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A comparative analysis of some Late Carboniferous basins of Variscan Europe

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Abstract – Vegetation diversity and pattern changes, and their relation to tectono-sedimentary histories are compared between selected Euramerican Late Palaeozoic coalfields, to understand better the controls on the dynamics of the Pennsylvanian terrestrial ecosystems and to demonstrate the problems with comparing data from various basins. The analysis is based on data from the following basins of different geotectonic and palaeogeographical positions: the cratonic Pennines Basin, the foreland South Wales and Upper Silesia basins, and the fault-related Intra Sudetic and Central and Western Bohemia basins. The analysis indicates that complex factors are responsible for changes in plant diversity and vegetation patterns. These are related to climate, tectonics, preservation potential, sampling biases and the current state of revision of the flora in each basin. Plant diversity patterns in the basins differ because of local controls and/or the character and detail of the available data. Maximum diversity varies among the basins within the Langsettian and Duckmantian substages. Two apparent step-like drops in diversity were detected within coal-bearing strata of most basins: at the Duckmantian/Bolsovian boundary and at the Bolsovian/Asturian boundary. Further and more prominent falls are related to transitions from coal-bearing to non-coal-bearing (mostly red bed) strata or vice versa during Stephanian times. Interpretation of climatic signals recorded in the sedimentary successions indicates that Westphalian and middle Stephanian times were wet intervals, whereas early and late Stephanian times were drier.

Keywords: Pennsylvanian, palaeobotany, biodiversity, Variscan Foreland.

1. Introduction

The Pennsylvanian Subperiod was one of the most important times of coal formation in Earth history. Extensive lowland areas of palaeotropical Euramerica and China were covered by extensive wetland forests dominated by arborescent lycopsids and sometimes marattialean tree ferns. For a combination of biological and taphonomic reasons, these forests resulted in the build-up of thick peat deposits which have since been converted into coal (Cleal & Thomas, 2005), hence these forests are often referred to as the coal forests. Our understanding of this general character of Pennsylvanian times has recently been improved by detailed palynological and floristic studies, which have revealed the existence of various environmentally or climatically controlled floral biomes, as well as biotic changes and extinctions through the time interval (e.g. Phillips, Peppers & DiMichele, 1985; DiMichele & Phillips, 1994; DiMichele, Pfefferkorn & Phillips, 1996; DiMichele, Pfefferkorn & Gastaldo, 2001; Falcon-Lang, 2003). Thus, we now know that the Euramerican coal forests were established to their maximum extent by the end of Early Pennsylvanian times, and began to disintegrate towards the end of Middle Pennsylvanian times, when most tree lycopsids were replaced by tree ferns. In Euramerica, the remains of coal forests had all but disappeared by the end of Late Pennsylvanian times, although extensive areas of forest persisted in China through into Permian times. This disintegration of the Euramerican coal forests was accompanied by global climatic warming (e.g. Cleal & Thomas, 2005; Gastaldo, DiMichele & Pfefferkorn, 1996) and in low palaeolatitudes by drying recorded by alternating coal-bearing and coal-barren red bed strata (e.g. Besly, 1987, 1988).

Mechanisms responsible for these changes in the Pennsylvanian terrestrial ecosystems are frequently discussed (e.g. Phillips, Peppers & DiMichele, 1985; Rowley *et al.* 1985; Besly, 1987; Parrish, 1998; Cleal & Thomas, 1999; Scotese, Boucot & McKerrow, 1999; Falcon-Lang, 2004; Opluštil, 2004). However, the importance of individual controls on the biotas still remains rather speculative, due to the general absence of broad comparative analyses of the tectono-sedimentary, climatic and fossil records between different basins. For example, Phillips, Peppers & DiMichele (1985) demonstrated an apparent change from arborescent-lycopsid to arborescent-fern vegetation around the Middle/Upper Pennsylvanian boundary, based on coal-ball and palynological analyses of mainly North American coals. However, it has since been shown that

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the change from lycopsid- to fern-dominated vegetation had already taken place elsewhere in Euramerica by late Middle Pennsylvanian times (Dimitrova, Cleal & Thomas, 2005), and that the diagrams presented by Phillips, Peppers & DiMichele (1985) over-generalize the floristic changes that took place across the Euramerican region as a whole. Clearly, locally driven environmental changes were taking place, probably driven by palaeogeographical position and geotectonic forces (Opluštil, 2004). Palaeogeographical position would have influenced a range of factors such as topography, altitude, latitude and distance from the sea/ocean, existence of rain shadows, and the character of winds responsible for local climatic condition, especially for annual precipitation rate and its distribution through the year. The geotectonic position of the basin would influence its tectonic regime and subsidence rate, which in turn is one of the most important controls on sedimentary environment including fluvial styles and water table stability/fluctuation resulting, for example, in uneven drainage and coexistence of poorly and well-drained floodplain (Calder, 1994).

In order to try to separate out these globally driven and locally driven causal mechanisms, we need to determine the differences in timing of these changes in basins in different palaeogeographical and tectonic settings. The main problem is that for each basin there is a lack of sufficient comparable data, and what is available is often scattered in a number of unpublished reports written in local languages. The present paper is a first attempt at such a comparative analysis of Euramerican coal basins of different tectonic regimes and palaeogeographies. The study mainly concentrates on late Middle Pennsylvanian times, when the coal forests were undergoing their most significant decline in aerial extent, and most arborescent-lycopsids were disappearing.

2. Methods

This paper focuses on a comparison of tectono-sedimentary, coal petrography and floristic records derived from the selected Euramerican Late Palaeozoic basins. The aim is to compare tectonic and climatic changes interpreted from the sedimentary records with floristic and palynological data, to recognize how the vegetation responded to changes in tectonic regime and climate. The success of such a comparative analysis depends heavily on the character and completeness of the data from the individual basins. For some areas, some types of data were unavailable to us, such as coal petrography and some palynology for the British coalfields, and plant diversity patterns for the Upper Silesia Coal Basin. Nevertheless, we believe that we have been able to assemble enough evidence to give a preliminary assessment of the general patterns of biotic and environmental change taking place.

In all, two foreland (Upper Silesia and South Wales), one cratonic (Pennines) and two strike-slip (Central and Western Bohemia and Intra Sudetic) basins were selected for comparison (Fig. 1). Besides the different geotectonic regimes, these basins are located in various parts of the former Euramerican floristic province, so that the local peculiarities (e.g. presence or absence of endemic species) related to palaeogeographical position can be also be traced. Cratonic basins, with long-lasting even subsidence which produced a continuous sedimentary record through most of the Carboniferous, seem to be suitable for the study of climatic changes and related eustatic sea-level oscillations. Their comparison with the tectonically more active foreland basins can thus help to distinguish between global and local controls.

Both published and unpublished data were gathered, and further processed to present them in two charts per basin, allowing easy comparison between basins: one chart for the whole Pennsylvanian, and one for the interval around the Westphalian/Stephanian boundary. In these charts, a lithostratigraphical column with the names of the stratigraphical units is plotted on a time axis, which masks the information about the thicknesses of the units, but stresses the continuity/discontinuity of the sedimentary record and the duration of deposition of particular lithostratigraphical units. These columns also reflect the prevailing lithologies and their changes through time, as well as the presence of coals, volcanites or other important horizons (e.g. marine bands). They also distinguish between continental, paralic (mixed marine and continental strata in basins located along the coast) and marine environments, which are further subdivided into alluvial, fluvial (braided or meandering river dominated), lacustrine, palustrine, deltaic and shallow marine intervals. Information on subsidence rate and its changes through time, or its difference between various parts of the basin, was calculated for uncompacted lithostratigraphical units.

The climatic record is interpreted mainly on the presence/absence of climatic indicators, such as basin-wide red beds or coal-bearing strata, differences in abundance of coal, character of soils, presence of evaporites or carbonate cement and nodules (caliche), freshwater limestones, etc. Formation of red beds coeval to coal-bearing strata due to local water-table drop may be related to various factors, including locally different climatic conditions as well as reduced subsidence rate. Their reliable interpretation was, therefore, possible only where detailed studies of these individual stratigraphical units have been carried out (Besly, 1987).

Floristic changes are examined using both palynological and macrofloral data. Three types of data were compared in both categories of biostratigraphical record: (1) plant diversity, (2) plant group pattern-changes expressed by percentage composition of

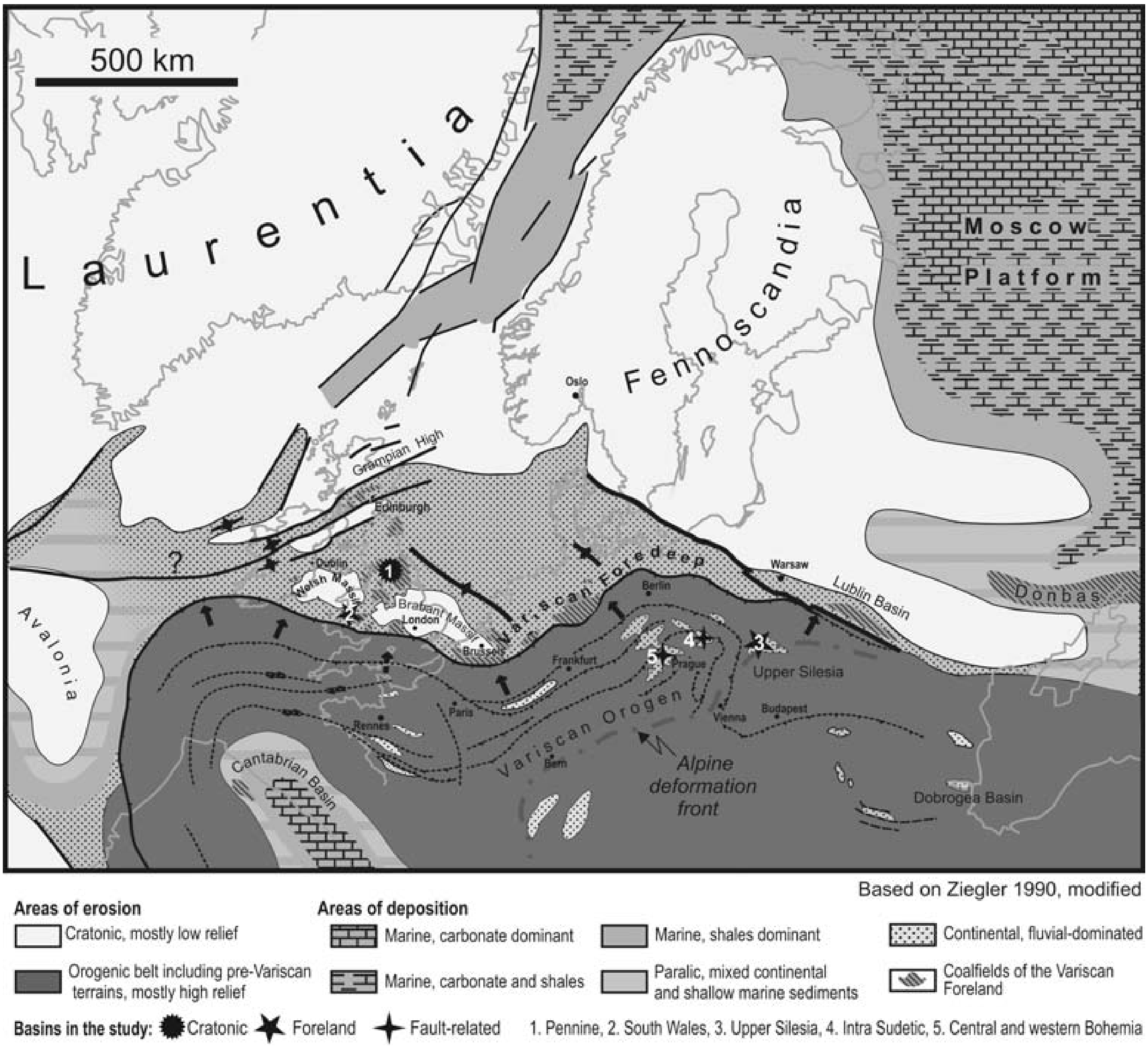

Figure 1. Palaeogeographical map of eastern part of the Euramerica during the Westphalian with position of the studied basins.

principal plant groups (lycopsids, calamites, sphenophylls, ferns, pteridosperms, cordaites and extrabasinal elements) and (3) stratigraphical ranges of key species. When trying to estimate plant diversity, we have mostly attempted to identify the number of biological species originally present, rather than the number of morphospecies present at a particular stratigraphical level, to avoid the problems of taxonomical inflation (see discussion in Cleal, 2005, and 2007, this issue). The estimation of a number of biological species is based mostly on foliage morphology for most plant groups, although in the ferns, sporangia were also used as a supporting criterion. The only exception is the Upper Silesia Coal Basin, where the number of morphospecies was the only type of data available in literature. Consequently, no direct comparison based purely on the number of species can be performed between the Upper Silesia Coal Basin and other coalfields involved in this paper. In most basins, diversity of lepidodendrids and sigillarias, considered as indicators of humid climate and high water table, was traced within the basin succession.

Since all of the coalfields studied in this paper are European, we have mostly used the Heerlen regional chronostratigraphical scheme, as this gives a better temporal resolution of the sequences than the IUGS global chronostratigraphy (Wagner, 1974; Wagner & Winkler-Prins, 1994, 1997).

3. Stratigraphical correlations

Clearly, stratigraphical correlation is critical in an analysis of this kind, as it is on the basis of this that the relative timing of events in the different basins is established. In Namurian and early Westphalian sequences in the European coalfields, there are numerous eustatically controlled marine bands that are well documented and effectively isochronous (e.g. Calver, 1968; Bless &

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Winkler-Prins, 1972; Ramsbottom, 1978, 1979) and provide a robust framework for such correlations.

The stratigraphically highest marine band in Europe is middle Bolsovian in age and so, above here, correlations must depend on biostratigraphy. The only fossil group found in any abundance and diversity in these strata are plant remains, either macrofossils or palynomorphs. Since the distribution of the plants may well have been influenced by the sort of environmental changes that we are investigating, there is a real risk of circular argument arising. We will therefore briefly outline the criteria used for identifying the main biozones on which our correlations are based.

3.a. *Linopteris obliqua* Zone

This Asturian Substage is the only subdivision of the Westphalian Stage whose base was not defined at a marine band and has been consequently the subject of considerable debate over the years. It was initially informally defined by Jongmans & Gothan (1937) at the lowest occurrence of the plant macrofossil species *Neuropteris ovata* Hoffmann. For a time, efforts were made to link formally the Westphalian D (now known as Asturian) Substage with the range of that species (e.g. Bode, 1970), but it soon became evident that this was impractical; in some areas (e.g. Canadian Maritimes Basin) *N. ovata* is absent from strata that on other criteria would be attributed an early Asturian age, whereas in others the species ranged up into strata which are clearly middle and late Stephanian in age.

To try to resolve this problem, Laveine (1977) established a much wider set of biostratigraphical criteria for identifying the base of the Asturian Substage, involving palynomorphs as well as macrofloras. The idea was that environmental factors would be unlikely to impact on all plant groups in the same way. Although the lowest occurrence of *N. ovata* remains the key criterion, it should only be used if it fits in with the other biohorizons in the scheme. The model has been successfully applied in South Wales (Cleal, 1978), Saarland (Cleal, 1984*a*), the Canadian Maritimes (Zodrow & Cleal, 1985) and northern Spain (Wagner & Alvarez-Vázquez, 1991). The results from Canada were particularly significant, as the model enabled the base of the Asturian to be identified at a level well below the lowest occurrence in that sequence of *N. ovata*, whose range seems to have been restricted there by local environmental factors. This biohorizon has subsequently been used to define the base of the *Linopteris obliqua* Zone by Wagner (1984), which now in effect has become the index to the base of the Asturian Substage (Wagner *et al.* 2002).

Particularly in Britain, the macrofloral record has tended to be ignored and other criteria instead used to identify the base of the Asturian Substage, notably the base of the non-marine bivalve *Anthraconauta tenuis* Zone and the base of the palynological OT (*Thymospora*) Zone (e.g. Ramsbottom *et al.* 1978). As pointed out by Cleal (1984*b*), however, the bases of these zones both occur below the base of the Asturian Substage, as indicated by macrofloral evidence.

3.b. Middle Asturian biozones

Laveine (1977) also noted a significant change in the macrofloral and palynological record in the middle Asturian Substage, involving the appearance of a number of abundant species of marattialean ferns and medullosalean pteridosperms. Wagner (1984) later formalized this as the base of the *Lobatopteris vestita* Biozone. Detailed records from South Wales (Cleal, 1978), Saarland (Cleal, 1984*a*), the Canadian Maritimes (Zodrow & Cleal, 1985), southern England (Cleal, 1986, 1987, 1997) and northern Spain (Wagner & Alvarez-Vázquez, 1991) have since shown that the change in fact occurs in two steps, which has resulted in a three-fold biostratigraphical division of the Asturian Substage (Cleal & Thomas, 1994; Cleal *et al.* 2006).

The lower of these two macrofloral biohorizons within the Asturian Substage coincides with a marked increase in marattialean palynomorphs, and in particular the base of the epibole of *Thymospora obscura* (Kosanke) Wilson & Venkatachala. It is often referred to in an abbreviated form as the base of the *Thymospora* epibole. As pointed out by Cleal *et al.* (2003), however, there are in fact two epiboles of species of *Thymospora*: one, the *T. obscura* epibole in the middle Asturian Substage, the other the *T. pseudothiessenii* epibole at or about the Asturian–Cantabrian boundary. There have been cases where these two epiboles have been confused (e.g. Peppers, 1985) and have resulted in errors in correlation, and care has to be taken when using these spores for correlation.

The upper of the macroflora biohorizons should be coincident with the appearance of *Vesicaspora* pollen, since the latter's parent plant (*Dicksonites*) is one of the macrofloral taxa that appear at this level. However, as *Vesicaspora* is rarely an abundant element in palynological samples of this age, its potential biostratigraphical importance has not been realized.

3.c. *Odontopteris cantabrica* Zone

The base of the Cantabrian Stage is defined in its stratotype so that it is coincident with the base of the *Odontopteris cantabrica* Biozone in the macrofloral record (Wagner & Alvarez-Vázquez, 1991). Cleal *et al.* (2003) analysed the macrofloral and palynological criteria for identifying the base of this zone, and found a consistent pattern of distribution across Euramerica of key taxa of sphenophylls, marattialean ferns, and mariopterid and medullosalean pteridosperms. Importantly, there was also a clear palynological signal at this boundary, notably with a sudden and marked increase in abundance of *Thymospora pseudothiessenii*

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and the first appearance of *Schiopfites dimorphus* (the latter spore of unknown affinities). Again, because of the consistent and widespread occurrence of this biohorizon, involving a range of different plant groups, it seems likely that it represents a reliable index for chronostratigraphical correlation.

4. Geological background of compared basins

The basins studied in this paper were formed as a result of the Variscan/Appalachian orogenesis. This in turn was caused by the collision of the southern margin of Laurussia (the 'Old Red Continent') and the north African part of Gondwana, together with the intercalated small plates of the Armorican Terrane Assemblage (Franke, 2000). The result was a W–E-striking belt of coal basins following the direction of the Carboniferous palaeoequator, traceable from the southern and eastern United States, through Atlantic Canada and the British Isles, northern France and the Benelux lands, to north Germany, Poland, the Czech Republic and probably into Bulgaria. It is characterized by an apparent transition from north to south (from west to east in parts of the United States) from stable platforms with cratonic basins, through a collision zone with foreland basins (North Variscan and Appalachian foredeeps), to the Variscan orogenic belt where usually smaller, fault-related intramontane basins arose. Cratonic and foreland basins occupied vast coastal lowlands, which underwent frequent marine incursions (paralic basins), whereas fault-related basins were filled only by continental sediments due to their high altitude (Fig. 2). Estimations of their altitudes vary between a few hundred metres and several kilometres (compare Opluštil, 2005, and Becq-Giraudon, Montenat & Driessche, 1996).

4.a. Cratonic basins

These basins, located on the consolidated parts of plates (cratons) with only weak tectonic activity, are characterized by slow subsidence rates produced by cooling of previously heated lithosphere, which, together with eustatic processes, are the most important controls on deposition and architecture of the sedimentary record.

The Pennines Basin was selected for this analysis as the example of a cratonic basin. It is situated in the English Midlands, north of the Variscan Foreland from which it is separated by the London–Brabant Massif (Eastern Avalonia terrane). In late Namurian times, deposition changed from purely marine to paralic, the latter lasting until the Duckmantian/Bolsovian boundary (Fig. 3). Coal-bearing deposition mostly spans the interval from Langsettian to Bolsovian times (Guion & Fielding, 1988; Guion, Fulton & Jones, 1995), although some economically unimportant coal seams also occur in the Asturian Substage. Red beds appear occasionally already during Langsettian times in the southernmost part of the basin adjacent to the northern slope of the London–Brabant palaeohigh but became more widespread in early Bolsovian times, and of basin-wide extent towards the end of Bolsovian and in Stephanian times (Besly & Turner, 1983; Besly, 1987, 1988; Glover, Powell & Waters, 1993).

4.b. Foreland basins

These basins are related to a compressional regime of collision zones along the plate margins, where the accommodation space is produced by lithosphere downflexure and by sediment load. Intensity of subsidence as well as deformation of basin fill significantly decreases away from the orogenic front (Gayer *et al.* 1993). Deposition in these basins is controlled mainly by tectonics and sedimentary influx. Purely eustatically driven cyclothems are usually overprinted by stronger tectonic influence (Heckel, 2002; Wagner & Winkler-Prins, 2002). Therefore, stratigraphical sequences of foreland basins cannot usually be correlated in a simple manner with those in cratonic basins, where there was slow and even thermal subsidence.

Two foreland basins are examined in this study: the South Wales and Upper Silesia coal basins. They represent the southern part of the North Variscan foreland (the North Variscan Foredeep) located along the active northern margin of the Armorican plate assemblage. Both basins therefore exhibit similar sedimentary histories with purely marine strata in the lower part of the succession, passing up into paralic and finally to continental deposits. The paralic and continental parts of the basin fill consist of coal-bearing strata. Both basins differ in the timing of onset of paralic and continental deposition, as well as in the thickness of the basin fill (Fig. 2). In the Upper Silesia Coal Basin, paralic deposition spanned early Namurian, that is, Arnsbergian and Chokierian times (Fig. 4), whereas in the South Wales Basin it ranged from late Namurian to middle Bolsovian times (Fig. 5; Jones, 1989, 1991; Hartley & Warr, 1990; Hartley, 1993*a*,*b*). Subsidence was much stronger in the Upper Silesia Basin, where up to 4000 m of Westphalian coal-bearing strata were deposited, compared to 2440 m of coal-bearing deposits of the same age in the South Wales Basin. In the Upper Silesia Basin, however, coal-bearing strata span the interval from early Namurian to late Westphalian times. The cumulative thickness of these sequences reaches 7500 m but, due to eastward syn-orogenic depocentre migration and consequently erosion of its western margin, the maximum thickness in any part of the basin does not exceed 4500 m. Palaeogeographically, these basins were about 1500 km apart.

Red beds in these two foreland basins are only known from Upper Silesia where they are represented by coarse-grained fluvial sediments of the Kwaczala Arkose overlying the coal-bearing Libiaz Beds of Asturian–Cantabrian age. The Kwaczala Arkose

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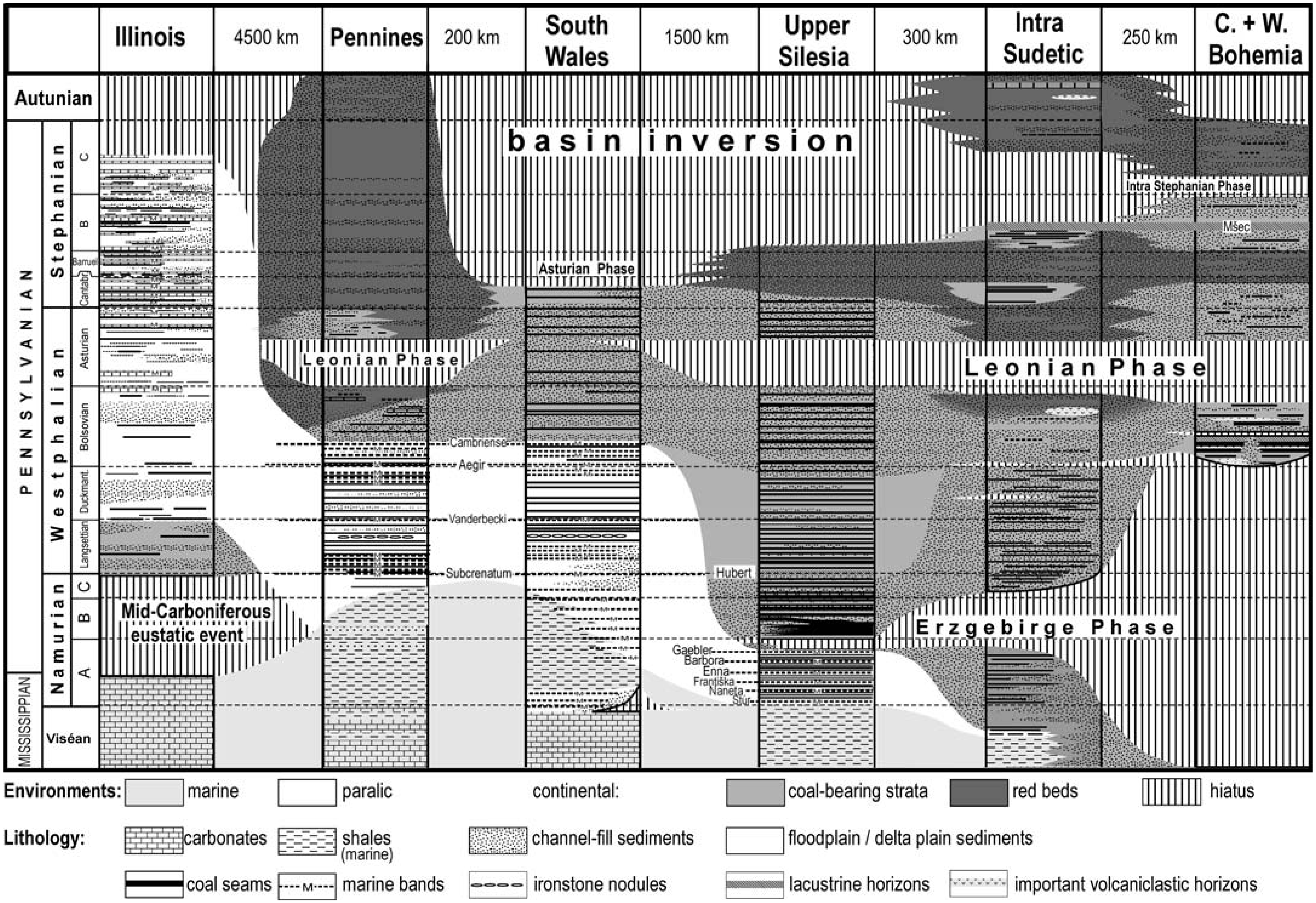

Figure 2. Comparison of tectono-sedimentary evolutions of selected basins of the North Variscan Foreland and adjacent continental basins in the Variscan hinterland. The cratonic Illinois Basin is added to note the differences between American and European parts of the Euramerican province.

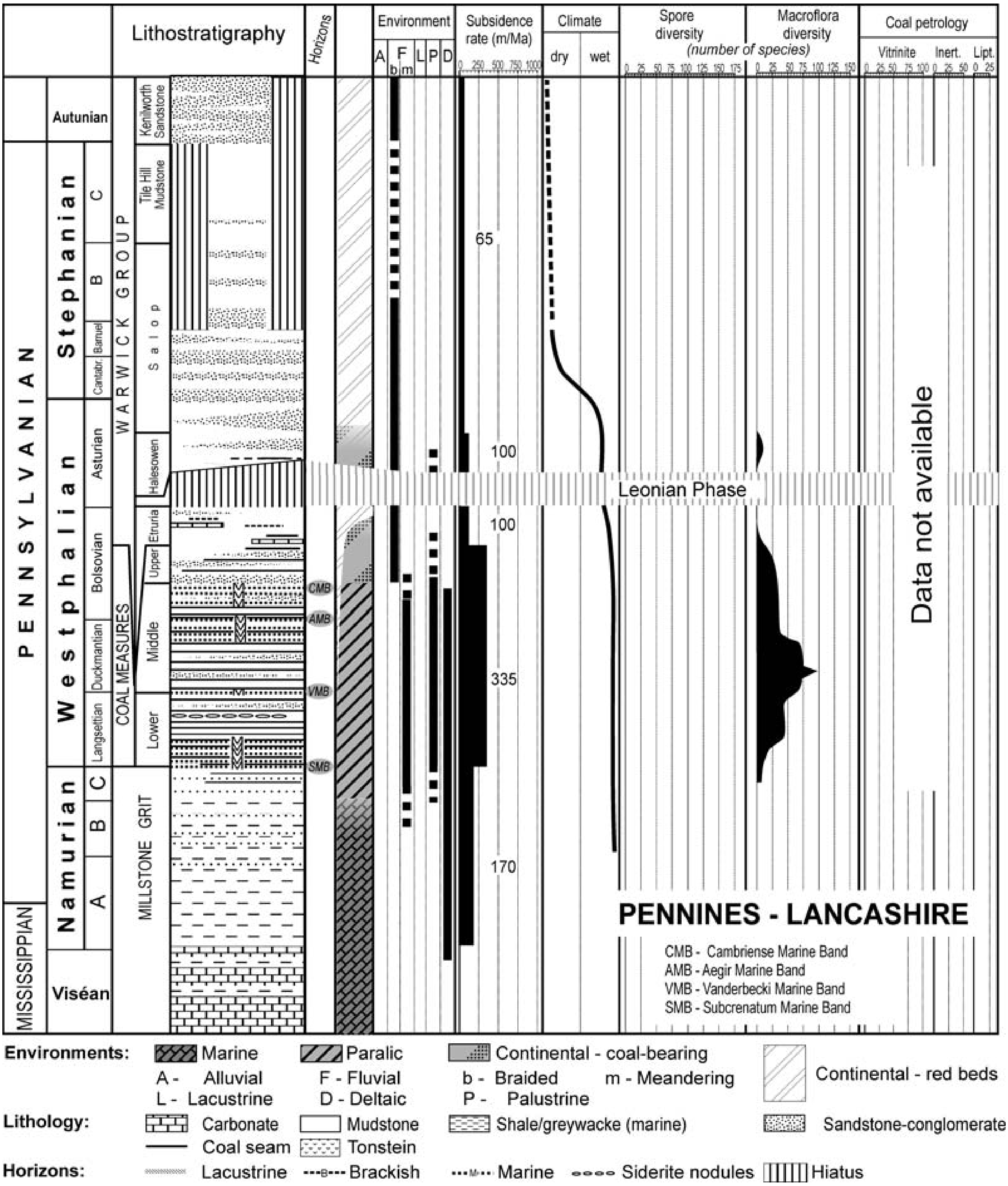

Figure 3. Tectono-sedimentary history and macrofloral diversity of the Pennines Basin.

contains only silicified woods and very rare red-coloured mudstone intercalations lacking any fossils.

4.c. 'Fault-related' basins

This category involves intramontane basins situated within the Variscan mountainous chain. Their formation was related to normal or strike-slip faulting, and basin fill is characterized by an alternation of red coal-barren and grey coal-bearing continental strata, interrupted by frequent hiatuses. The main controls on deposition are tectonics, climate and clastic input. Their intra-continental position and higher altitude could result in local climate conditions that are different from those in basins located along or near the coast in low altitudes. Two basins were selected for comparative

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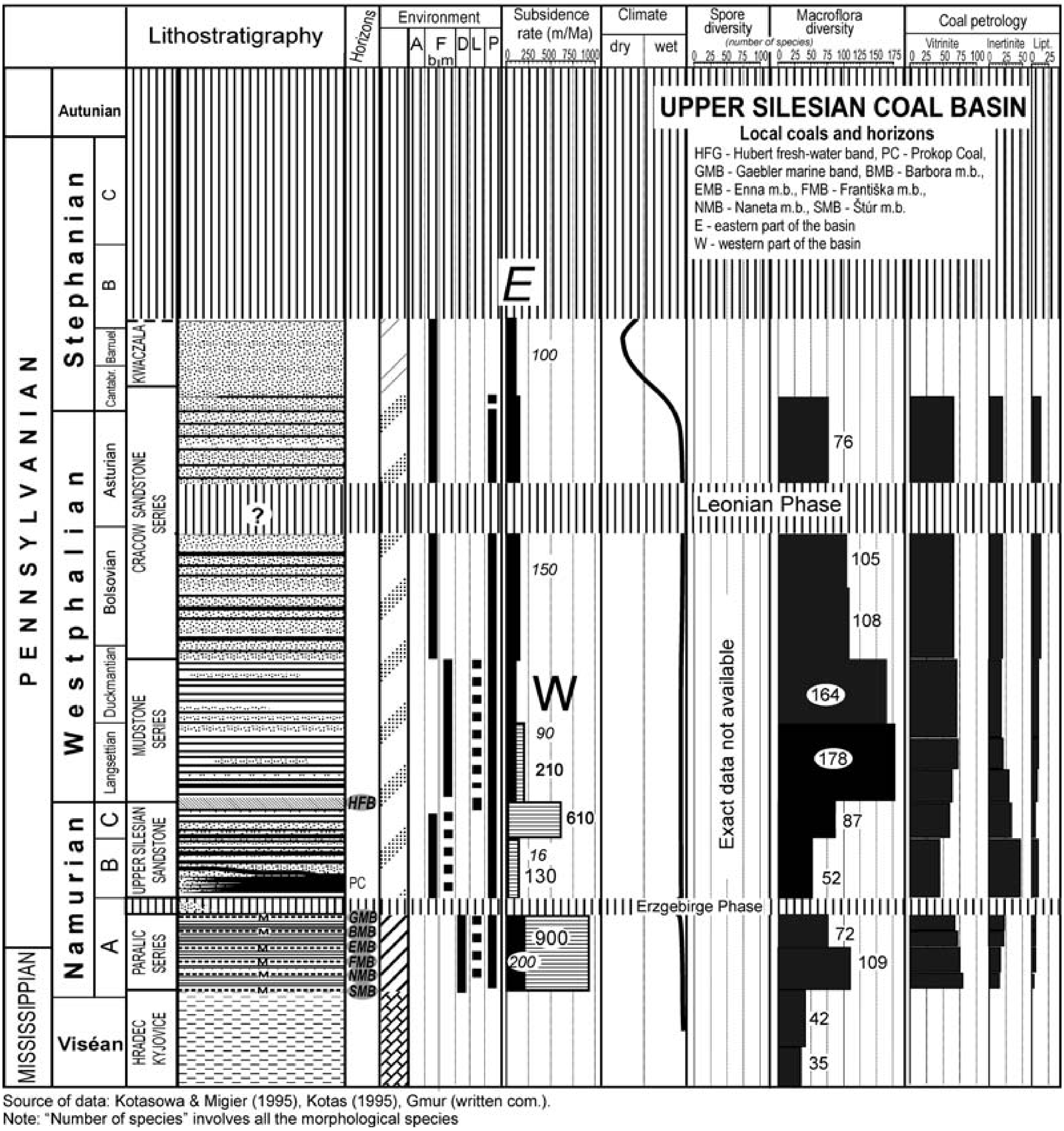

Figure 4. Tectono-sedimentary history, spore and macroflora diversity patterns, and maceral composition of particular sedimentary units of the Upper Silesia Coal Basin. For explanation see legend in Figure 3.

analysis: the Central and Western Bohemia Basin and the Intra Sudetic Basin along the Czech and Polish border in the Lusatian area. These basins are located on the Bohemian Massif, about 100 km apart. However, despite the close geographical position, they exhibit many differences, including stratigraphical range and thickness.

The Central and Western Bohemia Late Palaeozoic basins are formally subdivided into several subbasins, which exhibit very similar sedimentary records and share the same lithostratigraphical subdivisions. They originally formed a single basin, the stratigraphical range of which spans the interval from the Bolsovian Substage to the top of the Stephanian Stage (Fig. 6). A succession up to 1400 m thick is divided into four formations, based on an alternation of reddish coal-barren and grey coal-bearing deposits (Weithofer, 1896, 1902; Pešek, 1994). The most important coal-bearing unit, the Kladno (Lower Grey) Formation, consists of the Radnice (Bolsovian) and Nýřany (Asturian–Cantabrian) members, separated by a basin-wide hiatus. Caliche horizons appear in both red formations, being irregularly distributed within these units both stratigraphically and laterally (Skoček, 1993). Coal seams can occur but are

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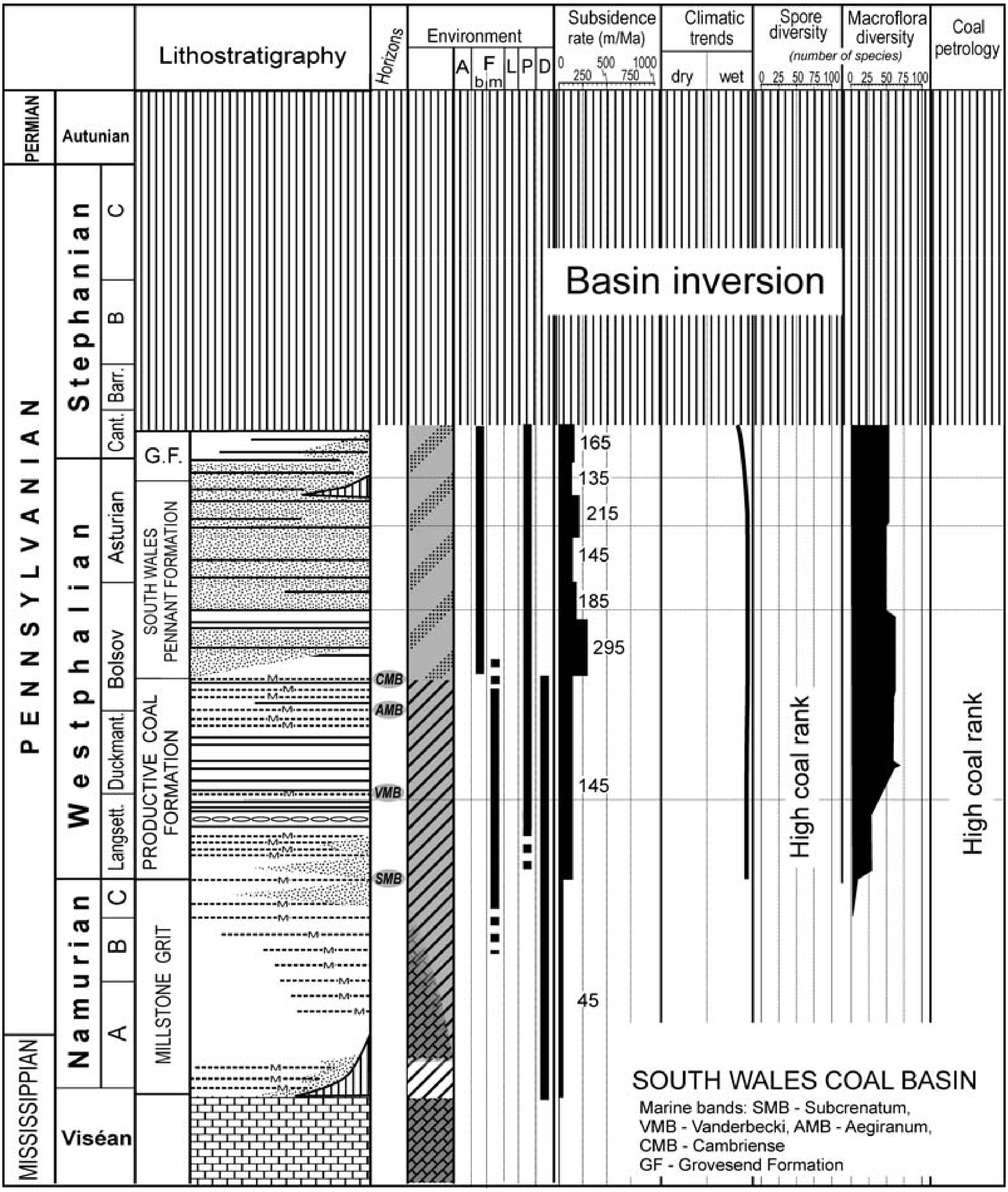

Figure 5. Tectono-sedimentary history and macroflora diversity of the South Wales Coal Basin. For explanation see legend in Figure 3.

very rare, thin, impure and locally developed in these red bed formations (Pešek, 1994).

The Intra Sudetic Basin exhibits a complex tectono-sedimentary history, which spans the interval from Mississippian to Triassic times. Except for the Viséan Stage, the remaining part of the succession consists purely of continental deposits. Pennsylvanian strata reach 2000 m in thickness, and are characterized by an alternation of red beds and coal-bearing sediments (Fig. 7). The transition from the Westphalian to the Stephanian stages is recorded in predominantly red bed development (Svatoňovice Member), with only a poor flora indicating an Asturian age. The upper part of this unit contains four coals of the Svatoňovice group, which are associated with a very rich macroflora of Cantabrian age (Šetlík, 1977; Pešek, 2004). This

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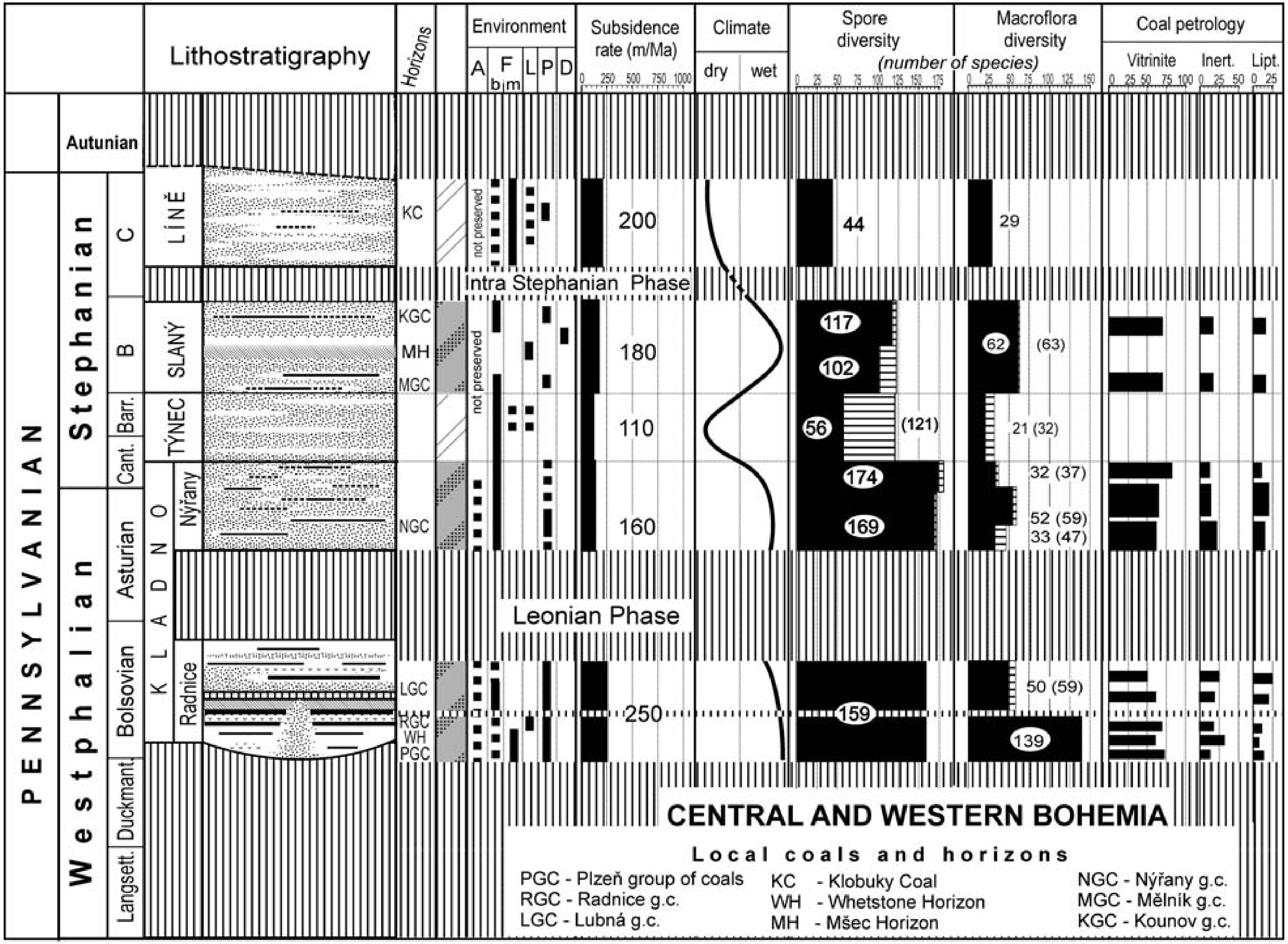

Source of data: Pešek 2005, Pešek et al. 2001

Figure 6. Tectono-sedimentary history, spore and macroflora diversity, and maceral composition of coal seams of the continental basins in Central and Western Bohemia. For explanation see legend in Figure 3.

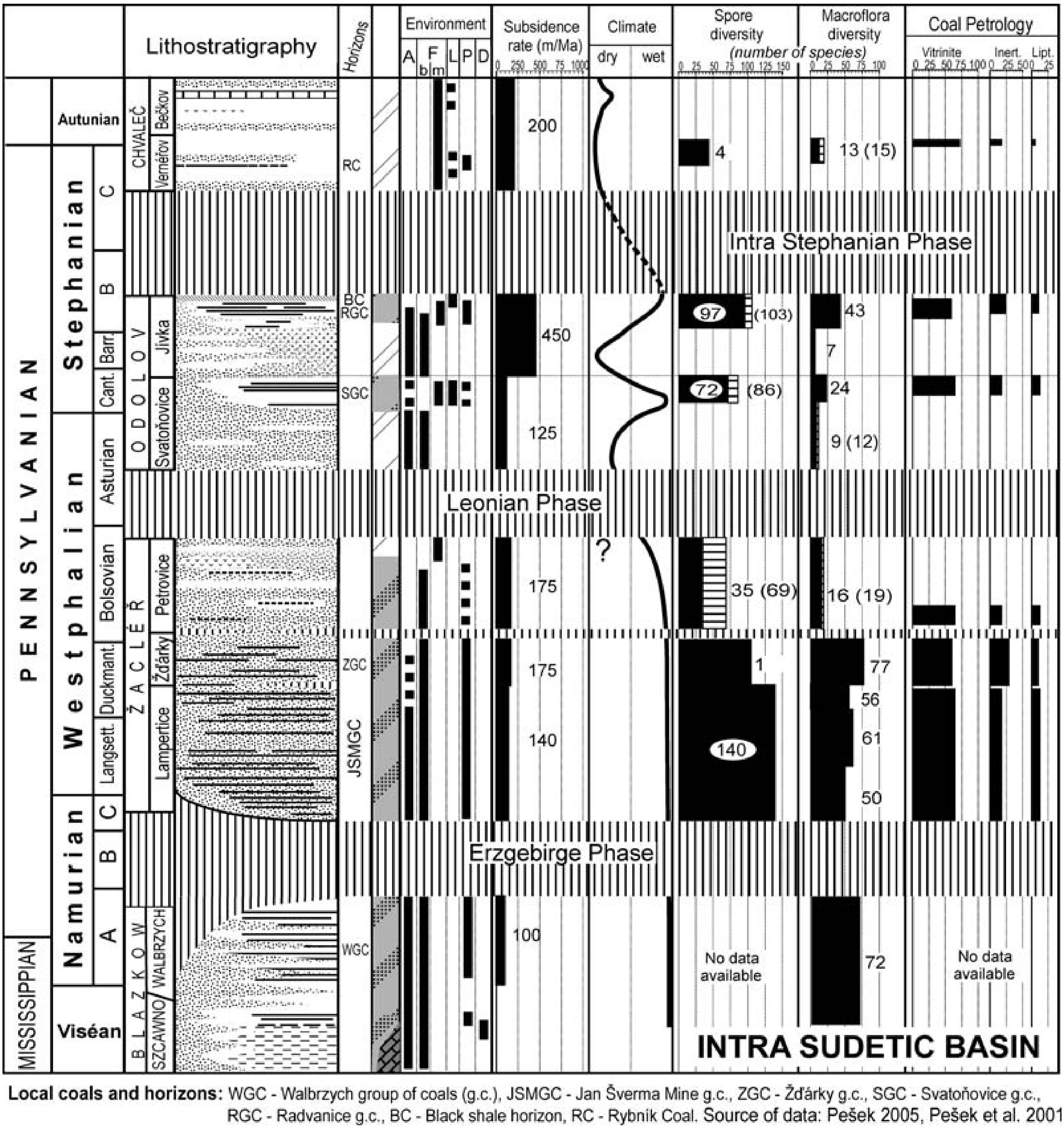

Figure 7. Tectono-sedimentary history and macroflora diversity of the Intra Sudetic Basin. For explanation see Figure 3.

coal-bearing horizon is, however, of regional extent and laterally passes into red-bed facies (Skoček, 1993). Tásler *et al.* (1979) also noted the presence of several thin brownish bituminous limestone beds in the Lower Svatoňovice Member. Some of these may be associated with the stratigraphically deepest coal of the Upper Svatoňovice Member or can occur between this and overlying coals. Red bed development is typical also for most of the following unit, the Jívka Member, composed predominantly of coarse-grained sediments with silicified woods and occasional mudstone intercalations. In the middle and upper parts, locally developed coal-bearing horizons of middle Stephanian age occur. The most important is the Radvanice group of coals containing several mineable but high-ash coals.

5. Comparisons and discussion

5.a. Sedimentary history and subsidence

The timing of the change from marine to paralic, and eventually to the continental environments typical for cratonic and foreland basins, has been compared between the different basins. Purely marine conditions predominated in these basins until the end of Viséan times. However, the onset of paralic conditions is apparently diachronous among the basins and migrates from east to west (Fig. 2). Highly coal-bearing paralic strata with frequent marine bands, deposited mostly on upper delta plains occupied by meandering to anastomosing fluvial systems, first appear in the Upper Silesia Coal Basin in the lower Namurian Stage (Pendleian/Arnsbergian). In the South Wales and Pennines basins, mixed marine and continental deposition also starts during Namurian times, but as a siliciclastic poorly coal-bearing lower delta plain succession with thin non-workable coal seams (the Millstone Grit) passing up into a highly coal-bearing one with economically important coals (Lower Coal Measures) in the Langsettian Substage (Hampson, 1998; Dreesen *et al.* 1995).

The duration of paralic deposition and the transition to purely continental environments without any marine influence is also highly diachronous, depending on geographical position. Generally, paralic deposition persisted much longer in the western part of Euramerica. In the Upper Silesia Basin, the last marine band (Gaebler MB) at the top of the Paralic Series (= Ostrava Formation in the Czech part of the basin) corresponds to the top of goniatite zone E_2, (Chokierian). A succession more than 3000 m thick, containing about 160 economically important coal seams and over 80 marine bands (Řehoř & Řehořová, 1972), was deposited during an interval of less than 3.5 Ma on a delta plain open to the sea located to the north. The average subsidence rate is estimated to be about 900 m/Ma but decreases substantially to less than 200 m/Ma eastward, away from the orogenic front. Further west, in the South Wales and Pennines coalfields, the youngest marine band (the Cambriense MB) is approximately of middle Bolsovian age. Coal-bearing paralic deposition here spans the interval from Langsettian to middle Bolsovian times, the duration of which was about 6 to 7 Ma (Menning, 1995).

A purely continental environment was first established in the Upper Silesia Coal Basin, as a response to the middle Carboniferous eustatic event, but also as a consequence of the Erzgebirge Tectonic Phase (Gothan, 1954; Kotas, 1995) resulting in a long-lasting hiatus between the paralic and continental parts of the basin-fill. Apart from a floristic break (Gothan, 1954) this hiatus is indicated by the significant and widespread silcrete palaeosol (ganister) horizon at the base of the continental part of the basin fill (Martinec & Kožušníková, 1997), as well as by the lithological contrast which reflects prominent changes in the sedimentary environment. In middle Namurian times, the mudstone-dominated delta plain was replaced by a braided fluvial system in which long-lasting mires developed (Gradzinski, Doktor & Slomka, 1995). This deposition lasted until the end of Namurian times, when it changed to a mudstone-dominated meandering/anastomosing fluvial system (Gradzinski & Doktor, 1995). Such fluvial styles also characterized coal-bearing deposition on the upper delta plain in the South Wales and Pennines basins during Langsettian to early Bolsovian times (Guion, Fulton & Jones, 1995). From the beginning of Bolsovian times in the Upper Silesia Coal Basin or middle Bolsovian times in the South Wales and Pennines basins, fluvial styles changed over most of the North Variscan Foreland and deposition took place in braided fluvial systems, which persisted in most coalfields until the end of Westphalian or early Stephanian times. This change in sedimentary environments in coalfields of the North Variscan Foreland in early to middle Bolsovian times corresponds to the onset of deposition in the Central and Western Bohemia Basin and with a short depositional break in the Intra Sudetic Basin (Figs 2, 7). Basin-wide lacustrine or lacustrine-delta horizons occur frequently in foreland and cratonic basins but are less common in intramontane fault-bounded basins. The Hubert fresh-water band of the Upper Silesia Coal Basin, located near the Namurian/Westphalian boundary, is approximately a time equivalent of the Subcrenatum Marine Band in western Europe (Fig. 2).

Continuity of the sedimentary record is generally more complete in basins located in lower altitudes near the sea, whereas frequent and prominent hiatuses are more typical for intramontane fault-bounded basins prone to substantial base level changes. The most important and widespread hiatus is at the boundary between the Mississippian and Pennsylvanian subsystems. It is well developed in the Upper Silesia Basin, whereas in the South Wales and Pennines basins it is marked by a transition from purely marine to paralic conditions. This hiatus is ascribed not only to the late early Namurian eustatic lowstand, but it is also related to the Erzgebirge Phase in the Upper Silesia Coal Basin. Another widespread hiatus detected in more than one basin corresponds to the Leonian phase of the Variscan Orogeny in late Westphalian times. It is more prominent in continental basins in the interior of the Variscan orogen. It has been identified biostratigraphically and lithologically in both the Intra Sudetic Basin and the basins of Central and Western Bohemia. In the Upper Silesia Coal Basin, there is only biostratigraphical evidence of the hiatus (Kotasowa & Migier, 1995); no sedimentological change has been observed here (M. Doktor & A. Kędzior, pers. comm.). In the Pennines Basin, there is widespread field evidence of a non-sequence at the base of the Halesowen Formation (late Asturian in age), indicating a hiatus equivalent in age to the Leonian Phase (Cleal, 1986; Besly & Cleal, 1997). In the South Wales Basin, this hiatus is only locally developed in the eastern part of the coalfield (Cleal, 1978). Another hiatus occurs between the Stephanian B and C substages, but has only been biostratigraphically proven (Pešek, 1994, 2004) in the continental basins where partly coal-bearing deposition continued to the end of Pennsylvanian times. It is marked also by a widespread erosional surface detected in seismic profiles (Pešek, 2005; Skopec & Pešek, 2005).

Thickness of basin fill also varies, not only among basins, but also within a single basin. Maximum thicknesses of Pennsylvanian strata in the compared basins varies between 1 km and more than 4 km. Subsidence rates were also significantly different, ranging from about 15 m/Ma to about 900 m/Ma for particular units. The average rate varies between 150 m/Ma and 250 m/Ma. Foreland basins and some fault-bounded continental basins typically had the highest rates of subsidence, and thus contain the thickest Pennsylvanian-aged sequences.

5.b. Coal petrography

The petrographical composition of coal provides basic information on hydrological conditions under which the peat accumulated, especially in relation to the water table and its fluctuations (e.g. Diessel, 1992). The hydrological regime of former mires (that is, whether they were rheotrophic or ombrotrophic) can depend on the character of the climate in the sense of seasonality and annual precipitation, as has been shown in modern tropical mires of SE Asia (e.g. Neuzil *et al.* 1993; Wüst & Bustin, 2004). To be formed, raised (ombrotrophic) mires require high annual precipitations (> 2500 mm) evenly distributed through the year to protect the peat from severe oxidation. In contrast, planar (rheotrophic) mires can develop in less wet areas where annual precipitation rate is lower, between 1300 and 2500 mm per year, and some seasonality occurs (Parrish, 1998; E. D. Gyllenhaal, unpub. PhD. thesis, Univ. Chicago, 1991). These primary conditions are reflected in petrographical composition as well as in ash and sulphur contents (Smith, 1962; Cecil *et al.* 1985; Eble & Grady, 1993). In this study, only data on maceral group composition of the major coal seams of the studied basins are compared, except for the South Wales and the Pennines coalfields from which no data were available. The purpose is to recognize any trends, either among basins or within the particular basins, as described, for example, by Harvey & Dillon (1985) in the Westphalian–Stephanian coal seams of the Illinois Basin. These authors observed increasing inertinite from about 9 % in older (Westphalian) to 16 % in younger (Stephanian) coal seams. They explain this trend as a result of increased peat oxidation as the weather became drier during Stephanian times. However, our available petrographical data does not reveal any prominent trend in maceral composition through time in any basin. Instead, compositional variations are irregular. Only in the Upper Silesia Coal Basin do they seem to follow changes in architecture and lithology of the sedimentary record (Fig. 4). Here, the inertinite-rich coals (locally more than 60 % of inertinite) occur in sandstone-dominated units (e.g. the Upper Silesia Sandstone Series), whereas vitrinite-rich coals are typical of mudstone-dominated parts of the succession. The formation of inertinite-rich coals is generally related to peat oxidation due to water table drop or fluctuation. This, however, may be caused by slow subsidence, but similar results can also occur in ombrotrophic mires formed in areas with regular and high precipitation rates, where the water table also quite often drops below the peat surface (Smith, 1962, 1968; Calder, 1993). The latter, however, can be distinguished by having a very low ash content where most of the mineral matter is represented by minerals of diagenetic and eolian origin.

5.c. Climatic indicators derived from the sedimentary record

The character of the Late Palaeozoic climate can quite often be inferred from the sedimentary record (e.g. Cecil *et al.* 1985; Donaldson, Renton & Presley, 1985; Besly, 1987, 1988). Coal, aeolian deposits, evaporites, red beds, palaeosols or mineralogy of clay minerals are usually used for such analyses since it is believed their formation is climate-dependent. However, climatic interpretations inferred from these rocks can in many cases be misleading without thorough sedimentological understanding of their genesis (e.g. see discussion in Besly, 1987 or Retallack, 2001). As an example, thin impure coal seams can form in places other than where a permanent water table allows peat accumulation, even in semi-arid conditions. Red beds are now agreed to form under climates ranging from humid tropical to arid, requiring only conditions that allow early diagenesis in an oxidative environment (Pye, 1980). Therefore, other indicators must be taken into account. Among them, palaeosols are often considered to be one of the most powerful palaeoclimatic indicators, since their chemistry and morphology is extremely sensitive to temperature and humidity (Besly, 1987; Retallack, 2001). The presence of laterites and calcretes has provided valuable information for distinguishing between particular red bed facies in the European Late Palaeozoic (van Houten, 1982; Besly, 1987, 1988). Comparison with recent calcretes suggests rainfall of 400–600 mm and mean temperatures of 16–22 °C (Goudie, 1973). Clay mineralogy can also potentially provide important data on palaeoclimate since the formation of clay minerals depends among other things on the amount of rainfall available to soil. Generally, the dominance of kaolinite is indicative of wetter climates (1000–2000 mm of annual rainfall), whereas more varied clay mineral associations are produced under more arid conditions (Besly, 1987; Retallack, 2001). Aluminium and iron oxides usually predominate in climates allowing ferallitic weathering under extreme conditions of high temperature and humidity.

In most basins selected for comparison, quite well-established sedimentary models exist. They allow the character of the climate to be interpreted based on basin-wide alternations of coal-bearing grey sediments and coal-barren red beds, on the type of palaeosols (podzolic, ferruginous or caliche), the spatial distribution of coals, and the composition of clay minerals. A comparison of the climate-indicative data from the studied basins displays a regional pattern traceable across the major part of Euramerica. Generally, there is a trend of increasing dryness and/or seasonality, marked by a decreasing number of coal seams and prolongation of periods of red bed formation, during Stephanian/Autunian times (Cecil *et al*. 1985; Besly, 1987; Gastaldo, DiMichele & Pfeffercorn, 1996). In detail, this trend appears to have been highly irregular, with alternating stratigraphical intervals containing indicators of drier and wetter conditions. However, there are also some irregularities in the stratigraphical and geographical distribution of these climatic indicators, possibly related either to poorly established correlations of the basins or local controls responsible for red bed formation. These local controls may include tectonic processes or possibly local climatic conditions related to orographical barriers (Rowley *et al.* 1985). Those ascribed to tectonics are typically of local extent, concentrated especially along the basin margin where better drainage resulted in a fall in the water table and the total oxidation of organic matter (Calder, 1994; Retallack, 2001). However, increasing dryness can be considered to be responsible for the formation of basin-wide red beds associated with pedogenic carbonates.

In the Upper Silesia Basin from middle Namurian times, and in most other basins from Langsettian times, extensive mires spread over the basin depocentres. Strata of this age are dominated by widespread coal-bearing deposits. Locally, notably along the margins of the Pennines Basin, some red beds of Langsettian and Duckmantian age can be found. These red beds, however, lack any caliche palaeosols and their mineralogy suggests re-deposited lateritic weathering profiles indicative of a humid climate in the source area (Besly, 1987). Red beds of similar genesis and age are also known from the Intra Sudetic Basin, where they can locally appear from latest Namurian times onwards. However, from middle Bolsovian times, red beds of the Etruria Formation become much more widespread and contain the stratigraphically earliest caliche deposits of the Pennines Basin. Coal seams become thinner and much less common. In the other basins dealt with in this paper, red beds are mostly absent in the upper Bolsovian Substage except for the Petrovice Member in the Intra Sudetic Basin. This latter unit consists mostly of arkosic sandstones with mottled to red mudstone intercalations, and locally also with thin, laterally impersistent and impure coals or carbonaceous claystones; conglomerates and arkosic sandstones with silicified woods occur in the marginal parts of the basin. No caliche has been reported (Skoček, 1993) in the Petrovice Member. In the Central and Western Bohemia Basin, locally developed red beds derived from re-deposited laterite weathering crusts can occur along the basin margin in Bolsovian strata (Skoček & Holub, 1968). They are coeval to coal-bearing strata associated locally with kaolinitic refractory claystones. In the Asturian and Cantabrian substages, red beds become more common in the Pennines Basin, where both ferruginous and caliche palaeosols occur (Besly, 1987). Similar red beds of this age are known also from the Intra Sudetic Basin. Here, the Lower Svatoňovice Member (Asturian age) can be compared with the Halesowen and lower part of the Salop formations (= lower part of the Keel Formation in Besly, 1987). Reworked caliche nodules sporadically occur in the Lower Svatoňovice Member, but the red beds are still dominated by kaolinite and other products indicating lateritic weathering in the source area (Havlena, 1964). In the Upper Svatoňovice Member (Cantabrian age), marginally occurring red beds are coeval with economically important coals. In other basins in which this interval has been proved, such as in the Central and Western Bohemia Basin (Pešek, 1994), coal-bearing strata dominate over locally developed red beds concentrated along the present-day basin margins. In the Upper Silesia Basin, economic coals are fairly common up to the Cantabrian Substage in the upper part of the Libiaz Beds. Similarly in the South Wales Basin, red beds are all but absent through the whole succession up to the Cantabrian Substage. However, they occur as a discrete interval in Bristol–Somerset (not included in this analysis), an adjacent satellite coalfield genetically related to the South Wales Basin (Cleal, Dimitrova & Zodrow, 2003). The first really regionally widespread red beds with abundant caliche palaeosols accompanied by an apparent drop in coal content (coals are either thin and impersistent or completely absent) occur in the Barrualian Substage. This interval is developed in all of the compared basins and can be traced also in other Euramerican coalfields including the Appalachians (Conemaugh Formation: Cecil *et al.* 1985; Donaldson, Renton & Presley, 1985) and the interior basins of central North America (e.g. lower part of the Douglas Group in the Western Interior Basin: Archer *et al.* 1994; Schutter & Heckel, 1985). In the basins dealt with in this paper, this interval is represented by the Týnec Formation in the Central and Western Bohemia Basin, by the lower part of the Jívka Member in the Intra Sudetic Basin, and possibly by the Kwaczala Arkose in the Upper Silesia Coal Basin and part of the Salop Formation in the Pennines Basin. Succeeding strata of middle Stephanian age can only be proved biostratigraphically in the continental basins, where they are characterized by the presence of widespread and economically important coals and an absence of red beds. This contrasts with the Pennines Basin, where no coal-bearing equivalent of this age occurs, the whole Stephanian succession consisting of red beds. Outside the studied set of basins, a coal-bearing interval of similar age can be traced in the

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Appalachians (level of the Pittsburgh Coal) and in the interior basins further west. Late Stephanian times is again a period of red bed formation, but this interval can be identified biostratigraphically only in the continental basins. In the Central and Western Bohemia Basin it is represented by an up to 1000 m thick sequence of the Líně Formation, in which caliche palaeosols are common, and pebbles of Lower Palaeozoic limestones occur locally in feldspathic conglomerates. However, at least three widespread and several tens of metres thick horizons of grey to mottled grey-red sediments occur (e.g. Pešek, 1994). They are locally accompanied by decimetre-scale lacustrine limestones and cherts, as well as by thin impersistent and impure coals. Mudcracks are common in some horizons, indicating repeated (?seasonal) drying (Pešek, 1994).

5.d. Fossil record

Interpretation of floristic (both palynology and macrofloral) records of individual basins heavily depends on how data were obtained and processed. Diversity changes are usually related to discordances, lithological changes (e.g. transition from mudstone- to sandstone-dominated units) or alternation of grey coal-bearing and red coal-barren strata. Important also is the current state of systematic revision of a flora or a particular plant group (see, for instance, the problems inherent within the Cordaitanthales if cuticular data are unavailable: Šimůnek, 2000; Zodrow, Šimůnek & Bashforth, 2000), which can significantly affect the number of taxa recorded. Macrofloral data in particular may be biased by sampling methods. Nevertheless, significant data have been obtained by comparing diversity changes within particular plant groups, especially in large paralic basins located along coastal flats of the Variscan foreland (e.g. Cleal, 2005; Cleal, 2007, this issue). Here, nearly unlimited migration of plants was possible, comparing this to the restricted migration potential between small continental basins of the Variscan hinterland (Cleal, Dimitrova & Zodrow, 2003). Consequently, some key taxa can be absent or their stratigraphical ranges can differ in continental basins. Thus, only a few stratigraphically important species could be found in all of the compared basins (e.g. *Sphenophyllum oblongifolium* (Germar & Kaulfuss) Unger).

A comparison of palynological and macrofloral data from a particular locality and stratigraphical horizon normally shows significant differences (Dimitrova, Cleal & Thomas, 2005). There are always more spore morphotaxa than plant taxa, and there are differences in the proportional representation of the different plant groups. In most cases, however, the general pattern of changes in diversity of the macrofloras is similar to that of palynomorphs.

Macroflora diversity varies significantly between coal-bearing and coal-barren strata. Generally, there is very low diversity in red beds due to a combination of the original vegetation being less diverse, the fossilization potential being lower due to oxidation, and the absence of mining activity. The number of whole plant species in coal-bearing strata usually varies between 25 and nearly 70, exceptions being the Lower Radnice Member (Central and Western Bohemia, Bolsovian), where about 140 species have been identified, and the Barnsley Seam in the Pennines Basin with nearly 90 species (Figs 3, 6).

The unusually high biodiversity in the Lower Radnice Member is probably related to several factors. One is the occurrence of fossiliferous tuff beds intercalated in coal seams, which preserve as compressions the vegetation of peat-substrate habitats, which in other basins is rarely seen in the compression record (e.g. *Spencerites*, *Omphalophloios*). The Lower Radnice Member compression record is also enhanced by the introduction of upland species derived from adjacent palaeohighs. The depocentre of the Lower Radnice Member had the character of a narrow river-valley system surrounded by up to 200 m high topography (Opluštil, 2005), from which it was easy to transport and deposit upland plants to the valley bottom. Many of the representatives of the genera *Triphyllopteris, Rhacopterisor* or *Noeggerathia* occur only at this stratigraphical level, and probably represent such upland taxa. However, the largest inflation of species is related to small sphenopteroid foliage morphotaxa (ferns and some pteridosperms), which are mostly poorly known and need taxonomic revision. This statement is supported by the fact that 49 of the 73 sphenopteroid species reported from this unit are known only from the Lower Radnice Member (Pešek, 2004) and many of them have been reported only once.

The enhanced diversity at the Barnsley Seam is less easy to explain. Wray (1932) suggested that it was because the Barnsley Thick Seam splits into two leaves in this area, and because many of the rarer and more unusual macrofloral taxa occur in the inter-seam beds (known as the 'Monckton Rock'). However, an analysis of the South Wales Basin macrofloral record shows a similar albeit less pronounced peak in diversity at a stratigraphically similar level, suggesting that this might not be just a local effect (Cleal, 2007, this issue).

Diversity patterns vary between different coalfields. In the central Pennines Basin, the number of species per horizon progressively increases through the Langsettian and lower Duckmantian substages, typically reaching 60 species (Cleal, 2005). This then falls to about 40 species in the upper Duckmantian Substage, and 15–20 species in the lower–middle Bolsovian Substage. This fall in diversity is characterized by proportional changes in all of the main plant groups, including lycopsids (Fig. 8). In the South Wales Coalfield, the number of species also progressively increases from the Namurian Stage to the lower Duckmantian Substage, reaching a typical diversity of

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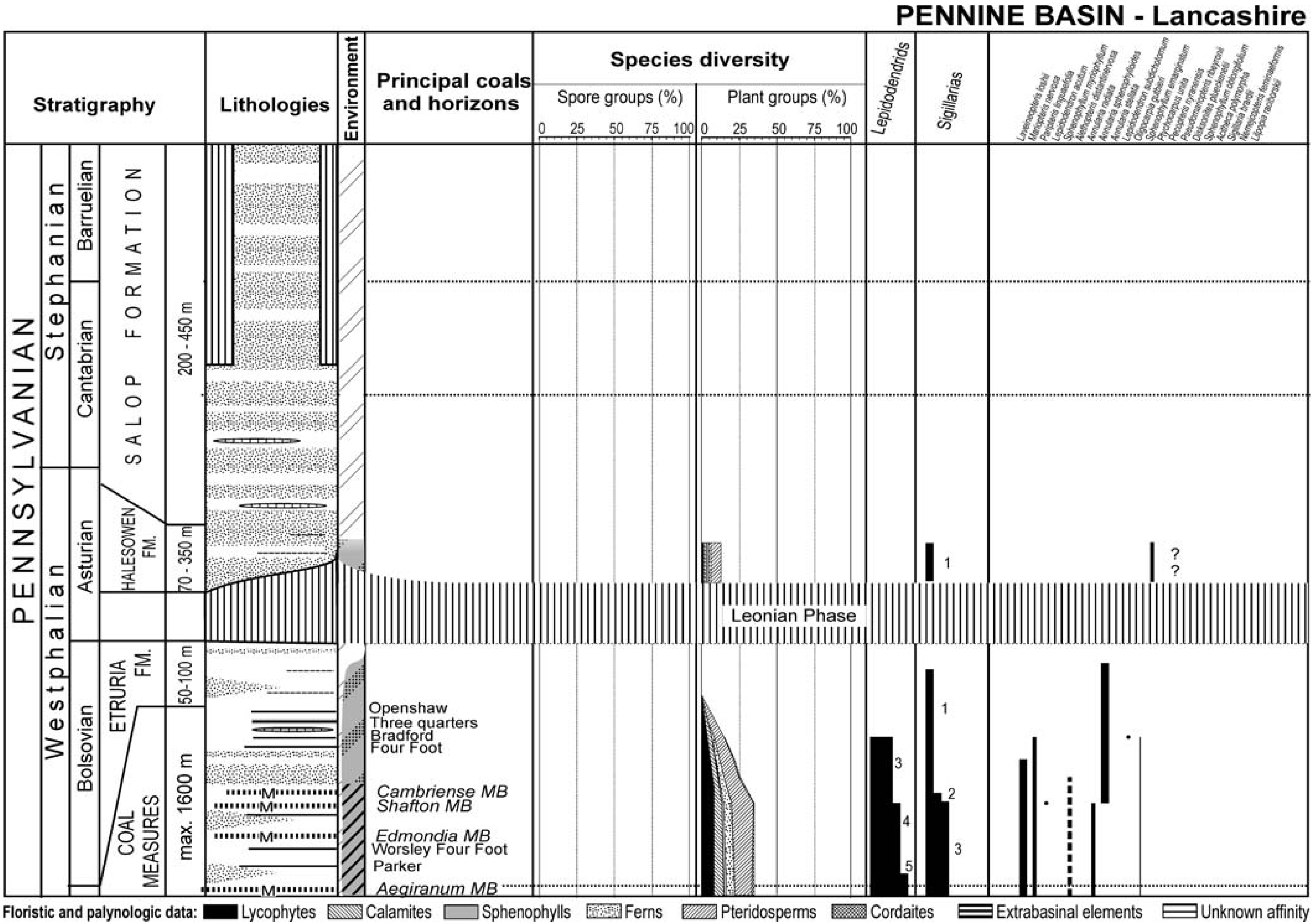

Figure 8. Detailed lithostratigraphy and plant diversity patterns of the late Westphalian–early Stephanian interval of the Pennines Basin, with the stratigraphical ranges of the key taxa. For lithology see Figure 3.

60 to 70 species (Cleal, 2007, this issue). As in the Central Pennines Basin, there is also a fall in species diversity to about 40 species per horizon in the South Wales Basin, but it occurs significantly later, in the Bolsovian Substage (Figs 3, 5). Whether and how these two drops in species diversity are related is at present unclear. The change in the Central Pennines Basin may be related to local changes in water table levels, as reflected in the development of red beds (Etruria Formation), but there is no equivalent change recognizable in the South Wales Basin. The influx of alluvial sands known as the Pennant Formation occurs significantly earlier and has no observable effect on species diversity. The onset of red beds deposition in the Pennines Basin may also explain lower lepidodendrid diversity in this basin compared to the South Wales Coalfield (Figs 8, 9). The South Wales Basin sees a subsequent rise in diversity to 50 species in the upper Asturian Substage, not reflected in the Central Pennines Basin, where diversity remains very low. This seems to correlate with the appearance of a range of new species of marattialean ferns, and callistophytalean and medullosalean pteridosperms, a vegetational change that has been widely recognized across Euramerica (e.g. Cleal, 1978, 1984*a*, 1987; Pfefferkorn & Thomson, 1982; Zodrow & Cleal, 1985; Wagner & Alvarez-Vázquez, 1991). This marked vegetational change coincided with Variscan tectonic activity known as the Leonian Phase (Wagner, 1966), which produced significant structural movement in the Iberian Peninsula and palaeogeograhical changes in northern and central Europe.

In the Upper Silesia Coal Basin, about 800 morphospecies have been identified (Kotasowa & Migier, 1995) within all the coal-bearing succession from the lower Namurian to the lower Stephanian. The use of morphotaxa inflates significantly the number of species (the Pennines and South Wales studies were based on estimates of original whole plant species numbers) but does not exclude the possibility of recognizing major changes in biodiversity. The number of morphotaxa irregularly increases from about 40 species in the Viséan Stage to 178 species in the Langsettian Substage. Two apparent drops in species diversity exist around the Duckmantian/Bolsovian and Bolsovian/Asturian boundaries, where their number decreases from 164 to 108 species or from 105 to 76 species, respectively (Figs 4, 10). Kotasowa & Migier (1995) explained these falls in diversity as being a result of short hiatuses and the lithological character of the strata. Bolsovian and Asturian successions are dominated by coarse-grained fluvial deposits which generally provide fewer plant fossils than mudstone-dominated units of Langsettian–Duckmantian age. It could be especially important when data are derived from cores that do not yield enough specimens from the plant-bearing lithologies, such as in the records of Kotasowa & Migier (1990). This can partly explain the decrease in diversity of species around the Duckmantian and Bolsovian boundary, but not another similar drop in diversity between the Bolsovian and Asturian substages where there is no lithological change and the sampling method remained the same. It indicates that the diversity curve of the Upper Silesia Coal Basin based on core data was only biased a little by lithological change and that it reveals real changes in original plant diversity. This assertion is supported by the observation in the South Wales Basin (Cleal, 2007, this issue), where a similar lithological change in the middle Bolsovian Substage is not associated with any significant drop in plant diversity (Fig. 5).

In the Central and Western Bohemia Basin, a sudden reduction in diversity occurs at the boundary between the Lower and Upper Radnice members, where the number of plant taxa decreases from 139 to 59 species (Fig. 6). This drop is explained by Havlena (1964) as being of an ecological nature, but we believe that less extensive mining activity in the Upper Radnice Member may be also partly responsible. Supporting evidence for a drop in original plant diversity might be obtainable from the palynological record, but to date we only have a list of palynotaxa for the whole Radnice Member. Another but less apparent drop in diversity is located at the boundary between the Radnice and Nýřany members, separated by a hiatus spanning the interval of the lower Asturian. The number of plant species decreases from 59 to 47 at the base of the Nýřany Member, where conglomerates and coarse-grained sandstones dominate the succession, but higher up increases to 59 species in the level of the economically attractive part of this unit (Nýřany and Chotíkov groups of coals). More apparent than the decrease of species is a floristic change marked by the disappearance of many species (e.g. lycopsids) and the appearance of new taxa, especially ferns. This compares in some ways to the observed pattern in the South Wales Coalfield, with a drop in diversity in the uppermost Bolsovian Substage, and a partial recovery in the middle Asturian Substage. However, the observed lower floral diversity in the economically unattractive basal and upper parts of the Nýřany Member is probably as much to do with sampling bias, as the macroflora was collected only from thin mudstone intercalations penetrated by deep boreholes and from few outcrops. This sampling bias is also indicated by palynological data, which display a similar diversity through the whole Nýřany Member (Figs 6, 11). The number of macrofloral species further decreases in the Barruelian Týnec Formation, the lithology of which is similar to the underlying Nýřany Member, but differs in the greater development of red beds (this further stresses the sampling bias caused when conditions become unfavourable for fossilization).

In the Intra Sudetic Basin, the highest diversity with rich arborescent lycopsids is in the coal-bearing Langsettian and Duckmantian strata, whereas it apparently

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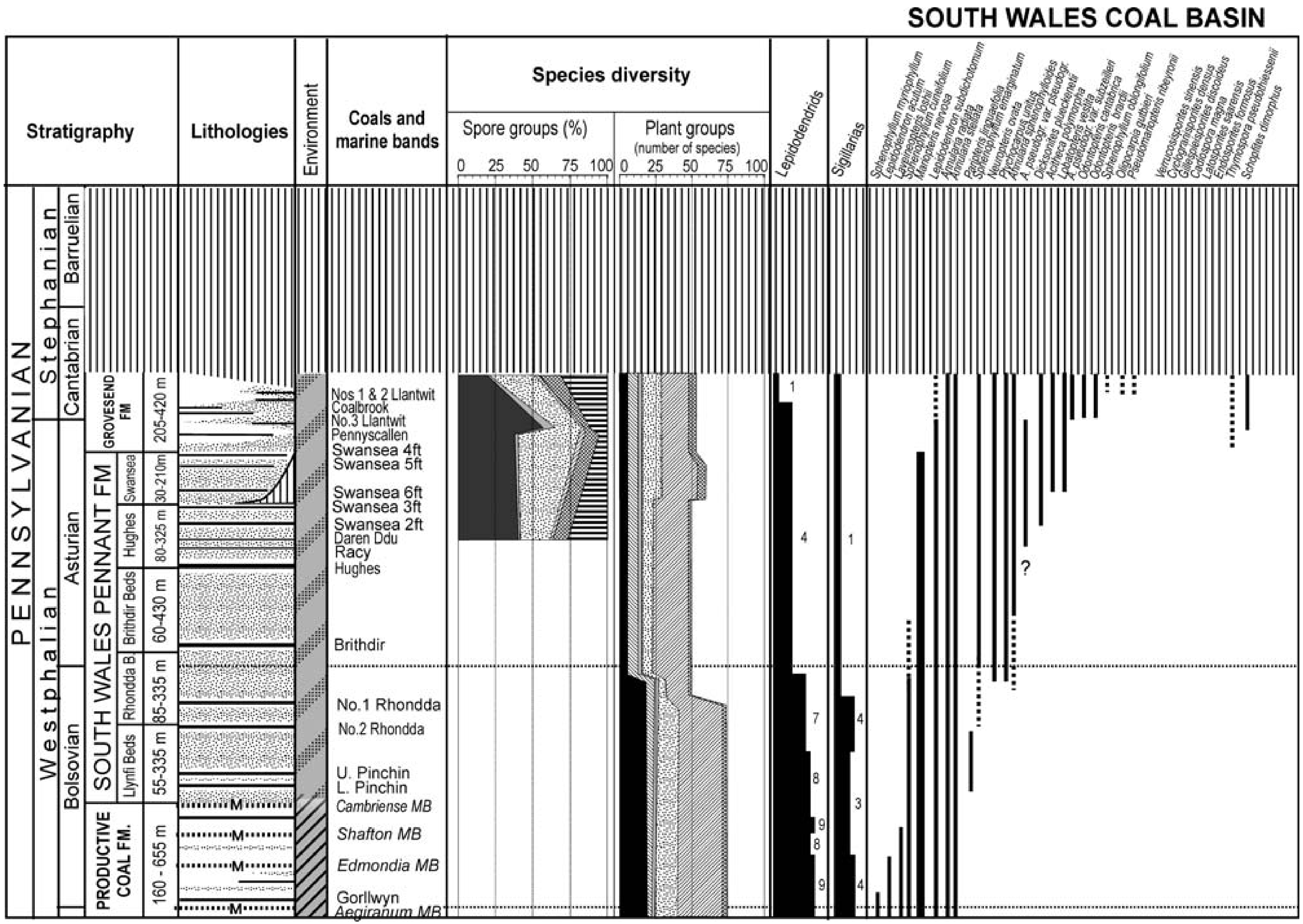
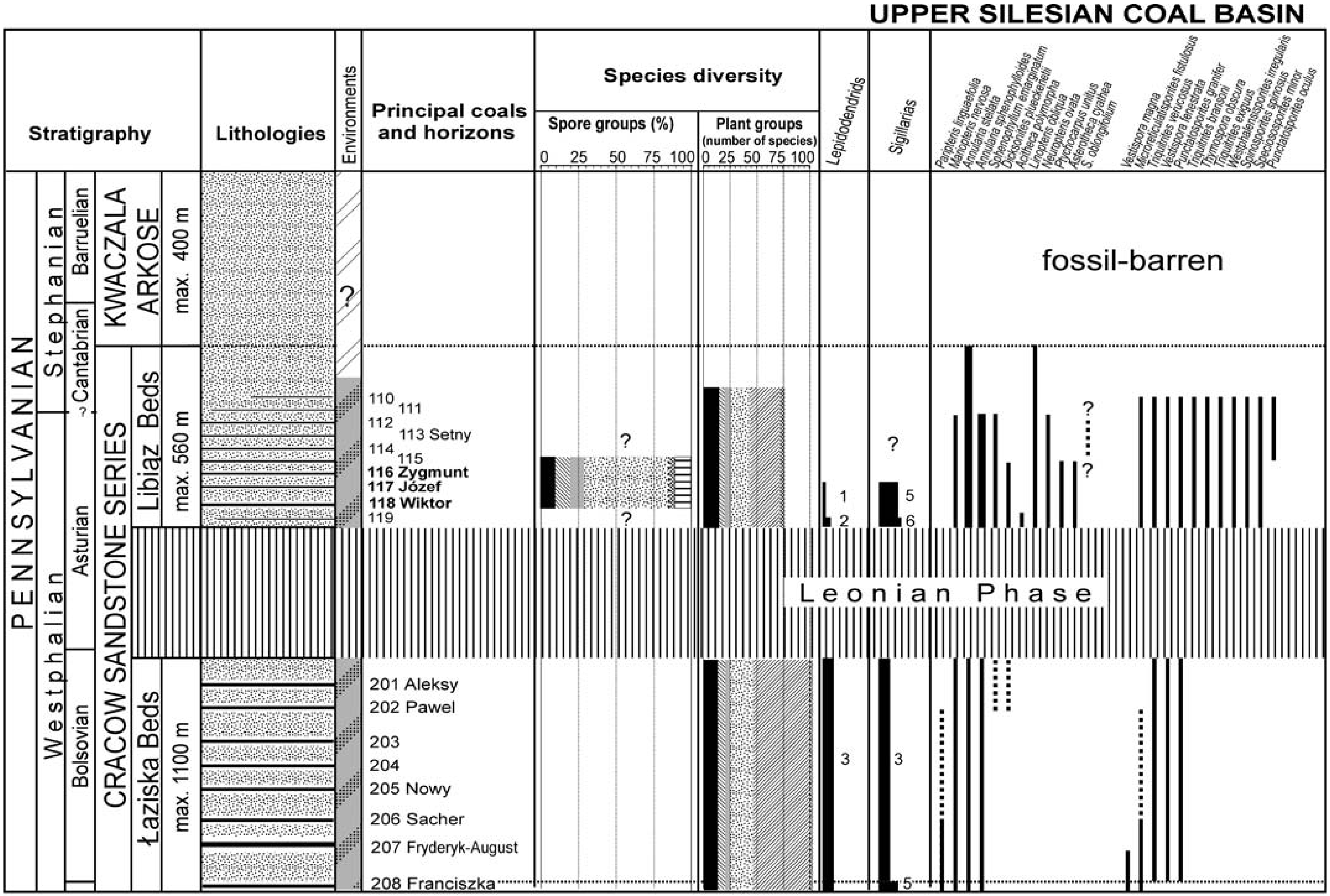


Figure 9. Detailed lithostratigraphy and plant diversity patterns of the late Westphalian–early Stephanian interval of the South Wales Coal Basin, with the stratigraphical ranges of the key taxa. For explanation see legend in Figure 8.



Source of data: Stopa (1957); Kotasowa & Migier (1995); Kotas (1995)

Figure 10. Detailed lithostratigraphy and plant diversity patterns of the late Westphalian–early Stephanian interval of the Upper Silesia Coal Basin, with the stratigraphical ranges of the key taxa. For explanation see legend in Figure 8.

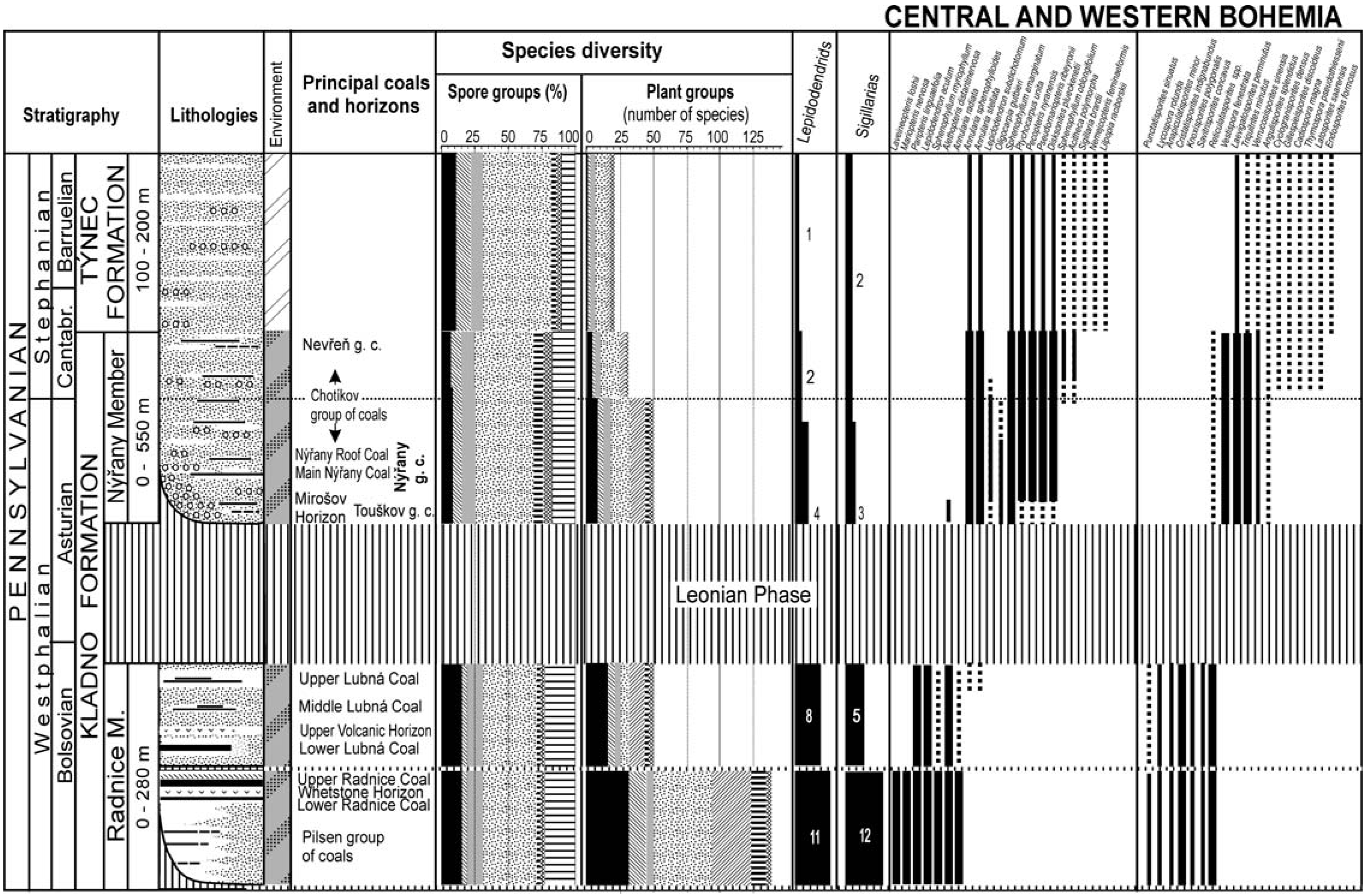


Figure 11. Detailed lithostratigraphy and plant diversity patterns of the late Westphalian–early Stephanian interval of the continental basins of Central and Western Bohemia, with the stratigraphical ranges of the key taxa. For explanation see legend in Figure 8.

decreases in the Bolsovian and Asturian sediments, probably as a result of the absence of economically important coal seams in this stratigraphical interval (Fig. 7). A sudden but not very prominent increase in diversity appears in the Svatoňovice (Cantabrian age), accompanied by a floristic change, and especially Radvanice (Stephanian B age) group of coals, whereas the thick sequence of coarse-grained fluvial to alluvial strata between these coal seams (Jívka Member, Barruelian age) has provided only permineralized woods. Palynomorph diversity displays a pattern similar to the macroflora (Figs 7, 12) and was probably affected by sampling technique as well as by poor conditions for preservation of spores in oxidizing environments during the deposition of red bed units. However, the number of spore taxa is two to three times higher than that of the plant species.

6. Interpretation

6.a. Tectono-sedimentary histories

The tectono-sedimentary history of the basins generally depends on their geotectonic and palaeogeographical positions, as well as on a conspicuous crustal heterogeneity in the European Variscides. Despite this, a certain degree of similarity in tectono-sedimentary history exists not only between particular foreland basins, but also between the cratonic and foreland basins of the North Variscan Foreland. In all these basins, sedimentation had already started in pre-Pennsylvanian times, and followed a progressive change from marine, to non-coal-bearing paralic, to coal-bearing paralic, to coal-bearing continental, and finally to coal-barren continental red bed deposition. Differences appear in timing of the major environmental changes, especially in the transition from paralic to continental deposition, and when sedimentation eventually ceased. However, a similar stratigraphical range of mudstone-dominated successions is apparent in the Upper Silesia, South Wales and Pennine basins (Fig. 2), which span the Langsettian to the end of the Duckmantian in the Upper Silesia Basin or to the late early Bolsovian in the South Wales and Pennines basins. In all three basins, this mudstone-dominated interval passes into sandstone-dominated sediments of a braided fluvial system. Similarities in the stratigraphical range of paralic deposition between the Pennines and South Wales basins may result from their close palaeogeographical position and therefore their similar response to tectonic processes in the Variscan orogen to the south.

In contrast, there is considerable autonomy in tectono-sedimentary histories displayed in the fault-related continental basins. They are characterized by deposition continuing through into Permian times, and sharing the same depositional breaks in late Westphalian (Leonian Phase) and late Stephanian times (Fig. 2). These breaks, marked by erosional discordances, are weak angular unconformities, and the resulting disruption in the continuity of the floristic record reflects the various peaks in the orogenic activity within the Variscides. The most widespread is the hiatus related to the Leonian Phase around the Bolsovian/Asturian boundary, which has been identified in all of the compared paralic basins.

6.b. Climatic trends

Based on our current observations and comparisons of data derived from selected basins of the north Variscan foreland (paralic basins) and its interior (intramontane basins), the following Pennsylvanian climatic history can be suggested for this area. Four regionally widespread intervals, defined by major basin-wide alternations of red and grey sediments related most probably to climatic changes, can be identified across the area during Westphalian and Stephanian times (Fig. 13): (1) a Namurian–Westphalian wet interval, (2) an early Stephanian dry interval, (3) a middle Stephanian wet interval and (4) a late Stephanian dry interval which continued until Autunian times where further increase of aridity is indicated by the associations of caliche with pedogenic sulphates (Skoček, 1993). This pattern is in agreement with the conclusion of Skoček (1974) and Besly (1987) and some other authors. The length of the particular intervals varied considerably. The Westphalian wet interval lasted for more than 6 Ma, but the others were much shorter in duration: the early Stephanian dry and middle Stephanian wet intervals lasted for about 1.5 and 2 Ma, respectively, and the late Stephanian dry interval for about 2 Ma, although continuing further into Permian times. These intervals, especially the longer ones, probably do not represent periods of uniform climatic conditions, but some oscillations are likely to have occurred (Phillips, Peppers & DiMichele, 1985; Besly, 1987; Cleal & Shute, 1995). In the Namurian–Westphalian wet interval, the climate was most probably ever-wet tropical with evenly distributed precipitation at least across the whole study area. A possible signal of forthcoming climatic change towards the end of this interval may be indicated by the sporadic occurrence of caliche palaeosols in the Pennines Basin from the middle Bolsovian onwards, which suggests short intervals of drier, probably seasonal climate (Besly, 1987). However, no caliche have been reported from the other basins in the Bolsovian Substage and only rarely do they occur in the Asturian Substage of the Intra Sudetic Basin (Skoček, 1993). This may suggest relatively drier climatic conditions in the Pennines Basin compared to other Variscan basins located further east. The climate in the early Stephanian (Barruelian) dry interval was probably characterized by reduced and (?)seasonal precipitation allowing the formation of common and widespread caliche palaeosols. A similar climatic record for this interval can be identified

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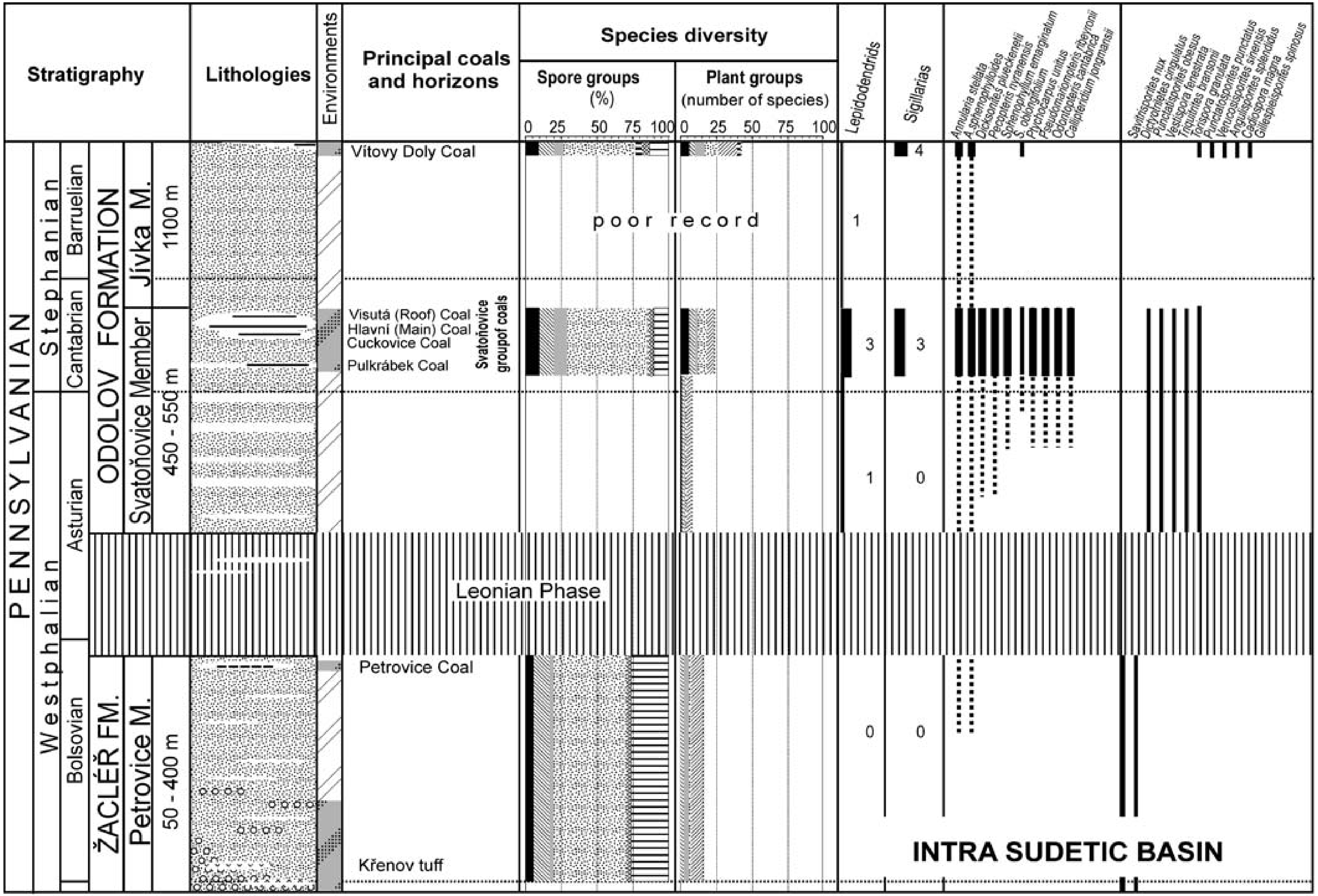

Source of data: Pešek 2005, Pešek et al. 2001

Figure 12. Detailed lithostratigraphy and plant diversity patterns of the late Westphalian–early Stephanian interval of the Intra Sudetic Basin, with the stratigraphical ranges of the key taxa. For explanation see legend in Figure 8.

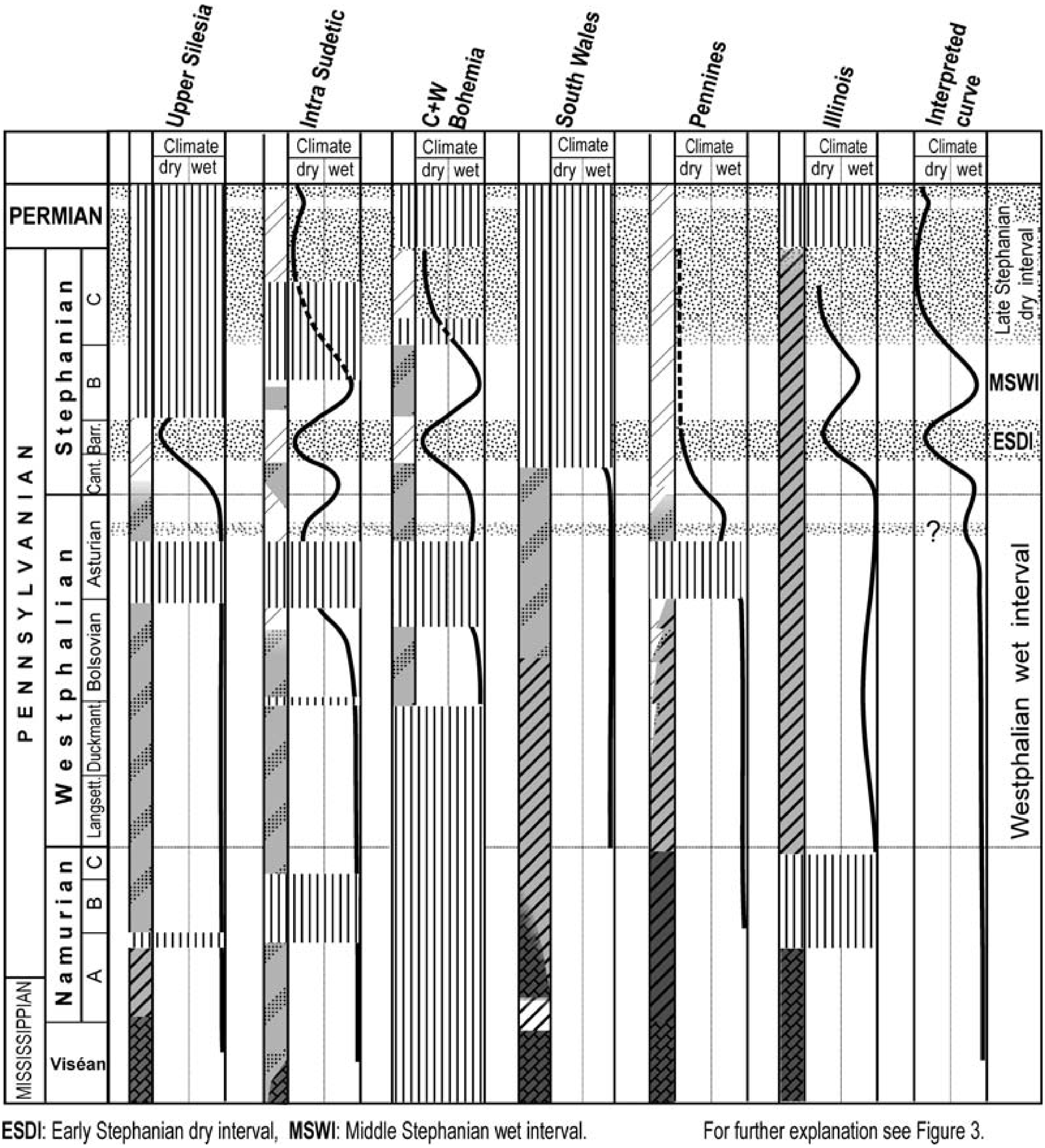

Figure 13. Climatic curves derived for the studied basins and interpreted climatic curve for the European part of the Euramerican province. For explanation see legend in Figure 3.

also in the Appalachians and the interior basins in North America. During the Middle Stephanian wet interval, the climate became humid enough to allow increased accumulation and preservation of peat and significantly restricted conditions favouring red bed formation except in the Pennines Basin. Here the evidence of this wet interval is absent. Instead, red beds with caliche occupied most of the Stephanian so the early and late Stephanian dry intervals merge together into one long dry period. The late Stephanian dry interval was a period of generally dry climate, but the existence of short humid periods is indicated by occasionally increased coal abundance, including some seams of economic importance. Mineable coal seams occur outside the compared basins, for example, in the Central French Massif, NW Spain, Saar-Lorraine (Breitenbach coals) and in the Boskovice Graben, coal seams are present in Stephanian C strata. Their formation might be related to high water tables in floodplains, probably maintained by poor drainage due

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to a high accommodation rate as well as to locally more humid climatic conditions.

The climatic history suggested above indicates that Pennsylvanian times were characterized by a transition from wetter to drier conditions, which is in broad agreement with the observations of previous authors (e.g. Besly, 1987; Gastaldo, DiMichele & Pfefferkorn, 1996; Scotese, Boucot & McKerrow, 1999). In detail, this transition was marked by alternations of shorter dry and wet intervals in which each consecutive dry interval was longer and more severe, whereas the wet interval became shorter and less pronounced. It indicates that controls on climatic change operated on several timescales, with short-term climatic fluctuations being imposed on a longer-term drying trend (Besly, 1987). The long-term drying trend has a periodicity on a scale of about several tens of million years. It already began in Mississippian times, and continued through Pennsylvanian times to peak in late Permian and early Triassic times (e.g. Gastaldo, DiMichele & Pfefferkorn, 1996). This long-term drying trend is often explained as a transition from icehouse to greenhouse climates due to fluctuation of CO_2 and O_2 levels in the atmosphere (Berner, 1990, 1991; Frakes, Francis & Syktus, 1992; Worsley *et al.* 1994). In this concept, the Pennsylvanian is a period of very high oxygen levels resulting in an icehouse climate which passed gradually into a late Permian and Triassic greenhouse world. A different explanation was proposed by Bless, Bouckaert & Paproth (1984), who explained the diachronous onset of drier conditions in Euramerica by a northwards drift of Pangea across the latitudinally controlled climatic belts. However, this assumption is not generally accepted for several reasons discussed by Besly (1987). A more sophisticated approach by Rowley *et al.* (1985) correlated the changes in atmospheric circulation to Pangea assembly and growth of mountains belts during the Variscan orogeny. The latter is believed to be responsible for a rain shadow across the Variscan and Appalachian forelands during the Stephanian (Rowley *et al.* 1985; Besly, 1987).

The origin of the short-term climatic cycles superimposed on longer ones is less clear. Their length varies between 1.5 and about 2 Ma, which excludes their connection with the orbitally driven climatic oscillations, especially with eccentricity, the periodicity of which is on the order of 100 and 400 ka. These cycles are recorded by regional-scale alternation of red coal-barren and grey coal-bearing intervals in the late Middle and Late Pennsylvanian in all the compared basins except the Pennines Basin. Here, an uninterrupted drying trend started in the late Westphalian and continued through the whole Stephanian. This indicates coexistence of drier sub-tropical climate in Central England with humid equatorial conditions in basins of the Variscan fold belt (Besly, 1987).

Quite separate from this cyclic and progressive increase in aridity were the eustatically driven climatic fluctuations operating on a scale of about 100 ka and usually connected with the orbitally driven Milankovitch cycles (e.g. Heckel, 1986; Leeder, Harris & Kirkby, 1998; Tandon & Gibling, 1994). These glacioeustatic oscillations produced a characteristic cyclicity in the sedimentary successions in the different basins, affected by local tectonics and sediment supply (Leeder, Harris & Kirkby, 1998; Gibling *et al.* 2004). However, while climatic changes related to these glacioeustatic cycles affected the migration of low-land ecosystems, they did not result in their extinction (Gastaldo, DiMichele & Pfefferkorn, 1996; Pfefferkorn, Gastaldo & DiMichele, 2000; Falcon-Lang, 2004) and so are not discussed further in this paper.

6.c. Floristic changes

Comparison of the macroflora and palynomorph diversity curves shows that the highest number of species occurs in Langsettian and Duckmantian strata (Fig. 14). This interval corresponds to the highest concentration of coals and maximum extent of the coal forests. Above this interval, diversity in most basins decreases except for the Central and Western Bohemia Basin, where deposition started in early Bolsovian times. The first apparent drop in biodiversity appears around the Duckmantian/Bolsovian boundary in the Pennines, Intra Sudetic and Upper Silesia basins, but a similar decrease occurs also in the Radnice Member (Central and Western Bohemia Basin) around the middle Bolsovian (Fig. 14). No such change has been identified in the South Wales Basin in this stratigraphical interval. In the Upper Silesia Coal Basin and Central and Western Bohemia Basin the decline in plant diversity coincides with a sedimentological change from a mudstone- to a sandstone-dominated fluvial succession, whereas in the Pennines Basin it coincides with a return of marine incursions (Cleal, 2005). In the Intra Sudetic Basin, an absence of mineable coal seams, probably related to a tectonically induced fall in the water table during Bolsovian times, seems to have been responsible for a prominent drop in plant diversity.

Another, but less prominent, decrease in biodiversity in the compared basins occurred between late Bolsovian and late Asturian times (Figs 14, 15). In most basins, this drop in diversity correlates with a depositional break, which makes it difficult to ascertain its precise stratigraphical location. However, in the South Wales Basin, the only basin with continuous deposition through this time, it corresponds approximately to the Bolsovian/Asturian boundary. In all the basins, this event is marked by an apparent decrease in diversity of the arborescent lycopsids, especially lepidodendrids (Figs 15, 16).

A local increase in diversity has been observed in the South Wales and the Central and Western Bohemia basins in late Asturian times, and in the Intra Sudetic

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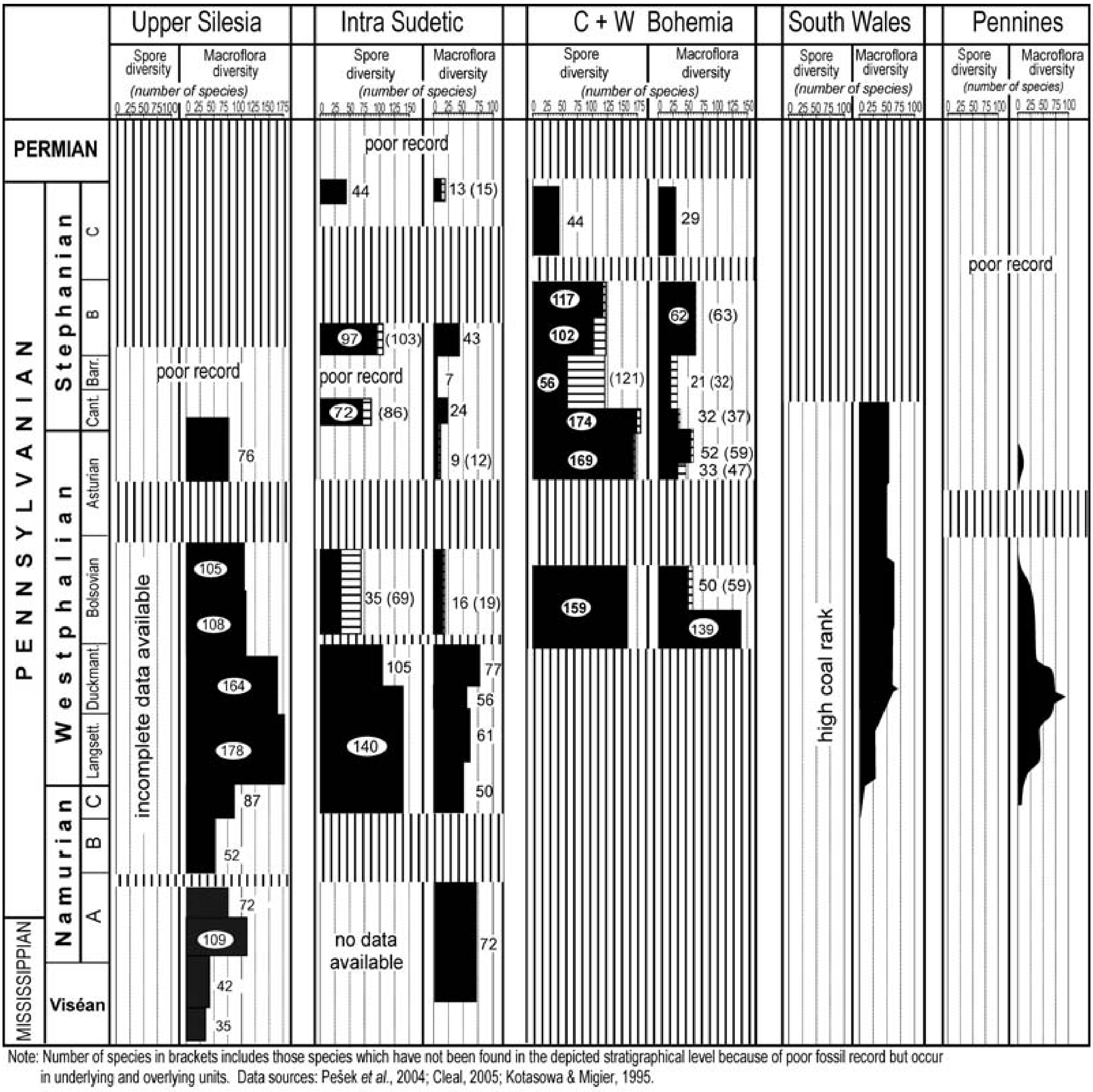

Figure 14. Comparison of Pennsylvanian-aged macroflora and palynomorph diversities in analysed basins.

Basin in Cantabrian times (Fig. 14). These gradual rises in the number of plant species reflect improved peat-forming conditions resulting from local rises of the water table. In the Intra Sudetic Basin, there is a reduction of both macrofloral and palynomorph taxa in the oldest Stephanian coal-bearing unit (Svatoňovice group of coals, Cantabrian age) similar to the latest Westphalian (early Bolsovian) decline. In the South Wales Basin, the basin succession ends in coal-bearing strata of early Cantabrian age, without any apparent decrease in plant species. Where deposition continued to the Stephanian, for example, in the Upper Silesia Coal Basin, Central and Western Bohemia Basin and in the Intra Sudetic Basin, another prominent drop in the number of plant taxa appears approximately around the Cantabrian/Barruelian boundary, when the climatic change related to the early Stephanian dry interval starts. In continental basins, this drop is followed by another increase in the number of plant species, related to the middle Stephanian wet interval. However, plant diversity of this interval reached only about 70 % of that of the Westphalian wetlands.

The diversity pattern of arborescent lycopsids copies the general macroflora diversity pattern. The highest number of species corresponds to the maximum concentration of coal seams in the Langsettian and Duckmantian, and starts to decrease in most basins during the Bolsovian (Fig. 16). Surprisingly, the highest

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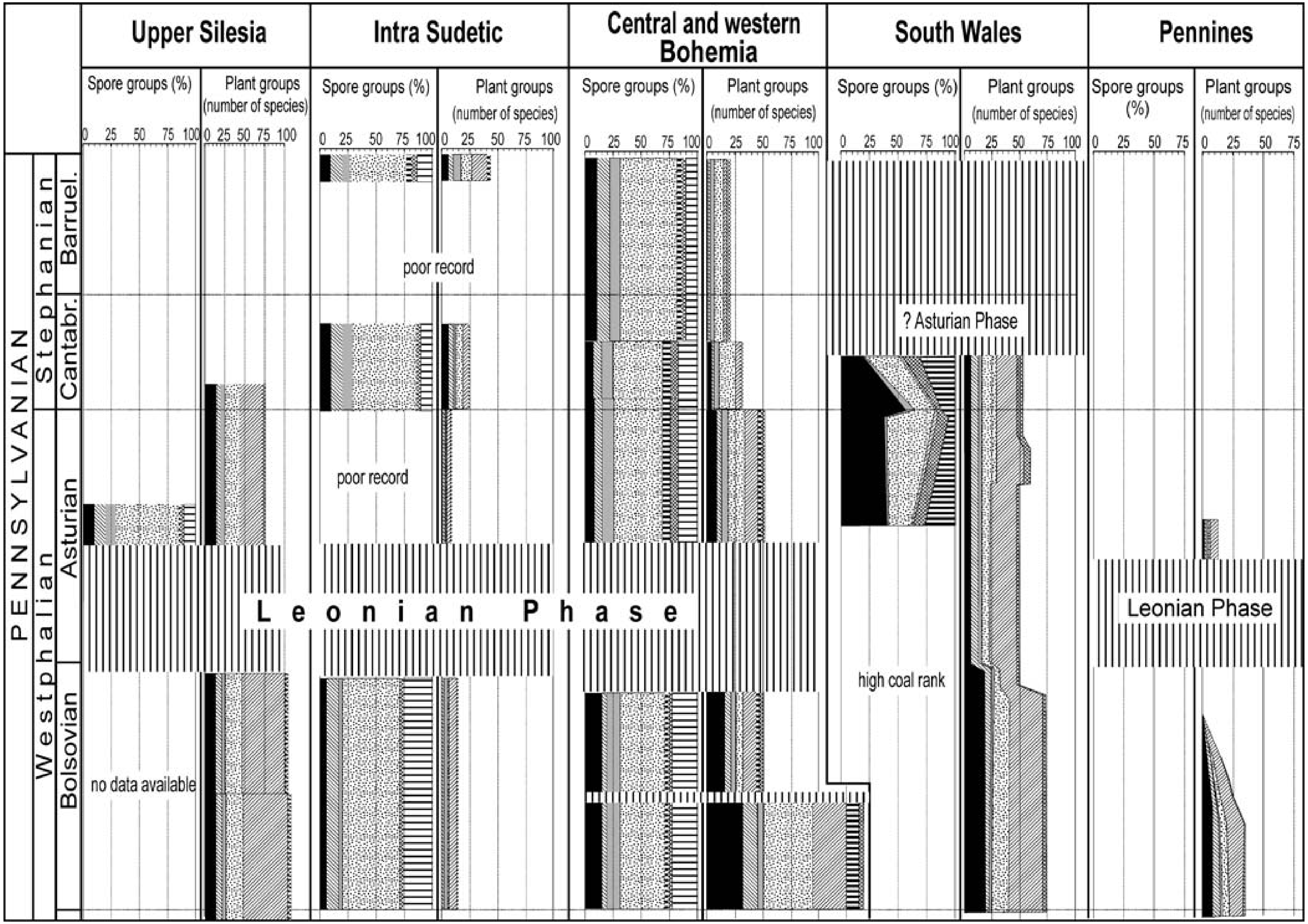

Figure 15. Comparison of late Westphalian–early Stephanian diversity patterns of macrofloras and palynomorphs of the studied basins. For explanation see legend in Figure 8.

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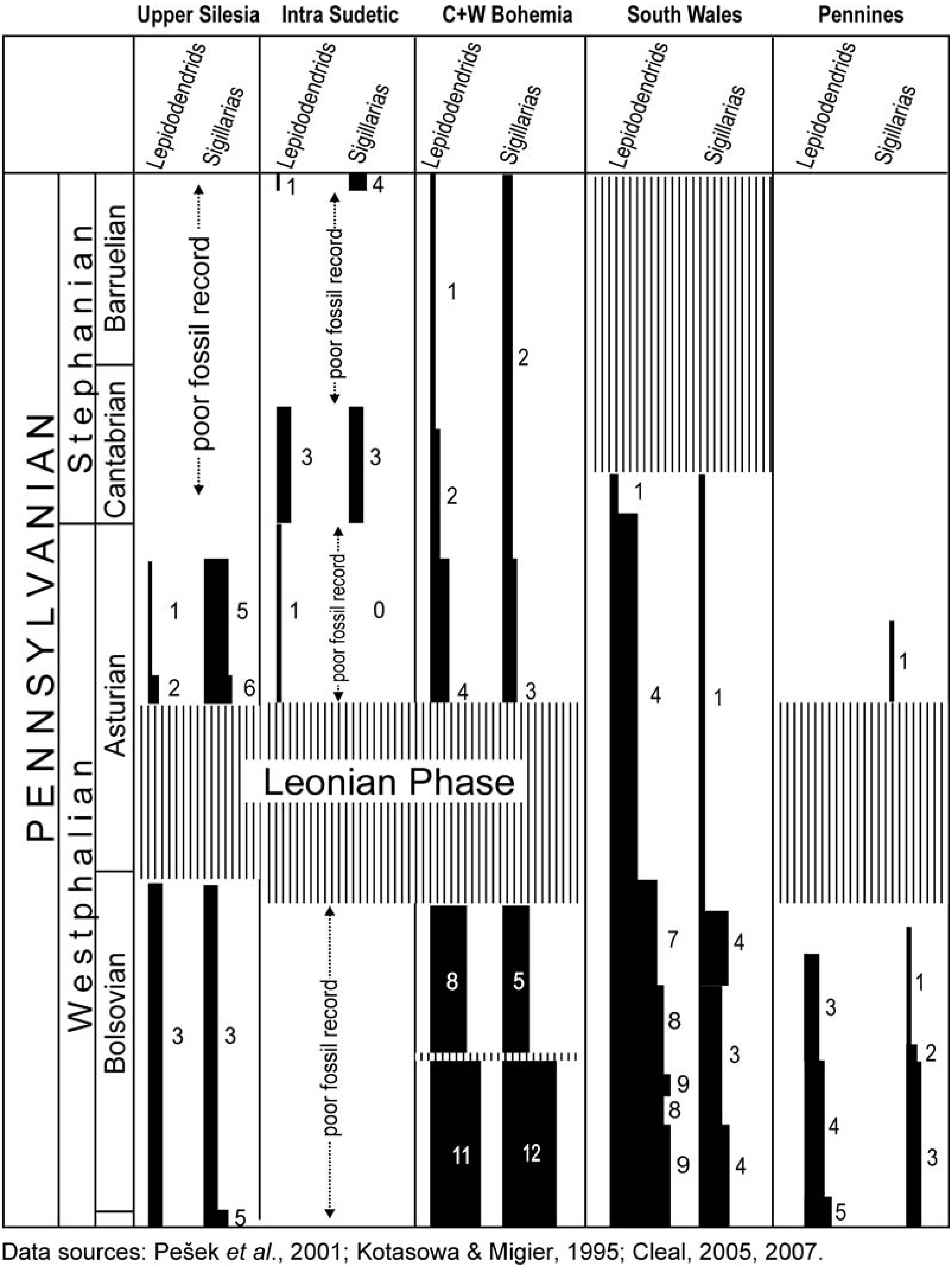

Figure 16. Diversity of lepidodendrids and sigillarians in the studied set of the basins.

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diversity of lepidodendrids is reported from Bolsovian strata in the Central and Western Bohemia Basin, probably as a result of their recent revision (Bek & Opluštil, 2004, 2006). The basic trend is a decrease in diversity of lycopsids, especially those of tree habit, and an increase in the number of fern species (mostly marattialean tree ferns) in the Asturian Substage. This trend, observed in the Central and Western Bohemia and South Wales basins, seems to be in contradiction with that in the Intra Sudetic and Upper Silesia coal basins, where the percentage of lycophytes increases. In detail, this increase is related mostly to the appearance of new sigillarian species, whereas the diversity of lepidodendrid lycophytes decreases (Fig. 16). Concerning the diversity of lepidodendrids and sigillarians, the apparent reduction by about 40 and 75 %, respectively, occurs near the Bolsovian/Asturian boundary. A further decrease took place during middle Cantabrian times when nearly all the Namurian–Westphalian lepidodendrids became extinct and only a few new but very rare species appear. This trend was identified in all of the compared coalfields (Fig. 16). Diversity of sigillarians in particular coalfields is more complicated. In the Central and Western Bohemia and South Wales basins, they follow the same pattern as the lepidodendrids (Fig. 16), but in the Intra Sudetic and Upper Silesia coal basins, the number of sigillarian species increases significantly (Pešek, 2004; Kotasowa & Migier, 1995).

7. Summary and conclusion

Comparison of vegetation diversity patterns derived from published data of several European coalfields of different geotectonic and palaeogeographical positions indicates that diversity in the macrofloral and palynological record is affected by the complex interaction of various controls. These involve primarily changes in original plant diversity, as well as in the fossilization potential related to groundwater table level and various sampling biases.

The most apparent diversity changes correlate with basin-wide alternations of coal-bearing strata and coal-barren red beds. Significantly reduced diversity in the red beds is undoubtedly a consequence of not only a less diversified vegetation due to unfavourable environmental conditions for spore-producing plants, especially arborescent lycopsids, but also of increased oxidation of organic matter during the deposition of red beds providing low potential for preservation of plant remains. Moreover, the absence of coal seams and thus of mining activity in red beds strata further increases the diversity contrast due to a sampling bias, since the plant remains from such strata can only be collected from borehole cores or natural outcrops.

Usually less apparent are diversity changes within the coal-bearing strata, where sampling bias and low fossilization potential are of only minor importance. Such changes therefore most probably represent primary changes in plant diversity. They were documented in most basins, both in cratonic and foreland basins north of the Rheno-Hercynian Zone of the Variscan Orogen and intramontane ones.

Diversity patterns in all of the compared basins differ to a greater or lesser extent, even between neighbouring basins, owing to local controls related mostly to local tectonic and palaeogeographical factors. Nevertheless, there are a number of events and/or trends which are common for most or all the basins and can be ascribed to regional or global controls, especially to those related to climate. Maximum diversity in all of the compared basins is in the Langsettian–Duckmantian interval, except in the Central and Western Bohemia Basin where the maximum is of early Bolsovian age.

The subsequent fall in plant diversity is very uneven and step-like. The first apparent decline occurs around the Duckmantian/Bolsovian boundary, which can be identified in all the basins except the South Wales Basin (where no change occurs) and the Central and Western Bohemia Basin (where a similar change occurs in the middle of the Bolsovian Substage). This drop can be explained in various ways, either as a consequence of a return of marine incursions (e.g. the Pennines Basin: Cleal, 2005), a lithological change from mudstone-dominated to coarse-grained fluvial facies (Kotasowa & Migier, 1995), or as a consequence of sampling bias (e.g. the disappearance of economically important coal seams in the Bolsovian strata of the Intra Sudetic Basin). Both environmental changes could result in water table changes fatal for survival of some plant species. The second, but less marked, diversity decline occurs near the Bolsovian/Asturian boundary, most obviously present in the Upper Silesia Coal Basin, the South Wales and the Pennines basins, and less obviously so in the Central and Western Bohemia Basin. In the Intra Sudetic Basin, strata of this age are in a non-coal-bearing facies, partly formed of red beds, with only a very poor plant fossil record.

The data presented in this paper indicate that the floristic changes observable in the upper Westphalian and Stephanian fossil record are related most probably to climatic changes marked by the onset of seasonal climate across most of Euramerica. This onset was probably not isochronous and was modified by various controls, including the existence of orographical barriers, distance of the basin from the coast, or sedimentary environment including water table level. Tectonics and basin inversion resulted in the collapse of coal measure forests only locally, that is, where they resulted in water table drop and improved drainage.

Nevertheless, the results of this analysis are considered as preliminary; more reliable conclusions will only be achieved by incorporating data from other basins of the North Variscan Foreland and the adjacent continental basins. This is the main objective of the project IGCP 469.

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