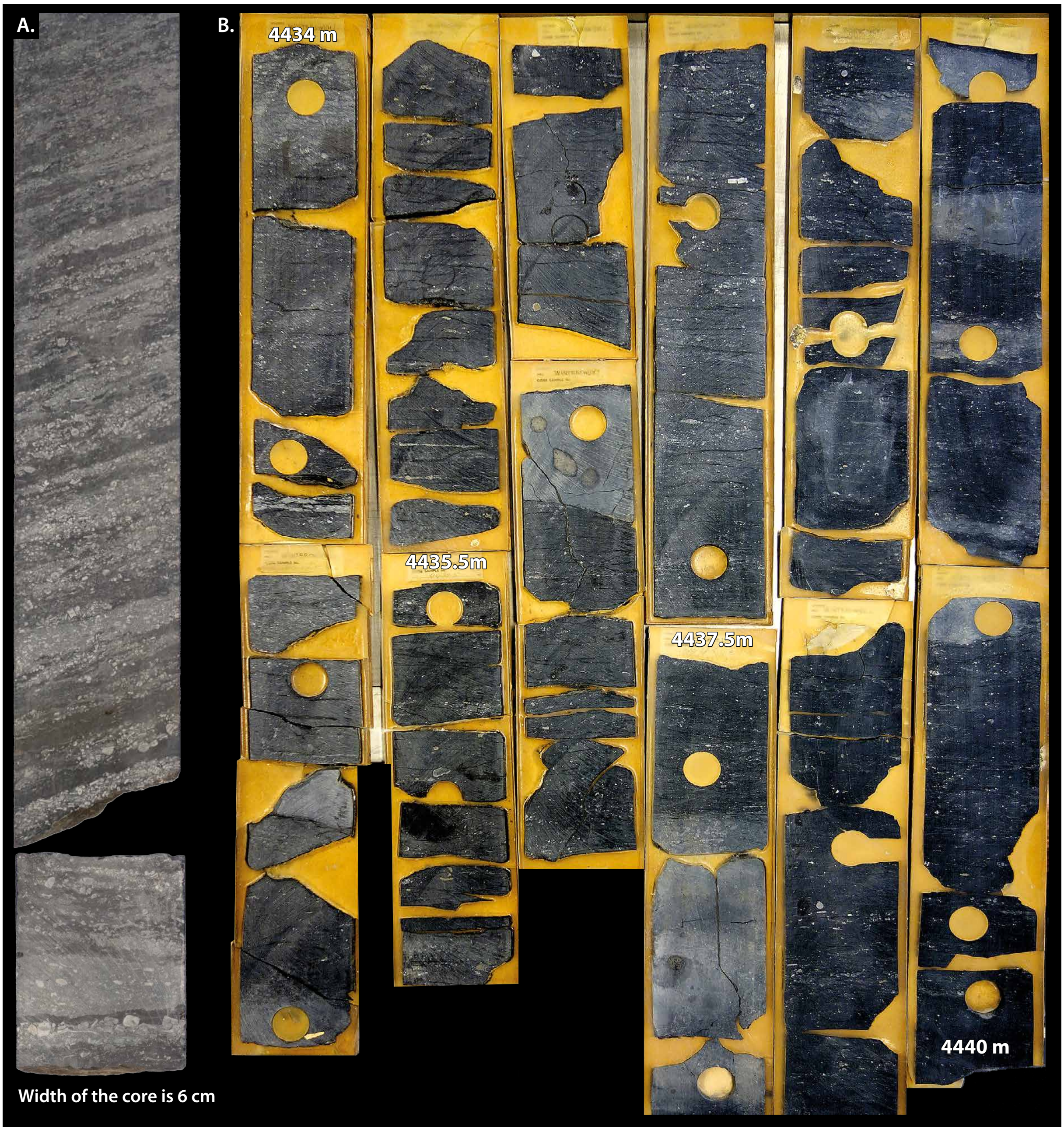


Figure 2.7. *a) Hastière formation. Dolomitized crinoidal bioclastic limestone (Kastanjelaan-2; photo taken from interval 427-435 m). b) Pont d'Arcole formation. Dark grey shale with crinoid fragments (Winterswijk-1, 4434-4440 m).*

at the top of the red-green coloured micaceous sandstones (i.e. at 483 MD) and subsequently included both the heterolithic as well as the crinoidal bioclastic limestone interval in the Bosscheveld Formation, which is then described as a transitional unit between the siliciclastics of the Famennian Condroz Group and the limestone-dominated strata of the Lower Carboniferous Limestone Group.

With the drilling of the California geothermal wells (CAL-GT-01 to CAL-GT-05) near Venlo, it became clear that the up-section partitioning of micaceous sandstones, bioclastic limestones and the elevated gamma-ray shales was clearly traceable on wireline-log profiles (Fig. 2.6a). This leads us to refer to the dolomitized bioclastic limestone (468-397 MD) that contains the Devonian-Carboniferous boundary as the Hastière formation (in analogy to the Ardennes where it is also applied, Denayer et al., 2021 and references therein). The Hastière formation in the Netherlands is now defined as a succession of predominantly claystone and (dolomitized) limestone, rich in crinoids, brachiopods and corals (Fig. 2.7a). The top forms at the transition to the fossiliferous dark shales, with their characteristic high-intensity gamma-ray signature. This interval is very well discernible on gamma-ray profiles. The type-section is well Kastanjelaan-2 (468-397 MD, Fig.

2.6c). The dark-grey fossiliferous shales were assigned to the Pont d'Arcole Formation by Laenen (2003). We concur with this interpretation and now propose to include this formation within the Carboniferous Limestone Group (Fig. 2.4).

Pont d'Arcole formation (proposed)

The Pont d'Arcole Formation is historically known from Belgium, where it is exposed in the type section of the Hastière Limestone, northwest of Hastière-Lavaux (Groessens, 1975; Paproth et al., 1983). In Belgium, parastratotypes are represented by wells 125E298 (Vieux-Leuze, SW Belgium) and 097E817 (Bossuit, SW Belgium) (Laenen, 2003). A third parastratotype was selected in well Kastanjelaan-2 in Maastricht (Laenen, 2003). This well has been studied in detail but the presence of an equivalent of the Pont d'Arcole Formation had not been interpreted previously, as it was estimated to be premature to establish a formal lithostratigraphic subdivision (Bless et al., 1981b).

In well Kastanjelaan-2 the proposed Pont d'Arcole formation occurs between 397 and 382 MD (Fig. 2.6c). It consists of dark grey fossiliferous (calcareous) mudstone, fissile pyritic crinoidal shale with finely dispersed plant debris and bioclastic wackestone intercalations and a generally increasing carbonate content towards the top. A crinoid-bryozoan-coral assemblage has been observed as well as the (Tournaisian) index brachiopod *Spiriferina peracuta* (Bless et al., 1981b; Poty et al., 2001; Laenen, 2003). Despite the dark colour, the organic carbon content is only 0.3-0.5 % (Bless et al., 1980). The Pont d'Arcole formation can be clearly recognized in the CAL-GT wells (Fig. 2.6a) and has been cored in well Winterswijk-1 (Fig. 2.7b). The depositional environment was open marine, below storm wave base. The base of the Pont d'Arcole formation correlates with the base of the Lower Alum Shale (Kahlenberg Formation in the Rhenish Massif area) which lies in a deeper water setting (Kulm facies) and marks a sudden flooding (Poty et al., 2001; Herbig et al., 2018). These lithofacies are fairly consistent over broad distances from southwest England to the Aachen region (Laenen, 2003; Amler & Herbig, 2006; Pharaoh et al., 2021).

The formation rests conformably on dolomitized bioclastic limestone of the Hastière formation in Kastanjelaan-2 (Fig. 2.6c). The upper boundary is placed at the base of the first massive carbonates of the Zeeland Formation, although Mozafari et al. (2019) stated that the upper boundary is gradual. The thickness is 20-30 m. This unit is widely recognized in the Dinant Synclinorium in Belgium and in Germany (North Rhine-Westphalia). It is not encountered in the southwestern Netherlands and in the Campine Basin in Belgium, where marine flooding of the London-Brabant Massif was delayed. The age is early Tournaisian.

Zeeland Formation

The Zeeland Formation (Fig. 2.4), was defined by Cameron (Cameron, 1993b) and Van Adrichem Boogaert and Kouwe (1994) for shallow water carbonate facies. However, this unit has also been used in literature for sediments in wells located in a proximal basinal environment, generating some confusion. Here we follow Mozafari et al. (2019) and apply the Zeeland Formation to generic carbonate units of lower Carboniferous age, comprising both platform and proximal basin environments. This unit is present in the Dutch onshore and offshore subsurface along the northeastern flank of the London-Brabant Massif (Fig. 2.8). The term London-Brabant Massif, sometimes also called London-Brabant Platform, refers to the late Paleozoic and younger elevated morphological feature that includes the former Anglo-Brabant Deformation Belt and the Midlands Microcraton (e.g. Pharaoh, 2018).

The base of this unit varies geographically from sharp to transitional. The contact with the underlying Pont d'Arcole formation may be transitional showing a gradual upward increase of the carbonate content, for example, in the wells Kastanjelaan-2 and Winterswijk-1. Often the Tournaisian part of this unit is missing and the Visean carbonates are in unconformable sharp contact with Devonian sedimentary rocks (e.g. Uithuizermeeden-2, Luttelgeest-1, Campine Basin and southwest Netherlands). This formation is found along the northeastern flank of the London-Brabant Massif in the Netherlands, Belgium and offshore UK, as well as in Germany close to Aachen. The same unit has been found in the isolated platforms in the central and northern Netherlands, for example, in wells Uithuizermeeden-2 and Luttelgeest-1 (Gutteridge et al., 2025). In Belgium and Germany, this unit is usually subdivided into several formations. In the Netherlands, the Zeeland Formation is subdivided into the Beveland, Schouwen and Goeree members (Fig. 2.4). The correlative basinal unit of the Zeeland shallow water carbonates is represented by the condensed Kulm facies encountered in the Rhenish Massif in Germany (Bless et al., 1976; Korn, 2010; Herbig et al., 2018; Arndt, 2021) and by proximal basin deposits found in Kastanjelaan-2, Heugem-1 and Winterswijk-1. In the subsurface of the North Sea, a transition towards the clastic units in the British offshore as found in well E06-01 is expected (Mozafari et al., 2019).

The top of the Zeeland Formation is placed at the transition from carbonate sediments to the clastics of the Epen Formation of the Limburg Group. The boundary either is represented by an unconformity with associated karst (e.g. O18-01), or, by a gradual transition to carbonaceous shales of the Geverik Member of the Epen Formation (Geverik-1 and Winterswijk-1 wells). The contact is sometimes difficult to define using gamma-ray logs, due to the similar values recorded in the upper portion of the Zeeland and

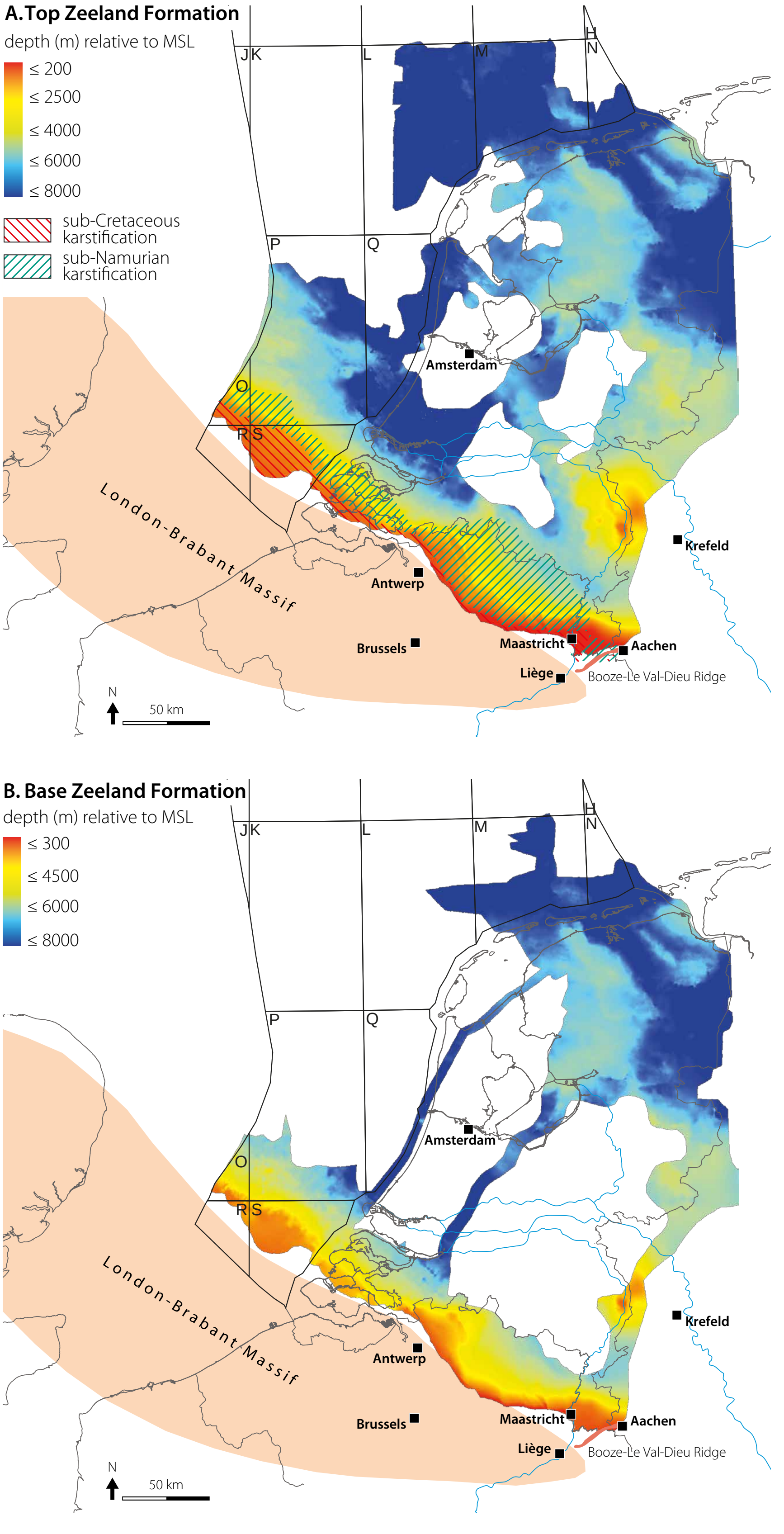


Figure 2.8. *Depth maps of the top (a) and base (b) of the lower Carboniferous (Dinantian) Zeeland Formation. White spots represent data-poor areas that have not been studied. The karstification at the base of the overlying clastic succession (sub-Namurian karstification) and the karstification related to the large hiatus between the Carboniferous succession and the overlying Upper Cretaceous succession (sub-Cretaceous karstification) are indicated in map (a). Depth data from Ten Veen et al. (2019); karstification from Mozafari et al. (2019).*

the Epen formations (Mozafari et al., 2019). In some cases, where younger unconformities erode deeply into the Carboniferous, the Zeeland Formation is also truncated (e.g. Saalian unconformity in well CAL-GT-02 or the Cretaceous Unconformity on the flank of the London-Brabant Massif). The definition of the upper boundary is also complicated by the presence of karst collapse and doline fills at the Visean to Serpukhovian contact (Dinantian to early Namurian) along the northern margin of the London-Brabant Massif (Dreesen et al., 1985, 1987; Mozafari et al., 2019; Gutteridge et al., 2025).

The age of the Zeeland Formation is Tournaisian to Visean. Depending on the location, the base of the formation is diachronous. On the isolated platforms of the central and northern Netherlands, the Tournaisian is absent and the Zeeland Formation is only Visean in age.

Beveland Member

This member consists of medium grey or brown grey to dark brown or brown-black, coarse-crystalline dolomites, often containing black organic intergrain residues. The dolomites are generally of secondary origin. In places minor grey to dark grey limestone intercalations occur, as well as minor quantities of dark brown to blackish siltstone and shaly claystone. In addition, dark beds of silicified dolomite are occasionally present (TNO-GDN, 2022a).

The onset and end of carbonate deposition appear to have taken place at different times in different parts of the basin. Both base and top of the Beveland Member are likely to be diachronous (Fig. 2.4), but their age is a matter of debate and needs to be re-examined (Mozafari et al., 2019). The top of the Beveland Member may cover a range from Tournaisian to middle Visean (late Hastarian to early Livian) in the north of the Netherlands (e.g. Uithuizermeeden-2).

The type section is defined in well S02-02 (2836-2604 MD). Van Adrichem Boogaert and Kouwe (1994) considered well Kastanjelaan-2 as an additional section (382-338 MD) but Mozafari et al. (2019) consider this well not to be representative of the Beveland Member, as it is very shaly and represents a proximal basin environment, deposited within the confined Visé-Puth Basin (Poty & Delculée, 2011).

Schouwen Member

This member comprises a succession of light to dark grey, dark- to yellowish-brown and brownish black, and light yellowish brown to dusty yellow brown limestones. The dense limestones are micritic, biosparitic to biomicritic, locally oolitic and fossiliferous. Reefal structures are known along structural highs in the Belgian Campine Basin and are also assumed to be present offshore the southwest Netherlands (e.g. Dusar et al., 2015). In places coarsely crystalline calcite veins occur. Intergranular bituminous, organic material is often present. Locally, the limestones are dolomitized, especially near fault zones, or are silicified in the leached zone in areas where later erosion truncated the Zeeland Formation (TNO-GDN, 2022d).

The age of this member is most likely early to middle late Visean (Fig. 2.4). Its boundaries are not defined, leaving uncertainty about the usage of this lithostratigraphic unit (Mozafari et al., 2019). The type section for the Schouwen Member is defined in well S02-02 (2604-2123 MD). Well Heugem-1 was considered as a parastratotype (502.7-114 MD) by Van Adrichem Boogaert & Kouwe (1994), but Mozafari et al. (2019) interpreted a rather different sedimentary environment in this well, possibly related to its position in the Visé-Puth Basin.

Goeree Member

This member comprises a succession of grey to dark grey and black limestones, thin- to thick-bedded and often partly silicified. The limestone beds often grade into calcareous and/or silicified black shales or black bedded chert towards the top. Very thin beds of tuffaceous rock occur, predominantly in the upper part of the member (TNO-GDN, 2022b).

The lower boundary is usually pinpointed by a gamma-ray increase compared to the underlying unit, due to an increase of shale and/or volcanic tuff. These gamma-ray spikes may be related to karst fills in wells along the northern margin of the London-Brabant Massif (Mozafari et al., 2019). The age of this member is mostly late Visean (Livian to Warnantian, Fig. 2.4) This member corresponds to the Goeree Formation of northern Belgium (Laenen, 2003).

Depositional history

Cambrian to Early Devonian

In the Netherlands, wells do not reach strata older than Silurian. Therefore, we outline here the older stratigraphy and geologic development based on information from surrounding countries (also see Fig. 2.9).

Deposition during the early Carboniferous, Devonian and older systems took place in a wide variety of sedimentary basins. While reconstruction of pre-Devonian sedimentary basins remains speculative for most of the area, they are better documented for the Late Devonian and early Carboniferous. Tectonostratigraphic comparisons and compilations, based on an integration of outcrop, well and geophysical data in the areas between the British Isles and the Danish-Polish Caledonides (e.g. Verniers et al., 2002; Winchester et al., 2002; Pharaoh et al., 2006; De Vos et al., 2010) serve as the main data source for interpre-

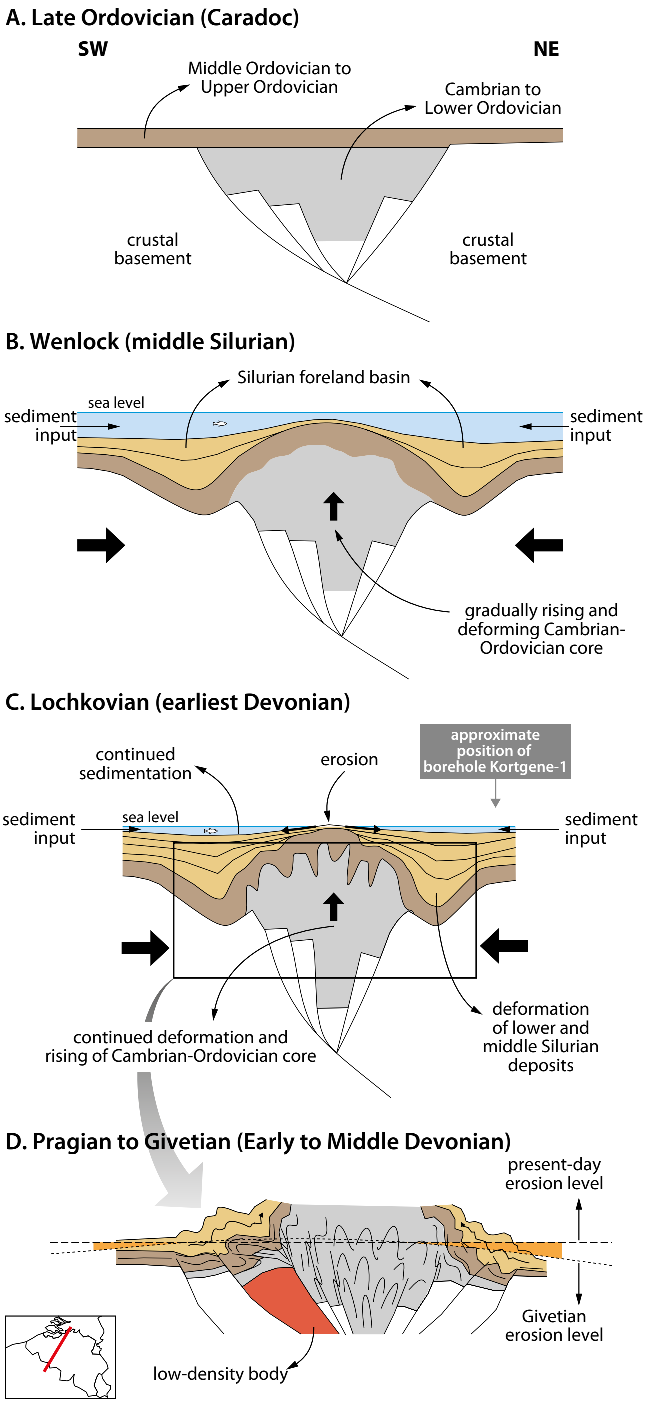


Figure 2.9. Schematic model of Debacker et al. (2005) for a four time-steps Late Ordovician to Middle Devonian evolution of the Brabant Massif, integrating the stratigraphic, sedimentological, metamorphic, geophysical, geochronological and structural data (not to scale). a) Late Ordovician (Caradoc): a basin, possibly a failed rift, filled with Cambrian to Lower Ordovician sediments overlain by more shallowly deposited Middle Ordovician sediments (cf. Verniers et al., 2002). b) Wenlock (middle Silurian): ongoing inversion of the basin, first affecting the core of the Brabant Massif. The shortening causes a rise of the core associated with a steepening of the deposits. Silurian foreland basin development is probably the result of the weight of the rising, deforming core. The sediment input is from the south in the southern foreland basin and from the north in the northern foreland basin. c) Lochkovian (earliest Devonian): continued shortening results in a continued rising and steepening of the core of the massif, which emerges, resulting in erosion of the condensed lowermost Devonian and upper Silurian deposits. The northern and southern foreland basins are now completely separated. Continued shortening leads to a spreading of deformation from the core towards the rims. Note the approximate position of borehole Kortgene-1. d) Pragian to Givetian (Early to Middle Devonian): during the Early Devonian continued shortening causes a further spreading of the deformation from the core towards the rims and results in the complete inversion of the Silurian to Early Devonian foreland basins during the late Pragian/Emsian, possibly Eifelian. At the same time, also the Cambrian core and the Ordovician sequences experienced further, final shortening. Note the lateral changes in thickness of the Silurian deposits (modified from Sintubin & Everaerts, 2002; Debacker et al., 2005).

tation of the pre-Devonian paleogeographic evolution of eastern Avalonia. Important updates and new insights on the tectonic driving mechanisms, largely based on these compilations, were published recently (e.g. Woodcock et al., 2007; Sintubin et al., 2009; Linnemann et al., 2012; Debacker et al., 2015; Smit et al., 2016, 2018; Herbosch et al., 2020).

Verniers et al. (2002) prepared a compilation of all available data on the "European Caledonides", stretching from the British Isles to the Danish-Polish Caledonides. Mostly based on observations in the Brabant Massif (Belgian part of Anglo-Brabant Deformation Belt) and surrounding areas, these authors distinguished three early Paleozoic 'megasequences' (not to be confused with sequence stratigraphic megasequences).

The Cambrian and lowermost Ordovician deposits of **megasequence 1** in the Brabant Massif are both marine and terrigenous, derived from Gondwana, as witnessed by acritarchs and other fossils (Cocks & Fortey, 1982; Cocks et al., 1997; Servais & Fatka, 1997). Their large sedimentary thickness may be related to deposition in a failed rift (e.g. Verniers et al., 2002). In Germany north of the Rhenish Massif, Lower Ordovician coarse-grained clastic schist debris with a pre-Variscan tectonic imprint was found in Middle Devonian nearshore conglomerates of the Viersen-1001 well (Ahrendt et al., 2001). This suggests that an eastern outlier of Brabant Massif-like pre-Devonian rocks may be present in the Krefeld High (Geluk et al., 2007; De Vos et al., 2010).

The core of the Ebbe Anticline in the northern Rhenish Massif contains a sedimentary record which started during the Middle Ordovician. It consists of shales and thin layers of siltstone and fine-grained sandstone followed by shales and greywackes. There are Ordovician rocks of similar composition in four isolated outcrops in the Remscheid Anticline about 50 km to the west (e.g. Verniers et al., 2002 and references therein). Within the Lippstadt High, about 60 km to the northeast of the Ebbe Anticline, graptolite-bearing pre-Devonian rocks have been identified in the Soest-Erwitte well (Clausen & Leuteritz, 1982). However, it remains unclear whether the stratigraphic gap between Ordovician 'deeper' water deposits and late Silurian shallow water shelf deposits such as the one found at the Stavelot-Venn Anticline, is sedimentary or tectonically induced, (De Vos et al., 2010). An hiatus separating uppermost Ordovician from Pridolian (upper Silurian) shales, marlstones and carbonates is known from the Ebbe, Remscheid and Müsen anticlines in Germany.

After separation from Gondwana, sediments of **megasequence 2** were deposited mainly on a shallow shelf. At the end of this sequence, in the Upper Ordovician (upper Caradoc and Ashgill), the faunal affinity of the eastern parts of Avalonia resembled that of Baltica (Giese et al., 1994, 1997; Vecoli & Samuelsson, 2001). Despite the inferred basin setting during megasequence 2, two episodes of wide-spread slumping have been documented (see Debacker, 2012 and references therein). The first took place during the Middle Ordovician (latest Dapingian-earliest Darriwilian, i.e. in the basal sequences of megasequence 2), and points to a south-dipping paleoslope (Debacker, 2012 and references therein). The second episode occurred during the Late Ordovician (late Sandbian-earliest Katian), shortly after onset of turbiditic sedimentation, and was associated locally with large-scale slumping (Debacker et al., 2001) (Fig. 2.9a). The inferred paleoslope is very poorly constrained, but is estimated to be ~north-dipping (Debacker et al., 2001; Debacker, 2012). This second episode points to basin instability around the time of Avalonia-Baltica docking (e.g. Verniers et al., 2002). Possibly, also the Asquempont Detachment System, an originally NE-dipping low-angle normal fault system that can be traced throughout the Brabant Massif (see Herbosch et al., 2008; Herbosch & Debacker, 2018, and original references therein), formed during this period. Roughly at the same time, or later during the Silurian, also the Ardennes were uplifted and eroded (Verniers et al., 2002). However, this so-called Ardennian phase, later overprinted by the Variscan deformation event (Carboniferous), cannot be tied to Avalonia-Baltica collision or another compressional plate-tectonic event, and may equally reflect Silurian extension along the northern margin of the Rheic Ocean (e.g. Sintubin et al., 2009).

North of the uplifted and eroding Ardennes Inliers, a rapidly subsiding foreland basin developed during the Silurian in the area of the future Brabant Massif (Fig. 2.9b) and its western prolongation in East Anglia (Woodcock & Pharaoh, 1993; Pharaoh et al., 1995). This occurred during **megasequence 3** of Verniers et al. (2002). Based on observations in the southern part of the Brabant Massif, it appears that during most of the Silurian turbidity currents supplied sediments from southern source areas (Verniers & Grootel, 1991; Debacker, 2001; Debacker et al., 2014 and references therein). Despite their fine-grained nature these Silurian turbidite deposits should not be regarded as distal, but rather as being sourced from a fine-grained source area, e.g. the eroding Ardennes Inliers (Debacker et al., 2014). During the Brabantian deformation phase (i.e. Acadian; e.g. Woodcock et al., 2007; Sintubin et al., 2009) inversion of this Brabant Basin formed the Anglo-Brabant Deformation Belt in the late Silurian to Early Devonian. The Cambrian core of this belt – the former possible failed rift – was progressively compressed and deformed (Fig. 2.9c), while sedimentation continued along the edges in the Silurian-earliest Devonian foreland basin (Debacker, 2001; Verniers et al., 2002; Debacker et al., 2005). Based on increased subsidence and foreland basin development, the Brabantian deformation phase already commenced around early Wenlock times (Debacker et al., 2005).

It is thought that deformation of the Cambrian-Ordovician core of the Brabant Basin was responsible for Silurian-earliest Devonian foreland basin development along its edges (Fig. 2.9c), eventually resulting in basin-wide deformation and cleavage development around 400 Ma (Sintubin & Everaerts, 2002; Verniers et al., 2002; Debacker et al., 2005). Foreland basin subsidence, degree of metamorphism and degree of Brabantian deformation decrease outwards from the Cambrian core. This is supported by observations in southwest Flanders and the Condroz Inlier, situated on the northern edge of the Midlands Microcraton, where diagenetic, undeformed Silurian to lowermost Devonian deposits occur (Verniers & Grootel, 1991; Verniers et al., 2002). Given the inferred more or less symmetrical geometry of the Brabant Basin inversion (e.g. Sintubin & Everaerts, 2002; Verniers et al., 2002; Debacker et al., 2005), a similar situation is expected in the Netherlands: non-metamorphic, very poorly deformed Silurian-lowermost Devonian deposits, except in the direct vicinity of inverted (Acadian-Brabantian) pre-existing basins. In

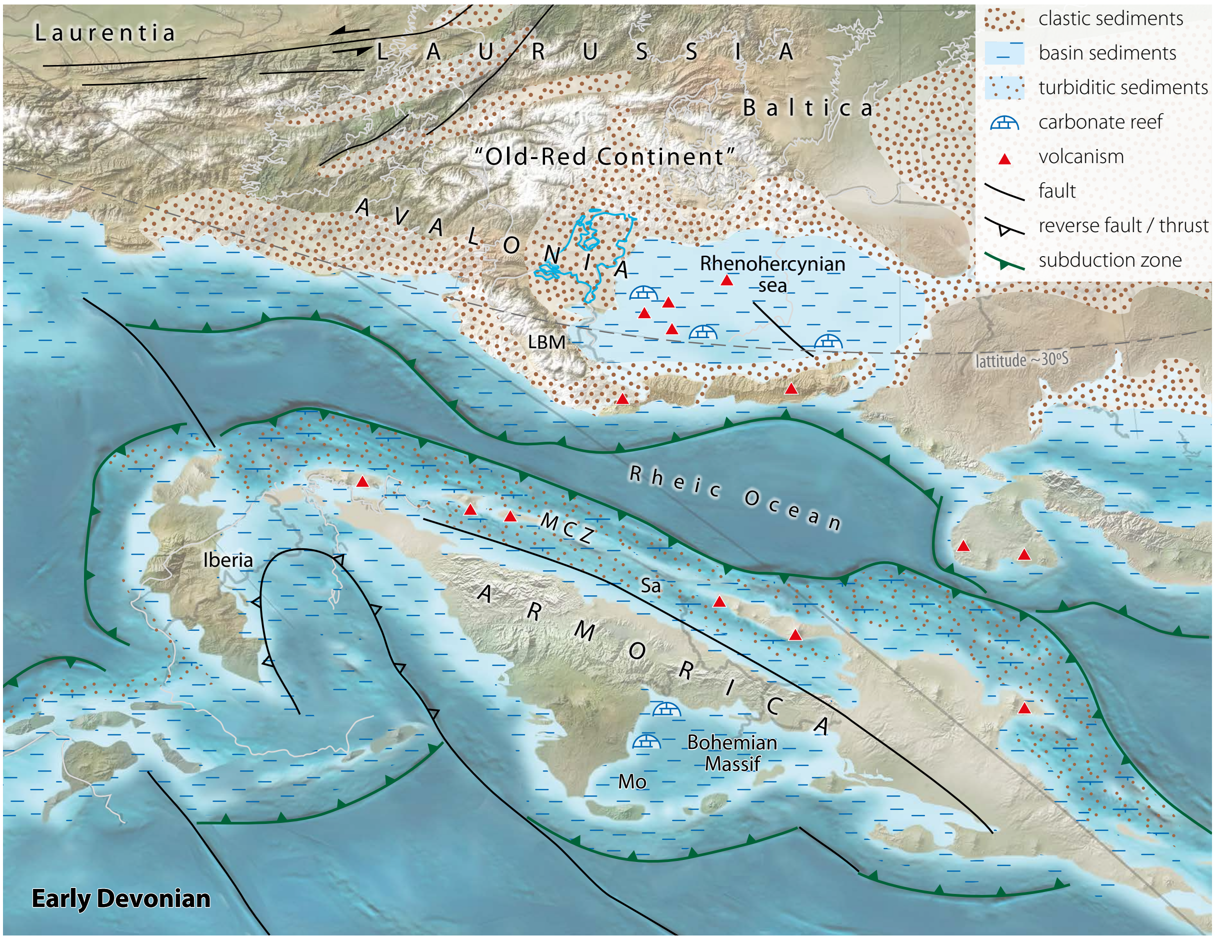


Figure 2.10. *Early Devonian paleogeographic reconstruction of the area surrounding the Netherlands (indicated with blue outline). LBM = London-Brabant Massif; MCZ = Mid-German Crystalline Zone; Mo = Moldanubian; Sa = Saxothuringian. Modified from Meschede and Warr (2019).*

fact, the diagenetically altered and (very) poorly deformed Silurian-lowermost Devonian strata of the Kortgene group encountered in wells Kortgene-1 and O18-01 (Houben & Vis, 2021) correspond to megasequence 3. In the palynological associations from well Kortgene-1 reworked Cambro-Ordovician elements were recognized, supporting the nearby orogen and development of a foreland basin over the southern Netherlands (Houben & Vis, 2021). The top of the Silurian-lowermost Devonian sequence is truncated by the Brabantian Unconformity (see Textbox 2).

Two wells with radiometric ages in the North Sea area are worth mentioning here. Well A17-01, situated on the Mid North Sea High, has a thick Upper Devonian succession overlying a granitic batholith. Seismic data show onlapping stratigraphic geometries on both sides of the granite, implying that it was in place before deposition of the sediments (see Fig. 1.6 in De Jager et al. (2025), this volume). The granite was dated using $^{40}Ar/^{39}Ar$ at 346±7 Ma (Frost et al., 1981), but is believed to have been overprinted during the early Carboniferous (Pharaoh et al., 1995), possibly as a result of burial during Late Devonian-early Carboniferous extension (Smit et al., 2018). Indeed, an unpublished U-Pb isotope age determination on zircons suggests that granite emplacement took place at 410±7 Ma, thus approximating the Silurian-Devonian boundary (Axel Gerdes, pers. com., 2022). This is further explained in Van Bergen et al. (2025, this volume). In northeast England, the Cheviot Granite was emplaced in Early Devonian time, as were many other granites in northern England (De Vos et al., 2010). The German offshore well Q1 terminated at a depth of 3804 m in muscovite-biotite schist that is probably a high-grade augengneiss showing retrograde metamorphism in greenschist facies (Frost et al., 1981). The absolute age of 415±8 Ma measured on muscovite with K-Ar is probably the age of greenschist-facies metamorphism (De Vos et al., 2010), which matches the Wenlock to Emsian age range of the progressive Brabantian deformation phase (Debacker et al., 2005).

Middle to Late Devonian

By the start of the Middle Devonian, the Brabantian deformation phase (Fig. 2.9d), which resulted in an uplifted

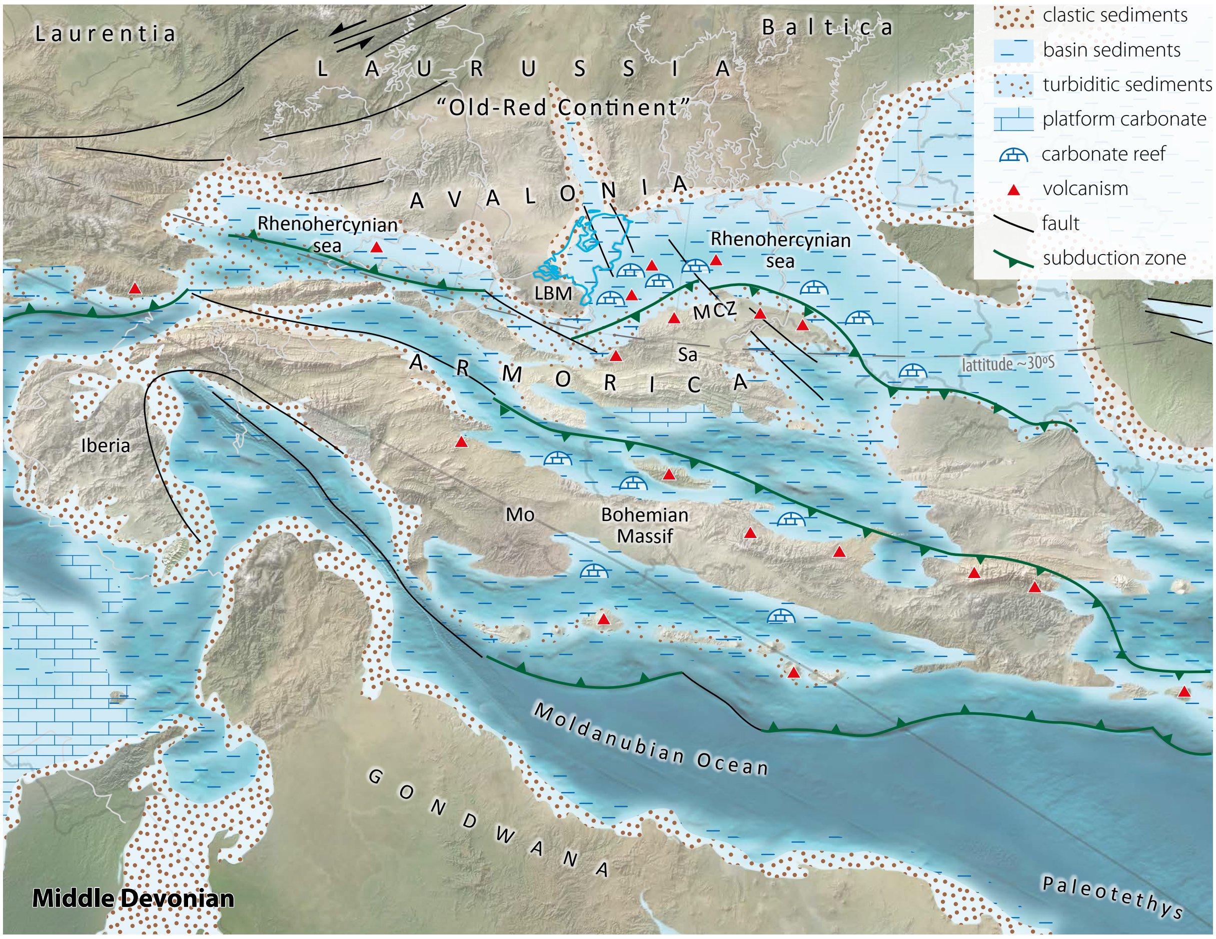


Figure 2.11. *Middle Devonian paleogeographic reconstruction of the area surrounding the Netherlands (indicated with blue outline). LBM = London-Brabant Massif; MCZ = Mid-German Crystalline Zone; Mo = Moldanubian; Sa = Saxothuringian. Modified from Meschede and Warr (2019).*

Brabant Massif had ended (Debacker et al., 2005). Over the course of the remaining Devonian, a complex of Gondwana-derived microcontinents such as Armorica moved northwards and docked onto Laurussia, on the southern side of Avalonia. This essentially contributed to the Variscan Orogeny in NW Europe (Figs 2.10, 2.11).

The Rhenohercynian sea opened up by Early Devonian times along the southern margin of eastern Avalonia (Figs 2.10, 2.11). The then eroding Anglo-Brabant Deformation Belt constituted a paleohigh (Fig. 2.9d), surrounded by different sub-basins of the Rhenohercynian sea (Fig. 2.11). The shape and orientation of these sub-basins were likely controlled by the pre-existing (Acadian/Brabantian and older) basement fabric. In the Campine Basin in Belgium, the wells Heibaart and Booischot (Fig. 2.1), provide insight in Middle and Upper Devonian stratigraphy (Lagrou & Coen-Aubert, 2017). In the Booischot borehole, a 400 m thick continental, mainly conglomeratic succession of the Booischot Formation overlies Caledonian (Silurian) basement rocks. The top of the Booischot Formation is dated as upper Givetian to upper Frasnian (Streel, 1965; Streel & Loboziak, 1987). This conglomeratic unit developed within a locally subsiding half-graben, close to an active fault, that coincides with the erosional boundary of the Brabant Massif. In the Heibaart borehole, located on a domal structure with stratigraphically delayed sediment cover and lower subsidence, the Caledonian basement is overlain by a few metres of the middle Frasnian Huccorgne Formation (Lagrou & Coen-Aubert, 2017). The Booischot and Huccorgne formations are overlain by the Upper Devonian marine Aisemont and Falisolle formations which are capped by Famennian sandstones of the Evieux Formation (Fig. 2.4). The Frasnian Aisemont Formation consists of nodular limestones containing corals and stromatoporoids, calcareous shales and oolitic ironstones. These reflect the transgression of the Frasnian sea over the edges of the Brabant Massif. The Falisolle Formation comprises a dominantly shaly lithology. It spans the Frasnian-Famennian boundary (Kimpe et al., 1978; Vanguestaine et al., 1983; Lagrou & Coen-Aubert, 2017). These formations have not been unequivocally recorded in the onshore Netherlands but based on seismic information, may be present.

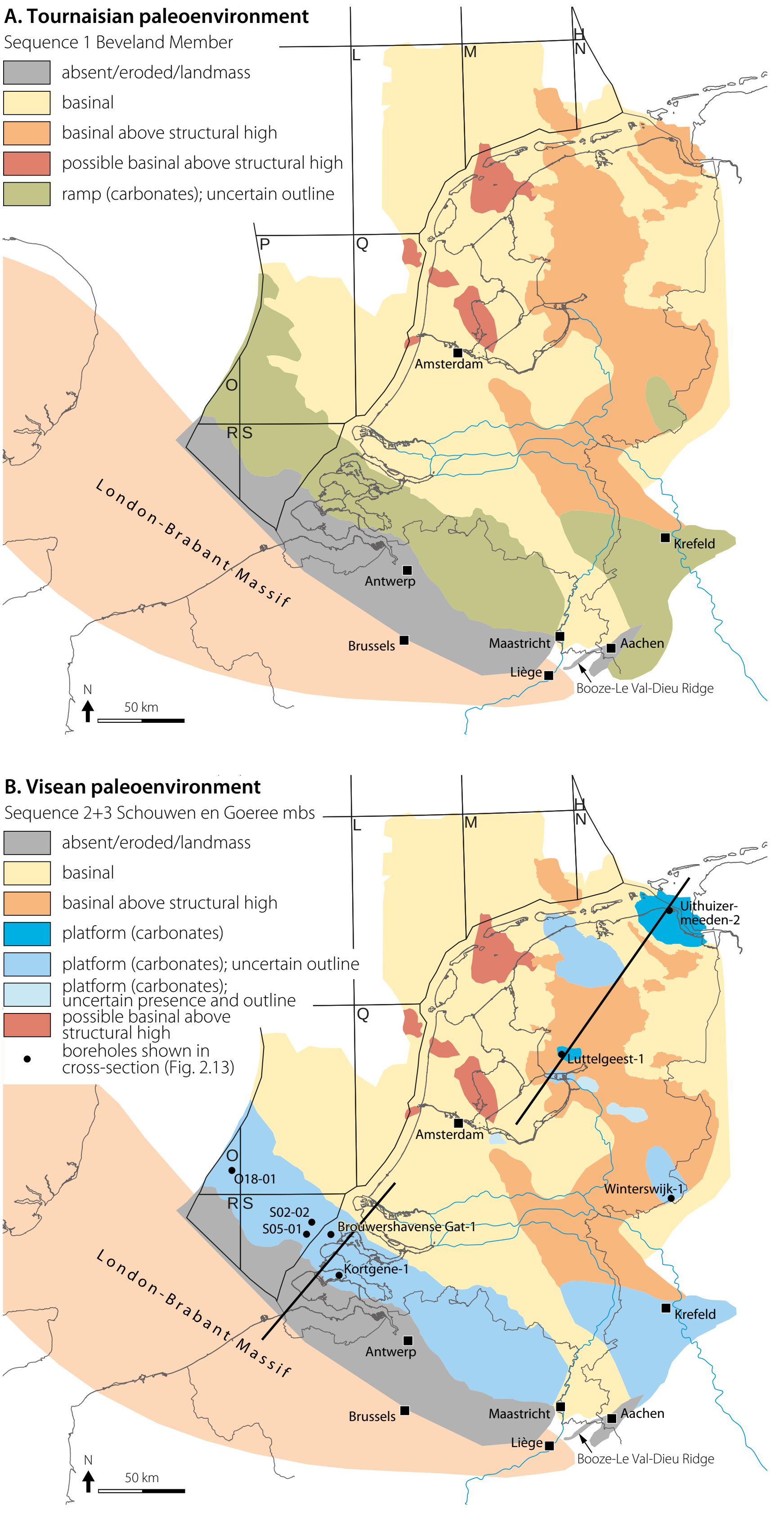


Figure 2.12. Maps depicting lower Carboniferous paleoenvironments of the Tournaisian (a) and Visean (b) during deposition of sediments representing the Zeeland Formation. Data from Mozafari et al. (2019). Lines and boreholes in (b) are used for the schematic cross section in Fig. 2.13.

In the southwest Netherlands (Zeeland area), the Middle to Upper Devonian Banjaard group rests unconformably on Silurian to lowermost Devonian turbiditic deposits of the Kortgene group (Fig. 2.4). The Bollen Claystone formation consists of shaly sediments with cyclic sandstone developments (tempestites). On wireline logs of wells in the Zeeland area, no prominent limestone can be recognized within the Banjaard group. It seems plausible that these marine strata were deposited as a shoreface complex on the northern margin of the Brabant Massif and that the reef ramp of the Brabant Massif was not located in the Zeeland area. Frasnian limestone accumulation may be expected elsewhere in the Netherlands, as illustrated by the Visé Block in Belgium (south and west of Maastricht). There, Frasnian limestones occur as strongly brecciated and karstified mogote (tower) karst. The towers are surrounded by solution collapse breccias and much younger Dinantian carbonates (Kimpe et al., 1978). Subvertical contacts between the karstified Frasnian limestones and onlapping Visean limestones are well-exposed near Visé (Poty, 1982). These Frasnian limestones are exposed at the surface because of the development of the Booze-Le Val Dieu Ridge (Ancion et al., 1943). These authors describe it as a ~5 km wide southwest to northeast trending anticlinal ridge between Visé and Aubel. Arndt et al. (2020) extend the structure towards Cottessen and Terziet in South Limburg, and further to Aachen. On this ridge, the Carboniferous Limestone Group is missing and upper Carboniferous (Namurian) or younger strata immediately overly upper Famennian strata. By Tournaisian times, the ridge formed a high separating the Visé sedimentation area (or Visé-Puth Basin in the Maastricht area) and the Namur sedimentation area to the south (Poty & Delculée, 2011).

The conglomerate in the Viersen-1001 well (see section on Cambrian to Early Devonian) on the Krefeld High stratigraphically more or less corresponds to the Booischot Formation in Belgium (Lagrou & Coen-Aubert, 2017). It is similarly covered by a Frasnian succession of reef limestones, dolomites and claystones (Birenheide, 1998). These are also encountered in well Münsterland-1 (Fig. 2.1; De Vos et al., 2010) and are regionally referred to as Massenkalk. In the east of the Netherlands (Achterhoek area), in well Winterswijk-1, 360 m of white-grey coloured sandstone is present in the lowermost part (TNO-NITG, 1998). The Frasnian reef limestones from well Münsterland-1, 25 km to the east, are absent here, leaving the Devonian stratigraphy of well Winterswijk-1 poorly understood.

A substantial Givetian to Frasnian limestone succession is suspected to be present at deeper levels below Mississippian carbonate build-ups in the Netherlands, as suggested by seismic data. More specifically, Devonian reefal build-ups are interpreted to be developed in the Groningen area and on the Luttelgeest platform (Van Hulten & Poty, 2009; Herber & De Jager, 2010; Van Hulten, 2012).

Devonian carbonates around the Mid North Sea High are referred to in UK lithostratigraphy as the Kyle Group (Cameron, 1993b; Marshall & Hewett, 2003; Kearsey et al., 2015), an interval characterized by limestone units separated by shale and sabkha facies. Two limestone units are seen in UK well 30/16- 5. The limestone is dominantly bioclastic, containing tabulate and rugose corals, ostracods, brachiopods, bivalves, gastropods and crinoids (Marshall & Hewett, 2003). The Kyle Group has been penetrated on the Auk Ridge and on the Dogger Granite High where it generates a distinct seismic reflector that can be correlated over large areas south and east of the Mid North Sea High (Arsenikos et al., 2015; Monaghan et al., 2017). This character is especially visible on relatively new long-offset seismic surveys (Ter Borgh et al., 2018b) and suggests that the Kyle limestones may be present in the Dutch northern offshore as well, despite the absence of well penetrations. In well A17-01, the carbonates were not encountered, and younger, clastic, Devonian rocks overlie granitic basement. These are assigned to the Patch Formation of Adrichem Boogaert & Kouwe (1994). In the German Q1 well, thin beds of marine Middle Devonian limestones are intercalated in the continental Old Red Sandstone (Best et al., 1983). The Kyle seismic facies can be observed only where the Devonian is not deeply buried below Mesozoic basins ((Dutch) Central Graben, Step Graben), but its distribution is thought to be widespread.

Overlying either the limestone of the UK Kyle Group or granitic basement, a thick succession of Middle Devonian to lower Carboniferous mainly fluvial, terrestrial sedimentary rocks is encountered in the Netherlands offshore, referred to as the Old Red Group (Fig. 2.4). The sediments were deposited in a massive southward prograding fluviodeltaic system, sourced from the Laurussian or Old Red Continent. Arid climate conditions due to a paleoposition in the southern hemisphere subtropical belt, caused extensive oxygenation and accumulation of red-beds. The Old Red strata have not been penetrated completely in one single well, but based on a compilation of well data thicknesses may locally be up to 1.5 km on the Mid North Sea High area (Geluk et al., 2007). To the south, the group extends most likely under large parts of the Netherlands offshore and onshore, grading laterally into the marine Banjaard group. Late Devonian transgressions were quite substantial given the intercalation of marine limestones of the Kyle Group with the Old Red fluviodeltaic system.

The nature of the lateral transition from the Old Red Group to the Banjaard group remains poorly understood. During the Famennian, the present-day onshore Netherlands was located within the Rhenohercynian sea. In this marine basin, shallow siliciclastic sediments accumulated,

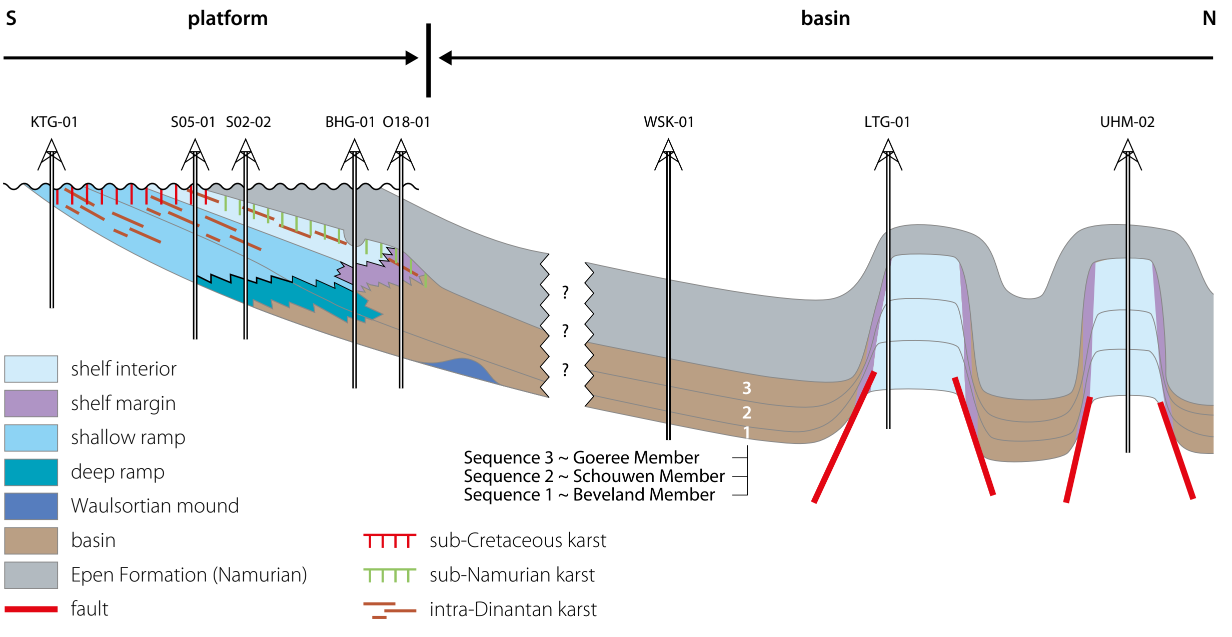


Figure 2.13. *Composite schematic north-south cross-section along projected wells (for approximate location see Fig. 2.12b), showing the sequence geometry and evolution of the attached and isolated early Carboniferous (Dinantian) carbonate systems of the Netherlands. Also note the three phases of karstification. Modified from Pickard & Gutteridge (1997). KTG-01 = Kortgene-1; BHG-01 = Brouwershavense Gat-1; WSK-01 = Winterswijk-1; LTG-01 = Luttelgeest-1; UHM-02 = Uihuizermeeden-2.*

which are thought to have been dispersed by longshore wind-driven currents (Paproth et al., 1986). The sediment is considered to be sourced from eastern source areas (Fig. 1 in Paproth et al., 1986), but Kołtonik et al. (2018) identified a Baltic provenance and also, a source from the nearby Brabant Massif cannot be excluded. These micaceous shallow marine sandstones were found in the Californië geothermal wells near Venlo and in recently drilled boreholes for the Einstein Telescope in South Limburg (Terziet-1 and -2 (B62D1112 & B62D1113) and Cottessen (B62D1181)). In wells Winterswijk-1, Uithuizermeeden-2 and Luttelgeest-1 similar strata were encountered, although lithofacies information from these wells is scarce. In well Kastanjelaan-2, near Maastricht, only the uppermost part of these Famennian sandstones was recovered, representing a regressive facies characterized by soil-development and plant-structures, possibly related to the Hangenberg regression (Kaiser et al., 2016). The nature of the transition to underlying older Devonian strata such as the carbonate-bearing Frasnian is unknown.

In wells Brouwershavense Gat-1 and S05-01, in the southwest of the Netherlands, a remarkable ~30 m thick transitional interval below the lower Carboniferous Zeeland Formation is present. This interval is characterized by a variable GR-pattern. The sedimentary rocks were cored in well Brouwershavense Gat-1 and yield an early late Famennian age (Bless, 1978). Consequently, we interpret this thin succession as corresponding to the Evieux formation. In South Limburg and wells in the eastern Netherlands, the Evieux formation transitions stepwise into the bioclastic limestone of the Hastière formation and an overlying dark grey marine shale (Pont d'Arcole formation). The latter two formations are absent in the southwest of the Netherlands, and in the Campine Basin (Belgium). In other wells in the southwest (Kortgene-1, O18-01 and S02-02), the Bollen Claystone formation is immediately overlain by the Zeeland Formation. This suggests a locally variable unconformable contact between the Middle to Upper Devonian and the lower Carboniferous successions along the London-Brabant Massif.

Early Carboniferous

During the Late Devonian to early Carboniferous extensional tectonics reactivated the pre-existing ('Caledonian') basement fabric and these reactivated basement structures controlled Late Devonian and younger sedimentation. Rifting and basement fabric reactivation took place in response to NE-SW/NNE-SSW extension that episodically occurred throughout the Late Devonian and early Carboniferous (Smit et al., 2016, 2018; Mozafari et al., 2019). In the present-day northern offshore, a large-scale coastal to fluviodeltaic system with a highly cyclic character developed at this time, first with the paralic marine Cementstone facies. This was followed during the early to middle Visean by stacked fluvial sandstone bodies and claystones, with minor intercalations of coal and limestones, that represent an overbank environment (Elleboog Formation). Towards the west, the amount of sandstone in the lower part of the formation increases and grades into the extensive fluvial sandstones of the Fell Sandstone Formation in the UK (Cameron, 1993b). Grain size decreases towards the top of the formation, the cause of which remains to be identified. The finer grained unit corresponds to the Scremerston Formation in the UK (Leeder et al., 1989, Cameron, 1993b). The shale-rich succession of the Cleveland Group in the UK (Visean to Serpukhovian) may represent the basinal environment between the fluviodeltaic system of the Farne Group and the carbonate system of the Zeeland Formation (see discussion in Kearsey et al., 2018). The well-known cyclic deposits of the Late Mississippian (late Visean) Yoredale delta occur as a lateral equivalent to the Cleveland Group; both are overlain by the coarse-grained sandstones of the Millstone Grit Formation (Kearsey et al., 2018).

The shallow marine areas around the Brabant Massif and the isolated highs in the onshore Netherlands remained outside of the influence of the fluviodeltaic system of the Farne Group and were the locus of a shallow-marine carbonate shelf environment during the Tournaisian and Visean (Fig. 2.12). The dark grey fossiliferous (calcareous) mudstone of the Pont d'Arcole formation (Fig. 2.7b) marks a major flooding (TST base, (Poty et al., 2001; Poty, 2016; Becker et al., 2021). The dark grey to black mudstones reflect an anoxic shelf setting. In a more distal setting in the Rhenohercynian sea to the east (Fig. 2.11), the base of the Pont d'Arcole formation correlates with the base of the Lower Alum Shale (Kahlenberg Formation in the Rhenish Massif) in a deeper water setting (Kulm facies). In the Netherlands, the Pont d'Arcole formation has been identified as far north as well Luttelgeest-1 and as far east as well Winterswijk-1.

The Pont d'Arcole formation is overlain by the Zeeland Formation, starting at the first massive carbonate beds. In the southeast of the Netherlands, the preservation of the lower Carboniferous succession is fragmentary and incomplete. In well Kastanjelaan-2, the transition from the Pont d'Arcole formation to the overlying carbonates of the Zeeland Formation appears to be a continuous record. Further east in South Limburg, in the area straddling the border with Belgium, the Devonian is overlain by middle Carboniferous (Namurian) sedimentary rocks. As recently demonstrated by the boreholes in Terziet (B62D1112 and B62D1113) Cottessen (B62D1181), the Tournaisian-Visean succession (Dinantian) is absent altogether, which is linked to their respective position on the Booze-Le Val Dieu Ridge (Ancion et al., 1943; Figs 2.8 and 2.9). In the Campine Basin, the Pont d'Arcole formation is absent and the base of the Tournaisian carbonates is assumed to become progressively younger (Laenen, 2003). In the southwest Netherlands and offshore towards the UK, upper Tournaisian carbonates are the oldest Carboniferous rocks present (Mozafari et al., 2019).

Recent studies of Tournaisian to Visean carbonates (Reijmer et al., 2017; Mozafari et al., 2019) have resulted in a schematic sequence stratigraphic geometry of the carbonate system (Fig. 2.13), built on previous work by Pickard & Gutteridge (1997). The correlative basinal unit of the Zeeland Formation shallow water carbonates is represented by the condensed, mostly starved siliciclastic Kulm facies encountered in Germany (Bless et al., 1976; Korn, 2010; Herbig et al., 2018; Arndt, 2021) and by proximal basin deposits found in wells Kastanjelaan-2, Heugem-1 and Winterswijk-1. In the North Sea, a transition towards marine predominantly clastic units (Farne Group) as found in well E06-01 is expected (Mozafari et al., 2019).

The carbonate system consists of three regional sequences (Fig. 2.13; Reijmer et al., 2017; Mozafari et al., 2019) that correspond to the Beveland, Schouwen and Goeree members. Within these sequences, ten depositional cycles have been recognized, which were also identified and correlated with basinal successions in the southeast and central Netherlands and with isolated carbonate platforms in the northern Netherlands (Mozafari et al., 2019). Based on biostratigraphic data, these depositional cycles are grouped into three lower order sequences (third order) of late Tournaisian to earliest Visean, early Visean and late Visean age.

Along the northern margin of the London-Brabant Massif carbonate ramps developed during late Tournaisian to mid Visean times (Fig. 2.13). Mozafari et al. (2019) proposed that deposition took place on a shallow ramp that includes cyclic peritidal carbonates sheltered behind carbonate shoals and algal reef mounds as observed on the Heibaart and Poederlee domes (e.g. Bless et al., 1981a; Muchez et al., 1987, 1990). Bioturbated packstones and wackestones were deposited in mid-ramp settings passing into a deep-water distal ramp depositional setting with inferred Waulsortian carbonate mud mounds (Fig. 2.13). The carbonate ramps sloped relatively gently at 8-9°. During the late Visean, these ramps developed into flat-topped carbonate platforms with more steeply-dipping margins at the uppermost slope (up to 30°) which passed downdip into basinal argillaceous carbonates (Kombrink et al., 2010; Ten Veen et al., 2019).

In the northern Netherlands, a number of isolated carbonate platforms (Uithuizermeeden and Luttelgeest, Fig. 2.12b) were initiated at various times during the late Tournaisian to early Visean over structural highs in an area of overall basinal sedimentation (Fig. 2.12b). These isolated platforms had relatively high depositional slopes of up to 12°, sometimes even locally 30°. Most of the platforms were built upon pre-existing tilted Devonian fault blocks, on which Bahama-type shallow platform carbonates were deposited (Mozafari et al., 2019).

An important aspect of the carbonate succession, and especially interesting for the exploration for geothermal and hydrocarbon energy sources (Mijnlieff et al., 2025, this volume), are karstification episodes. Three phases (Fig. 2.13) were identified based on core and seismic data (Mozafari et al., 2019): (1) karstification which is more or less contemporaneous with Tournaisian or Visean carbonate accumulation (intra-Dinantian karst); (2) karstification at the base of the overlying clastic succession (sub-Namurian karst) (e.g. Dusar & Lagrou, 2008); and (3) karstification related to the large hiatus between the Carboniferous succession and the overlying Upper Cretaceous succession (sub-Cretaceous karst). The vertical extent of the latter karst system reaches some 170 m beneath the Base Cretaceous Unconformity (Mozafari et al., 2019). Some wells in South Limburg (e.g. Kastanjelaan-2 and Heugem-1 and `s-Gravenvoeren in Belgium; Fig. 2.1) demonstrate complete dissolution of the upper part of the Dinantian carbonate matrix, resulting in porosities of up to 50% (Bless et al., 1981b; Batten et al., 1987; Dusar & Hogenhuis, 1997; Dusar, 2001). A well-developed paleosol including a kaolinite layer is found in association with the hiatus. The hiatus was caused by the progressing Variscan uplift in the south and the consequent development of a large foreland basin north of the emerging mountains. Early Namurian sea-level rise caused the foreland basin to be initially filled with basinal shales and turbidites of the Geverik Member. Increasing siliciclastic influx from the Variscan mountains in the south and the Fennoscandian Shield in the north resulted in a large-scale progradation, depositing a thick succession of deltaic sediments during the late Carboniferous (Westphalian). This is discussed in detail in the next chapter (Huis in 't Veld & Den Hartog Jager, 2025, this volume).

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Digital map data

Spatial data of figures in this chapter for use in geographical information systems can be downloaded here: https://doi.org/10.5117/aup.28162985.

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